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GR focus review

Tectonic units of the Alpine collision zone between Eastern Alps and western Turkey

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ABSTRACT

We present a map that correlates tectonic units between Alps and western Turkey accompanied by a text providing access to literature data, explaining the concepts used for defining the mapped tectonic units, and first-order paleogeographic inferences. Along-strike similarities and differences of the Alpine-Eastern Mediterranean orogenic system are discussed. The map allows (1) for superimposing additional information, such as e.g., post-tectonic sedimentary basins, manifestations of magmatic activity, onto a coherent tectonic framework and (2) for outlining the major features of the Alpine-Eastern Mediterranean orogen. Dinarides-Hellenides, Anatolides and Taurides are orogens of opposite subduction polarity and direction of major transport with respect to Alps and Carpathians, and polarity switches across the Mid-Hungarian fault zone. The Dinarides-Hellenides-Taurides (and Apennines) consist of nappes detached from the Greater Adriatic continental margin during Cretaceous and Cenozoic orogeny. Internal units form composite nappes that passively carry ophiolites obducted in the latest Jurassic—earliest Cretaceous or during the Late Cretaceous on top of the Greater Adriatic margin successions. The ophiolites on top of composite nappes do not represent oceanic sutures zones, but root in the suture zones of Neotethys that formed after obduction. Suturing between Greater Adria and the northern and eastern Neotethys margin occupied by the Tisza and Dacia mega-units and the Pontides occurred in the latest Cretaceous along the Sava-İzmir-Ankara-Erzincan suture zones. The Rhodopian orogen is interpreted as a deep-crustal nappe stack formed in tandem with the Carpatho-Balkanides fold-thrust belt, now exposed in a giant core complex exhumed in late Eocene to Miocene times from below the Carpatho-Balkan orogen and the Circum-Rhodope unit. Its tectonic position is similar to that of the Sakarya unit of the Pontides. We infer that the Rhodope nappe stack formed due to north-directed thrusting. Both Rhodopes and Pontides are suspected to preserve the westernmost relics of the suture zone of Paleotethys.

1. Introduction

The system of Alpine orogens in the Mediterranean region, i.e. orogens that formed after Mesozoic rifting and opening of new oceanic domains following the Variscan cycle, is notoriously known for being of bewildering complexity. This orogenic system forms a very broad and diffuse boundary between the African and European plates that includes intervening microcontinents, most notably Adria, remnants of closed ocean basins (ophiolites and subduction mélanges) as well as still open (Eastern Mediterranean) or newly opened (Western Mediterranean) oceanic domains (Cavazza et al., 2004; Jolivet and Brun, 2010; Handy et al., 2010;

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Facenna et al., 2014; Jolivet et al., 2013; Gürer and van Hinsbergen, 2019). It serves as an excellent natural laboratory to study a wide variety of geodynamic scenarios active in the past and still on going at present not only at crustal scale (e.g. Şengör et al., 1984; Jolivet et al., 2003; Cavazza et al., 2004; Schmid et al., 2004; Carminati et al., 2012; Nocquet, 2012), but also in terms of mantle dynamics driving orogeny (e.g. Wortel and Spakman, 2000; Facenna et al., 2003; Kissling and Schlunegger, 2018).

The western and eastern parts of the Mediterranean regions contain blocks and orogens that are part of mostly one or two countries. The Pyrenees, Central Iberian Ranges, and Betics are part of Spain, the Apennines are entirely in Italy, and the Anatolian orogen is almost entirely part of Turkey. However, data on the geology and tectonics of the area from the Alps to the Aegean region, which is in the focus of this contribution, have been collected in a multitude of different countries and published in different languages, often by using concepts and terminology that abruptly change across national boundaries. This complexity combined with the unusual large amount of papers focusing on local details, makes a correlation of tectonic units over larger areas difficult. Fortunately, a number of publications synthetizing data and focusing on a larger area are available such as, mentioning just a few relevant ones: Froitzheim and Schuster (2008) for Alps and Western Carpathians; Csontos and Vörös (2004), Vozár et al. (2010), Kovács et al. (2011a) and Haas et al., 2012) for the Circum-Pannonian region; Săndulescu (1994), Burchfiel and Bleahu (1976) and Maţenco (2017) for the Romanian Carpathians; Pamić et al. (2002), Karamata (2006) and Schmid et al. (2008) for the Dinarides; Papanikolaou (2009) for the Hellenides; Vangelov et al. (2013) and Burchfiel and Nakov (2015) for the Balkan orogen; Jahn-Awe et al. (2010) and Burg (2011) for the Rhodopes; Okay (2008) and Pourteau et al. (2016) for western Turkey. These turned out to provide access to relevant publications offering details that are relevant for compiling a unified map of the main tectonic units of the Alpine collision zone between Eastern Alps and western Turkey.

The principal aim of this contribution is to provide a regional overview of the major tectonic units located between the Alps and western Turkey in the form of a map and accompanying text. Of course, mapping of tectonic units always goes together with a personal interpretation of data according to a chosen concept. When describing the individual units such concepts will be presented. Where needed, i.e. in case of particularly controversial issues, our concept will also be discussed in the light of alternatives. According to Roberts and Bally (2012) the purpose and scope of regional geology "…is to judiciously reconcile insights obtained by studies and surveys of the various geoscience disciplines and to end up with a coherent, observation/data-based narrative that explains the geologic evolution of larger regions…" Needless to say, that modern science is ever more process-oriented and hence the development of such a narrative is a potentially useful tool attempting to provide a factual basis for modern process-oriented studies.

The map presented in Plate 1 (also available at high resolution in the Supplementary material) provides a southward and eastward extension of an earlier map compilation limited to the area north of latitude 42° and east of longitude 30° by Schmid et al. (2008), which was moderately modified for this work. This contribution primarily provides a base map of the area between Alps and western Turkey that may be used in studies focusing on a particular topic or process in the future. The earlier compilation (Schmid et al., 2008) served as a base map for the kinematic restoration of the Circum-Pannonian area for the last 20 Ma (Ustaszewski et al., 2008). The map presented here, available on ResearchGate in provisional and unpublished forms already since 2010, already served as a base map for a kinematic restoration of the Aegean-West Anatolian accretion and extension since the Eocene (van Hinsbergen and Schmid, 2012). Seghedi et al. (2013) used a provisional version of this map in their review on the Miocene-Quaternary volcanism in the context of the geodynamic evolution of the area from the Pannonian basin to the Menderes massif. Horváth et al. (2015) and Balázs et al. (2016) used this map to discuss extension and associated faulting in the Pannonian basin. Finally, the definite version presented here also served as a base map for the kinematic restoration of Mediterranean tectonics since the Triassic over a wider area and time frame (van Hinsbergen et al., 2019).

We emphasize that the map presented in Plate 1 represents the core of this contribution. This map is the result of a long-lasting effort of the authors to attempt a compilation of tectonic units of a larger region across national boundaries. Accordingly, the text accompanies this map compilation, and is devised such as to explain in great detail the sources of information, concepts and interpretations used during compilation. The reader is encouraged to make use of a map that indicates the geographical location of all the local names used in the text that is available in the Supplementary Material only. Fig. 1 provides a simplified overview introducing the reader into major groups of tectonic units mapped in Plate 1. We emphasize that this compilation is open to discussion and partly based on interpretations and assumptions. Hence, we consider it important to offer detailed, and hence occasionally rather lengthy, descriptions of the individual tectonic units. Particular attention will be given to an appropriate definition of the mapped units and a short discussion of the concepts chosen for grouping together local tectonic units, often labelled with a multitude of different names, into a mappable larger unit. The text also includes many references to published work that are considered as the most relevant. We also embed a series of figures that represent detailed extracts of the map presented in Plate 1 in the text. Moreover, figures presenting crustal-scale profiles across orogens will also accompany the text. These profiles provide a three-dimensional picture of this complex system of orogens, which formed as the result of a long-lasting orogenic evolution, starting in the Late Jurassic or even before and ending with very substantial Neogene displacements and rotations. This series of crustal-scale profiles is very schematic and to some extent conceptual but will hopefully help for the visualization of the geometry of the mapped units in profile view.

2. Methods of map compilation

For the tectonic map of the Alpine collision zone between Eastern Alps and western Turkey (Plate 1) we used a Lambert conformal conic projection with a central meridian at 20°E, two standard parallels at 38°N and 48°N and a latitude of origin at 30°N. This choice came about because we originally (Schmid et al., 2008) used a map of the Carpathian-Balkan Mountain Systems edited by Mahel (1973) that was convenient and initially served as a base map covering a large part of the area of Plate 1. Our map is based on several larger-scale (>1:100,000) maps cited below that were georeferenced in ArcGIS and then laid out in Adobe Illustrator. Due to the general lack of information about the projections and geodetic datum's on the maps used for compilation, the accuracy of the full-scale version of Plate 1 (see also Electronic Appendix) is variable and amounts to several hundreds of metres for the largest part of the map. Smaller-scale maps (≤1:100,000) were directly imported into the Adobe Illustrator file and scaled and rotated without further distortion so to fit a 20′ (latitude) by 30′ (longitude) georeferenced grid. The accuracy resulting from this procedure is limited to a few hundreds of metres.

The method of compilation was supported by consulting detailed geological maps and explanations available for the different countries. These maps are by far too numerous to be cited

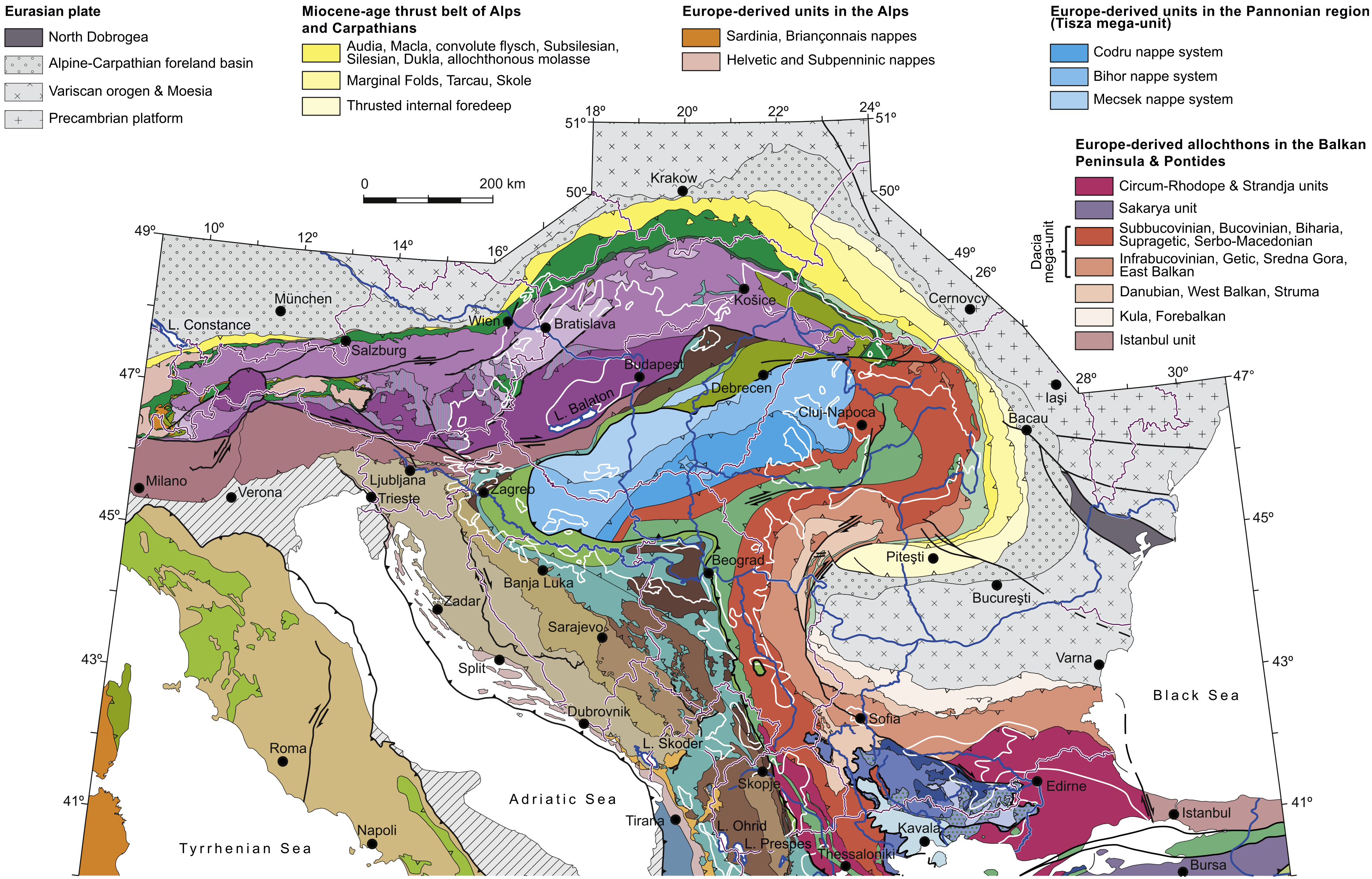

Plate 1. Tectonic units of the Alpine collision zone between Eastern Alps and western Turkey (an integral version of plate 1 can be downloaded in the Supplementary Material).

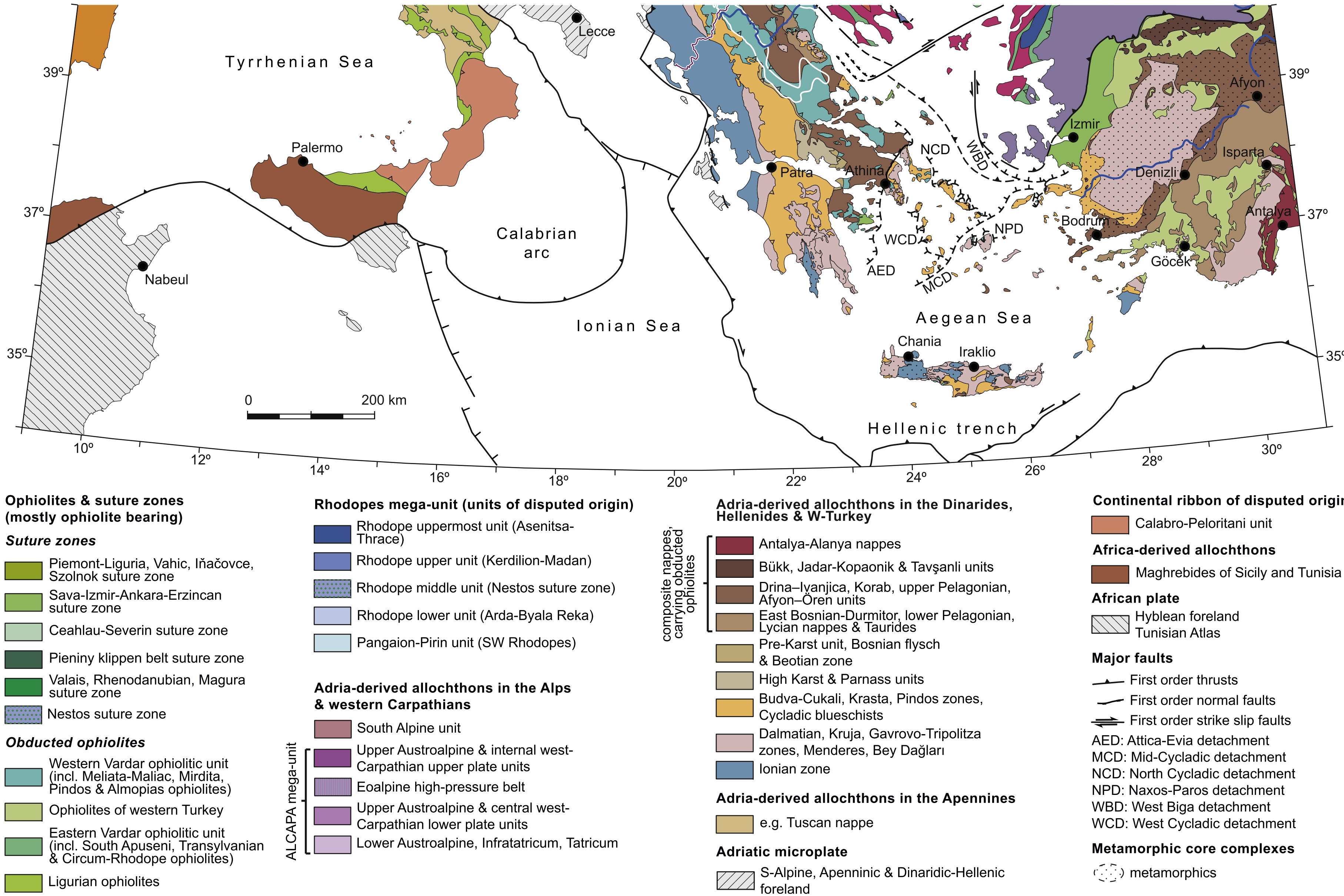

Plate 1. (*continued*).

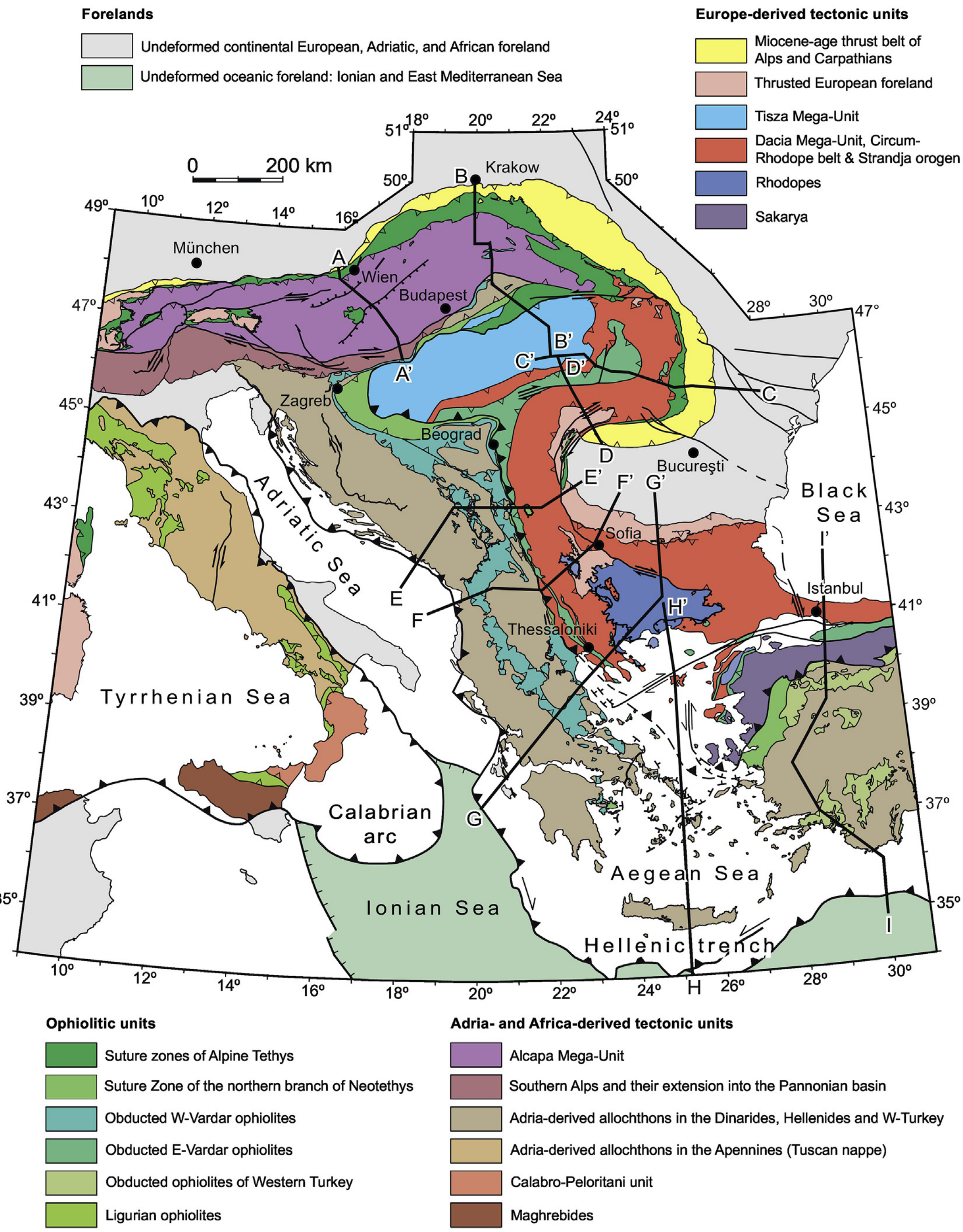

Fig. 1. Simplified overview of the major units showing the traces of the profiles presented in figures.

under the names of the authors. Here we only mention most important large-scale maps that we used. For the Alps we relied on the compilations by Schmid et al. (2004, 2013) and Bousquet et al. (2012). The main map sources used for the Western Carpathians are the 1: 500,000 sheets Geologická mapa Slovenska by Biely et al. (1996) and the Geological map of the Western Carpathians and adjacent areas by Lexa et al. (2000) (online geological map of Slovakia at the scale 1:50,000 available at https://apl.geology.sk/gm50js/). Particularly important for the Hungarian territory is a subsurface map of the Pannonian basin containing information about the pre-Cenozoic formation below the Pannonian basin fill based on numerous drill holes (Haas et al., 2010). The territory of Romania is covered by the impressive Geological Map of Romania 1:1,000,000 by Săndulescu et al. (1978) as well as by numerous cross sections (Ştefănescu, 1988), and the country is completely covered by 1:200,000 maps, and partially, by 1:50,000 maps published by the Geological Institute of Romania. The area of former Yugoslavia is completely covered by 1:100,000 sheets (Osnovna geološka karta SFRJ) of excellent quality. For Greece a 1:500,000 map is available (Bornovas and Rontogianni-Tsiabaou, 1983) and a large part of the country is covered by 1:50,000 sheets. Neighbouring Albania is covered by an excellent 1:200,000 map (Xhomo et al., 1999). A 1:500,000 geological map is available for Bulgaria (Cheshitev et al., 1989), and detailed maps on the scales 1:100,000 are available for the entire territory; more recent ones in 1:50,000 are available for large parts of the country. In Turkey we used the 1:500,000 map sheets Istanbul (Türkecan et al., 2002), İzmir (Konak et al., 2002a), Denizli (Konak et al., 2002b), Zonguldak (Aksay, 2002), Ankara (Turhan and Aksay, 2002) and Konya (Senel, 2002).

A second important data source was of course the study of the existing literature, which, in view of a vast number of publications, can only be selectively cited during the description of the mapped units. Particularly useful was a monograph by Vozár et al. (2010) that contains a wealth of data and references regarding the entire circum-Pannonian region resulting from a longstanding international cooperation. Thirdly, many key areas were visited by numerous reconnaissance-type or in-depth research field campaigns by the authors over more than a decade. Such fieldwork turned out to be absolutely essential for a better understanding and assessment of the available literature data that often offer controversial views and that are of highly variable quality.

The principle aims of the compilation, namely to elucidate the main architecture of the orogenic system between Eastern Alps and western Turkey, has to face the problem that vast areas (e.g. Pannonian basin, Transylvanian basin, Thrace basin) are covered by sedimentary basins that had to be "removed" to reveal the large-scale structural grain. Such basin fills are termed "post-tectonic" in the sense that they unconformably overly major structures of the Alpine-Eastern Mediterranean orogen. Note, however, that often these sediments were later deformed again by displacements associated with basin formation and inversion. The outlines of such major basins, and those of numerous "inselbergs" within them, exposing the underlying units, are schematically represented by white lines on our map (Plate 1). It is needless to mention that the attribution of bedrocks underneath such basins to particular tectonic units and the exact localisation of tectonic boundaries remain rather uncertain in many places.

In case of the Pannonian basin, the post-tectonic sedimentary cover rocks mostly consist of Miocene sediments that reach a thickness of up to some 6 km. This cover obscures crucial first-order tectonic boundaries of mega-units such as Tisza, Dacia and ALCAPA (Csontos and Vörös, 2004; Schmid et al., 2008) However, the sedimentation of the post-tectonic cover of the Pannonian basin above the basal unconformity is older and typically starts with either Upper Cretaceous or Paleogene strata (see overviews given in Horváth et al., 2006; Haas et al., 2010 and 2012). The Upper Cretaceous to Miocene fill of the Transylvanian basin (e.g. Huismans et al., 1997; Brouker et al., 1998; Maţenco et al., 2010; Tiliţă et al., 2013) also obscures many of the more internal units of the Eastern Carpathians and Apuseni Mountains and particularly the Transylvanian ophiolites. There, a wealth of information regarding lithology and structure of the bedrocks buried in the subsurface fortunately became available from hydrocarbon exploration (e.g. Krézsek and Bally, 2006). Large Cenozoic basins such as the East Rhodope basin (Pleuger et al., 2011), the Thrace basin (Elmas, 2012) and the Thermikos basin (Burchfiel et al., 2008) cover substantial areas of the Rhodopes and surrounding units.

Tectonic boundaries were also interpolated between large intrusions. These are very widespread in the Rhodopes where they cover large areas and many of them are associated with extensional events. Their ages vary between latest Cretaceous and Miocene (Burg, 2011).

The construction of crustal-scale cross-sections, in parallel to the map compilation (Plate 1), led to additional insight that significantly improved map compilation and vice versa. Due to the lack of seismic information these crustal scale profiles that accompany the text remain highly schematic and conceptual. The quality of the information regarding Moho depth is highly variable.

3. Overview of the major groups of mapped tectonic elements

The map legend of Plate 1 displays a total of 50 units mapped between Alps and western Turkey. Attention was paid to keep the number of mapped units at a minimum. To gain an overview and facilitate the reading of the map we group these units into 19 groups in a first step. This chapter represents an overview of these major groups of mapped tectonic elements. Note that for practical reasons the headings of these subchapters are not always identical with the categories of tectonic units as displayed in the figure legend of Plate 1. The groups of tectonic units used in the subchapters presented below are displayed in Fig, 1, which is a simplified version of Plate 1. Fig. 1 also displays the traces of the crustal-scale profiles embedded in the text.

3.1. Undeformed foreland

3.1.1. Criteria used to map the front of the Alpine-type orogens of the Eastern Mediterranean

The decision as to where exactly to place the boundary between orogen and undeformed foreland is not always easy and subject to debate. It is well known that, for example, thick-skinned manifestations of deformation propagated far into the European foreland of northern Germany (Ziegler, 1987) in the Late Cretaceous. This coincided with an important change in relative motion between the European and African plates (Kley and Voigt, 2008).

In the case of the Alps and Carpathians, adjacent to the European and Moesian platforms, we chose the frontal-most thin-skinned thrusts onto the flexural foreland basin (foredeep). In the case of the Eastern and Southern Carpathians, we took the projection of the Miocene thin-skinned front, buried at depth, to the earth's surface as marking the boundary (e.g. Ştefănescu, 1988; Maţenco et al., 2010). However, this front is often crosscut by subsequent thick-skinned thrusts and the boundary becomes more problematic westwards in the Southern Carpathians foredeep due to the interplay between the gradually decreasing thrusting and transcurrent motions (e.g. Maţenco et al., 2007; Răbăgia et al., 2011; Krézsek et al., 2013). Furthermore, this boundary is more difficult to trace in the case of the Carpatho-Balkan orogen of eastern Serbia and Bulgaria that either lacks a pronounced foredeep or exhibits a frontal thrust buried at depth.

In the case of the Dinarides-Hellenides, thrusting the Adriatic foreland, it is also problematic to place a clear orogen-foreland boundary. In the area of the Ionian Islands the frontal thrust of the Ionian unit over what is referred to as Pre-Apulian (or Paxi) unit in Greece was taken as the boundary because the latter is only weakly deformed. A decision is also problematic in Albania. The geometry of the frontal thrust zone is also very complex north of the Kephalonia transform zone and a second transform zone NW of Corfu. We took the Sazani unit of Albania (Prifti and Uta, 2012) of the Sazani Peninsula as an "undeformed" area although its Mesozoic cover is tilted and forms the roof of a triangle zone (Roure, 2008). In the area of the Periadriatic depression located in the coastal area of Albania around and northwest of Tirana, characterized by a thick and largely synorogenic accretionary prism of Neogene age deforming the foredeep (Roure et al., 2004), we took the projection of the front of the Mesozoic thrust sheets at depth to the earth's surface as marking the boundary. All the way to Istria, the frontal thrust of the Dinarides runs in the offshore area (Picha, 2002; Korbar, 2009). Below the Po Plain of Italy the orogenic front of Dinarides, Southern Alps and Apennines is well defined by subsurface data (Pieri and Groppi, 1981). Mapping of the frontal thrust of the Apennines onshore and offshore the Adriatic coast and in Sicily are largely after Bigi et al. (1992) and Finetti (2005). The southern limit of the Alpine chain in Tunisia was mapped along the front of the Tell units over the Atlas (Frizon de Lamotte et al., 2000).

Southeast of the Kephalonia transform the southern boundary of the Hellenides and the Taurides of western Turkey is marked by the still active Hellenic trench and its eastward continuation. This trench is not always easy to define and we mapped it largely parallel to the southern boundary of the so-called backstop in the sense of Le Pichon et al. (2002). This backstop consists of a pile of Hellenic nappes that migrated outward from the Aegean fold-thrust belt within the adjacent Mediterranean basins during late middle Miocene. It is adjacent to the still open oceanic lithosphere of the Eastern Mediterranean Ocean of which the Ionian Sea is a part, subducted beneath the backstop and covered by a thin Miocene-age accretionary wedge referred to as Mediterranean ridge (e.g., Kastens, 1991).

3.1.2. European and Adriatic foreland

For the sake of simplicity, the foreland of the European plate was very roughly categorized into two areas. A first one comprises the **Variscan orogen** of Central Europe and the **Moesian Platform** as well as the Scythian platform located north of the **North Dobrogea orogen** (see figure legend of Plate 1). A second one comprises the **pre-Cambrian East European platform** located east of its SW boundary - the Tornquist-Teisseyre line (Ziegler, 1981). Note, however, that this is an oversimplification since parts of the foreland west of the Tornquist-Teisseyre line such as the Małopolska block north of Kraków, in Poland (Bula et al., 1997), and most of the Moesian Platform (Tari, 2005; Seghedi et al., 2005), also remained largely undeformed during Variscan orogeny. On the other hand, the Scythian platform located north of the North Dobrogea orogen and extending to Crimea, i.e. located along and east of the Tornquist-Teisseyre line is considered to have been part of the late Paleozoic Scythian orogen linking the Variscides with the Ural Mountains (Nikishin et al., 2011), in the continuation of significant latest pre-Cambrian to early Paleozoic deformations recorded at the SW margin of the East European platform (Stephenson et al., 2004; Saintot et al., 2006). Additionally, we mapped two smaller foreland areas: The **Alpine-Carpathian foreland basin**, characterized by flexuring of the lithosphere at the front of the orogen, and very pronounced in case of the bending area in the southeastern Carpathians, partially related to the effects of slab pull exerted by the Vrancea slab (e.g. Tărăpoancă et al., 2003 and 2004; Maţenco et al., 2007; Ismail-Zadeh et al., 2012). The second one is the **North Dobrogea orogen** meriting special attention and described below in more detail.

The **North Dobrogea orogen** (Săndulescu, 1984 and 1994; Seghedi, 2001; Segehdi, 2004; Gradinaru, 2000) is delimited to the south by parts of the Moesian Platform (Central and South Dobrogea areas, only weakly deformed during the Variscan orogeny) along the Peceneaga-Camena fault zone. To the north, the Sfântu Gheorghe fault forms the tectonic boundary to the late Variscan Scythian platform. The area is known for the exotic nature of its hemipelagic to pelagic Triassic deposits that contrast with the epicontinental facies Triassic deposits of the surrounding cratons since end of the 19th century (Peters, 1867; Haas et al., 1995), and additionally, for its Early Triassic volcanics. It represents a NE-facing early Alpine intra-continental orogen cutting into the cratonic foreland. The North Dobrogea orogen consists of three major nappes, which are, from SW to NE and structurally higher to lower: the Măcin nappe, Niculiţel nappe and Tulcea nappe.

While some authors (e.g. Stampfli, 2000) claim that the area of the North Dobrogea orogen hosted Early Triassic mid-ocean ridge basalts, geochemical evidence (Saccani et al., 2004) and field observations (Seghedi, 2001) suggest that the Early Triassic volcanic and subvolcanic rocks (Măcin and Nicuţel basalts) originated in a continental intra-plate extensional tectonic setting that also affected the entire length of the Tornquist-Teisseyre line. In the North Dobrogea orogen this rifting was preceded by late Variscan shortening affecting the Pre-Mesozoic basement (Seghedi, 2012). The first contractional pulses during the Alpine cycle initiated with an "Early Cimmerian" event taking place in the Late Triassic to Early Jurassic ("Early Cimmerian tectogenesis" of Săndulescu, 1994) associated with Upper Triassic to Middle Jurassic terrigeneous flysch-type deposits of the Nicuţel and Tulcea nappes. A second phase of shortening ("Late Cimmerian tectogenesis" of Săndulescu, 1994) is mainly responsible for N-directed intense folding and nappe formation and is of latest Jurassic to Early Cretaceous age (Seghedi, 2001; Hippolyte, 2002). It involved verticalization of Oxfordian to early Kimmeridgian strata at the southern contact of the North Dobrogea orogen with the adjacent Central Dobrogea that is considered a part of the Moesian Platform (Gradinaru, 1988). The orogen is sealed by horizontal Albian and younger post-tectonic deposits (Gradinaru et al., 2006). Hence, in the Albian the Moesian Platform, North Dobrogea orogen and Scythian platform were welded together, from that time onward representing a segment of the undeformed foreland of the Eastern Carpathians. There is no important kinematic link of the Eastern Carpathians with the North Dobrogea orogen at the earth's surface but at the lithospheric scale the heterogeneities in the foreland play a major role during the shaping of the bending zone in the southeastern Carpathians (e.g. Cloetingh et al., 2004). Only minor reactivation (up to 35–40 km) of pre-Albian faults (thick black lines in Plate 1) occurred in the Miocene-Quaternary (Tărăpoancă et al., 2003; Maţenco et al., 2003 and 2007; Leever et al., 2006), primarily along the Intramoesian fault, the Peceneaga-Camena fault (southern boundary of the North Dobrogea orogen), the Trotuş fault (within the Scythian platform) and the Bistriţa fault (northern margin of the Scythian platform).

The area mapped as **Adriatic microplate** refers to those parts of a lithospheric plate or "subplate" that remained undeformed by thrusting, including Istria, Gargano, and Puglia. Adria is at present, and has since the early Miocene, been a microplate that rotated slowly relative to Africa accommodated by extension in the Strait of Sicily (van Hinsbergen et al., 2014a; Le Breton et al., 2017). During this time, Adria acted as a rigid indenter during final Alpine collision (e.g. Vialon et al., 1989; Schmid et al., 2017). Opening in the Ionian Sea between northern Africa and Apulia in Triassic (e.g. Speranza et al., 2012) and/or later Mesozoic time (e.g. Frizon de Lamotte et al., 2011) led to the motion of Adria towards the north with respect to Africa (Handy et al., 2010). In fact, the Ionian Sea is part of a larger oceanic domain that we refer to as Eastern Mediterranean Ocean, also called Mesogea by some authors (i.e. Biju-Duval et al., 1977; Berra and Angiolini, 2014), that also contains even older, Carboniferous oceanic crust (Granot, 2016). The Adria microplate acted as a rigid indenter in Cenozoic time and during final Alpine collision, again decoupled from Africa (e.g. Vialon et al., 1989; Schmid et al., 2017). However, note that the term "Adriatic" is also used for denoting the paleogeographical affiliation of structural entities, which originally were part of this Adria continent (Dercourt et al., 1986), termed "Greater Adria" by Gaina et al. (2013). Hence, in this contribution the term "Adriatic" will also be used for denoting the entire paleogeographical realm originally located south of the Alpine Tethys Ocean in the case of the Alps and Western Carpathians, south of the northern branch of Neotethys in the case of the Dinarides, Hellenides and Anatolides-Taurides, and north of the Eastern Mediterranean (or Mesogea) Ocean.

3.2. Europe-derived units

3.2.1. Miocene-age thrust belt of Alps and Western Carpathians

The Miocene thin-skinned thrust belt of the Alps and Carpathians is built up by Cretaceous to Cenozoic flysch units. It is the only structural feature that is common to the Alps and the entire Carpathian chain. This curved belt rims the site of a former partly oceanic embayment thought to consist of the last remnants of the Alpine Tethys (Balla, 1987) and is followed all the way from the Alps to the Southern Carpathians, forming the frontal most thrust sheets (Plate 1 and Fig. 2). These most external parts of the Carpathian fold-and-thrust belt (e.g. Săndulescu et al., 1981a and Săndulescu

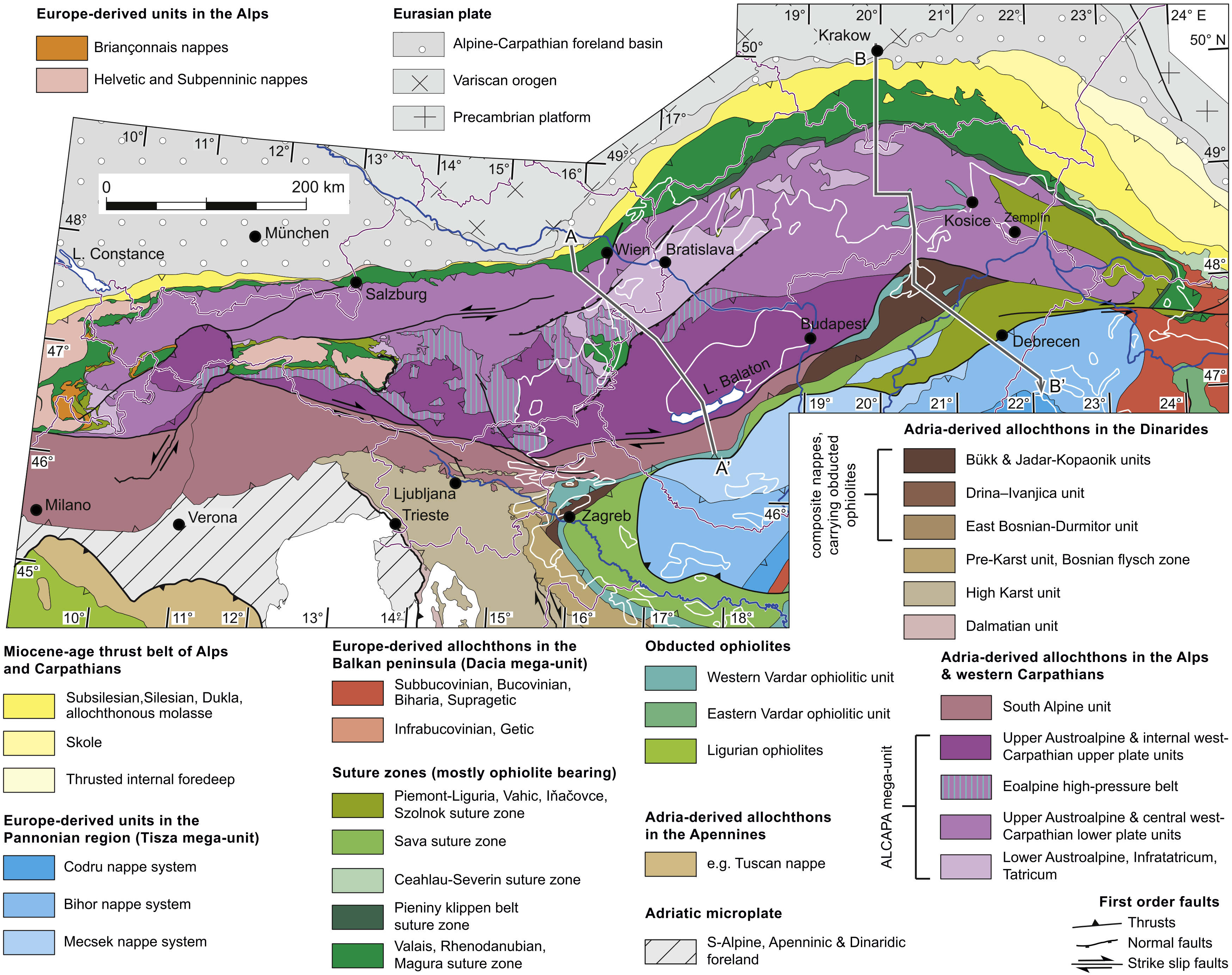

Fig. 2. Tectonic map of the Eastern Alps and Western Carpathians and adjacent areas; extract of Plate 1.

et al. 1981b; Morley, 1996; Maţenco and Bertotti, 2000; Oszczypko, 2006; Ślączka et al., 2006; Golonka et al., 2011 and 2018) formed late during the tectonic history, i.e. predominantly during the Neogene and in the context of the invasion of the ALCAPA and Tisza-Dacia mega-units into the Carpathian embayment. Their soft collision with the European foreland (e.g. Balla, 1987) is thought to have been triggered by a combination of lateral extrusion from the Alps (e.g. Ratschbacher et al., 1991a and 1991b), and more importantly, by the retreat (roll-back) of the subducting European lithospheric slab (in the sense of Royden, 1988 and 1993) leading to major back-arc extension in the Pannonian basin (Balázs et al., 2016). Calc-alkaline and alkaline magmatism in the Pannonian basin is interpreted to be related to subduction and rollback-related extension (Seghedi et al., 2004). The Southern Carpathians contain no major foreland basin but were instead juxtaposed with the Moesian Platform during the Paleogene-Miocene by a combination of strike-slip movements along curved fault systems and oblique thrusting in the context of eastward displacement and dextral rotation of the Dacia mega-unit (Ratschbacher et al., 1993; Răbăgia and Maţenco, 1999; Fügenschuh and Schmid, 2005). This Alps-Carpathians flysch belt was subdivided into a most external group of units referred to as **thrusted internal foredeep,** an intermediate group comprising the **Skole** nappe of Poland and Ukraina and the **Marginal Folds** and **Tarcau** nappes of the Eastern Carpathians, and finally, a most internal group comprising the **allochthonous molasse** in the Alps**,** the **Subsilesian, Silesian** and **Dukla** nappes in the Western Carpathians, and the **Audia**, **Macla, Convolute flysch** nappes of the Eastern Carpathians (see Golonka et al., 2011, Ślączka et al., 2006, Maţenco and Bertotti, 2000 for overviews).

3.2.2. *Allochthons derived from the European foreland of Alps and Western Carpathians*

This very heterogeneous group of tectonic units represents frontal slices or nappes that were derived from the European foreland. In the Alps such units tectonically underlie the ophiolitic remnants of the Piemont-Liguria Ocean, and consist of detached sediments (**Helvetic nappes**) or of metamorphosed basement and cover nappes referred to as **Subpenninic nappes** (Milnes, 1974; Schmid et al., 2004 and 2013). Europe-derived nappes structurally located between the remnants of the Valais and Piemont-Ligurian Ocean are referred to as **Briançonnais nappes** (Figs. 2 and 3). The latter are derived from the Briançonnais continental fragment, a spur of continental lithosphere that split away from Europe during the opening of the Valais Ocean (Frisch, 1979; Schmid et al., 2004; Handy et al., 2010). The Variscan foreland of Corsica occupies a similar paleogeographic position and was hence given the same attribution in Fig. 1 and Plate 1. Units of this group are not exposed in the easternmost Alps and Western Carpathians, but their presence is suspected in the subsurface (Figs. 3 and 4, profiles A and B, respectively), based on the interpretation that some of the sedimentary units found within the Pieniny Klippen Belt must have been derived from Europe-derived continental crust (the conceptual Oravic ribbon; Plašienka, 2018).

Similar allochthons derived from the European foreland that were formerly adjacent to the Moesian Platform are also found in the Southern Carpathians, the Balkan orogen and the Pontides (Plate 1 and Fig. 1) and will be further discussed in chapter 3.2.4.. They are found in the Danubian window of the Southern Carpathians (**Danubian nappes**; Berza et al., 1994; Iancu et al., 2005a and b; Seghedi et al. 2005), in a tectonic window in the Kraishte area of western Bulgaria (**Struma unit**: Kounov et al., 2010) and in the **West Balkan unit** (or Stara Planina unit) of the western part of the Balkan orogen (Kräutner and Krstić, 2006; Vangelov et al., 2013; Burchfiel and Nakov, 2015). The **Kula** and **Forebalkan** units of the Balkan orogen are in a very external position adjacent to the

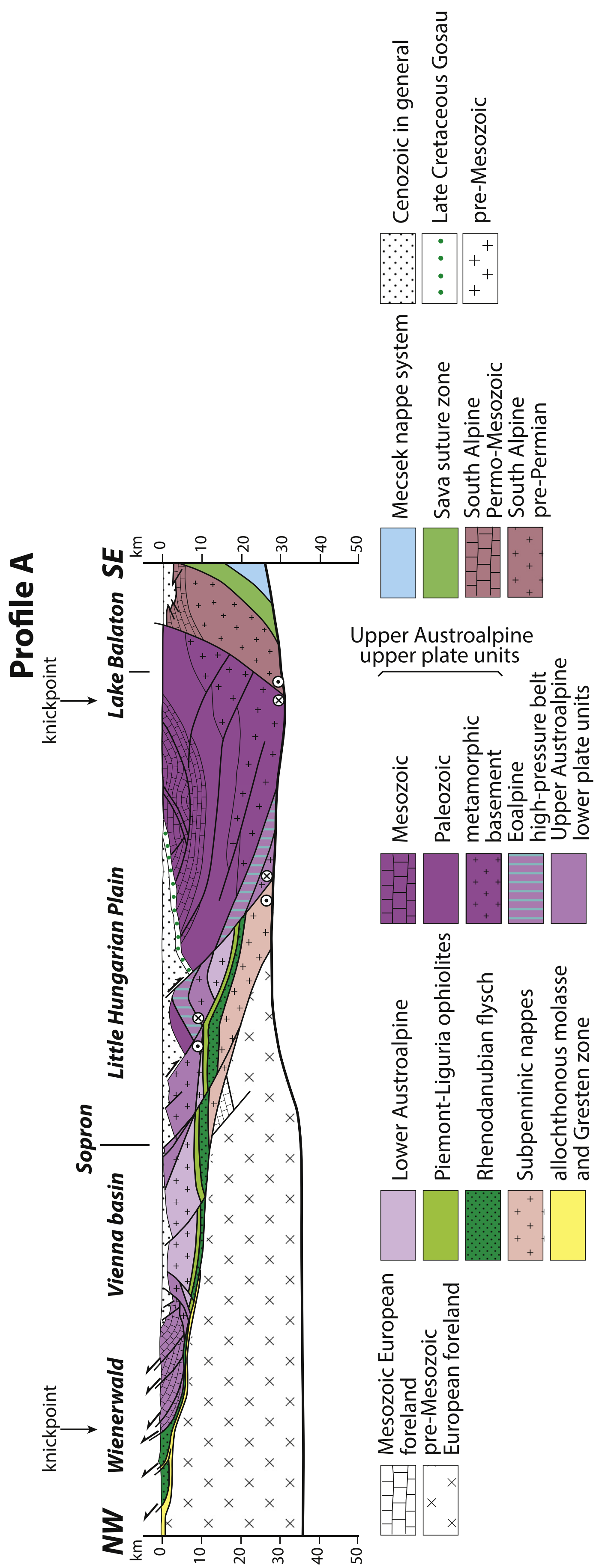

Fig. 3. Crustal-scale profile across the easternmost Alps and the western Pannonian basin (profile A). See Figs. 1 and 2 for the trace of profile A. Profile construction partly based on Wessely (1987), Tari (1994 and 1996), Szafián et al. (1999). Moho depth after Horvath et al. (2006), Hrubcová and Środa (2015).

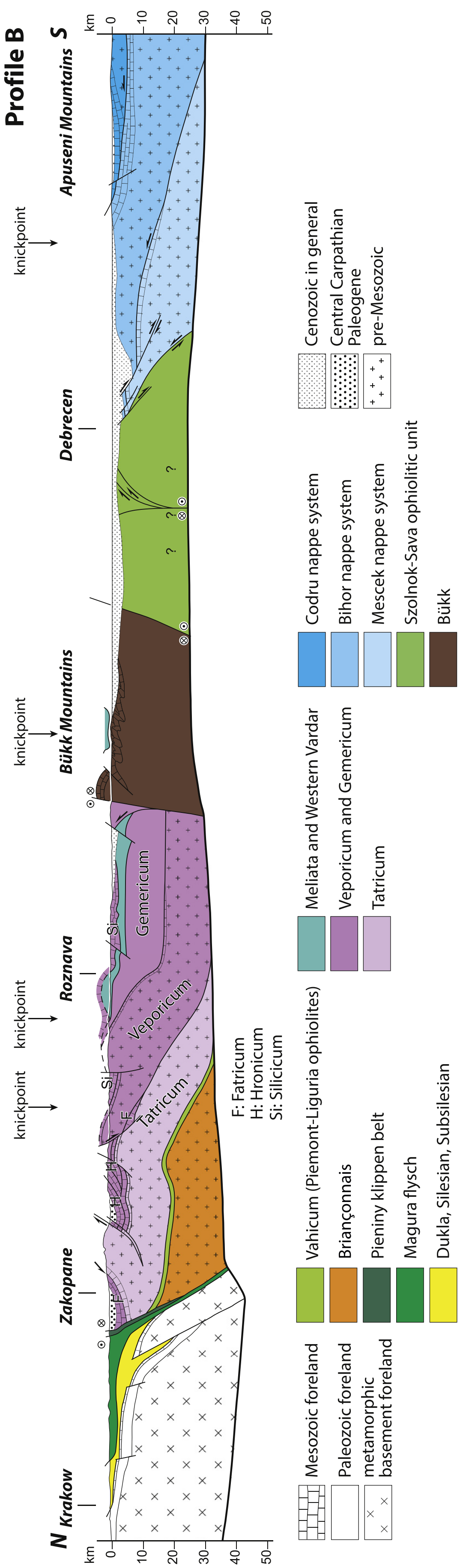

Fig. 4. Crustal-scale profile across the Western Carpathians and the eastern Pannonian basin (profile B). See Figs. 1 and 2 for the trace of profile B). Profile construction partly based on Fülöp and Dank (1987), Tomek (1993), Roca et al. (1995), Plašienka et al. (1997a and b), Sperner et al. (2002), Less and Mello (2004), Windhoffer et al. (2005), Golonka et al. (2018). Moho depth after Horváth et al. (2006), Hrubcová and Środa (2015).

Moesian Platform. According to many authors the **Istanbul unit** of the Pontides in Turkey has been interpreted as a lateral equivalent of the Forebalkan unit that has been laterally displaced during the opening of the Black Sea in the Cretaceous. The thrust fronts of the Forebalkan and Istanbul units were active during the Cenozoic (Okay et al., 1994; Okay et al., 2006; Okay, 2008; Munteanu et al., 2011). However, although Istanbul unit and Moesian Platform show a similar Paleozoic-Mesozoic stratigraphy (e.g. Okay, 2008) the two probably were never contiguous (see van Hinsbergen et al., 2019). Hence, we mapped the Forebalkan and Istanbul units as two separate tectonic units (see Plate 1).

3.2.3. *Tisza mega-unit*

The term mega-unit is widely used in the Pannonian basin and the circum-Pannonian area to denote groups of tectonic units, sometimes also referred to as composite terranes (e.g. Balla, 1987; Csontos, 1995; Csontos and Vörös, 2004; Kovács et al., 2011a and 2011b; Haas et al., 2011 and 2012; Schmid et al., 2008), which underwent contrasting and characteristic sedimentary facies and orogenic evolutions during the Mesozoic. Their final emplacement into the present-day configuration (Plate 1, Figs. 1 and 5) took place during the Paleogene-Miocene, i.e. after their internal structuring that occurred mostly during Cretaceous-age orogeny. This final emplacement was associated with large displacements and paleomagnetically documented rotations (Balla, 1987; Patrascu et al., 1994; Márton, 2000; Ustaszewski et al., 2008). Along the northern boundary of Tisza is the Mid-Hungarian fault zone formed in the Miocene (Csontos and Nagymarosy, 1998; Tischler et al., 2007).

Much of the tectonic units of Tisza (from bottom to top: **Mecsek, Bihor and Codru nappe systems**) are only exposed in isolated and rather small inselbergs within the Pannonian basin (e.g. Mecsek Mountains of Hungary; Haas et al., 2012 or Slavonian Hills in Croatia; Balen et al., 2017) and correlations between the rare outcrops have to be based on the subsurface maps of the Pannonian basin (Haas et al., 2010; Schmid et al., 2008). A coherent pile of nappes is only found in the north Apuseni Mountains of Romania (Balintoni, 1994; Kounov and Schmid et al., 2013; Figs. 4 and 6, profiles B and C, respectively). Note that the highest and most internal nappe system of the north Apuseni Mountains (the Biharia nappe system; Balintoni, 1994), traditionally taken as a constituent of Tisza (Csontos and Vörös, 2004), is attributed to the Dacia mega-unit in our compilation (Fig. 1 and Plate 1) for reasons discussed later.

The nappe stack of basement and cover rocks of the Tisza mega-unit formed in the Albian to Turonian (ca. 110–88 Ma). This is inferred from the age of the youngest deformed sediments and the age of post-tectonic Gosau-type cover (e.g. Haas and Pero, 2004; Schuller et al., 2009), and additionally, from radiometric dating of metamorphic events associated with nappe formation (e.g. Arkai et al., 2000; Lelkes-Felvary et al., 2003 and 2005; Balen et al., 2017; Reiser et al., 2017 and 2019). The direction of tectonic transport in present-day coordinates is top-N to top-NE (Fig. 4, profile B), which after retro-rotation of nearly 90° of clockwise rotation in Cenozoic time (Patrascu et al., 1994; Panaiotu, 1998), is originally top-W to top-NW. Basement and cover rocks rifted and drifted off Europe in the Middle Jurassic along the Alpine Tethys Ocean and the Ceahlau-Severin Ocean whose opening is interpreted as being kinematically linked to that of Alpine Tethys (Fig. 7). The break-away of Tisza from Europe is documented by petrological data and age determinations suggesting similarities in the origin and age of granitoids of the Mecsek Mountains (a part of the Tisza mega-unit) with those of the South and Central Bohemian Batholiths of the European foreland (Klötzli et al., 2004), and additionally, by the facies of the Triassic and Lower Jurassic sequences (Haas and Pero, 2004). After the break-away of Tisza from Europe, the facies of the Tisza Mesozoic cover became comparable to that of the Austroalpine or South Alpine realm; the post-rift sediments, such as

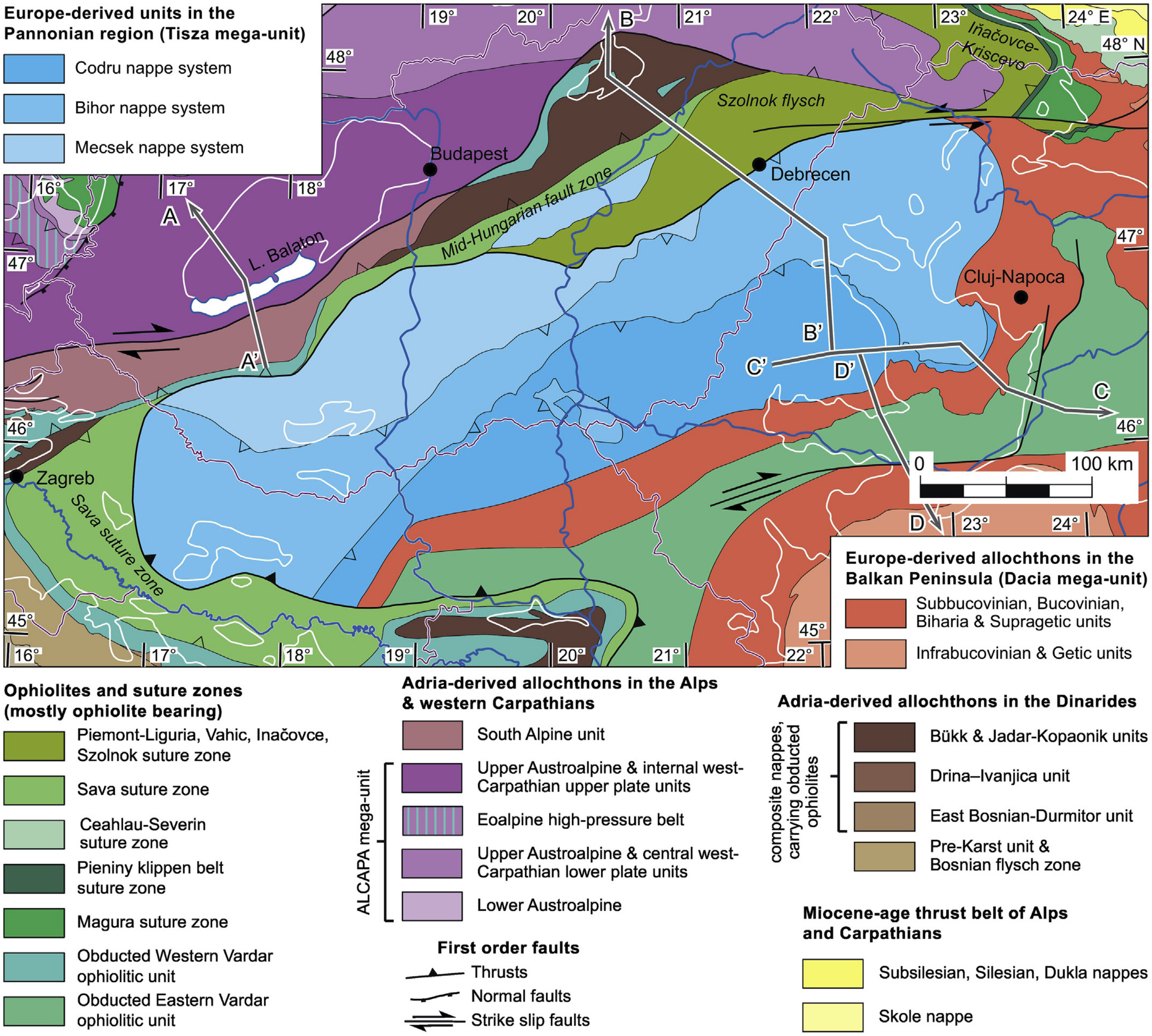

Fig. 5. Tectonic map of the Tisza mega-unit and adjacent areas; extract of Plate 1.

radiolarites and pelagic Maiolica-type limestones, exhibit Adriatic ("Mediterranean" in the sense of faunal provinces) affinities (e.g. Vörös, 1977 and 1993; Lupu, 1984; Haas and Pero, 2004; Schmid et al., 2008).

3.2.4. *Dacia mega-unit, Circum-Rhodope unit, Moesia-derived units and Istanbul unit*

This chapter discusses a group of units that, in a paleogeographic sense, rim the northern and eastern boundary of the northern branch of Neotethys (Fig. 7). The term **Dacia mega-unit** in the strict sense denotes a large block consisting of far travelled Cretaceous-age basement-cover nappes found in the internal Eastern and Southern Carpathians, in parts of the Apuseni Mountains as well as in the subsurface of the Transylvanian basin (e.g. Csontos and Vörös, 2004;), displaced and rotated in the Miocene (Ustaszewski et al., 2008 and references therein) (see Figs. 8 and 9). The Moesia-derived units (**Danubian, West Balkan, Struma, Kula** and **Forebalkan units**, as well as the **Istanbul unit** of the Pontides; see Plate 1) had a different evolution during the Miocene and are hence not part of the Dacia mega-unit in the way originally defined. Also, the **Circum-Rhodope unit** and its lateral equivalent, the **Strandja** unit, located far south of the Pannonian basin and the Moesian Platform are not really part of the Dacia mega-unit and underwent a distinct tectono-metamorphic evolution during the latest Jurassic as will be discussed below.

The Dacia mega-unit in the strict sense corresponds to the Median Dacides as defined by Săndulescu (1994) comprising the **Bucovinian, Subbucovinian**, and **Infrabucovinian** nappes (see Fig. 6, profile C) and lateral equivalents of these nappes in the Southern Carpathians (**Getic** and **Supragetic** nappe systems) (see Fig. 9, profile D). However, parts of this same nappe pile can be followed much further around the oroclinal bend located in the area of the Danube gorge (Săndulescu, 1994; Fügenschuh and Schmid, 2005; Schmid et al., 2008). Across the Iron Gate of the Danube the **Getic nappe system** can be followed into the internal units of the E-W striking Balkan Mountains (**Sredna Gora** and **East**

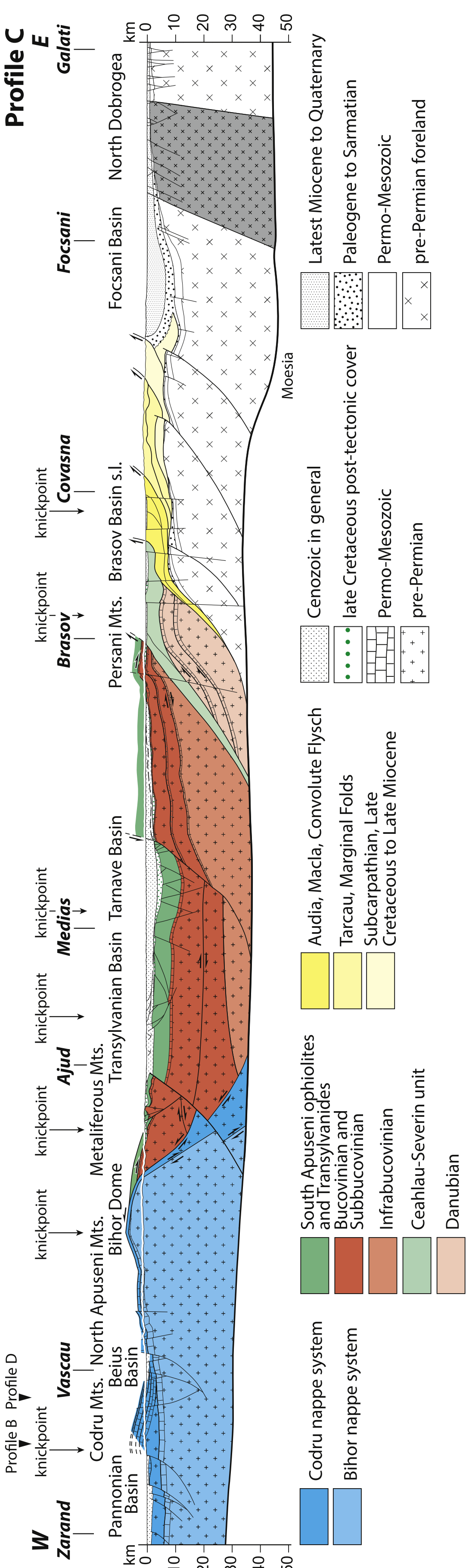

Fig. 6. Crustal-scale profile across the Transylvanian basin and the Eastern Carpathians (profile C). See Figs. 1, 2 and 8 for the trace of profile C. Profile construction largely based on Visarion et al. (1988), Ştefănescu (1988), Maţenco and Bertotti (2000), Bocin et al. (2005), Leever et al. (2006), Maţenco et al. (2007 and 2010). Moho depth after Martin et al. (2005) and Hauser et al. (2007).

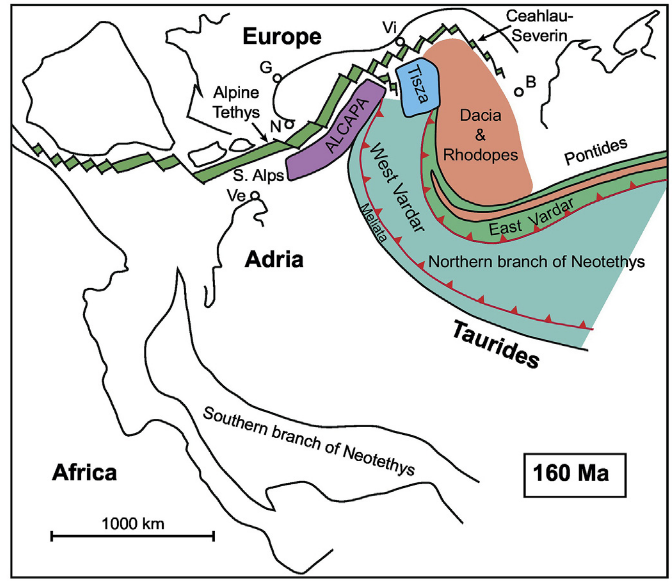

Fig. 7. Paleogeographic sketch map depicting the oceanic domains of the northern branch of Neotethys at 160 Ma ago (after Gallhofer et al., 2017 and van Hinsbergen et al., 2019).

Balkan units of Plate 1 and Fig. 8) and towards the Black Sea (Burchfiel and Nakov, 2015). The Supragetic nappe system, on the other hand, can be followed southwards into the most internal Dinarides and Hellenides all the way to North and Greek Macedonia changing name to **Serbo-Macedonian** unit (Kockel et al., 1971; Kockel and Mollat, 1977a, 1977b) (see Fig. 8).

In the present compilation we grouped the **Circum-Rhodope unit** (Kauffmann et al., 1976) together with the Paikon unit (Brown and Robertson, 2003) since both these units share a common tectonometamorphic event during the Late Jurassic as will be discussed below. The Circum-Rhodope unit almost completely surrounds the Rhodopes (Burg, 2011), together with the **Strandja unit** of southeastern Bulgaria and western Turkey (Okay et al., 2001) we consider as a lateral equivalent (see Figs. 1 and 8). We are aware that this part of our compilation is controversial to some extent because the attribution of the units surrounding the Rhodopes to the west, south and east is much debated in the literature. For example, the southern prolongation of the **Serbo-Macedonian unit** (Vertiskos unit of northern Greece) that actually thrusts the Circum-Rhodope unit from southern Kosovo all the way to Greek Macedonia (Plate 1; Fig. 8) along a Cenozoic W-facing thrust is often taken as the westernmost part of the Rhodopes (e.g. Kydonakis et al., 2014). In a nutshell, the Serbo-Macedonian unit, together with the Circum-Rhodope unit define the western boundary of the Europe-derived units located east of the Sava suture zone that we interpret as representing a Neotethys suture zone between Adria and Europe (Ustaszewski et al., 2010). The units discussed in this chapter taken together comprise all units east of the Sava suture zone and west of the Ceahlau-Severin suture zone of the Southern Carpathians. Below we discuss these units in more detail.

The nappe system in the Eastern Carpathians (Fig. 6; profile C) consists of, from bottom to top, the **Infrabucovinian, Subbucovinian** and **Bucovinian nappes**); the lateral equivalents in the Southern Carpathians are referred to as **Getic** (= Infrabucovinian) and **Supragetic** (= Subbucovinian and Bucovinian) nappes (Fig. 9, profile D); the **Biharia nappe system** of the Apuseni Mountains occupies a tectonic position corresponding to that of the Supragetic nappe system (Schmid et al., 2008 and references therein). The age of stacking of these nappes in the Eastern and Southern

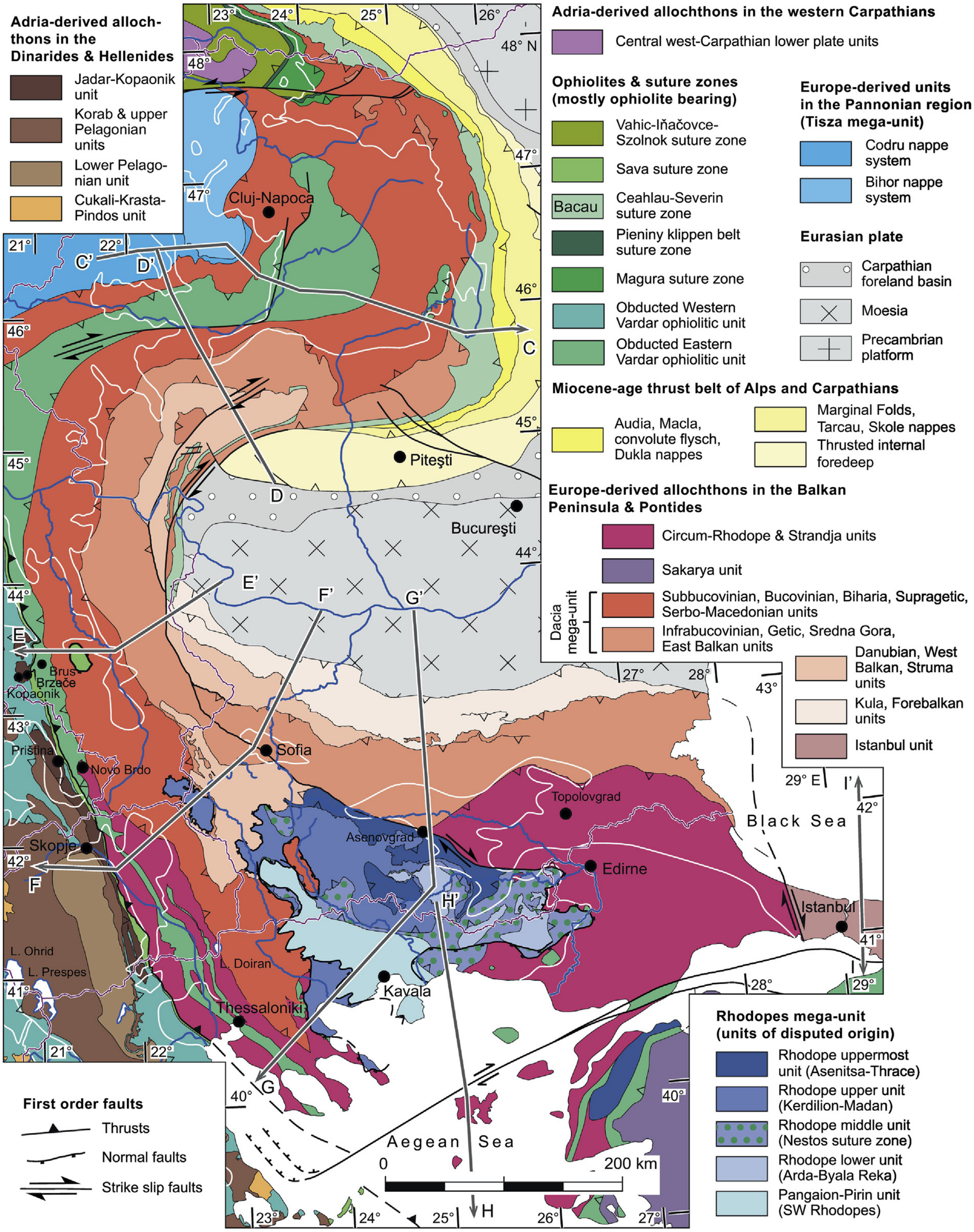

Fig. 8. Tectonic map of the Dacia mega-unit and adjacent areas; extract of Plate 1.

Carpathians is constrained by stratigraphical and radiometric data. Orogeny started with Late Jurassic obduction of the Eastern Vardar ophiolites (e.g. Nicolae and Saccani, 2003; Gallhofer et al., 2017). This was followed by Hauterivian to Barremian (about 130 Ma) orogeny as indicated by the onset of syntectonic sedimentation. It peaked in the Albian and continued until the Cenomanian (between 110 and 90 Ma) as documented by the age of *syn*- and post- tectonic basins (Săndulescu, 1984 and 1994; Kounov and Schmid et al., 2013; Gröger et al., 2013; Olariu et al., 2014). This tectonic event resulted in a second step of emplacement of the Eastern Vardar ophiolites postdating obduction during the latest Jurassic, occupying. During the Lower Cretaceous ("Austrian") orogeny they occupied a highest structural position and where thrusted over a large distance spanning from the Apuseni

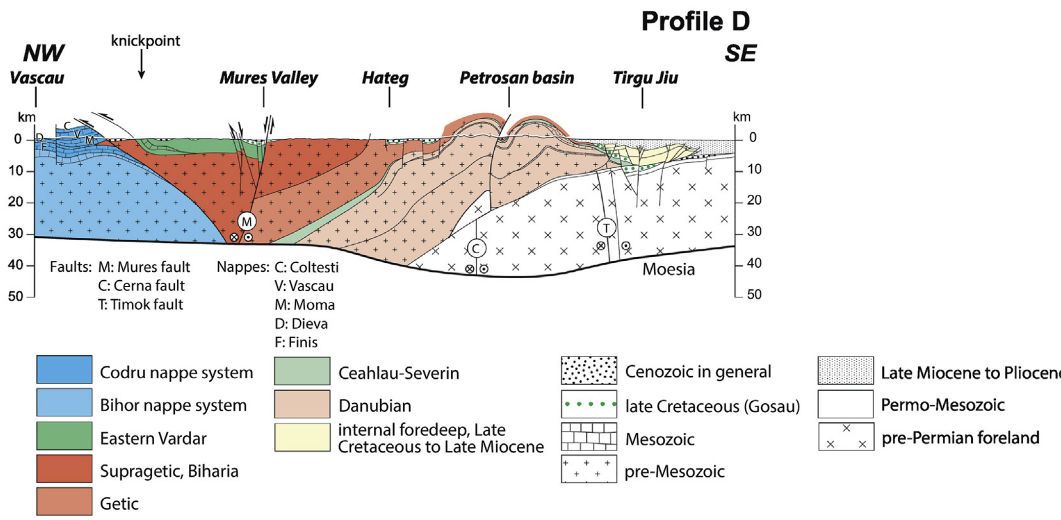

Fig. 9. Crustal-scale profile across the Southern Carpathians (profile D). See Figs. 1 and 8 for the trace of profile D). Profile construction using data from: Ştefănescu (1988); Săndulescu (1989); Fügenschuh and Schmid (2005); Răbăgia et al. (2011). Moho depth after Horváth et al. (2006).

Mountains, over the pre-Miocene basement and cover of the Transylvanian basin all the way to the Eastern Carpathians (Săndulescu and Visarion, 1977; Schmid et al., 2008; Ionescu et al., 2009) (see Fig. 6, profile C). In present-day coordinates, nappe stacking is top-NE to E (Săndulescu, 1994; Reiser et al., 2017) in the Eastern Carpathians and Apuseni Mountains. Since this area underwent the same clockwise rotation by up to 90° in the Miocene, the Cretaceous-age tectonic transport must have been top-NW to N. Evidence based on less precise radiometric data regarding regional metamorphism of the Biharia nappe system in the Apuseni Mountains (Dallmeyer et al., 1999; Reiser et al., 2017 and 2019), locally reaching amphibolite grade complies with the 130–90 Ma time bracket. In summary, the age of initial nappe stacking in the Dacia mega-unit of Romania is older than orogeny in the Tisza mega-unit. Both pre-date a second, less important event of latest Cretaceous (late Campanian - early Maastrichtian) thrusting in the Dacia mega-unit with W-E to NNW-SSE oriented shortening observed in present-day orientations in the Apuseni Mountains, Transylvanian basin, Eastern and Southern Carpathians (Schmid et al., 1998; Maţenco and Schmid, 1999; Schmid et al., 2008). Since Tisza and Dacia dextrally rotated by roughly the same amount in the Cenozoic, thrusting directions within the two mega-units must also have differed in Cretaceous time (originally top-W to top-NW in Tisza; originally top-NW to N in Dacia). The main arguments for attributing the Biharia nappe system of the Apuseni Mountains (e.g. Săndulescu, 1994; Balintoni, 1994) to the Dacia unit mentioned earlier are (see also Kounov and Schmid et al., 2013): (1) timing and thrusting directions in the Biharia nappe system comply with those observed in the rest of the adjacent parts of the Dacia mega-unit (Reiser et al., 2017), and (2), the Biharia nappe system was affected by obduction of the Eastern Vardar ophiolites and associated island arc magmatism in the latest Jurassic (Schmid et al., 2008; Gallhofer et al., 2017) like other parts of Dacia such as the Serbo-Macedonian unit, while the Tisza mega-unit remained unaffected by ophiolite obduction.

As is the case for Tisza, also basement and pre-rift Mesozoic cover of the Dacia mega-unit broke off Europe during the Jurassic rifting and opening of the narrow **Ceahlau-Severin** Ocean, an ocean that can be considered an easternmost branch of Alpine Tethys that became closed at the end of Early Cretaceous orogeny in the Romanian Carpathians (Săndulescu, 1994; Schmid et al., 2008) (see Fig. 7). However, no relics of this ocean in terms of a Ceahlau-Severin suture zone have been recognized in the Balkan Mountains, last relics being preserved in the bending area between Southern Carpathians and the Balkan Mountains (Figs. 1 and 8).

The internal top NE- to E Bucovinian nappe pile of the Eastern Carpathians (Fig. 6, profile C) bends around into the E-W-striking (and now top SSE; Schmid et al., 1998; Maţenco and Schmid, 1999) nappe pile of the Southern Carpathians (**Supragetic** nappe system over the **Getic** nappe system; see Fig. 9, profile D) (Iancu et al., 2005a, 2005b), bending back to the N-S-striking top-E nappe pile observed in the bending zone of eastern Serbia (Krstekanić et al., 2017). In the Kraishte area of eastern Bulgaria the largest part of the Getic nappe system progressively bends into the E-W strike typical for the Balkan Mountains, changing name and referred to as **Sredna Gora** unit in Bulgaria (e.g. Kounov et al., 2018 and references therein). Note however, that top-N thrusting within the Sredna Gora unit and neighbouring areas in the Balkan orogen occurred repeatedly, the main orogenic events taking place in the Early to Middle Jurassic, the late Early Cretaceous, the latest Cretaceous and the Paleogene (Vangelov et al., 2013; Burchfiel and Nakov, 2015). A tectono-metamorphic event associated with top-NE thrusting, starting in the Valanginian and peaking before 112 Ma, is contemporaneous with nappe stacking in the internal Eastern and Southern Carpathians and documented by radiometric dating for the contact area between a small part of the much larger so-called "Morava nappe" (a term used by Kounov et al., 2010) that we regard as a part of the Getic nappe system in this work, and the **Struma** unit in the Kraishte area of eastern Bulgaria (Kounov et al., 2010) that we regard as a part of the Danubian nappe system (Fig. 8).

The Supragetic nappe system, laterally changing name to **Serbo-Macedonian** unit, across the Romanian-Serbian border continues southward into the area west of the Rhodopes, maintaining its NNW-SSE strike into eastern Serbia, North Macedonia and northern Greece (Fig. 8). Ductile deformation and metamorphism in the central segment of the Serbo-Macedonian unit formed during pre-Alpine events (Antić et al., 2017). During Early Cretaceous thrusting of the Serbo-Macedonian unit over the Struma unit (Kounov et al., 2010), the Serbo-Macedonian unit was at a high structural level, Alpine deformation being of brittle nature (Antić et al., 2016). However, for the southern continuation of the Serbo-Macedonian unit in northern Greece, changing name into "Vertiskos unit", Alpine metamorphism and ductile deformation are reported. Metamorphic ages exhibit a very large spread of Late Jurassic to early Late Cretaceous ages (Kydonakis et al., 2016). Ductile deformation is dominantly top-WSW but a Cretaceous age of this top-SW tectonic transport is problematic. Deformation is suspected to be of Cretaceous age by Kydonakis et al. (2015), but reworking in Cenozoic time is more likely (Kilias et al., 1999). The ESE-directed basal thrust of the Serbo-Macedonian unit over the Circum-Rhodope unit is marked by lower greenschist facies mylonites with top-WSW senses of shear exposed in southern Kosovo (near Podgrade; Osnovna geološka karta SFRJ 1:100,000 sheet Vranje by Vukanović et al., 1977) and in Greek Macedonia (basal thrust of the Vertiskos gneiss and underlying Permo-Triassic Examili Formation over the Circum-Rhodope unit west of Lake Volvi; Meinhold et al., 2009). We interpret this basal thrust of the Vertiskos unit over the Circum-Rhodope unit to have formed after the closure of the Sava Ocean in the Cenozoic and during the main phase of nappe stacking in the Hellenides (see Ferrière and Stais, 1995; their Fig. 6).

Following Kauffmann et al. (1976) we extended the **Circum-Rhodope** unit of northwestern Greece, which presently occupies a structural position below the Vertiskos unit but above the **Rhodopes mega-unit** that we consider a mega-core complex, all the way to northeastern Greece (Meinhold and Kostopoulos, 2013; Bonev and Stampfli, 2008) (see Fig. 8). In northeastern Greece they are characterized by Late Jurassic-age top-N thrusting (Bonev and Stampfli, 2011). Similar ages of deformation (latest Jurassic to earliest Cretaceous) and kinematics (top-N) are also described for the **Strandja** unit (Banks, 1997; Okay et al., 2001; Sunal et al., 2011; Bedi et al., 2013; Cattò et al., 2018; Bonev et al., 2019). The Sakar sub-unit (the western part of the Strandja unit) near the SE Bulgarian town of Topolovgrad exhibits a substantially higher degree of Late Jurassic metamorphism, probably due to combined effect of regional metamorphism and pluton driven thermal pulses (Gerdjikov, 2005). The pre-Mesozoic basement of the Strandja unit is dominated by latest Carboniferous to earliest Permian (305 Ma - 295 Ma) granitoids interpreted to represent a magmatic arc that formed north of the N-dipping and still open Paleotethys oceanic slab (Bonev et al., 2019). The Triassic succession of the western part of the Strandja unit (Sakar sub-unit) exhibits amphibolite facies conditions. Neubauer et al. (2010) present $^{40}Ar/^{39}Ar$ amphibole and white mica dating that yields ages ranging between 144 and 136 Ma, constraining the age of the main tectonic event of ductile deformation to the Jurassic/Cretaceous boundary. Earliest Cretaceous sedimentation extended along the whole length of the Circum-Rhodope unit across the north Aegean region and post-dates tectono-metamorphism in the Circum-Rhodope unit s. str. located south of the Strandja unit (Ivanova et al., 2015). Rapid cooling post-dating Late Jurassic orogeny is also inferred for the Strandja unit based on fission track dating (Cattò et al., 2018). The Triassic succession of the lower parts of the Strandja unit are characterized by Germanic-type facies. However, the structurally higher, so-called allochthonous tectonic units exhibit deep marine and often flysch-type Triassic deposits (Chatalov, 1988 and 1990; Tchoumatchenco and Tronkov, 2010; Cattò et al., 2018). Sedimentation terminated in Middle Jurassic (Bathonian) time, presumably due to the onset of the Late Jurassic to earliest Cretaceous top-N nappe stacking referred to as “Neo-Cimmerian” or “Paleo-Alpine” by some authors (e.g. Cattò et al., 2018). Since the Strandja unit exhibits degree and age of metamorphism as well as kinematics (top-N) closely similar to those of the Circum-Rhodope unit in northeastern Greece and southeastern Bulgaria we mapped it as a part of the Circum-Rhodope unit (Plate 1 and Fig. 8). The Ezine Group of the western Biga Peninsula in Turkey (Beccaletto et al., 2005) was taken as the easternmost extension of the Circum-Rhodope unit in our compilation (Fig. 1, Plate 1).

The E-W striking Balkan Mountains (see overviews by Vangelov et al., 2013; Burchfiel and Nakov, 2015) merit a more detailed discussion since they exhibit major along-strike changes in terms of age of deformation, and they are also relevant for a better understanding of the southerly adjacent and enigmatic Rhodope Mountains. The Balkan Mountains have repeatedly been deformed, always with a vergence to the north. Shortening affected the Balkan Mountains as early as during the Late Triassic to Middle Jurassic (“Early Cimmerian event” of Săndulescu, 1994), in the Kimmeridgian-Berriasian, the late Early Cretaceous, the latest Cretaceous to Early Paleogene and finally in the Late Eocene to middle Miocene. The so-called Trojan flysch of the central part of the Forebalkan of Tithonian-Berriasian indicates orogenic activity nearby also during this additional time interval (see discussion below). Shortening was only interrupted in the Late Cretaceous (Cenomanian to Middle Campanian) when a magmatic arc was installed in the area of the **Getic, Sredna Gora and East Balkan units** related to the N-directed subduction of the northern branch of Neotethys. This Apuseni-Banat-Timok-Sredna Gora Late Cretaceous magmatic arc was associated with extensive co-magmatic sedimentary basins (von Quadt et al., 2004 and 2005; Gallhofer et al., 2015 and references therein) that covered most of the pre-Cenomanian deformation features.

There are many different schemes regarding the subdivision of the Balkan orogen into individual units in the Bulgarian literature, most of them being purely geographic, using terms such as Sredna Gora and Stara Planina, or alternatively, West, Central and East Balkan units (e.g. Bergerat et al., 2010; their Fig. 2 drawn after Ivanov, 1988). Regarding our mapping of the Balkan Mountains (Plate 1 and Fig. 8) we made the attempt to follow eastwards the well-defined tectonic units in Romania, eastern Serbia and westernmost Bulgaria, being aware that there are important along-strike changes. Note that this resulted in a nomenclature that departs from that proposed by Ivanov (1988). The Kula and Forebalkan units comprise moderately deformed parts of the Moesian Platform and its sedimentary cover that have no equivalent in Serbia and Romania. The most frontal and small **Kula unit**, limited to westernmost Bulgaria (Kräutner and Krstić, 2006), exhibits folded Upper Cretaceous flysch unconformably covered by upper Neogene cover only documented by boreholes. The westernmost part of the **Forebalkan unit** in Bulgaria was parallelized with the Lower Danubian nappe by Kräutner and Krstić (2006). We do not follow this correlation, because the Danubian nappes of the Southern Carpathians formed at the end of the Cretaceous and are allochthonous with respect to the Moesian Platform, which is not the case for the Forebalkan unit that stratigraphically rests on the basement of the Moesian Platform. The morphologically visible thrust front of the Forebalkan unit, stretching all the way to the Black Sea is formed by thick-skinned thrusts affecting basement and Mesozoic cover of the Moesian Platform, associated with moderate amounts of shortening (10–20 km; e.g. Vangelov et al., 2013; Kounov et al., 2018). They are of late to post-Eocene age since they also affect an angular unconformity at the base of the Eocene, which is testimony of pre-Eocene deformations (Burchfiel and Nakov, 2015). The so-called Trojan flysch of the central part of the Forebalkan consisting of terrigeneous, mostly turbiditic basinal sediments of Tithonian-Berriasian age with a thickness reaching >1 km is a remarkable sedimentological feature (Cherni Osam Formation of Minkovska et al., 2002). Most authors interpret this flysch trough to have resulted from extension in a graben located between the eroding Rhodopes mega-unit in the south and the Getic, Sredna Gora and East Balkan units and the Moesian Platform in the north. However, as suggested by Georgiev et al. (2001; their Fig. 17) we regard a contractional scenario leading to the instalment of a foreland basin in front of the Rhodopes mega-unit and the Strandja unit to be more likely because of (1) the extreme asymmetry of this trough with siliciclastic bodies composed of polygenic conglomerates, coarse-grained sandstones and locally reworked blocks of Late Jurassic reef limestones limited to the southern border of the basin (Minkovska et al., 2002) and (2) the absence of any evidence for a Late Jurassic to Early Cretaceous extensional event in the Rhodopes mega-unit and in the northerly adjacent Getic, Sredna Gora and East Balkan units. Intense deformation of the Trojan flysch is associated with a slaty cleavage and is of latest Cretaceous to early Paleogene age, as inferred from the fact that the unconformity at the base of the Paleogene is much less deformed.

The next higher unit, referred to as **West Balkan unit** (Bergerat et al., 2010), is regarded as an equivalent of the Danubian nappes and restricted to western Bulgaria. It cannot be traced farther east into the Central Balkan area (Plate 1). Hence, it laterally corresponds to the Danubian nappes of the Southern Carpathians, displaced by the N-S-striking Timok left-lateral strike slip fault during the invasion of the Dacia mega-unit into the Carpathian embayment (Fügenschuh and Schmid, 2005; Kräutner and Krstić, 2006) (see Fig. 8). Shortening resumed in this Bulgarian equivalent of the Danubian nappes in the Cenozoic, partly in connection with the Cenozoic activity of the Timok fault. The eastern end of the Danubian unit in the central part of the Balkan Mountains is cut by the frontal thrust of the next higher unit, the Sredna Gora and East Balkan units, which we parallelize with the Getic nappe system of Romania and eastern Serbia.

The frontal thrust of the **Sredna Gora** and **East Balkan units** originally formed in the Early Cretaceous, as is known from its lateral equivalent, the Getic nappe system of eastern Serbia (e.g. Kräutner and Krstić, 2006) and southern Romania as described earlier. However, it was reactivated during Cenozoic shortening in the case of the Balkan orogen. Thick-skinned Cenozoic thrusting is spectacularly exposed in the Botev Vrah Mountain of the central segment of the Balkan Mountains (Balkanska and Gerdjikov, 2010; Kounov et al., 2018). A thrust brings pre-Mesozoic basement of the Sredna Gora unit over Paleocene strata in the footwall that in turn unconformably overlie previously Mesozoic strata of the Forebalkan unit. Eastward this Botev Vrah thrust changes geometry, bifurcating in a number of smaller thrusts and eventually changing from a thick-skinned to a thin-skinned style typical for the eastern Balkan orogen (Georgiev et al., 2001; Vangelov et al., 2012). In the eastern Balkan orogen the Cenozoic thrust at the base of the East Balkan unit, not distinguished from the Sredna Gora unit in Plate 1 and Fig. 8, appears to gradually accommodate more displacement and amounts to some 30 km near the Black Sea (Doglioni et al., 1996; Sinclair et al., 1997; Georgiev et al., 2012).

A belt of Early to Middle Jurassic flysch, known as "Kotel belt" (Georgiev et al., 2001) or "Kotel zone" (Burchfiel and Nakov, 2015) is testimony of Early to Middle Jurassic tectonics. This flysch belt, located immediately behind the thrust front of the East Balkan unit over the Forebalkan unit, is unconformably transgressed by Cenomanian deposits of the Sredna Gora co-magmatic sedimentary basin that is part of the Apuseni-Banat-Timok-Sredna Gora Late Cretaceous magmatic arc. This basin grades into the Campanian-Maastrichtian Emine flysch and overlying Paleogene flysch that forms a piggy-back basin behind the thrust front of the East Balkan unit over the Forebalkan unit (Georgiev et al., 2001; their figs. 3 and 4). The Kotel Formation of the Kotel belt is of Aalenian to middle Bathonian age and built up by dark or black shales containing many Triassic and Jurassic olistoliths, and it is considered a typical "wildflysch". It stratigraphically overlies Upper Triassic to Toarcian siliciclastic flysch of the Sinivir Formation (Tchoumatchenco et al., 2004 and 2009; Budurov et al., 2004; Tchoumatchenco and Tronkov, 2010). Tchoumatchenco et al. (2004) analysed a series of N-facing folds in the Luda Kamchia valley region that they consider as elements of a larger thrust sheet transported northward in pre-Late Cretaceous, most probably Jurassic time. This is supported by close stratigraphical similarities with units of similar age within the allochthonous Strandja unit characterized by Jurassic tectonics (Tchoumatchenco and Tronkov, 2010). Hence, it is likely that the Strandja unit was directly connected with the East Balkan orogen in Jurassic time, the connection being now buried by the extensive Upper Cretaceous to Paleogene post-tectonic cover with respect to Jurassic deformation, as was proposed by Gočev (1986). The North Dobrogea orogen exhibits a similar stratigraphy (Tchoumatchenco et al., 2006). It is also characterized by Late Triassic to Early Jurassic (early Cimmerian) N-directed thrusting (Săndulescu, 1984 and 1994). During the Late Triassic numerous folds also formed in the intervening Moesian Platform, where fault-bend folds involve various Paleozoic decollement levels, locally inverting (Permo-) Triassic grabens filled with *syn*-rift magmatic rocks (Paraschiv, 1978; Georgiev, 1996; Tari et al., 1997). In summary, the Circum-Rhodope unit including the Strandja unit, the Sredna Gora early Cimmerian orogen with its Kotel Formation, the mildly deformed Moesian Platform and the North Dobrogea orogen together represent the frontal part of the Mediterranean Cimmerides, northwards propagating into the European foreland (Tari et al., 1997; Banks, 1997). Note that a major along-strike change occurs eastwards towards Crimea where deposition of similar flysch and thrusting at this same time is S-directed (Nikishin et al., 2015).

3.2.5. Rhodopes mega-unit

The gneissic massif of the Rhodope Mountains (see Burg, 2011 for a review) undoubtedly comprises the most enigmatic group of tectonic units in the Balkans for many reasons. First of all, different authors draw its lateral boundaries differently. Secondly, the Rhodopes occupy an intermediate position between the NE to N-facing Sredna Gora and East Balkan units and Forebalkan units located to the north and the SW to S-facing Dinarides-Hellenides (Plate 1, figs. 1, 10 and 11) located to the south. Accordingly, Kober (1928) referred to them as a "Zwischengebirge". Some authors (e.g. Ricou et al., 1998; Burg, 2011), returning to the original concept of Kossmat (1924), considered also the Serbo-Macedonian unit (called Vertiskos unit in Greece) a part of the Rhodopes because it is also located east of the Vardar unit as originally defined by Kossmat (1924). However, when coming southward from the Romanian and Serbian Carpathians it becomes clear that the Serbo-Macedonian unit is an integral part of the Carpathians (i.e. part of our Dacia mega-unit). Froitzheim et al. (2014) additionally considered the Circum-Rhodope unit as a part of the Rhodopes mega-unit.

In our view the Serbo-Macedonian unit, together with the Circum-Rhodope unit, forms the hanging wall of a first-order extensional detachment, namely the Kerdilion detachment (Fig. 10) formed in the middle Eocene (ca. 42 Ma) to latest Oligocene (ca. 24 Ma) (Brun and Sokoutis, 2007; Kounov et al., 2015). Farther north, in SW Bulgaria much of the western boundary of the Rhodopes mega-unit is formed by the Miocene (24–12 Ma) Strymon detachment (Dinter, 1998) that, in map view (Fig. 10), cuts the older Kerdilion detachment and exhumes the central parts of the SW Rhodopes (Georgiev et al., 2010). This, together with the observation that the Circum-Rhodope and Strandja units, i.e. units that we consider as parts of the Dacia mega-unit, tectonically overlie the Rhodope massif along top-N Jurassic-age thrusts (Bonev and Stampfli, 2011) later overprinted by Cenozoic extension starting in Paleocene time (Bonev et al., 2013), leads us to define the Rhodopes mega-unit as a Cenozoic mega-core complex exposing originally deep-seated units of disputed origin. The northern boundary of the Rhodopes mega-unit is formed by the Late Cretaceous (Campanian) dextrally transpressive Maritsa strike slip fault zone (Naydenov et al., 2013; Georgiev et al., 2014; Fig. 10) separating the Sredna Gora unit (part of the Dacia mega-unit) to the N from the Thrace tectonic sub-unit (Naydenov et al., 2013) representing the northernmost part of the Rhodopes mega-unit to the south. The discrete jump in grade of Alpine metamorphism across the subvertical Maritsa shear zone from greenschist to lower

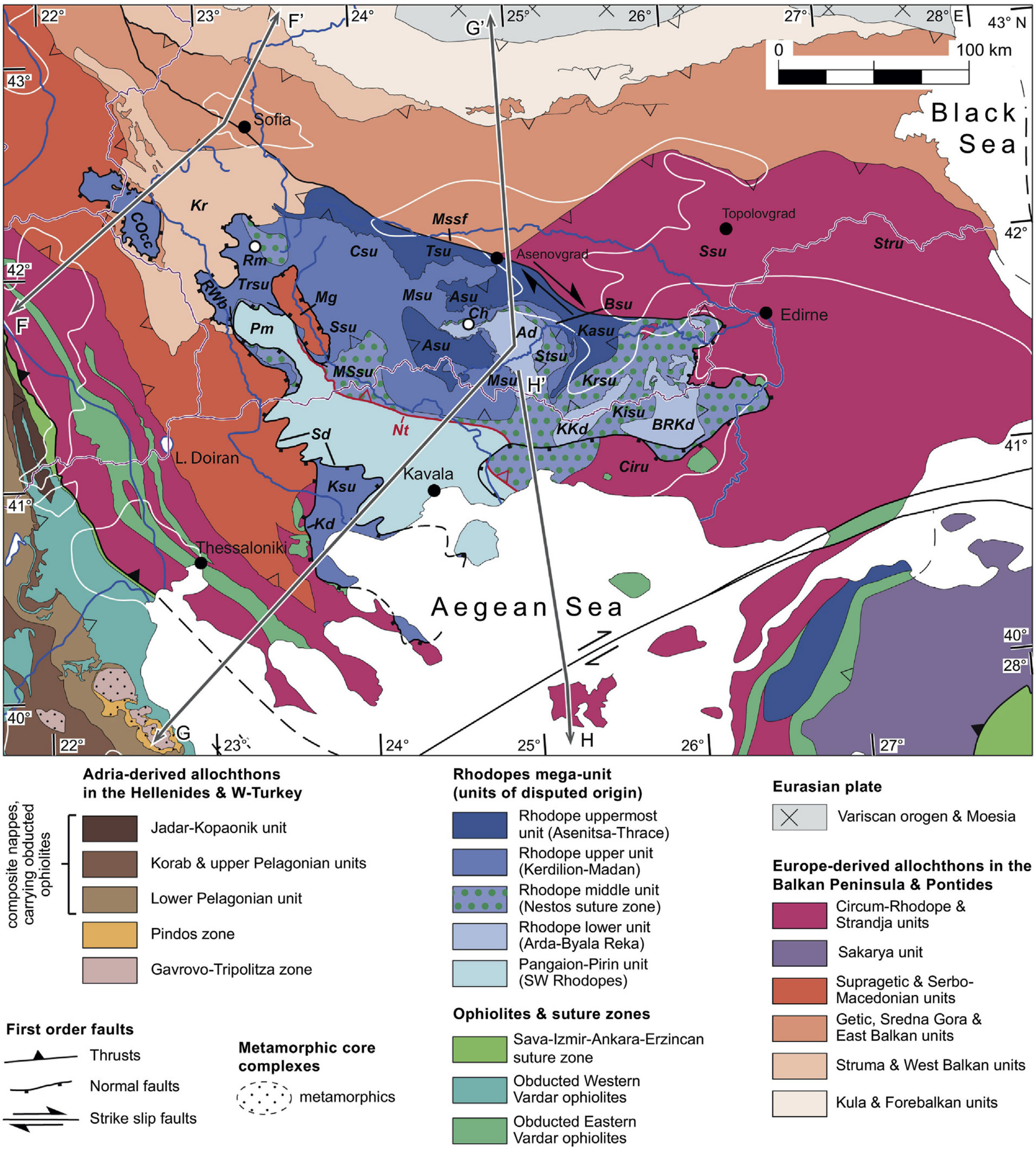

Fig. 10. Tectonic map of the Rhodopes mega-unit and adjacent areas; extract of Plate 1. Abbreviations are as follows: Ad: Arda dome; Asu: Asenica sub-unit; BRKd: Byala Reka-Kechros dome; Bsu: Borovitsa sub-unit; Ch: Cepelare; Ciru: Circum-Rhodope unit; Csu: Chepinska sub-unit; COcc: Crnook-Osogovo-Lisets core complex; Kasu: Kardžali sub-unit; Kd: Kerdilion detachment; Kisu: Kimi sub-unit; KKd: Kesebir-Kardamos dome; Kr: "Kraishte" units; Krsu: Krumovitsa sub-unit; Ksu: Kerdilion sub-unit; Mg: Mesta half-graben; Mssf: Maritsa strike slip fault zone; MSsu: Mesta and Slashten sub-units; Msu: Madan sub-unit; Nt: Nestos thrust; Pm: Pirin Mountains; Rm: Rila monastery; RWb: Rila-West Rhodopes batholith; Ssu: Sarnitsa sub-unit; Stsu: Starcevo sub-unit; Sd: Strymon detachment; SSu: Sakar sub-unit; Stru: Strandja unit; Trsu: Troskovo sub-unit; Tsu: Thrace sub-unit.

amphibolite grade prevailing in the "Thrace lithotectonic unit" of the northernmost Rhodope mega-unit as defined by Naydenov et al. (2009) to non-metamorphic conditions in the northerly adjacent Sredna Gora unit is only partly due to the N-dipping so-called North Rhodopean detachment (Naydenov et al., 2013). This suggests that this Late Cretaceous fault located north of the above-mentioned detachments also accommodated a substantial vertical offset differentially exhuming the northernmost unit of the Rhodope mega-unit (our Rhodope uppermost unit, Plate 1 and Fig. 10) with respect to the Dacia mega-unit along the Maritsa strike slip fault. The northern corner of the Rhodope mega-core complex is again bounded by various Paleogene normal faults that juxtapose what we consider a Danubian unit (the so-called "Kraishte units" of Kräutner and Krstić, 2006) in the hanging wall with the Rhodopes core complex in the footwall. We propose that the smaller and isolated Crnook-Osogovo-Lisets core complex (Kounov et al., 2004; Antić et al., 2016) located NW of the NW corner of the Rhodopes mega-unit should also be considered as a part of the Rhodopes mega-unit (Plate 1; Figs. 1 and 10). This core complex rapidly cooled as a result of exhumation between 48 and 35 Ma (Kounov et al.,

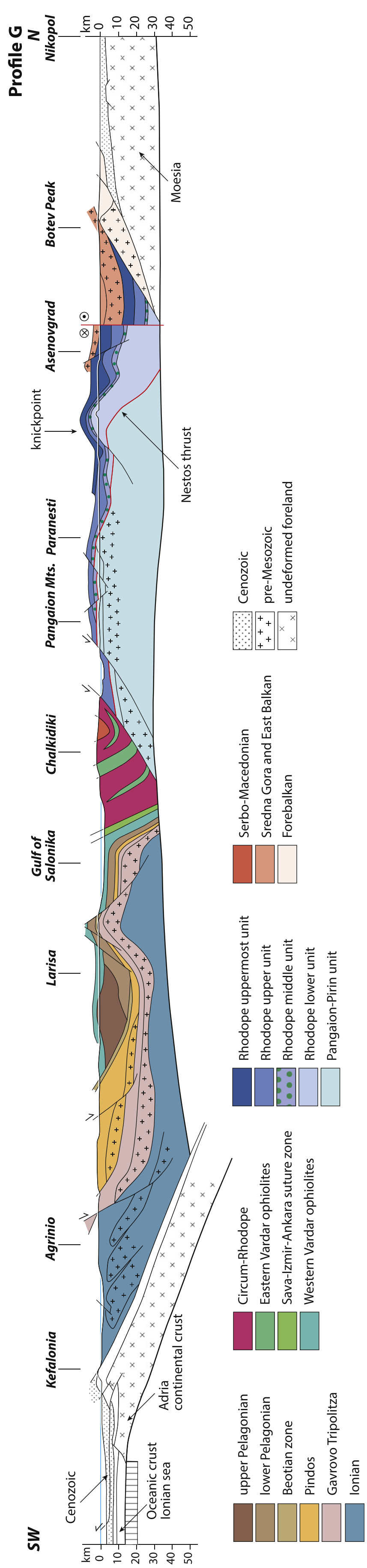

Fig. 11. Crustal-scale profile across the central Hellenides, Rhodopes and Balkan orogen (profile G). See Figs. 1, 10 and 14 for the trace of profile G. Profile construction based on Bornovas and Rontogianni-Tsiabaou (1983), 1:50'000 map sheets of Greece, 1: 50'000 map sheets of Bulgaria and a profile across the southern Rhodopes by Ivanov (1981). Moho depth after Finetti (2005), Sachpazi et al. (2007), Grad et al. (2009), Pearce et al. (2012), and Delipetrov et al. (2016).

2004). This roughly coincides with the timing of the onset of exhumation along the Kerdilion detachment (42 Ma "at the latest"; Kounov et al., 2015). The onset of cooling in the Eocene is also indicated by the deposition of Upper Eocene sediments in the Mesta half-graben (Burchfiel et al., 2003), the synkinematic intrusion of ca. 40 Ma old granitoids intruding the predominantly Cretaceous-age granitoids of the Rila-West Rhodopes batholith (Peytcheva et al., 2007) and top-SW extensional exhumation at the southern margin of this batholith (Balkanska and Dimov, 2009).

In summary, the available structural and fission track data indicate that the Rhodopes mega-unit represents a mega-core complex, extension starting as early as around 48 Ma ago and ending at around 12 Ma ago. As discussed by Kounov et al. (2018) extension in the eastern Rhodopes could even have started earlier, i.e. at the end of the Cretaceous, inferred from the presence of Maastrichtian and Paleocene to early Eocene age sediments unconformably overlying some of the metamorphic units in the eastern part of the massif (Goranov and Atanasov, 1992; Boyanov and Goranov, 1994; Dimitrova et al., 2001).

The internal structure of the Rhodopes mega-unit is of bewildering complexity and evolved over a long period of time, starting during or before the Jurassic and ending with core complex formation from the Paleogene onward. The Rhodopes mega-unit contains a series of stacked nappes, although the number as well as the direction of emplacement is debated. These nappes include imbricated slivers of eclogitic meta-ophiolites (e.g. Froitzheim et al., 2014) and relicts of a suspected Jurassic magmatic arc (Turpaud and Reischmann, 2010). In view of the predominantly top-S to SW shear senses recorded by the gneisses deformed under amphibolite-facies conditions of medium pressure type metamorphism most authors consider the Rhodopes mega-unit basically as a S- to SW verging orogen (e.g. Burg, 2011). It has to be noted, however, that not all sense of shear criteria indicate top-S or SW-ward shearing (Burg, 2011). In fact, many of the top-S to SW senses of shear are associated with extension rather than nappe imbrication. Radiometric data on HP and UHP metamorphism in the Rhodopes mega-unit, if one takes the geological interpretations at face value, suggest multiple subduction-related events (at around 200 Ma, 150 Ma, 120 Ma, 100–50 Ma and 50-40 Ma; Miladinova et al., 2018; their Fig. 9). We find it difficult to accept such multiple events of subduction in such a thin nappe stack and cast doubt on the validity of the geological interpretation of some of the isotopic data.

Concerning the paleogeographic affiliation of the eclogitic meta-ophiolites there are two options. A first option is to attribute them to the Meliata-Maliac-Vardar ophiolitic units issued from the northern branch of Neotethys (e.g. Jahn-Awe et al., 2010; Froitzheim et al., 2014). The second alternative option regards them as issued from a separate oceanic tract located north of the Neotethys Ocean, named the Nestos Ocean (e.g. Turpaud, 2006), possibly representing a part of the Paleotethys Ocean (Turpaud and Reischmann, 2010). This ocean would separate a southerly adjacent continental block (Drama block of Ricou et al., 1998; "Thracia terrane" of Turpaud and Reichmann; Pangaion-Pirin unit of Plate 1 and Fig. 10) from the southern margin of Europe prior to the Middle Jurassic. According to this second option the Drama block (Pangaion-Pirin unit of Plate 1) could represent a Cimmerian terrane (Şengör and Yılmaz, 1981) that drifted away from Gondwana in the Triassic, closing the Paleotethys (Nestos) Ocean and opening Neotethys behind this NW-ward drifting terrane.

The tectonic scenario outlined in the following is built on the second option. It implies that the Rhodopes mega-unit, for Late Jurassic and younger times, entirely consists of Europe-derived units that were located north of the northern branch of Neotethys, together with the originally overlying Dacia mega-unit. We are fully aware that this scheme departs from subdivisions by previous authors (e.g. Burg, 2011; Froitzheim et al., 2014). The proposed scheme is favoured by the first author (and also by the companion paper by van Hinsbergen et al., 2019). But it is not the preferred one for all co-authors. It is based on ideas and hypotheses that certainly need to be tested by future investigations. It rests on two working hypotheses: (1) The interpretation that the Rhodopes mega-unit represents the underpinnings of the nappes of the Balkan orogen that were exhumed in a Cenozoic mega-core complex, and (2) the postulate that, in analogy to the originally overlying Balkan orogen and the Circum-Rhodope unit, N-directed rather than S-directed shearing was active during pre-Cenozoic stages of orogeny in the Rhodopes mega-unit. In summary, we place the entire Rhodopes mega-unit north of the suture zone of the northern branch of Neotethys (Sava-İzmir-Ankara-Erzincan suture zone).

However, top-S thrusting in the Rhodopes mega-unit is unequivocally documented for Cenozoic time (Jahn-Awe et al., 2010 and references therein) as will be discussed later, implying a change from the original top-N transport prevailing in pre-Cenozoic time. The exact timing for the onset of top-S tectonic transport that overprints older top-N tectonic transport in the Rhodopes mega-unit according to the chosen working hypothesis presented above is difficult to assess. The existence of the Apuseni-Banat-Timok-Sredna Gora Late Cretaceous magmatic arc installed at around 92 Ma (Gallhofer et al., 2015) above the N-dipping Aegean slab allows for a rough estimate. Some 100's of km of slab must have been formed to trigger the onset of magmatism in this magmatic arc (see discussion in Gallhofer et al., 2015). This means that N-directed subduction possibly linked to top-S thrusting must have initiated perhaps some 10 Ma earlier at the latest, given Africa-Europe convergence rates of 2.5 cm/a for the time period around 90 Ma (e.g., van Hinsbergen et al., 2019). Hence, in view of the fact that the Circum-Rhodope and Strandja units were thrust towards north in the Early Cretaceous, a change from top-N to top-S thrusting in middle Cretaceous time at around 100–110 Ma (Albian) appears to be a reasonable estimate.

In the following we feel it necessary to discuss the subdivision of the Rhodopes mega-unit into individual tectonic units mapped in Plate 1 in more detail. This is because our subdivision not only departs from previously proposed schemes (e.g. Burg, 2011; Froitzheim et al., 2014) but will remain controversial for some time given the complexity of the geology of the Rhodopes Mountains.

Top S to SW thrusting is well documented and indisputable in the case of the Cenozoic Nestos thrust (Nagel et al., 2011; Fig. 10). This thrust is interpreted as a backthrust relative to the rest of the nappe stack of the Rhodopes mega-unit and is associated with a minimum estimated transport distance of some 70 km (Fig. 11), and it formed later than 55 Ma (Jahn-Awe et al., 2010) but before the onset of extension in the Rhodopes mega-unit some 42 Ma ago (Kounov et al., 2015). The **Pangaion-Pirin unit** of the SW Rhodopes Mountains, as mapped in Plate 1 and Fig. 10, represents the footwall of the Nestos thrust fault. We distinguish this Nestos thrust fault from the term “Nestos suture zone” (Turpaud and Reischmann, 2010) or “Nestos shear zone” (Krenn et al., 2008, 2010) that denotes a suture zone of finite width and up to >1 km thick. We specifically mean that the Nestos thrust is a Cenozoic top-S to SW thrust that cross cuts an earlier top- N to NE-facing Jurassic orogen and associated nappe stack (see profile G of Fig. 11). In map view (Fig. 10) the Nestos thrust follows the base of the Nestos suture zone in the sense of Turpaud and Reischmann (2010) and overlies the Pangaion-Pirin unit. Further north it is interpreted to root below the **Rhodope lower unit** (Arda dome; Figs. 10 and 11). The Nestos shear zone defines what we map as **Rhodope middle unit,** extending into the area north of the Pangaion-Pirin unit (see Fig. 10 and discussion below). This Nestos suture zone is interpreted to represent a suture zone of Paleotethys as will be discussed below).

The lower pressure (greenschist to amphibolite facies) **Pangaion-Pirin unit**, also called Drama continental element (Ricou et al., 1998; Nagel et al., 2011), is exposed in the southwestern part of the Rhodopes core complex where it forms the footwall of the Nestos thrust along its northeastern margin and the footwall of the Strymon detachment along its southwestern margin (Fig. 10). The top-S to SE thrusting along the Nestos thrust goes hand in hand with deformations that postdate the closing of the Sava Ocean and took place during the Paleogene (post-56 Ma according to Jahn-Awe et al., 2010). However, this thrusting predated the onset of core complex extension at or after 48 Ma (Kounov et al., 2004). Note that an analogous change in thrusting direction also affected the originally NE-facing Serbo-Macedonian and Circum-Rhodope units of North Macedonia and Greek Macedonia and led to the presently NE-dipping orientation of tectonic contacts in this region, as will be described later (see Ferrière and Stais, 1995; their Fig. 6; see profile G of Fig. 11).

Some authors (e.g. Jahn-Awe et al., 2010) consider the Pangaion-Pirin unit, characterized by a very prominent and thick sequence of presumably Mesozoic marbles, as a window of the (Greater) Adria-derived units, equivalent to those exposed the Olympos window, thrusted by the rest of the Rhodopes mega-unit north of the Nestos thrust in Cenozoic time over several hundreds of kilometres. As explained above, we prefer the view that it is part of a separate nappe stack that formed on the European side of the Neotethys (Sava) Ocean.

Most authors agree that the hanging wall of the Cenozoic Nestos thrust contains a ‘mixed unit’ (our **Rhodope middle unit** that we, following Turpaud and Reischmann, 2010, consider to represent the **Nestos suture zone**, see Plate 1). This zone is made up of very high-grade rocks, amongst them ultra-high-pressure rocks including meta-ophiolitic sequences. Eclogitic remnants were metamorphosed in the Early Jurassic according to zircon U—Pb geochronology (Bauer et al., 2007). Based on zircon textures, zoning and chemistry, these authors suggest that the high-temperature peak associated with the ultra-high-pressure event occurred at or before ca. 160 Ma and that the zircons were disturbed by later events at around 115 and 79 Ma. The Rhodope middle unit contains a lithologically heterogeneous stack of various lithologies that include metamorphosed slices of an ophiolitic sequence metamorphosed under eclogite facies conditions and sandwiched between higher and lower thrust sheets that lack such meta-ophiolites, as described in more details below. We interpret these oceanic rocks as derived from a separate oceanic basin, i.e. the Paleotethys Ocean, and later amalgamated in a suture zone located north of the Sava suture zone. This suture zone is interpreted to have closed in the Early Jurassic, i.e. the time when post-orogenic cover sealed the Karakaya accretionary complex of the Sakarya unit in Turkey (see e.g. Okay and Whitney, 2010 and the following chapter discussing the Sakarya unit). It is important to mention that according to our concept this Rhodope middle unit is not limited to the immediate hanging wall of the Nestos thrust (Fig. 11, profile G). Rather we use what we mapped Rhodope middle unit as a kind of Ariadne's thread for subdividing the originally top-N nappe stack of the Rhodopes mega-unit located north of the Nestos out-of-sequence backthrust into four units that are, from bottom to top (Fig. 10): (1) the Rhodope lower unit (Arda-Byala Reka) exposed in the Arda, Kesebir-Kardamos and Byala Reka-Kechros core complexes (Burg, 2011), (2) the ophiolite-bearing Rhodope middle unit (Nestos suture zone), (3) the Rhodope upper unit (Kerdilion and Madan sub-units) and (4) the Rhodope uppermost unit (Asenitsa- and Thrace sub-units). These four units are located exclusively in the hanging wall of the Nestos thrust (Fig. 5).

The deepest partly migmatitic high-grade metamorphic **Rhodope lower unit** (Arda, Kesebir-Kardamos and Byala Reka-Kechros domes or sub-units; Fig. 10) is dominated by quartzo-feldspatic gneisses. It forms Cenozoic core complexes exposed in windows below the next higher ophiolite-bearing eclogitic Rhodope middle unit zone that rims these domes. The domes of the Rhodope lower unit such as the Arda dome north of the Nestos thrust represent extensional core complexes formed in the middle to late Eocene. Most of the top-SW senses of shear seen in these domes are hence related to extensional unroofing rather than to thrusting. The high-grade Rhodope lower unit has been parallelized with the marble-rich greenschist to lower amphibolite-facies Pangaion-Pirin unit by most authors (e.g. Turpaud and Reischmann, 2010, their "Thracia terrane"; Burg, 2011, his "Lower Terrane", Froitzheim et al., 2014, their "Lower Allochthon") in spite of the obvious large lithological differences both in terms of lithology and metamorphic grade. It is correct that both the Rhodope lower unit and the Pangaion-Pirin unit are characterized by Permo-Carboniferous protolith ages for the granitoids (Turpaud and Reischmann, 2010) but this on its own cannot serve as a criterion for the tectonic affiliation. The granitoids of the Rhodope middle, upper and uppermost units, predominantly exhibit Jurassic to Early Cretaceous protolith ages (Turpaud and Reischmann, 2010). However, the proposal of Turpaud and Reischmann (2010) that the granitoids with Late Jurassic protolith ages are the product of subduction magmatism related to the same subduction channel that caused the UHP metamorphism of the Rhodope middle unit (Nestos suture zone) lacks supporting evidence. Yet, this criterion is taken as the major one for parallelizing the Rhodope lower unit and the Pangaion-Pirin unit into one and the same unit (Lower terrane of Burg, 2011; Lower Allochthon of Jahn-Awe et al., 2010 and Froitzheim et al., 2014). We propose that the Arda dome of the Rhodope lower unit became thrusted southward later in the Paleocene, i.e. after high-pressure eclogite facies metamorphism, over the lower grade Pangaion-Pirin unit along the out-of-sequence Nestos backthrust (see profile G of Fig. 11). This proposal is supported by occurrences of Pangaion marbles in drill holes beneath the Arda dome (Ivanov, 1981; Benderev et al., 2015). These marbles imply that the Nestos thrust cuts down into the base of the Arda unit and no longer follows the pre-existing Nestos shear zone (Fig. 11). Based on lithological composition and metamorphic grade we propose to parallelize the Pangaion-Pirin unit in the footwall of the Cenozoic Nestos backthrust with another marble-bearing unit, which overlies the Rhodope middle unit: the Rhodope uppermost unit (Asenitsa and Thrace sub-units) (see below). This implies that the Pangaion-Pirin unit would originally have represented a high structural level of the Rhodopes mega-unit that became the deepest tectonic unit due to Paleogene backthrusting along the Nestos thrust well after the formation of the Sava-İzmir-Ankara-Erzincan suture zone.

The **Rhodope middle unit** (Nestos suture zone) rims the domes of the Rhodope lower unit (Fig. 5) and consists of high-grade variegated series that contain, amongst many other lithologies, ophiolitic material and eclogites (occasionally preserving ultra-high-pressure metamorphism). This unit may be interpreted as a mélange zone (Turpaud and Reischmann, 2010) or a zone of intense imbrications. The reported protolith ages for the ophiolitic remnants are as follows: Liati et al. (2005) postulated Anisian ages based on SHRIMP U/Pb data on zircon (245.6 ± 3.9 Ma) for mafic rocks of the Rhodope middle unit. From the eastern Rhodopes Mountains Late Permian protolith ages (255.8 ± 2.1 Ma) are reported, again based on zircon SHRIMP-data (Liati and Fanning, 2005; Liati et al., 2011). Bauer et al. (2007) postulate a minimum age of a gabbro protolith based on the oldest apparent age of 288 ± 6 Ma (early Permian) obtained on zircon U/Pb dating with various methods. On the other hand, Froitzheim et al. (2014) obtained Jurassic protolith ages around 160 Ma (Oxfordian) on meta-plagiogranites and a metagabbro from the Satovcha ophiolite, inferred from U–Pb zircon dating using laser ablation.

Regarding ages of eclogite facies metamorphic overprint within the Rhodope middle unit, the Jurassic ages reported by several authors are the oldest ones available out of a wide spectrum that includes also Cretaceous and Cenozoic ages (see Liati et al., 2011 and 2016; Miladinova et al., 2018 and references therein). Some of these data, i.e. those covering the time interval 70.5–92.7 Ma determined with Sm–Nd geochronology on garnets from metapelites in the Chepelare shear zone (part of Rhodope middle unit) by Collings et al. (2016) almost perfectly overlap with the age of the activity of the Apuseni-Banat-Timok-Sredna Gora Late Cretaceous magmatic arc (~92–67 Ma; Gallhofer et al., 2015), magmatism related to the subduction of Neotethys located south of the Rhodopes mega-unit rather than to subduction within the Rhodopian mega-unit. The problems with dating high-pressure metamorphism may arise because the Rhodope middle unit probably remained under high-grade metamorphic conditions in terms of temperature all the time since the Jurassic until exhumation in Cenozoic time. Bauer et al. (2007) provide evidence for such a long-lasting post-eclogitic overprint under high temperatures lasting from around 171 Ma to 95 Ma. Didier et al. (2014) obtained two groups of monazite ages in two shear zones attributed to the Rhodope middle unit: a first group at around 140 Ma indicating granulite facies dehydration melting and a second one around 36 Ma indicating fluid-assisted ductile shearing in the presence of anatectic melts. This indicates that high-grade conditions probably prevailed until some 36 Ma ago, i.e. well into the period of core complex formation in the Rhodopes that led to final cooling by exhumation.

Given the evidence for a Triassic to Middle Jurassic N-facing tectonic event documented for the Balkan Mountains, presently located adjacent to and north of the Rhodopes mega-unit, but originally on top of the Rhodopes mega-unit (see discussion in the previous chapter), we regard the oldest Jurassic group of ages as the most likely ones to indicate the approximate age of eclogite facies metamorphism related to the subduction of Paleotethys. These fall into an Early Jurassic age range, namely ca. 184 Ma (Nagel et al., 2011), ca. 200 Ma (Petrik et al., 2016a) and pre-171 Ma (Bauer et al., 2007).

We included the following units into the Rhodope middle unit (see Fig. 10 for the locations of these units): Variegated series found rimming dome-like structures in the northwestern part of the Rhodope lower unit in the Pirin Mountains (Georgiev et al., 2010), the Mesta and Slashten sub-units located southeast of the Pirin Mountains (Sarov et al., 2008; Froitzheim et al., 2014), the Malyovitsa sub-unit in the area upstream from Rila monastery (Gorinova et al., 2019), the Chepelare shear zone (Petrik et al., 2016a), the Starcevo and Borovitsa sub-units (Pleuger et al., 2011), and finally, the Krumovitsa and Kimi sub-units in southeastern Bulgaria and adjacent Greece rimming the Kesebir-Kardamos dome (Mposkos and Krohe, 2006; Bonev and Stampfli, 2011; Georgiev et al., 2016; Moulas et al., 2017).

The **Rhodope upper unit** (Kerdilion and Madan sub-units), although of heterogeneous lithological composition, may be best defined as a medium- to high-grade unit (amphibolite facies to anatectic at its base) that structurally overlies the Rhodope middle unit but rests in the footwall of the lower grade (lower amphibolite to greenschist facies Rhodope uppermost unit (see below). The Rhodope upper unit is intruded by massive volumes of Late Cretaceous to Cenozoic granitic intrusions. Granitoid rocks predominate, but marbles, biotite schists, amphibolites, metagabbros and ultramafics are also found. The boundary to the underlying Rhodope middle unit is often not a sharp one and problematic to map exactly. The Rhodope upper unit comprises the following sub-units (see Fig. 10 for the locations of these units): the Kerdilion sub-unit in the footwall of the Kerdilion detachment (Brun and Sokoutis, 2007), the Sarnitsa sub-unit east of the Mesta Graben (Georgiev et al., 2010), the Troskovo (Zagorchev, 2001) and Kabul (Gorinova et al., 2019) sub-units downstream from Rila Monastery, the Crnook-Osogovo-Lisets core complex (Antić et al., 2016), the

Chepinska sub-unit (Naydenov et al., 2013), the Madan sub-unit, including the Bachkovo-Dobralak metamorphics (Raeva and Cherneva, 2009; Naydenov et al., 2006), and finally, the Borovitsa sub-unit (Pleuger et al., 2011).

The **Rhodope uppermost unit** (Asenitsa-Trace unit) is of particular interest because it is characterized by a lower grade of metamorphism (upper greenschist to lower amphibolite grade (Naydenov et al., 2009, 2013). Moreover, it typically exhibits top-NW, -N, or -NE senses of shear (Burg, 2011; and own observations). The Rhodope uppermost unit comprises a series of sub-units that are known under different names in the literature (see Fig. 10 for the locations of these units). The Asenitsa sub-unit, also referred to as gneiss-marble imbricate (Burg, 2011; his Fig. 4) predominantly consists of variegated meta-sedimentary rocks, including very thick series of marbles that are very reminiscent of the marbles of the Pangaion-Pirin unit of the southwestern Rhodopes Mountains (e.g. exposed along the main road southwest of Asenovgrad). The Asenitsa sub-unit turns around an E-plunging antiform exposing the Rhodope upper unit (orthogneisses of the Bachkovo-Dobralak metamorphics that are part of the Madan sub-unit) in its core near Asenovgrad. West of Asenovgrad it ends in a N-dipping orientation, associated with top-N senses of shear (Naydenov et al., 2013). The top-N senses of shear in this N-dipping belt west of Asenovgrad can be interpreted in two different ways. Either they represent later tilted top-N thrusting of the Asenitsa sub-unit over the Rhodope upper unit (Madan and Bachkov-Dobralak sub-units), and/or, they indicate normal faulting associated with extensional exhumation of the southerly adjacent Rhodope upper unit along a N-dipping detachment. Between this N-dipping part of the Asenitsa sub-unit and the Maritsa shear zone, defining the northern boundary of the Rhodopes mega-unit, Naydenov et al. (2013) described what they refer to as the Thrace “unit”. The lithological composition of this Thrace sub-unit is similar to that of the Asenitsa sub-unit and hence we also map the Thrace sub-unit as a part of the Rhodope uppermost unit. The dominantly carbonatic meta-sedimentary cover yielded a Mesozoic age based on strontium isotope studies (Naydenov et al., 2013), while the intruding magmatism associated with the Late Cretaceous Kapitan Dimitrievo pluton, which marked the last deformation in the Maritsa shear zone at ~78 Ma (Henry et al., 2012), indicates a minimum age of metamorphism. In the Besapara hills Late Jurassic magmatic rocks intruded an alternation of Triassic-Lower Jurassic sandstones, marls and limestones (Naydenov et al., 2009). Apart from the marbles and other metasediments the Thrace sub-unit also includes a poly-metamorphic sequence of gneisses, kyanite-bearing metapelites, metagabbro and large ultramafic bodies (Ichev and Pristavova, 2004). This succession is regarded as a piece of Variscan medium- to high-grade basement of Europe and was correlated with the Sredna Gora type metamorphic basement by Ivanov et al. (2000). von Quadt et al. (2006) and Naydenov et al. (2009) obtained Carboniferous zircon ages (about 334 Ma) for orthogneisses from this basement, which allows for a correlation with the Sredna Gora unit (high-grade metamorphism at ~336 Ma, Carrigan et al., 2003, 2005 and 2006) according to these authors. This strongly suggests a close link between the northernmost part of the Rhodope uppermost unit and the Sredna Gora unit of the Dacia mega-unit of the Balkan Mountains, the two only being separated by the Late Cretaceous Maritsa shear zone (Naydenov et al., 2013). The Maritsa shear zone is dominated by dextral strike slip kinematics with unknown displacement, but is also associated with relative uplift of the northern most Rhodopes mega-unit with respect to the Sredna Gora unit in Late Cretaceous time.

The Kardžali sub-unit located east of the Arda dome tectonically overlies the Borovitsa sub-unit of the Rhodope upper unit along a normal fault (Kardžali shear zone; Pleuger et al., 2011) and is also considered a part of the Rhodope uppermost unit. The Kardžali sub-unit consists of various mylonitic ortho- and paragneisses, schists, marbles, quartzites mafic and ultramafic rocks that were deformed under greenschist-facies conditions and which are commonly more fine-grained than rocks of the underlying units. The shear senses of the Kardžali shear zone are top-to-the-N-NW (Pleuger et al., 2011).

Interestingly, both Rhodope upper and Rhodope uppermost unit are seen to laterally wedge out eastward in map view (Plate 1 and Fig. 10). They appear to be laterally replaced by the Circum-Rhodope unit all around the eastern termination of the Rhodopes mega-core complex. But it is unclear if this is a primary feature or due to very substantial tectonic omission during Cenozoic normal faulting at the eastern termination of the Rhodopes mega-core complex.

3.2.6. Sakarya unit

The **Sakarya unit** comprises the southern part of the Pontides and is located adjacent and north of the Sava-İzmir-Ankara-Erzincan suture zone (Okay, 2008). In this sense the Sakarya unit occupies a position that is similar to that we envisage for the Rhodopes mega-unit and the Circum-Rhodope unit of northern Greece (Plate 1 and Fig. 1). In the north the Sakarya unit is separated from the Istanbul unit, the second constituent of the Pontides, by what is commonly referred to as the Intra-Pontide suture zone (Akbayram et al., 2013; see Plate 1). This ophiolite-bearing zone formed probably during the Early Cretaceous (Akbayram et al., 2013); it contains, amongst other constituents, ophiolitic slivers of an enigmatic Intra-Pontide Ocean that is supposed to have opened in Middle Jurassic time (Akbayram et al., 2013). It remains unclear if this ophiolite-bearing unit should be considered a suture zone. According to Okay (2008) the Intra-Pontide Ocean reopened a former Variscan suture zone between the Istanbul and Sakarya units that formed in Carboniferous time. Some authors consider oceanic parts of the Intra-Pontide suture zone as an eastward extension of the Eastern Vardar ophiolites (Şengör and Yılmaz, 1981; Marroni et al., 2013). Maffione and van Hinsbergen (2018), instead, suggested a disconnection of the Balkan system and the Pontide system along a long-lived transform fault. They regarded the ophiolites of the Intra-Pontide suture zone as issued from a basin that formed in the Middle Jurassic above N-dipping Neotethys subduction below the Pontides.

The Sakarya unit is of heterogeneous composition (Okay, 2008; Okay et al., 2006; Okay and Whitney, 2010). On our map, we include occurrences of high-grade metamorphic Variscan basement metamorphosed during the Carboniferous and overlying non-metamorphic sedimentary cover on the one hand, and subduction-accretion complexes (Karakaya accretionary complex) that probably formed during subduction and closing of Paleotethys in Triassic to Early Jurassic times on the other hand. The latter include eclogites with Triassic $^{40}Ar/^{39}Ar$ ages of 215–203 Ma (Okay and Monié, 1997; Okay, 2002) and other metamorphic units (Nilüfer Formation of the Lower Karakaya accretionary complex), but also non-metamorphic and strongly deformed Early to Middle Triassic sediments (Upper Karakaya accretionary complex). The latter contain blocks of Carboniferous and Permian neritic limestones, basalts, as well as Devonian to Triassic radiolarian cherts. This clearly indicates that Paleozoic oceanic crust (Paleotethys) was consumed during the early Mesozoic (Okay and Mostler, 1994; Okay et al., 2011; Sayit et al., 2011). The blueschists and eclogites of the Sakarya unit are thought to have been exhumed in a subduction channel and show that deep oceanic underthrusting below the Sakarya unit has occurred either northward (Okay and Nikishin, 2015) or southward (Şengör and Yılmaz, 1981; Eyuboglu et al., 2018) below the Variscan parts of the Sakarya unit. Note, however, that the view of a Triassic back-arc basin in connection with southward subduction of Paleotethys (Eyuboglu et al., 2018) is difficult to reconcile with the findings of Devonian ribbon cherts in the Karakaya accretionary complex (Okay et al., 2011).

This subduction was active until at least the latest Triassic given the Late Triassic age of the eclogites (Okay and Monié, 1997; Okay et al., 2002). The northward subduction hypothesis of Okay and Nikishin et al. (2015) would imply that there was no Cimmerian block at all since both Paleotethys and Neotethys would have been closed south of the Sakarya unit. Southward subduction would imply a very narrow Cimmerian block in the form of the Sakarya unit.

Tentatively, we also consider the Carboniferous Karaburun-Chios accretionary complex as a part of the Sakarya unit. This accretionary complex formed during the closing of Paleotethys (Meinhold et al., 2007 and 2008). Note, however, that Okay et al. (2012) considered the Karaburun-Chios accretionary complex as a part of the Bornova flysch zone, arguing that this complex represents a semi-intact huge block that is part of a Late Cretaceous mélange, namely the Bornova flysch zone. For these authors the Bornova flysch zone is not a part of the Sava-İzmir-Ankara-Erzincan suture zone but a mélange zone formed along and adjacent to a linear feature they refer to as İzmir Ankara suture, with the implication that this suture is a linear feature rather than a wide zone proposed here (see chapter 3.3.7 on the suture zone of the northern branch of Neotethys (Sava-İzmir-Ankara-Erzincan suture zone).

The unifying feature of the heterogeneous Sakarya unit is the presence of an unconformity below little deformed Lower Jurassic terrigeneous to shallow-marine, clastic sedimentary rocks (e.g. Okay and Whitney, 2010; Dokuz et al., 2017, and references therein). Şengör and Yılmaz (1981) considered the Sakarya unit as a part of their “Cimmerian” terranes, i.e. microcontinental blocks that rifted and drifted away from the northern margin of greater Adria (at this time still part of Gondwana) in the Triassic and during ongoing southward subduction of Paleotethys, opening Neotethys south of the Sakarya unit. According to this hypothesis Sakarya would have collided with the southern margin of Eurasia in the Early Jurassic, thereby closing the Paleotethys Ocean. Many other authors, however, assume that Paleotethys and Neotethys formed within the same basin and place Sakarya and the rest of the Pontides along the southern European margin, primarily based on the presence of Variscan basement (e.g., Topuz et al., 2013; Rolland et al., 2016; Yilmaz et al., 2000; Okay and Nikishin et al., 2015). In any case, as discussed in more detail below, we do not follow Stampfli and Borel (2004) who extended the Cimmerian continent all the way across Greece, i.e. across units we mapped as Adria-derived allochthons, and across Adria into the Apennines of southern Italy, placing the suture zone of Paleotethys between the Pindos and Tripolitza units of Greece.

3.3. Oceanic units, ophiolites, suture zones

3.3.1. Suture zones and ophiolites of Alpine Tethys

The term Alpine Tethys (also called Penninic Ocean; e.g. Frisch, 1979) was introduced by Stampfli (2000) to denote the extension of the Central Atlantic Ocean breaking eastward into the area of the present-day Circum-Mediterranean orogens during rifting and drifting in the Middle Jurassic (Fig. 7). In the Western Alps two branches of Alpine Tethys can be reconstructed: (1) The **Piemont-Liguria Ocean** that was paleogeographically located south of the Briançonnais continental fragment and opened in the Middle Jurassic. Its remnants now form the suture zone between the Adria-derived Austroalpine nappes and the Europe-derived Briançonnais continental fragment, and (2) the **Valais Ocean** that was paleogeographically located north of the Briançonnais continental fragment and opened in the Early Cretaceous (Schmid et al., 2004; Handy et al., 2010). Its remnants form a suture zone between the Briançonnais-derived and Europe-derived nappes.

However, east of the Lower Engadine window, i.e. in the Eastern Alps, no equivalents of the Briançonnais-derived nappes are found and hence the spreading associated with the younger Valais Ocean must have occurred inside the pre-existing Piemont-Liguria Ocean (Liati et al., 2005). Nevertheless, a subdivision into units derived from the Piemont-Liguria Ocean or the Valais Ocean, respectively, can also be made in the Eastern Alps based on the following criteria (Schmid et al., 2013): (1) age of rifting and passive margin formation, (2) presence or absence of radiolarites and aptychus limestone deposited on exhumed mantle rocks, (3) rock assemblages that are diagnostic for an ocean-continent transition towards either an Adriatic or a European continental margin, and, (4) age of accretion of the oceanic assemblies (Cretaceous for the Piemont-Liguria derived oceanic relics vs. Cenozoic for the Valais-derived oceanic relics).

In the Western Carpathians outcropping relics of serpentinites, gabbros or basalts that formed during the opening of Alpine Tethys are missing. The famous Meliata ophiolitic fragments of the innermost Western Carpathians are part of the northern branch of Neotethys. The Rhenodanubian flysch, however, attributed to the Valais branch of Alpine Tethys (Schmid et al., 2004) can be traced into the Magura flysch of the Outer Western Carpathians according to most authors (e.g. Oszczypko et al., 2015; Plašienka, 2018). The southerly adjacent **Pieniny Klippen Belt**, only several km wide, but at least 500 km long (Plašienka, 2018) undoubtedly represents a complex suture zone of Alpine Tethys dividing the Central Western Carpathian nappe stack (equivalent of the Adria-derived Austroalpine nappes) and the Outer Western Carpathians (equivalent of the Valais Ocean and Europe-derived flysch units) although it lacks outcrops of ophiolitic basement rocks. Going from north to south it contains (1) slivers that, according to many authors, represent transitions into the Magura basin (Grajcarek Unit; Jurewicz, 2018), (2) equivalents of the Briançonnais units of the Western Alps (Czorsztyn ridge; Birkenmajer, 1977; Golonka et al., 2006) and (3) slivers known as Kysuca Unit (or Pieniny s.str.) that are derived from the slope grading into pelagic equivalents of the Piemont-Liguria Ocean (=South Penninic Ocean) characterized by pelagic sediments deposited below the CCD (Plašienka, 2018). According to Plašienka (2012) sediments interpreted to be related to the opening and closure of the Piemont-Liguria Ocean proper, known as the **Vahic** Ocean, are only preserved in a very small window located in the Považský Inovec Mts. (Vahic unit) located ca. 100 km northeast of Bratislava (see the very small window mapped in Fig. 2). This window structurally underlies tectonic units derived from the Tatric and Infra-Tatric units representing the Adria passive continental margin (see Pelech et al., 2016 for an alternative view regarding this locality). The Miocene-age East Slovakian basin east of Kosice extending into Ukraine is underlain by a core complex made up of ophiolitic slivers and Bündnerschiefer-type lithologies, diagnostic for the ophiolitic suture zones representing the remnants of the Alpine Tethys in the Alps proper (see Fig. 5). These lithologies are only known from boreholes in eastern Slovakia and constitute what is known as the **Iňačovce-Kriscevo** sub-unit (Sotak et al., 1994, 1997).

The Pieniny Klippen Belt, together with the Magura unit, marking the suture zone of Alpine Tethys, follows the northern rim of the East Slovakian basin and can be followed eastwards across western Ukraine all the way to the Maramures area of north-western Romania (Ślączka et al., 2006) where the easternmost occurrences of the Pieniny Klippen Belt are found (Poiana Botizei area; Săndulescu et al., 1979/1980; Bombita et al., 1992; see Fig. 5). In the Maramures area the easternmost occurrence of the Magura flysch, the Klippen Belt and the internally adjacent Iňačovce-Kriscevo unit form a tight arc in map view (Tischler et al., 2007). This tight arc formed in the Burdigalian, in connection with the lateral extrusion of the ALCAPA mega-unit (Ratschbacher et al., 1991b;

Tischler et al., 2007; Ustaszewski et al., 2008). It delimits the Western Carpathians whose internal units are built up by the ALCAPA mega-unit, from the Eastern Carpathians built up by the Dacia mega-unit in its internal parts (see Fig. 5). Around this tight arc the Magura flysch, the Pieniny Klippen Belt and the Iňačovce-Kriscevo unit, both forming a suture zone of Alpine Tethys, line up with the ENE-WSW-striking Mid-Hungarian fault zone marked by the **Szolnok flysch** belt in eastern Hungary. The uppermost Cretaceous to Paleogene flysch of the Szolnok flysch belt shows affinities to the Pieniny Klippen Belt and adjacent units in Maramures (Nagymarosy and Baldi-Beke, 1993; Kováč et al., 2016). Hence, the >1600 m thick Szolnok flysch of Late Cretaceous to Eocene age known from numerous drill holes over a length of 130 km along the Mid-Hungarian fault zone (Haas et al., 2010 and 2012) structurally connects the suture zone of Alpine Tethys with the **Sava suture zone** of the Dinarides (see Fig. 5) that closed Neotethys (Schmid et al., 2008; Ustaszewski et al., 2010). Hence the suture zone of Alpine Tethys (with the exception of the Ceahlau-Severin branch of Alpine Tethys; see discussion below) and the suture zone of Neotethys were probably once connected along the length of the Mid-Hungarian fault zone (see Fig. 7) which today marks the boundary between the ALCAPA and Tisza-Dacia mega-units (Csontos and Nagymarosy, 1998).

The young Mid-Hungarian fault zone must formerly have been the site of a major lateral change in the polarity of subduction that existed in pre-Miocene time. In the Alps and Western Carpathians, the European plate was subducted southward beneath the Alpine Tethys. In the Dinarides, however, it was the Adria margin that was subducted beneath Neotethys, which is in turn overlain by the Europe-derived Tisza and Dacia mega-units (Schmid et al., 2008; Ustaszewski et al., 2010; Handy et al., 2015).

While the above-described tight arc in Maramures that bends the Pieniny Klippen Belt structurally overlies the westernmost part of the Bucovinian nappes of the Dacia mega-unit (Tischler et al., 2007; Marmarosh massif of the Ukrainian authors; e.g. Hnylko et al., 2015) rocks from another oceanic basin are found north of this arc and structurally below the Bucovinian nappes. This additional oceanic branch is denoted **Ceahlau-Severin suture zone** (see Plate 1 and Fig. 8) and opened in the middle Jurassic, i.e. contemporaneous with the opening of Alpine Tethys as will be outlined below.

The Ceahlau-Severin suture zone (Outer Dacides of Săndulescu, 1994) comprises a series of nappes in the Eastern and Southern Carpathians (Fig. 8). These are the Black flysch nappe of the northernmost Eastern Carpathians (Săndulescu, 1975 and 2009) that are equivalent to three slices of the so-called "Fore-Marmarosh suture zone" in Ukraine (Hnylko et al., 2015), the Ceahlau, Bobu and Baraolt nappes of the rest of the Eastern Carpathians and the so called "Severin suture zone" of the Southern Carpathians (Săndulescu, 1984 and 2009). Strongly dismembered ophiolitic lithologies such as harzburgitic ultramafics, gabbros and pillow basalts are mainly preserved in the Severin unit of the Southern Carpathians (Savu et al., 1985). The Black flysch nappe is characterized by mafic basalts, spilites and tuffs, ranging in age from Middle Jurassic to Berriasian (Săndulescu, 2009). This indicates that the Ceahlau-Severin Ocean did not open before the Middle Jurassic, simultaneously with the Alpine Tethys to the northwest, but probably also with the Intra-Pontide Ocean to the southeast. The large Ceahlau nappe of the Eastern Carpathians is primarily made up by Upper Jurassic radiolarites, interlayered with basic igneous rocks and deep-water deposits (Azuga beds) at the base, followed by the Tithonian to Lower Cretaceous Sinaia beds, mostly shaly and calcareous flysch-type deposits. These are followed by Barremian to Aptian proximal turbidites (Săndulescu, 1984; Melinte-Dobrinescu and Jipa, 2007). The contact with the overlying Getic nappe system is sealed by lower Albian massive sandstones and conglomerates that are syn- to postkinematic (Bucegi conglomerate; e.g., Patrulius, 1969; Olariu et al., 2014; Jipa and Olariu, 2018). The existence of these conglomerates implies that the inner parts of the Ceahlau-Severin suture zone were accreted to the Dacia mega-unit (e.g. the Getic nappe system) already in latest Early Cretaceous time. In contrast, the external parts of the Ceahlau-Severin suture zone cannot have overridden the Danubian nappe pile (Fig. 8) before the latest Cretaceous because Turonian to late Senonian flysch is present in the Danubian units underlying the Severin suture zone in the Southern Carpathians (Schmid et al., 1998 and literature therein). In the southernmost Eastern Carpathians, the external contact of the Ceahlau-Severin suture zone with sediments of the more external flysch units also formed during latest Cretaceous shortening, as indicated by upper Campanian - Maastrichtian syn- to postkinematic cover in the contact zone (Săndulescu, 1984; Melinte and Jipa, 2005). During the Cenozoic the Ceahlau-Severin suture zone was displaced eastward and rotated together with the Dacia mega-unit (Fügenschuh and Schmid, 2005), progressively closing the partly oceanic embayment of the Alpine Tethys (Balla, 1987) and the entire external flysch belt (Mațenco and Bertotti, 2000).

3.3.2. Ligurian ophiolites

The ophiolites in the Apennines were mapped based on the maps of Bigi et al. (1992). These ophiolites were emplaced during the growth of the Apenninic-Maghrebian orogenic system by thrusting and nappe stacking in the area of the African and Adriatic continental paleo-margin, starting in the late Oligocene and reaching the external foredeep of the Apennines in the latest Miocene (Vai and Martini, 2001). The **Ligurian ophiolites** occupy the structurally highest position in the Apennines except for the Calabro-Peloritani unit in southernmost Italy (Plate 1). From the Oligocene onwards, these ophiolites were emplaced to the northeast over the Adria-derived allochthons in the Apennines. This late-stage ophiolite emplacement was roughly contemporaneous with oroclinal bending in the southernmost Western Alps (Schmid et al., 2017), the 50° counterclockwise rotation of the Corsica-Sardinia block (Advokaat et al., 2014) and W-directed subduction and subsequent roll back of the Adria mantle slab below the Apennines (Spakman and Wortel, 2004). The ophiolites represent relic parts of the southwestern part of Alpine Tethys between Adria and Iberia that did not close during Alpine collision in the Eocene and stayed open until the late Oligocene (Molli et al., 2010; Handy et al., 2010; Schmid et al., 2017).

In the northern Apennines such ophiolitic units and off-scraped accretionary prism rocks are known as the Ligurides, subdivided into Internal and External Ligurides (Elter and Pertusati, 1973; Molli et al., 2010). Note that we also mapped the External Ligurides of the northern Apennines under the label "Ligurian ophiolites" although they actually represent a fossil ocean-continent transition zone between the Piemont-Liguria Ocean and continental Adria (Marroni et al., 2001). In the southern Apennines and the Maghrebides of Sicily we also mapped the sedimentary units of the Sicilides under the same label since they represent the lateral equivalent of the External Ligurides of the northern Apennines (Elter et al., 2003). These ophiolitic and ocean-continent transition units are non-metamorphic except for the Calabrian ophiolites that structurally underlie the continental Calabro-Peloritani unit and suffered Alpine-age deformation and metamorphism (Liberi et al., 2006 and references therein).

3.3.3. Controversies regarding the Neotethys Ocean(s)

The term Neotethys is connected with the idea that this Mesozoic ocean opened while Paleotethys closed north of the Neotethyan spreading site. Paleotethys stayed open longer in the areas east of the Variscides of Europe and did not close before Triassic to

Early Jurassic times when a continental fragment known as Cimmeria is supposed to have collided with Laurasia. This classical idea postulates northward drift of Cimmeria leading to rifting in "Northern Gondwanaland", opening Neotethys at the same time (Şengör, 1979). This scheme, successfully developed in Iran (Stöcklin, 1968 and 1974) was first extended westward into Turkey by Şengör and Yılmaz (1981) and later further west all the way to the Ionian Sea (Stampfli, 2000; Stampfli and Borel, 2004). However, as discussed earlier, the existence of a Cimmerian continental fragment is under debate. Stampfli and Borel (2004) follow this original concept of Şengör and Yılmaz (1981) and propose that a series of different oceanic Alpine-Eastern Mediterranean orogen back-arc basins (named "Meliata", "Küre", "Maliac", "Pindos", "Vardar", "Izanca") successively opened within Cimmeria, due to northward roll back at the northern margin of Paleotethys in an area located north of the major strand of Neotethys running along the present-day Eastern Mediterranean basin. Their reconstruction leads to an extremely complicated configuration with multiple back-arc oceans (Stampfli and Borel, 2004; Stampfli and Hochard, 2009), which we consider primarily conceptional and not really supported by field data.

Another variety of multi- vs. single-ocean controversy exists in regard to the ophiolites of the Dinarides and Hellenides (see discussion in Cvetković et al., 2016). These ophiolites are seen as either issued from a single oceanic domain (e.g. Bernoulli and Laubscher, 1972; Schmid et al., 2008; Bortolotti et al., 2013; Tremblay et al., 2015; Ferrière et al., 2012 and 2016; Gawlick et al., 2008 and 2017a) located east of the Pelagonian massif, or alternatively, as being issued from two oceans originally located east and west of a Pelagonian-Korab-Drina-Ivanjica microplate, often referred to as Vardar and Pindos Oceans, respectively (e.g. Smith, 1993; Robertson et al., 1991 and 2009a; Dilek and Flower, 2003; Rassios and Dilek, 2009). Some authors adhering to the terrane concept even postulated up to four oceanic terranes within the Dinarides and Hellenides (Karamata, 2006; Papanikolaou, 2009). Regarding the Dinarides and Hellenides, the present compilation (following Schmid et al., 2008) advocates the one-ocean concept in the sense that the Dinaric and Hellenic ophiolites were obducted westward over the Pelagonian-Korab-Drina-Ivanjica units that represent the passive margin of Adria, rooting in the "Vardar zone" located east of the Pelagonian massif, a term originally defined by Kossmat (1924). Note that here we use the term "Sava suture zone" in order to denote this root (Plate 1). This one-ocean concept logically follows from restoring the modern structural architecture of the Dinarides-Hellenides orogen (see van Hinsbergen et al., 2019).

Unfortunately, the situation is complicated regarding the so-called "Vardar zone" of Kossmat (1924). This "Vardar zone" as defined by Kossmat (1924) is a very heterogeneous 40–70 km wide zone, wedged between the Pelagonian massif to the west and the Serbo-Macedonian unit to the east (Plate 1 and Figs. 11, 12, 13, 14 and 15). It is composed of subvertically oriented slices made up of Upper Cretaceous hemipelagic or flysch-type cover of the Pelagonian massif that is post-orogenic with respect to Early Cretaceous orogeny, pre-Triassic basement and dismembered Mesozoic cover of the Jadar-Kopaonik unit, and finally, often rather thin slices of ophiolitic rocks (serpentinites, basalts) (Pamić et al., 2002). Unfortunately, and contrary to the original definition, the Vardar zone gradually became a synonym for the so-called "Vardar Ocean" and "Vardar ophiolites". As will be discussed later, the Vardar zone of Kossmat (1924) consists of what Schmid et al. (2008) referred to as, from west to east: Upper Cretaceous cover of the Pelagonian massif, remnants of the Western Vardar ophiolitic unit, slices of the continental Jadar-Kopaonik unit, the Sava suture zone, continental units of the Circum-Rhodope unit interleaved with ultramafic slices and occasionally larger bodies of the Eastern Vardar ophiolitic unit such as the Demir Kapija ophiolite (Plate 1; Fig. 4. Profile F; Božović et al., 2013; Prelević et al., 2017). In the present compilation we keep the terminology proposed by Schmid et al. (2008) but point out that the Eastern Vardar ophiolitic unit has recently been considered a short-lived Jurassic oceanic back-arc basin, separated from the Western Vardar ophiolites by an intra-oceanic subduction zone (Fig. 7; Gallhofer et al., 2017). However, other interpretations are possible (e.g. van Hinsbergen et al., 2019). Following the terminology proposed by Şengör and Yılmaz (1981) we consider the Western Vardar ophiolitic unit, Sava-İzmir-Ankara-Erzincan suture zone and Eastern Vardar ophiolitic unit as units that evolved from plate tectonic processes within the Neotethys.

The situation changes in western Turkey, i.e. in the area east of the Balkan Peninsula, where most southward obducted ophiolites, rooting in the Sava-İzmir-Ankara-Erzincan suture zone, are of Cretaceous rather than Jurassic age (Parlak, 2016; Plate 1). Note, however, that a belt of Jurassic age supra-subduction zone ophiolites is present to the north of the Cretaceous-age obducted ophiolites. These are part of the Sava-İzmir-Ankara-Erzincan suture zone east of our area of investigation (e.g. Topuz et al., 2013), documenting the former presence of a Jurassic ocean that was completely closed in Anatolia.

A reconstruction explaining the appearance of Cretaceous ophiolites in Turkey has recently been proposed by Gürer et al. (2016). These authors argue that the oceanic realms that are part of the Neotethys Ocean in Turkey and closed in western Turkey at the end of the Cretaceous along the Sava-İzmir-Ankara-Erzincan suture zone belonged to two separate plates during the Late Cretaceous. The southern plate is in the Late Cretaceous the African/Adriatic plate and contains the southern part of the Sava Ocean of the Dinarides and Hellenides (a remnant of the Northern branch of Neotethys that closed at the end of the Cretaceous along the Sava suture zone) that was attached to the northward subducting Adriatic-Tauride continental margin. The northern plate became, since the initiation of intra-oceanic subduction at or before ~95 Ma, part of a separate mostly oceanic tectonic plate within the northern branch of Neotethys referred to as Anadolu plate. This intra-oceanic subduction zone soon afterwards led to S-directed obduction of the ophiolites of western Turkey onto the Anatolides-Taurides starting in the late Santonian (~90–85 Ma; Robertson et al., 2009b; van Hinsbergen et al., 2016; Pourteau et al., 2019).

The Eastern Mediterranean Ocean separated the Anatolide-Tauride platform that we consider the lateral equivalent of the Adria-derived allochthons in the Dinarides and Hellenides from the main body of Gondwana, i.e. northern Africa and the Arabian platform. It runs offshore western Turkey and Crete, ending in the Ionian Sea (see Fig. 7; Southern branch of Neotethys). It entered the trench around 35 Ma ago south of Anatolia (van Hinsbergen et al., 2010a; McPhee et al., 2018b), whilst to the west in the Aegean region, continental subduction of Adria continued progressively younger, ~20 Ma south of Crete, and Pliocene towards Zakythos and Kefallonia (Jolivet and Brun, 2010; van Hinsbergen and Schmid, 2012). Structural and stratigraphic evidence demonstrates that the Cretaceous-age ophiolites overlying the Antalya-Alanya nappes (mapped near the SE edge Plate 1) were not issued from the Northern branch of Neotethys. They were emplaced northward over the Tauride platform in the latest Cretaceous, and their frontal thrust is sealed by Eocene sediments (Okay and Özgül, 1984). Their emplacement direction clearly shows that they were derived from a separate subduction segment. They root in the southern branch of Neotethys, together with the ophiolites of Cyprus and Syria (Maffione et al., 2017; McPhee et al., 2018a, 2018b).

3.3.4. Western Vardar ophiolitic units

The **Western Vardar ophiolitic unit** comprises ophiolitic units that were obducted onto the Adriatic margin in Late Jurassic to earliest Cretaceous times. This was before this margin became involved in orogeny forming the nappe stack of the easternmost Alps, Western Carpathians, Dinarides and Hellenides (Schmid et al., 2008 and references therein). They do not represent ophiolitic suture zones but rather represent the structurally upper parts of composite nappes that consist of Adria-derived continental units and previously obducted ophiolites, which formed by out-of-sequence thrusting (with respect to obduction) during Cretaceous and Cenozoic orogeny. As a result of this, in present-day map view, they form belts or isolated ophiolitic domains in the Dinarides-Hellenides (Figs. 1, 11, 12, 13, 14 and 15). When mapping these ophiolitic units (Plate 1) the occasionally thick layers of ophiolitic mélange underlying the obducted ultramafics and their metamorphic sole were also included, as well as the post-obduction sedimentary cover of latest Jurassic to Cenozoic age stratigraphically overlying the ophiolites. The crustal rocks of the obducted ophiolites are of Jurassic age as is demonstrated by the upper Bajocian to lower Oxfordian (168–166 Ma) radiolarites stratigraphically overlying a completely preserved and, as inferred from geomagnetic anomalies (Kane et al., 2005), some 10 km thick ophiolitic sequence in Albania (Mirdita ophiolite; Bortolotti et al., 2005, 2013 and references therein). This is confirmed by radiometric dating of gabbros and basalts in the Dinarides and Hellenides (ca. 174 Ma - 158 Ma; Liati et al., 2004; Dilek et al., 2008; Ustaszewski et al., 2009; Slovenec et al., 2011; Lugović et al., 2015). The age range of the metamorphic soles welded to the base of the ophiolitic sequence (ca. 174 Ma - 157 Ma; Lanphere et al., 1975; Okrusch et al., 1978; Dimo-Lahitte et al., 2001; Borojević Šoštarić et al., 2014) dates the approximate time of intra-oceanic subduction initiation that always precedes Tethyan-style obduction (Edwards et al., 2015; Maffione et al., 2015; Maffione and van Hinsbergen, 2018). Given the precision of the data, this range of ages is identical to that available regarding the formation of the obducted ophiolites. The eastern part of the Western Vardar ophiolites has supra-subduction zone geochemical signatures, in contrast to a western belt in Albania and Greece, showing that parts of the ophiolitic crust formed by spreading above a subduction zone (Dilek et al., 2008; Maffione et al., 2015). The coincident ages of crust and sole shows that ophiolite formation must have closely followed intra-oceanic subduction.

Preserved ophiolitic remnants of Triassic-age ocean floor along the Adria passive margin are rare and almost exclusively found as blocks in the sub-ophiolitic mélange, except for the Fourka unit in the Othrys Mountains of Greece. There, a 200–300 m thick slice of Triassic basalt (Ladinian-Carnian) covers a surface of about 35 km by 35 km sandwiched between obducted Jurassic peridotite above and sedimentary nappes of the western continental margin of Adria below (Ferrière et al., 2016). It is this occurrence that coined the term "Maliac Ocean" used by Stampfli and Borel (2004) to denote a separate oceanic domain. Numerous occurrences of blocks of basalts that preserve direct stratigraphical contact with radiolarites dated to be of Triassic age, embedded in Jurassic-age mélange formations, have been found in the Hellenides (e.g. Ozsvárt et al., 2012), Dinarides (e.g. Halamić et al., 1999; Babić et al.,

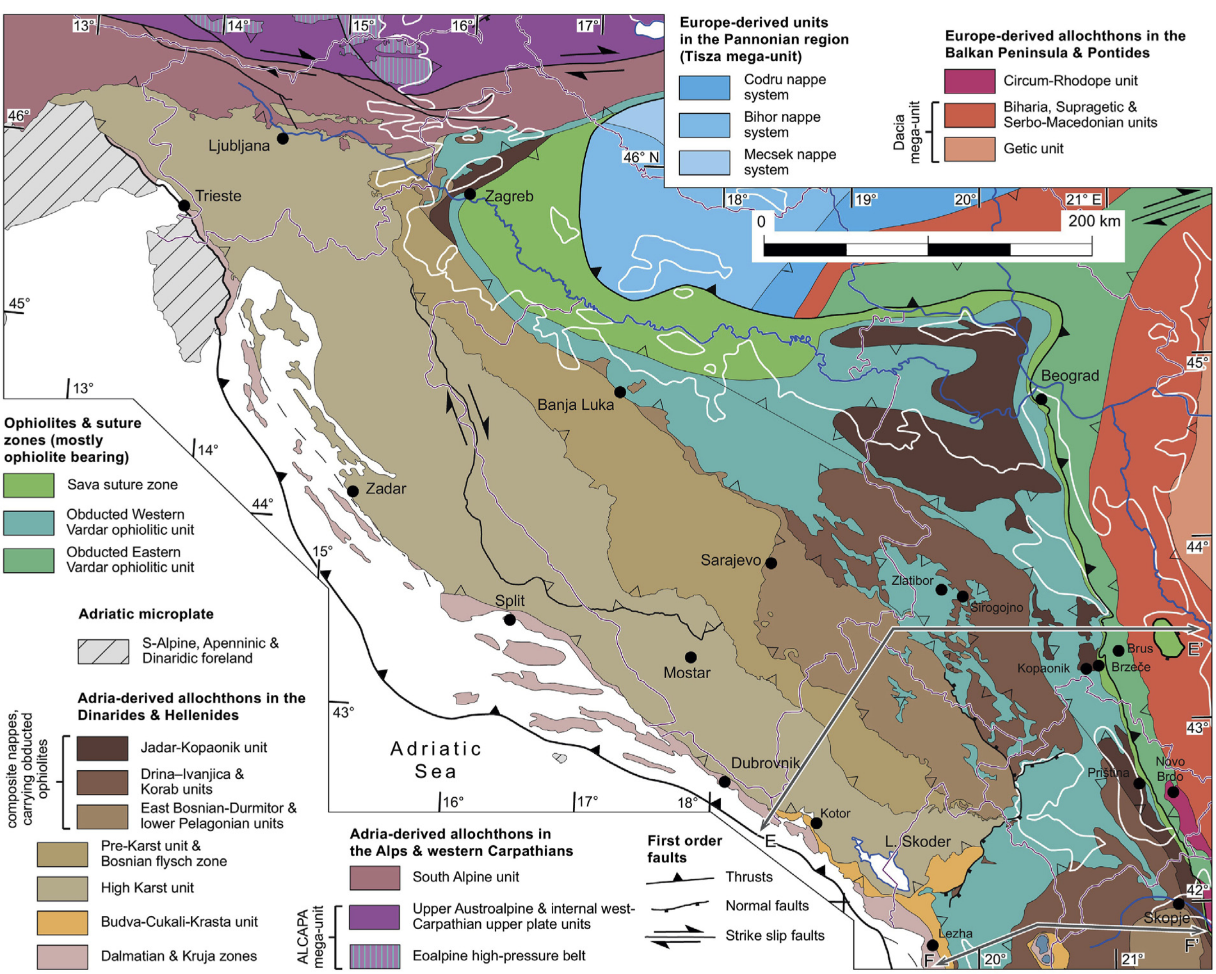

Fig. 12. Tectonic map of the Dinarides and adjacent areas; extract of Plate 1.

2002; Goričan et al., 2005; Vishnevskaya et al., 2009) and Western Carpathians (e.g. Kozur, 1991) where they are known as being derived from the so-called "Meliata Ocean". We propose that blocks of Triassic pillows were scraped off the older (Triassic) part of the northern branch of Neotethys fringing the lower plate Adriatic margin and subducted beneath the Jurassic ophiolite sequence of the obducted upper plate (Schmid et al., 2008; their Fig. 5). Such findings and their interpretation lead to the concept of a single ocean including the former Western Vardar, Meliata and Maliac oceans that opened during Triassic to Middle Jurassic times (Fig. 7). Interestingly such mélanges often also contain Jurassic ophiolites and associated radiolarites, which must have been shed as olistoliths into an olistostrome that later became incorporated into the same tectonic mélange that contains Triassic basalts (Vishnevskaya et al., 2009). This leads to the conclusion that some blocks of these mélanges represent olistoliths while others were scraped off the distal-most Adriatic margin (Gawlick et al., 2017a) including adjacent Triassic ophiolites and their cover (Kozur, 1991).

The age of mélange formation and obduction is loosely constrained by four categories of data: (1) Obduction post-dates intra-oceanic subduction and is hence constrained not to have occurred before Aalenian to Oxfordian times based on the radiometric data concerning metamorphic sole formation. (2) The ages of the youngest sediments found beneath the mélange, radiolarites deposited onto the Adriatic margin immediately underlying the mélange, commonly reach the Kimmeridgian or Tithonian, but in some cases even the lowermost Cretaceous (Berriasian; Jotanovici–Lipnje Section in Bosnia of Vishnevskaya et al., 2009). This indicates that at least in some places obduction continued beyond the Jurassic/Cretaceous boundary. (3) The age of the youngest Jurassic cherts incorporated into the mélange revealed Callovian to early Kimmeridgian ages, implying a Kimmeridgian or younger age of obduction of the Western Vardar ophiolites at this particular locality in Serbia (Gerzina and Djerić, 2016). (4) The age of the base of overstepping carbonates, Tithonian in Albania (Schlagintweit et al., 2008), Berriasian in Bosnia (Hrvatović and Pamić, 2005) postdates the final stages of obduction since they were deposited in a shallow water environment. Some authors (e.g. Gawlick et al., 2016 and 2017b) claim to have dated mélange formation as Middle to early Late Jurassic and contemporaneous with intra-oceanic subduction. However, this does not apply to the majority of the sub-ophiolitic mélanges in the Dinarides that include off-scraped elements of the Adria margin and that therefore must have formed during obduction rather than during intra-oceanic subduction. Moreover, given the fact that the age of mélange formation refers to the age of deformation by definition (Hsü, 1968 and 1974) fossil findings only provide an estimate for the maximum age of mélange formation. Although a precise age cannot be provided on the basis of these data it appears that obduction might have been a rather long-lasting event, starting in the Kimmeridgian and ending in the Berriasian.

The time gap between the radiolarites capping the pillow lavas of the well-preserved Mirdita ophiolites in Albania (168–166 Ma) and the estimated onset of obduction at the beginning of the Kimmeridgian (about 157 Ma) only leaves some 10 Ma between the formation of the ophiolites and their arrival at the Adriatic continental margin. The total length of the obducted Western Vardar ophiolite amounts to some 200 km measured along a transect from Albania to North Macedonia (see profile in Fig. 15).

There is a much-discussed question as to whether intra-oceanic subduction really started at a former mid-ocean ridge, such as suggested by Nicolas and Boudier (1994) for Oman and also in the case of the Dinarides-Hellenides (Nicolas and Boudier, 1999; Schmid et al., 2008; Maffione et al., 2015). An alternative and presently most popular model is the so-called "supra-subduction" (structurally, all obducted ophiolites are upper plate ophiolites!)

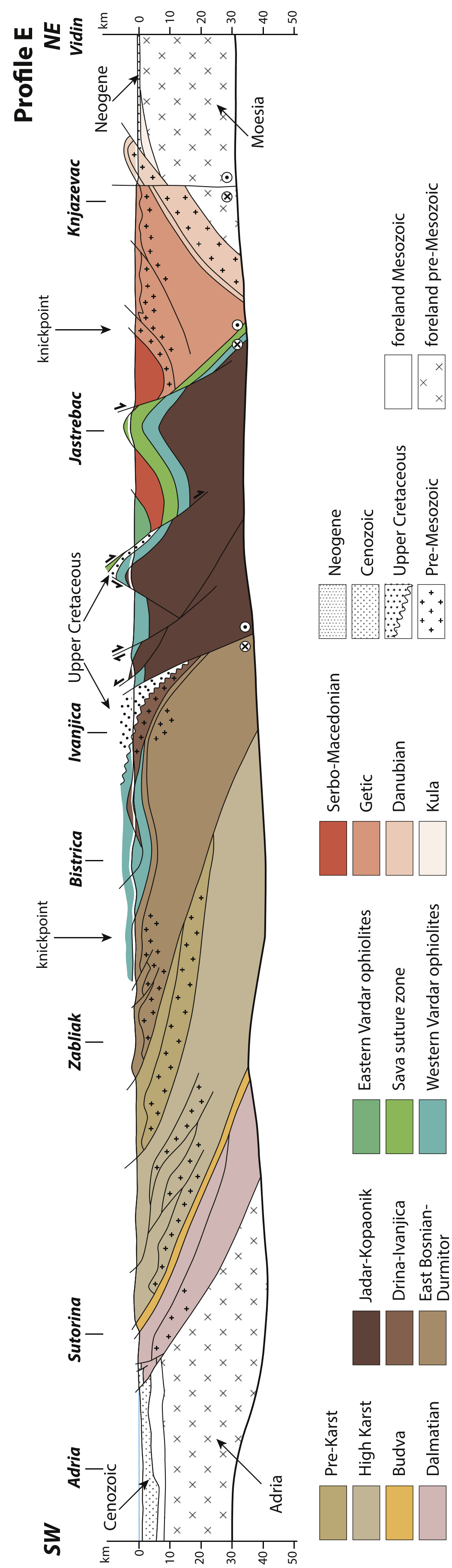

Fig. 13. Crustal-scale profile across the Dinarides (profile E). See Figs. 1 and 12 for the trace of profile G. Profile construction is after Schmid et al. (2008) and Maţenco and Radivojević (2012). Moho depth after Finetti et al. (2005), Grad et al. (2009), and Stipčević et al. (2011).

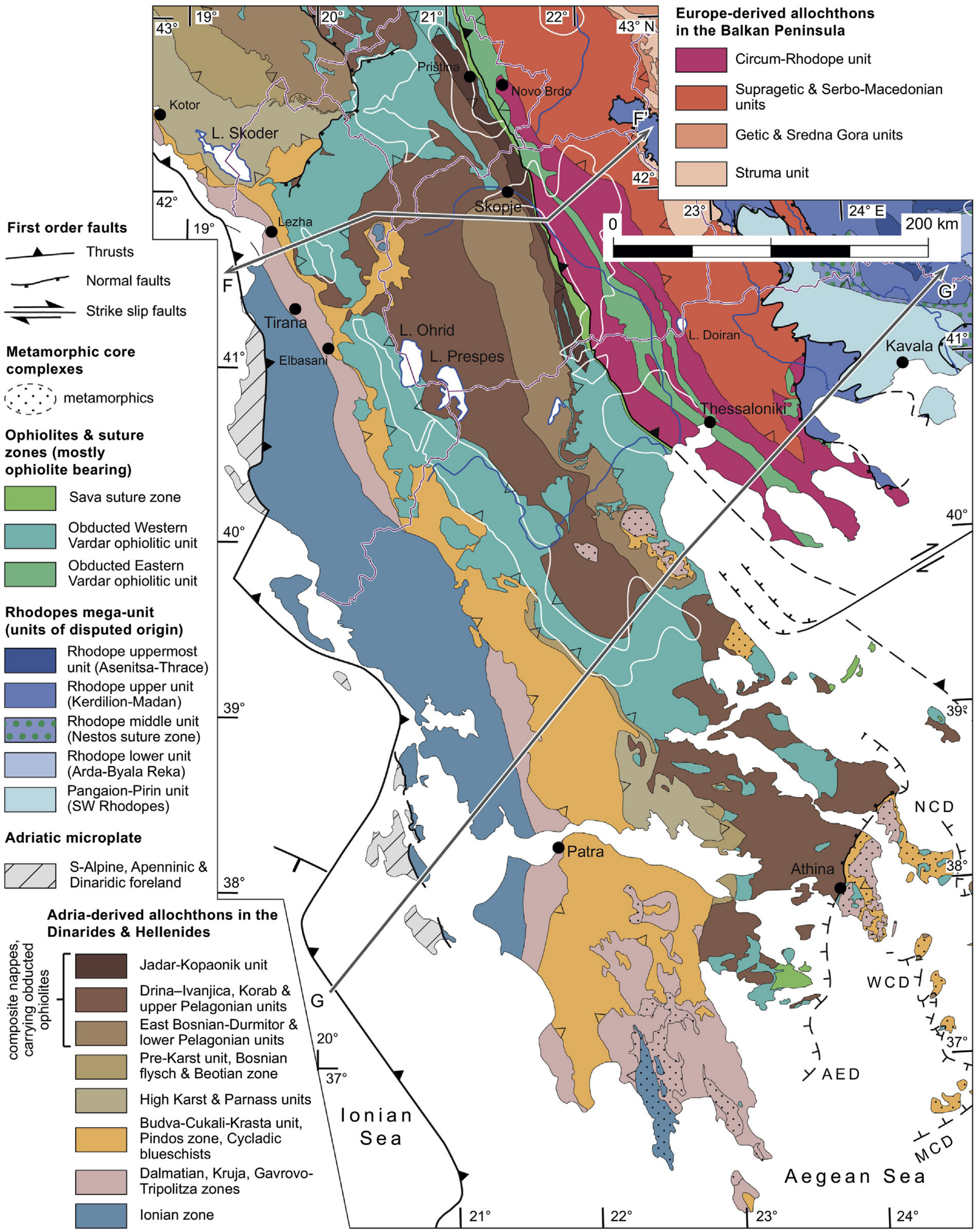

Fig. 14. Tectonic map of the Hellenides and adjacent areas; extract of Plate 1.

scenario postulating in its extreme form that the genesis of the entire later obducted ophiolitic sequence is governed by slab dehydration and accompanying metasomatism of the mantle, melting of the subducting sediments, and repeated episodes of partial melting of metasomatised peridotites in a forearc setting (Dilek and Furnes, 2009). The geochemical signature of the Mirdita ophiolites turns out to be hybrid given the fact that the western Mirdita ophiolites have a MORB affinity, whereas the eastern Mirdita ophiolites predominantly, but not exclusively, have a supra-subduction geochemistry (Dilek and Furnes, 2009). The situation regarding the Mirdita ophiolites, which represent the best-preserved and -investigated part of the Western Vardar ophiolites, can be readily explained by a scenario similar to that proposed by Maffione et al. (2013a, 2015). They proposed renewed magmatic

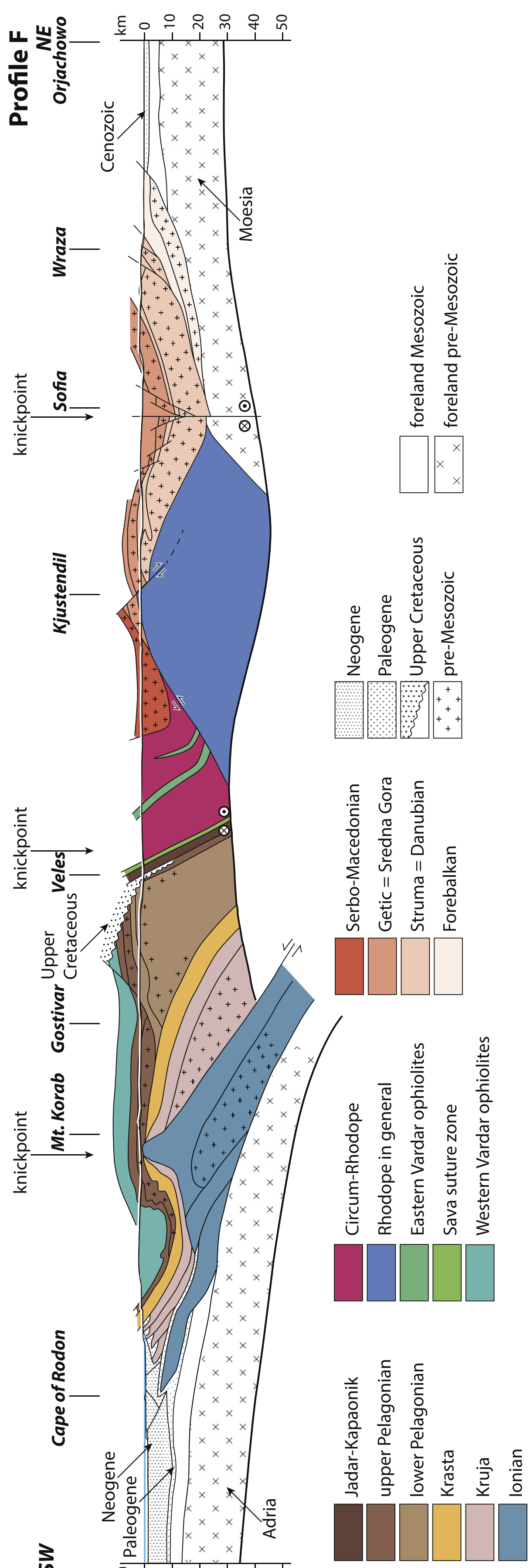

Fig. 15. Crustal-scale profile across the northernmost Hellenides (profile F). See Figs. 1, 12 and 14 for the trace of profile F (note that the profile across the central Hellenides is found in Fig. 11). Profile construction is based on the available geological maps of Albania (particularly on the profiles accompanying the 1:200.000 geological map of Albania (Xhomo et al. 2002), the geological maps of former Yugoslavia (Osnovna geološka karta SFRJ) and Bulgaria, as well as on own field work (see also Handy et al., 2019). Moho depth after Finetti (2005), Sachpazi et al. (2007), Grad et al. (2009), Pearce et al. (2012), and Delipetrov et al. (2016).

activity during intra-oceanic subduction that took place not far from the mid-ocean ridge, generating supra-subduction type magma in the forearc area exhibiting island-arc tholeiitic and boninitic signatures. Thereby a narrow slice of MORB-type lithosphere would remain trapped between the trench along which obduction was nucleated (located west the present-day erosional front of the Mirdita ophiolites) and the forearc (present-day Eastern Mirdita massifs). This scenario could readily explain the well-known fact that two ultramafic belts, Western-type and Eastern-type, are distinguished in the Mirdita ophiolites (Nicolas and Boudier, 1999; Hoxha, 2001; Tremblay et al., 2009; Meshi et al., 2010 and 2014). The Western-type ultramafic massifs are characterized by a lherzolitic-harzburgitic mantle sequence covered by a discontinuous and rather thin sequence of troctolites, gabbros and N-MORB basalts, as well as by oceanic core complexes that are typical for slow-spreading environments. The Eastern-type ultramafic belt, on the other hand, consists of a mostly harzburgitic mantle sequence grading up into an ultramafic cumulate sequence and a well-developed thick gabbro section covered by a sheeted dike complex and a sequence of basalts showing MORB, island arc tholeite and boninitic geochemical affinities.

The present-day orientation of the former spreading axis of the Western Vardar ophiolites was reconstructed to be NNE-SSW oriented based on the consistent orientation of sheeted dykes in the Mirdita ophiolite (Maffione et al., 2013a, 2013b). Paleomagnetic data allowed for reconstructing the original orientation of the spreading axis as having been oriented N-S, i.e. at a very high angle to the present-day strike of the Dinarides (Maffione et al., 2015; Maffione and van Hinsbergen, 2018), but more or less parallel to the restored passive margins of Neotethys in Jurassic time (van Hinsbergen et al., 2019). This orientation is consistent with numerous own measurements in Serbia, Kosovo, Albania and Greece regarding the direction of shearing in metamorphic soles and tectonic mélanges immediately underlying the metamorphic sole. These are invariably top-W to top-WNW and hence approximately perpendicular to the inferred N-S orientation of the former spreading axis. This also agrees with the reconstruction of the former plate configuration shown in Fig. 7.

3.3.5. Eastern Vardar ophiolitic units

The term **Eastern Vardar ophiolitic unit** comprises ophiolitic units that were emplaced onto units derived from the northeastern margin of the Neotethys (Dacia mega-unit) in the latest Jurassic (Schmid et al., 2008). During and after obduction they became an integral part of the Cretaceous-age Carpatho-Balkan orogen (Fig. 8). During the subduction of the Sava Ocean in the Cretaceous, ending with collision at the end of the Cretaceous, they formed the upper plate unit of the Dinarides and Hellenides, together with the Dacia mega-unit that they tectonically overlie since the end of the Jurassic (Schmid et al., 2008; profiles in Figs. 13 and 15). In map view, north of Belgrade (Fig. 12), they connect across the subsurface of the southernmost Pannonian basin in northernmost Serbia (Kemenci and Čanović, 1999; Čanović and Kemenci, 1999) with the ophiolites of the south Apuseni Mountains (Săndulescu, 1984 and 1995; Balintoni, 1994; Nicolae and Saccani, 2003; Maţenco and Radivojević, 2012; Kounov and Schmid et al., 2013; Reiser et al., 2017; Gallhofer et al., 2017) rather than further following the Sava suture zone and the Western Vardar ophiolitic units westwards towards Zagreb (Figs. 1 and 12; Plate 1). In the Apuseni Mountains these ophiolites tectonically overlie the Biharia nappe system (Balintoni, 1994; see profile of Fig. 6). Between Belgrade and Pristina, the Eastern Vardar ophiolites are known as “Central Vardar Subzone” (Dimitrijević, 1997) or “Main Vardar Zone” (Karamata, 2006) and here they structurally overlie the Serbo-Macedonian unit (Schmid et al., 2008; Figs. 12 and 13). As is the case for the Eastern Vardar ophiolites of the south Apuseni Mountains, also the

Belgrade-Pristina stretch of Eastern Vardar ophiolites is stratigraphically overlain by uppermost Jurassic to Berriasian reef limestones overstepping these ophiolites after their emplacement onto the Serbo-Macedonian unit (Şerban et al., 2004; Bucur and Săsăran, 2005; Karamata, 2006; Toljić et al., 2018). This is followed by the deposition of Valanginian to Aptian "paraflysch" sediments (Dimitrijević, 1997) that resemble the cyclic deposition of turbidites but lack their gradational and internal organization characteristics (Toljić et al., 2018). In the Belgrade area this paraflysch grades into Albian to Cenomanian shallow water and continental deposits, overlain by Turonian to Maastrichtian turbidites (Toljić et al., 2018 and references therein). In contrast to the Western Vardar ophiolites, metamorphic soles and mélange formations are only occasionally found between Belgrade and Pristina; in the south Apuseni Mountains they have not been found at all.

An important along-strike change occurs east of Pristina in the area of Novo Brdo (Pavić et al., 1974). In this area the Eastern Vardar ophiolitic unit is laterally replaced by the northernmost part of the Circum-Rhodope unit, namely by a greenschist facies unit derived from continental crust and composed of schists and marbles known as Veles Series in the literature (Fig. 12). The Veles Series are dated as Devonian to middle Carboniferous by palynomorphs in the area around Veles located SE of Skopje (Fig. 14) in North Macedonia (Grubić and Ercegovac, 2002). A NW-dipping tectonic contact separates the Belgrade-Pristina segment of the Eastern Vardar ophiolitic units from the Veles Series. Both of them are tectonically overlain by the Serbo-Macedonian unit along a NE-dipping thrust of Cenozoic age. South of Pristina this Veles series basement complex is imbricated with ophiolitic slices we attribute to the Eastern Vardar ophiolites (Figs. 11, 14 and 15) and this zone of imbrication can be followed all the way from Kosovo to the border between North Macedonia and Greece near Lake Dojran, joining what is known under the names "Circum Rhodope belt" or "Peonias zone" in Greek Macedonia (Kockel et al., 1971; Kockel and Mollat, 1977a, 1977b; Mercier and Vergély, 2002). In northern Greece the stratigraphy of the Permo-Mesozoic sediments that stratigraphically overlie the Veles Series is well established (e.g. Mercier, 1966; Kockel, 1986; Meinhold et al., 2009 and 2010): This Permian to Middle Jurassic sequence, characterized in its upper part by flysch-type Jurassic deposits (Svoula flysch, also known as Melissochori Formation) and metamorphosed under lower greenschist facies conditions, is unconformably overlain by non-metamorphic cover with uppermost Jurassic often terrestrial, occasionally shallow-water sediments (Kauffmann et al., 1976). The character of the Jurassic sediments and the Upper Jurassic unconformity are testimony of a Late Jurassic orogenic event that also affected the westerly adjacent Paikon unit including the Tzena massif, exposed in the Greek-North Macedonian border area (Mercier and Vergély, 2002, and references therein). Note that we also included the Paikon unit and Tzena massif into what we mapped as Circum-Rhodope unit in Plate 1 and Fig. 14. This same latest Jurassic event also affected the Eastern Vardar ophiolites since they are also unconformably overlain by Upper Jurassic overstep sediments in the case of the Demir Kapija ophiolite in North Macedonia close to the border with Greece (Kukoč et al., 2015), representing the northern continuation of the well-known Guevgueli ophiolite on Greek territory (Saccani et al., 2008). Hence the Eastern Vardar ophiolites, together with slices of continental crust of the Circum-Rhodope unit, were affected by Late Jurassic orogeny. This makes it likely that the imbrication of ophiolitic and non-ophiolitic units, typical for the Circum-Rhodope unit all the way to Thrace in NE Greece, took place during that same Late Jurassic tectonic event. Note that in Plate 1 and Fig. 14 we mapped occurrences of continental crust as Circum-Rhodope (part of the Dacia mega-unit, see Fig. 1) but kept the name Eastern Vardar ophiolite for the interleaved ophiolitic slivers. It is important to point out that while, in present-day coordinates, Late Jurassic orogeny was top-E, the Circum-Rhodope unit and interleaved ophiolites are presently E-dipping (see profiles in Figs. 11 and 15) since they became reworked by top-W imbrication during the Cenozoic and after closure of the northern branch of Neotethys as first pointed out by Ferrière and Stais (1995).

The Eastern Vardar ophiolites are distinct from the Western Vardar ophiolites in many respects. They not only consist of ultramafic and mafic rocks but also include large bodies of granitoids and volcanics (Šarić et al., 2009; Nicolae and Saccani, 2003). These yielded Oxfordian to Kimmeridgian U–Pb ages (159–156 Ma for ophiolites and 158–153 Ma for granitoids) in the case of the South Apuseni ophiolites (Gallhofer et al., 2017). The ophiolites, as well as the granitoids further south in Greek Macedonia and Thrace yield older ages (as old as Middle Jurassic) according to the literature (see compilation in Gallhofer et al., 2017; their Fig. 9). These ages are confirmed by the ages of radiolarites stratigraphically covering the ophiolites (late Bathonian to early Callovian in case of Demir Kapija, Kukoč et al., 2015; Callovian to Oxfordian in the case of the Apuseni Mountains, Lupu et al., 1995). The Upper Jurassic granitoids and their subvolcanic and volcanic equivalents originated in an island arc setting as indicated by geochemical evidence (Savu et al., 1981; Bortolotti et al., 2002 and 2004; Nicolae and Saccani, 2003; Božović et al., 2013; Boev et al., 2018). This is supported by field evidence from the Paikon anticlinorium (Brown and Robertson, 2003), which is westerly adjacent to the Guevgueli ophiolite (Saccani et al., 2008). The Mesozoic sediments of the Paikon area are characterized by interlayered Jurassic andesitic volcanics and volcanoclastics. These are in term part of the stratigraphic cover of a continental basement, which indicates an island arc scenario installed on continental crust (Vergély and Mercier, 2000; Mercier and Vergély, 2002; Brown and Robertson, 2004; Saccani et al., 2015). We mapped this continental crust as a part of the larger Circum-Rhodope unit as outlined (Plate 1 and Fig. 14). A possible explanation was suggested by Brown and Robertson (2004; their fig. 13): The Almopias ophiolites (mapped as a part of the Western Vardar branch of Neotethys) were subducted beneath the margin of the European continent (Circum-Rhodope unit of the Dacia mega-unit), generating the subduction-related arc volcanism observed in in the Paikon area. This was associated with the opening of an oceanic back-arc basin in which the ophiolites of the Eastern Vardar ophiolitic units formed, such as, for example, the Guevgeli ophiolite presently located east of the Paikon island arc sequences. In the case of the south Apuseni Mountains the back-arc ocean behind the island arc opened within pre-existing oceanic lithosphere (see Fig. 2 and Gallhofer et al., 2017, their Fig. 10). Soon after the opening of this back-arc ocean in the Middle Jurassic this narrow oceanic domain must have closed in late Kimmeridgian to Tithonian times when the Eastern Vardar ophiolites were obducted onto the Dacia mega-unit. This is indicated by the age of the overstepping shallow water limestones that are widespread in the Apuseni Mountains (Bortolotti et al., 2002; Şerban et al., 2004) and in North Macedonia (Kukoč et al., 2015). Alternatively, the Eastern Vardar ophiolites may be explained by a W-dipping intra-oceanic subduction zone that culminates in eastward obduction and subsequently on-going continental subduction forming the Balkan orogen (van Hinsbergen et al., 2019).

In summary, the Eastern Vardar ophiolites were issued from a short-lived back arc basin, intra-oceanic in the north and intra-continental in the south, which closed soon after its generation due to top-E obduction onto the Biharia nappe system in the Apuseni Mountains, the Serbo-Macedonian unit in Serbia and the Circum-Rhodope unit in North Macedonia and Greece, which are all part of the Dacia mega-unit). Note that in the case of North Macedonia and Greek Macedonia the present-day architecture is generally characterized by steeply E-dipping and W-facing imbrications (see profiles of Figs. 11 and 15) that are due to later deformations in Cenozoic time, i.e. after the Sava Ocean closed. It is also important to mention that the northern Romanian segment of the Eastern Vardar ophiolites, which includes the Apuseni Mountains, was subsequently emplaced during the Early Cretaceous in the highest structural position overlying the Bucovinian nappes of the Eastern Carpathians (see profile C of Fig. 6). The Sava suture zone marking the suture zone of the Sava Ocean, located west of the Eastern Vardar and Circum-Rhodope units, on the other hand, is located between the Almopias (Western Vardar) and Guevgelija (Eastern Vardar) ophiolites in the border area between North Macedonia and Greece.

3.3.6. Anatolide-Tauride Ophiolites of western Turkey

The **ophiolites of western Turkey** located south of the Sava-İzmir-Ankara-Erzincan suture zone were obducted onto Mesozoic sediments of the Anatolide-Tauride carbonate platforms (Fig. 16). As is the case for the obducted ophiolites of the Dinarides-Hellenides also the Anatolide-Tauride ophiolites often exhibit a metamorphic sole at the base of the mantle rocks, underlain by an ophiolitic mélange. However, the age of the metamorphic sole and overlying ophiolites differs being Late Cretaceous instead of Jurassic (Parlak, 2016; Çelik et al., 2006). The majority of these ophiolites was issued from the Sava-İzmir-Ankara-Erzincan suture zone (Parlak and Robertson, 2004), i.e. from Neotethys, and transported southward over considerable distance over the Tauride margin (some 200 km in the case of western Turkey; Plunder et al., 2013 and 2016). The northernmost continental units of Greater Adria (the Tavşanlı unit) and subducted adjacent oceanic units became metamorphosed together with these sediments under high-pressure metamorphic conditions related to the N-directed subduction of the Anatolide-Tauride block closely following intra-oceanic subduction and incipient obduction (Plunder et al., 2016). However, the ophiolites of Antalya, together with the Alanya nappes, are part of a thrust belt that was emplaced from south to north (Okay and Özgül, 1984); i.e. they originate from the Eastern Mediterranean Ocean (referred to as southern branch of Neotethys or Southern Neotethys by some authors; e.g. Robertson et al., 2006) along with ophiolites farther east (e.g., Göksun Ophiolite) (Maffione et al. (2017). This northward obduction followed westward radial rollback of an intra-oceanic subduction zone, and also emplaced ophiolites southward onto the northern African and northwestern Arabian margin (e.g., Cyprus, Hatay, and Syria) (Stampfli and Hochard, 2009; Maffione et al., 2017; McPhee and van Hinsbergen, 2019), located east of our area of interest. According to most authors this southern ophiolite belt was issued from the southern branch of Neotethys (see chapter 3.7.7).

Within the area covered by Plate 1 and Fig. 16 there is a northernmost group of ophiolites overlying and/or imbricated with sediments that are part of the Tavşanlı unit. The age of intra-oceanic subduction is inferred from Lu/Hf garnet formation ages of the metamorphic soles of ca. 104 Ma in case of the Halilbağı ophiolite (Pourteau et al., 2019), an age similar to those reported farther east in Turkey (Peters et al., 2018) as well as for Oman (Guilmette et al., 2018), and approximately 10–14 Ma older than the exhumation age of the crust of the suprasubduction zone ophiolites, and older than the cooling ages of the soles measured with Ar/Ar geochronology (van Hinsbergen et al., 2016, and references therein). The age of blueschist to eclogite facies metamorphism of non-ophiolitic rocks of the Tavşanlı unit is estimated at ca. 90–80 Ma (Okay et al., 1998; Pourteau et al., 2019; Seaton et al., 2014; Mulcahy et al., 2014). Farther south, the obducted ophiolites are associated with non-ophiolitic rocks of the Afyon-Ören unit that underwent high-pressure metamorphism at around 65 Ma (Pourteau et al., 2013, 2014; Özdamar et al., 2013). A southern group of ophiolites referred to as Lycian ophiolites exhibits metamorphic soles dated at ca. 93–90 Ma old (Çelik et al., 2006). These Lycian ophiolites overlie non-metamorphic ophiolitic mélange and carbonates of the Lycian nappes. This trend extends eastwards towards the Central Taurides, where non-metamorphic nappe accretion continued until the middle to late Eocene (McPhee et al., 2018a, 2018b).

It thus appears that the age of the intra-oceanic subduction initiation inferred from the ages of the metamorphic sole remains practically unchanged going from north to south, while the age of subduction and accretion of the underlying sediments decreases going in the same direction, the sediments of the Lycian nappes being non-metamorphic. This suggests continuous accretion of the Taurides underlying these ophiolites from ca. 85 to 35 Ma. There is no compelling reason to postulate that the ophiolites root in an “Intra-Tauride” oceanic domain postulated to have existed between Tavşanlı and Afyon-Ören units (e.g. Dilek et al., 1999; Robertson et al., 2009b; Parlak et al., 2013; Menant et al., 2016). Candan et al. (2005) showed that the sediments of the Tavşanlı and Afyon-Ören units represent the proximal and distal parts of the same margin. The reader is referred to the discussion regarding this controversy presented in Pourteau et al. (2016), van Hinsbergen et al., (2016), and Maffione et al. (2017).

There is a general consensus amongst authors that the ophiolites along the Tauride ophiolite belt formed in supra-subduction zone (SSZ) settings in the sense that the ophiolites that later became obducted were generated above an intra-oceanic subduction zone (see Parlak, 2006 for a review and references). However, the generation of the supra-subduction ophiolites issued from the northern branch of Neotethys must have taken place within a pre-existing older oceanic crust characterized by a different geochemical signature since the geochemistry of the metamorphic sole amphibolites suggests magma derivation from seamount-type alkaline basalt, mid-ocean ridge basalt (MORB) and island arc basalt (Parlak, 2006). In fact, Early to Middle Jurassic intra-oceanic subduction is known from metamorphic sole amphibolites found within the Sava- İzmir-Ankara-Erzincan suture zone NE of Ankara, and thought to have formed in the southern Eurasian forearc (Çelik et al., 2011; Topuz et al., 2014). This strongly supports the model proposed by Gürer et al. (2016) claiming that a separate and (mostly) oceanic plate, referred to as Anadolu plate, formed above an intra-oceanic subduction zone, became isolated upon Late Cretaceous intra-oceanic subduction initiation (ca. 105 Ma ago) within the pre-existing northern branch of the Neotethys (the so-called Sava Ocean that denotes the Cretaceous-age relict ocean of the northern branch of Neotethys, as is discussed in the following chapter).

3.3.7. The suture zone of the northern branch of Neotethys (Sava-İzmir-Ankara-Erzincan suture zone)

The term **Sava suture zone** was used by Schmid et al. (2008) and Ustaszewski et al. (2009, 2010) to denote a belt of magmatic and metamorphic rocks, Cretaceous flysch-type sediments and locally also gabbros and basalts that stretches from Zagreb over Belgrade and along the Vardar-Axios river all the way to the Gulf of Salonika. It corresponds to the Sava-Vardar zone of Pamić et al. (2002), who wrote: “The Sava-Vardar zone (SVZ), about 1000 km long, represents the most internal tectonostratigraphic unit of the Dinarides and Hellenides”. This suture zone divides the Europe-derived Tisza and Dacia mega-units to the NE and E from the Adria-derived Dinarides-Hellenides to the SW and W (Figs. 12 and 14). The latter represent the lower plate with respect to the Tisza and Dacia mega-units during subduction in Late Cretaceous time that was followed by lower amounts of deformation into the Paleogene (Ustaszewski et al., 2010; Gallhofer et al., 2015; Stojadinović et al., 2017; Toljić et al., 2018).

The northwestern end of the Sava suture zone sharply turns into the SW-NE strike of the Mid-Hungarian fault zone in the area of Zagreb (Fig. 12) but outcrops are rare. This suture zone is buried below the Cenozoic cover of the Pannonian basin. A few exposures (e.g. Slovenec et al., 2011) and information from drill holes (Haas et al., 2000) clearly show that sediments of the Sava suture zone, together with remnants of the Western Vardar ophiolites and equivalents of the Mesozoic of the Jadar-Kopaonik unit that represent the distal-most Adriatic margin, turn in the area of Medvednica Mountain by >90° (Tomljenović et al., 2008; van Gelder et al., 2015) into parallelism with the Mid-Hungarian fault zone (Fig. 5). Ophiolites are also found in rare exposures and boreholes along the SW end of the Mid-Hungarian fault zone (Fig. 12; Haas et al., 2000; their Kalnik Zone) providing a link of the Western Vardar ophiolites with ophiolites found in the area of the Bükk Mountains (Szavarskö and Darno ophiolites and mélange) in northern Hungary (Aigner-Torres and Koller, 1999; Fodor et al., 2005; Kovacs and Haas et al., 2010; Kiss et al., 2011; Haas et al., 2012). The striking similarity of the Permo-Carboniferous and Triassic formations of the Bükk Mountains that structurally underlie the ophiolites with the corresponding formations of the Dinarides (Jadar-Kopaonik unit) has been reported by many authors (e.g. Csontos, 1999 and 2000; Dimitrijević et al., 2003; Filipović et al., 2003; Haas et al., 2012; Petrik et al., 2016b). The trace of the Sava suture zone along the Mid-Hungarian fault zone must run between these Dinaric elements displaced eastward during Miocene lateral extrusion and located to the north of this suture zone, and the Tisza mega-unit SSE of it. Mapping of the Sava suture zone that joins the suture zone of Alpine Tethys along the Mid-Hungarian fault zone (Plate 1 and Fig. 5) is of course rather schematic and solely based on the subsurface map of Haas et al. (2010) constructed on the basis of borehole data that are not always unambiguous.

A first outcrop belonging to the Sava suture zone in the Dinarides is found southeast of Zagreb in Moslavačka Gora (Fig. 12). We included this inselberg tentatively in the Sava suture zone because it consists of a Cretaceous S-type granite pluton intruding a Cretaceous low-pressure/high-temperature (LP/HT, emplacement at ~720 °C, Balen and Broska, 2011) metamorphic envelope whose age is estimated at ca. 90–100 Ma using the method of electron microprobe-based monazite dating (Starijaš et al., 2010). The Central Granite was dated at 82 ± 1 Ma (LA-SF-ICP-MS zircon age) while Balen et al. (2003) dated a gabbro at 109 ± 8 Ma. The Cretaceous high heat flow regime recorded in the Moslavačka Gora is unique in the subcrop of the Pannonian basin and may be a local feature triggered by a mafic intrusion in the lower crust.

The best outcrops of the Sava suture zone are found in inselbergs sticking out the Pannonian fill in northern Bosnia (North Kozara, Prosara, Motajica) and in the area around and south of Belgrade (Fig. 12). In inselbergs that are surrounded by Miocene deposits Upper Cretaceous siliciclastic sediments crop out as parts of the Sava suture zone, deposited on the Adriatic continent and incorporated into an accretionary wedge that evolved during the initial stages of continent-continent collision. Structurally deeper parts of the accretionary wedge, exposed in northern Bosnia, became exhumed during core complex formation in the Miocene. These underwent metamorphism reaching amphibolite-grade metamorphism affecting uppermost Cretaceous sediments at around 65 Ma (Ustaszewski et al., 2010). This area also exposes a bimodal igneous succession comprising isotropic gabbros, doleritic dikes, basaltic pillow lavas and rhyolites (Ustaszewski et al., 2009). Pelagic limestones, intercalated with pillow lavas, yielded a Campanian globotruncanid association, consistent with concordant U–Pb ages on zircons from dolerites and rhyolites of ca. 81 Ma. This intra-oceanic magmatism, together with pelagic sedimentation onto possibly oceanic crust in the Sava suture zone of northern Bosnia indicates that a deep-marine seaway persisted at least until the Campanian (Karamata et al., 2000 and 2005; Grubić et al., 2009; Ustaszewski et al., 2009; Vishnevskaya et al., 2009). Based on geochemical evidence the origin and geodynamic significance of this magmatism have been considered as intra-oceanic and associated with an ocean island or back-arc basin setting (Pamić et al., 2002; Pamić et al., 2002; Ustaszewski et al., 2009). Cvetković et al. (2014) also concluded that these ophiolites of northern Bosnia are not indicative for an N-MORB and suggested that the ophiolitic segment of the North Kozara represents part of an anomalous mid-ocean ridge or, alternatively, an island plateau that was located not far from the ridge.

The Sava suture zone can be followed farther into a narrow strip north of Fruška Gora (northwest of Belgrade; Fig. 12) whose location is only loosely constrained by subsurface information (Kemenci and Čanović, 1999; Čanović and Kemenci, 1999; Maţenco and Radivojević, 2012). Fruška Gora exposes a metamorphic core containing a typical Triassic to Jurassic sequence of the distal Adriatic margin that is overlain by remnants of the Western Vardar ophiolites and Upper Cretaceous to Paleogene sediments. All this is preserved in the core of an antiform formed during late Miocene inversion of the Pannonian basin (Toljić et al., 2013). South of Belgrade, the Sava suture zone abruptly bends into the N-S strike (Fig. 12) observed widely in the Belgrade area and southwards (Toljić et al., 2018). There the Sava suture zone is, however, not easy to define since the Upper Cretaceous sediments covering Western Vardar ophiolites and Eastern Vardar ophiolites adjacent to the suture zone are virtually indistinguishable in the field from those that define the Sava suture zone. Starting from the alternative interpretation of a main contact along the large-offset Bela Reka fault (Toljić et al., 2018), Bragin et al. (2019) have reinterpreted the exact location of the Sava suture zone along a diffuse boundary between the location of scarce remnants of the Western and Eastern Vardar ophiolites that rim the Sava suture zone to the west and east. As proposed by Toljić et al. (2018) the Upper Cretaceous sediments, exhibiting a substantial thickness of 3–4 km in the Belgrade area, were deposited, from west to east, over the subsiding passive margin of Adria including the overlying Western Vardar ophiolites, a trench representing the Sava suture zone. Additionally, a Cretaceous to earliest Paleogene basin was installed on the upper “European” plate, i.e. the Eastern Vardar ophiolites and underlying Serbo-Macedonian unit. A Campanian to Maastrichtian pelagic cover, well dated with foraminifera, is also reported for the Eastern Vardar unit north of Belgrade in the subsurface of the Pannonian basin (Duncić et al., 2017), supporting the idea of a forearc setting of the Sava suture zone and adjacent units proposed by Toljić et al. (2018). Mantle-derived lamprophyres intrude Upper Cretaceous sediments in a quarry south of Belgrade (Tešića Majdan; see Toljić et al., 2018 for location). The intrusion ages (85 Ma; Sokol et al., 2017) are very close to those reported for dolerites and rhyolites from the Sava suture zone in northern Bosnia by Ustaszewski et al. (2009). The lamprophyres most probably intruded in a forearc setting.

South of Belgrade, the Sava suture zone and the Cretaceous to earliest Paleogene forearc basin can be followed by the discrimination between deep-water turbidites in the former and a shallower water clastic-carbonatic sequence in the later (Toljić et al., 2018). In the Kopaonik area (Fig. 12), the Sava suture zone is defined by strongly deformed and isoclinally folded sub-greenschist facies metamorphic post-Turonian flysch-type sediments referred to as Senonian flysch. Pebbly mudstones, turbiditic and hemipelagic “Scaglia”-type hemipelagic sediments (Brzeće unit) are exceptionally well exposed along the road between Brzeće and Brus (Zelic, 2004; Zelic et al., 2010; Schefer et al., 2010; Schefer et al., 2011). These sediments contain large olistoliths, including ophiolites and metamorphic rocks. Some olistoliths yielded Globotruncanides yielding a Campanian to Maastrichtian and possibly Paleogene age. Sigma clasts indicate top-W-oriented shear senses, later overprinted by E-dipping normal faults. These post-Turonian sediments unconformably transgress the Western Vardar

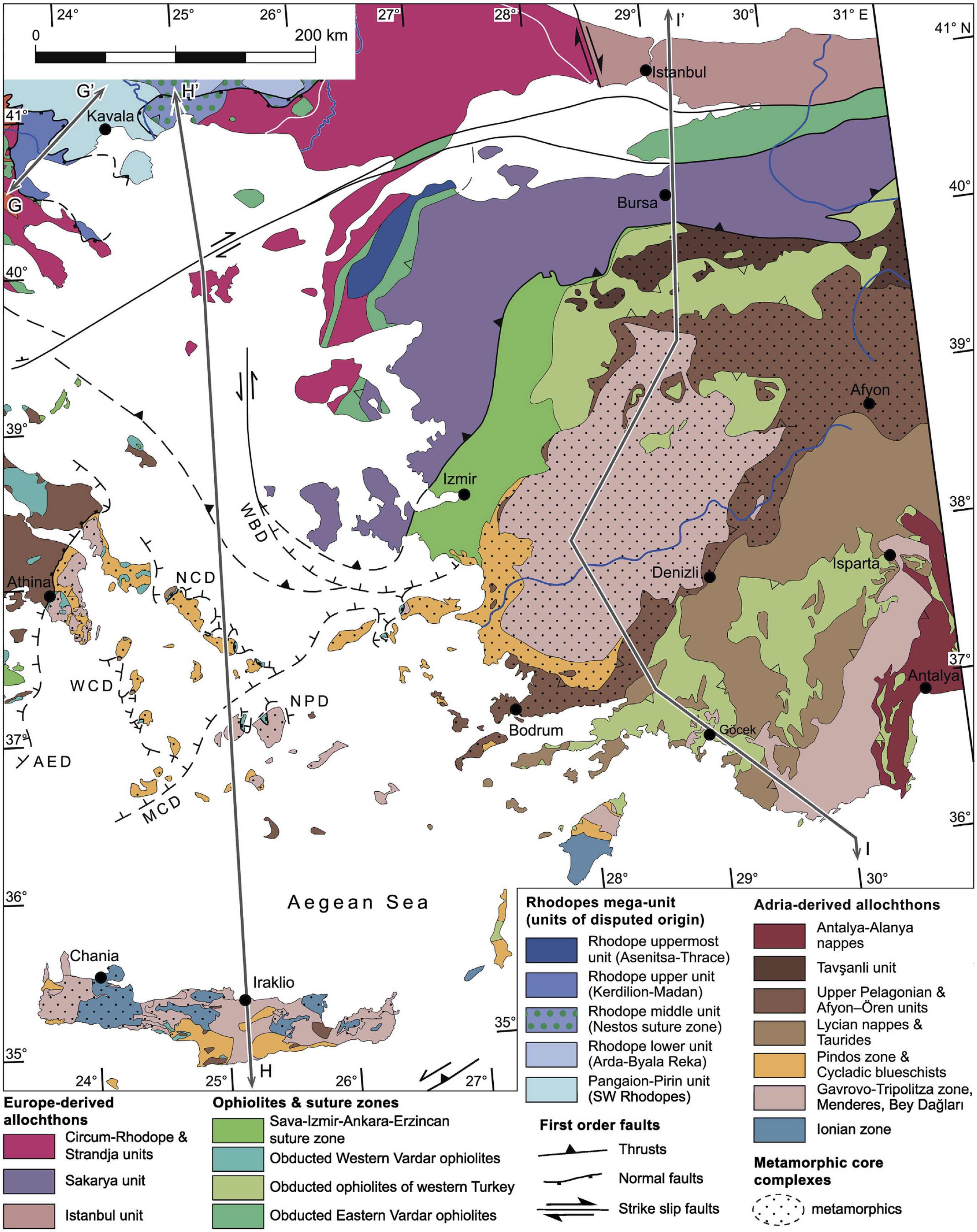

Fig. 16. Tectonic map of the Aegean islands and western Turkey; extract of Plate 1. Abbreviations: AED: Attica-Evvia detachment; MCD: Mid-Cycladic detachment; NCD: North Cycladic detachment; NPD: Naxos-Paros detachment; WBD: West Biga detachment; WCD: West Cycladic detachment.

ophiolites and underlying sediments of the Jadar-Kopaonik unit with a basal breccia containing ophiolitic detritus as well as previously metamorphosed components. A tectonic window exposing rocks belonging to the Sava suture zone below the Serbo-Macedonian unit in the hanging wall is exposed in the Jastrebac window located some 180 km SSE of Belgrade Fig. 12). This window exposes Upper Cretaceous to Paleogene psephites, psammites and pelites, which are regionally metamorphosed under sub-greenschist to lower greenschist facies conditions (Maţenco et al., 2007), separated from the overlying Serbo-Macedonian unit by a strongly tectonized zone, too thin to be mapped in Plate 1 and made up of lithologies parallelized with those occurring in the Western Vardar ophiolites and underlying distal-most Adriatic margin (Erak et al., 2017). This core complex that presently forms a window following final exhumation during Miocene extension, nicely documents the allochthony of the Serbo-Macedonian unit as a part of the Dacia mega-unit (upper plate) over units of the Dinarides (lower plate) and the Sava suture zone.

The Late Cretaceous flysch trough adjacent to the Sava suture zone can be followed southward into North Macedonia (Fig. 14) where the Upper Cretaceous sediments are seen to directly transgress the pre-Triassic basement of the Pelagonian massif, starting with a Turonian basal conglomerate (Robertson et al., 2013; Prelević et al., 2017; Rakicević et al., 1965, map sheet Prilep). Interestingly, for Kossmat (1924) this Upper Cretaceous flysch, a post-orogenic deposit with respect to Early Cretaceous deformation in the Pelagonian massif (his "Gosau und Flyschfazies"), formed an integral part of what he defined as Vardar zone. The Sava suture zone follows the eastern boundary of this belt of Upper Cretaceous flysch and is marked by subvertical, heavily tectonized and dismembered slices of Mesozoic rocks that are attributed to the Jadar-Kopaonik unit by along-strike correlations. These slices are imbricated with the ca. 10 km long and 1–2 km wide so-called Klepa block (Prelević et al., 2017) consisting of ca. 80 Ma old pillows, sheet flows, dikes and cumulates, locally overlain in stratigraphical contact by sheared reddish clay-rich Scaglia Rossa hemipelagic limestones of presumably Campanian age, by analogy with the rocks of the Sava suture zone in Bosnia (North Kozara Mts.; Ustaszewski et al., 2009 and 2010). These are covered by sandy (possibly Maastrichtian) conglomerates. Based on detailed geochemical analyses, Prelević et al. (2017) discussed two possible geodynamic scenarios for the intrusion of the Klepa basalts: (1) magmatism along this zone would be situated within the forearc region of the subducting Sava Ocean, triggered by ridge subduction (2) extension-driven partial melting at the asthenosphere- lithosphere boundary and intrusion into intra-continental crust post-dating an earlier closure of the Neotethys Ocean between Adria and Europe. An island plateau or seamount origin within the subducting Sava Ocean, as proposed by Cvetković et al. (2014) for the North Kozara Mountains in northern Bosnia, discussed earlier, is a third possible scenario. In view of the absence of any evidence for such intra-continental settings along the other segments of the Sava suture zone (see above) and strong evidence for Late Cretaceous subduction of oceanic lithosphere below the Dacia mega-unit (Gallhofer et al., 2015; see discussion below) we prefer to exclude the second hypothesis of Prelević et al. (2017) although it is tenable from a purely geochemical point of view. Additionally, the observation of a forearc basin in the Belgrade area linked to similar Santonian to early Campanian magmatism (Toljić et al., 2018) favors magmatism related to the subduction of the Sava oceanic domain (first hypothesis of Prelević et al., 2017, and model proposed by Cvetković et al., 2014). The idea of the Klepa block representing the Sava suture zone is also compatible with the observation that along its eastern boundary the Klepa Block is in direct tectonic contact with mylonitized schists and marbles of the Veles series that are part of the Circum-Rhodope unit, and hence part of the northeastern margin of the Neotethys.

Farther south the Sava suture zone is again well exposed in the border area between North Macedonia and Greece (Fig. 14), i.e. in the Voras Mountains (Brown and Robertson, 2004; Katrivanos and Kilias, 2013; Katrivanos et al., 2016). An Upper Cretaceous sequence of conglomerates, grading upward into flysch (Fidopetra flysch) transgressively overlies the peridotites and basalts of the Ano Garevi ophiolite (Saccani et al., 2015), which is an E-dipping slice of Western Vardar ophiolites known as a part of the so-called "Almopias ophiolites". This strongly tectonized flysch marks the Sava suture zone since it occurs in the footwall of a major E-dipping thrust at the base of the Paikon unit (Pinovon unit; Brown and Robertson, 2004), i.e. a unit we map as part of the Circum-Rhodope unit. Southwards this flysch that marks the Sava suture zone, is followed into a Turonian to Maastrichtian flysch. This flysch is dated with radiolaria (Sharp and Robertson, 1993; Brown and Robertson, 2003; Katrivanos et al., 2016) and separates the Paikon unit from the so-called Krania and Mavrolakkos units (easternmost slices of the Almopias ophiolites) along an E-facing backthrust of the Almopias ophiolites over the Paikon anticlinorium (Mercier and Vergély, 2002; Sharp and Robertson, 1993). This backthrust overprinted the previously formed Sava suture zone.

Some authors deny (e.g., Csontos and Vörös, 2004) or cast doubt on (e.g. Prelević et al., 2017) the existence of a Sava oceanic domain that closed and formed the Sava suture zone at the end of the Cretaceous as proposed by Schmid et al. (2008), claiming that the northern branch of Neotethys already closed in the Late Jurassic. This is because (1) ophiolitic remnants are very rare within the Sava suture zone and (2) because the geochemical signature of these remnants is definitely different from that of the Western Vardar or Eastern Vardar ophiolites (see discussion above and in Prelević et al., 2017). Apart from the clear deep-water character of the Cretaceous sediments in the Sava suture zone separating two zones of distinct Adriatic and European Middle Triassic to Jurassic depositional facies (Schefer et al., 2010; Ustaszewski et al., 2010; Maţenco and Radivojević, 2012; van Gelder et al., 2015), the most compelling argument in favour of the existence of a Late Cretaceous ocean is an indirect one. The Late Cretaceous Apuseni-Banat-Timok-Sredna Gora magmatic arc in southeastern Europe is the westernmost arc in the Alpine-Himalayan orogenic system is related to subduction of Neotethys according to most authors (e.g., Berza et al., 1998; Popov et al., 2002; von Quadt et al., 2005). This magmatic arc extends over 1000 km length from the Apuseni Mountains of Romania, through Serbia and Bulgaria to the Black Sea. It most probably formed during NE to N-directed (in present-day coordinates) subduction of oceanic crust below the Dacia mega-unit in an Andean-type scenario (e.g., Gallhofer et al., 2015). Arc magmatism, associated with co-magmatic sedimentation, was active for 25 Ma (~92–67 Ma) and stopped before the formation of the Sava suture zone at the Cretaceous/Cenozoic boundary. Gallhofer et al. (2015) estimated a width of some 300 km for the Cretaceous remnant of the northern branch of Neotethys referred to as Sava Ocean. However, as discussed above, the oceanic remnants preserved along the trace of the Sava suture zone indicate that these remnants either formed in an island plateau (Cvetković et al., 2014) or forearc setting (Prelević et al., 2017, his scenario 1; Toljić et al., 2018). This implies that all remnants of the Cretaceous Sava Ocean that may have formed in a MORB-type environment, or in a supra-subduction scenario, must have been subducted without being accreted in the suture zone, which is actually normal for an Andean-type subduction system. An additional indirect line of evidence for the existence of a Cretaceous oceanic crust within the northern branch of Neotethys is of course provided by the intra-oceanic subduction initiating at ~105 Ma in western Turkey, leading to the obduction of the ophiolites of Anatolia (Gürer et al., 2016; see discussion in the previous chapter).

The exact trace of the Sava suture zone across the Aegean Sea is unknown due to the severe post-orogenic displacements and rotations during late Cenozoic extension in the Aegean region (van Hinsbergen and Schmid, 2012). In Plate 1 and Fig. 16 we map the northern rim of the Sava suture zone such as to run immediately north of the islands of Alonissos (Jacobshagen and Matarangos, 2004) and Skyros (Karkalis et al., 2016), built up of Western Vardar ophiolites and transgressive Upper Cretaceous formations, and south of Chios island (Meinhold et al., 2007, 2008) and the Karaburun Peninsula in western Turkey (Robertson and Pickett, 2000) that we attribute to the Sakarya unit. The Adheres unit outcropping as the uppermost tectonic element in the southernmost tip of Argolis (Bortolotti et al., 2003) may represent an Upper Cretaceous to lower Paleogene mélange unit that we mapped as a part of the Sava suture zone (Plate 1), displaced to its present position during the opening of the Aegean Sea.

In western Turkey, the Sava suture zone connects with the SW-NE striking Bornova flysch unit that we consider an integral part of the same suture zone we refer to as **Sava-İzmir-Ankara-Erzincan suture zone** (Fig. 16). According to our interpretation this suture zone separates the Karaburun Peninsula in the northwest, which we attribute to the Sakarya unit (see Okay et al., 2012 for an alternative view), from the Cycladic and Menderes massifs to the SE. Note that, in contrast, Okay et al. (2012) consider the İzmir Ankara- Erzincan suture a discrete fault located NW of the Bornova flysch zone. The Bornova flysch zone represents a mélange unit containing large blocks of Mesozoic limestone and Middle Triassic to Cretaceous oceanic units enclosed in a Maastrichtian to lower Paleocene matrix of sheared sandstone and shale (Okay et al., 2012; Moix and Goričan, 2013). NE-ward this non-metamorphic mélange formation wedges out and terminates between the HP/LT metamorphic Tavşanlı unit to the south that consists of sediments of the northernmost passive continental margin of the Anatolide–Tauride Block overlain by Cretaceous ophiolites (Okay et al., 1998; Plunder et al., 2013) and the northerly adjacent Karakaya accretionary complex that is an integral part of Sakarya. In this northeastern area Taurides and Pontides are merely separated from each other by a very thin veneer of tectonites that represent the İzmir-Ankara-Erzincan suture zone that turns into an E-W strike all the way to central Anatolia, representing the eastward continuation of the Sava suture zone. Like most authors, we consider the Sava-İzmir-Ankara-Erzincan suture zone that separates the Sakarya zone and the rest of the Pontides from the Taurides, including the Kırşehir block that is only exposed east of the area mapped in plate 1, as a suture zone that closed the northern branch of Neotethys (e.g. Topuz et al., 2013). Note however, that some authors working in the eastern Pontides and analysing magmatic rocks related to Tethys subduction in an area far outside the area mapped in plate 1 (e.g. Eyuboglu et al., 2015 and 2018), cast doubt on this interpretation and postulate a persisting S-dipping Tethyan Ocean located north of the Pontides. It is important to note that the Sava-İzmir-Ankara-Erzincan suture zone of western Turkey not only preserved relics of Cretaceous age ophiolites but also Jurassic ophiolites (Topuz et al., 2013) providing testimony of a Jurassic part of the ocean that formed the northern branch of Neotethys. Furthermore, Tekin et al. (2002), found Triassic radiolarians along the Sava-İzmir-Ankara-Erzincan suture zone in the vicinity of the eastern margin of the map presented in Plate 1, suggesting that the northern branch of Neotethys already opened in Triassic times, analogous to the relics of a Triassic Neotethys found along the Sava suture zone of the Dinarides and Hellenides.

3.4. Adria-derived units

3.4.1. South Alpine unit and its eastern extension into the Pannonian basin

The **South Alpine unit** of northern Italy, located south of the Periadriatic fault system (Fig. 2), formed during dominantly Neogene top-S thrusting of basement and cover over the undeformed foreland of the Adriatic continent. The geometry of this stack of imbricates is thin-skinned style (Pfiffner, 2006) whereby thrust faults also cut down into the crystalline basement and level off some 5 km beneath the basement-cover interface (Schönborn, 1992 and 1999; Nussbaum, 2000). Variscan basement is unconformably overlain by Upper Carboniferous to Permian sediments followed by a thick sequence of Mesozoic to Cenozoic cover sediments similar to those of the external Dinarides in many places. West of the Miocene Giudicarie belt, sinistrally displacing the Periadriatic fault by some 80 km (Frisch et al., 2000; Linzer et al., 2002; Scharf et al., 2013), some of the top-S deformation in the South Alpine unit is pre-Neogene (Schönborn, 1992). Dominantly S-vergent Neogene thrusting is considered to represent a retro-wedge with respect to the N-vergent main body of the Alps (Schmid et al., 1996).

East of the Giudicarie belt Neogene S-vergent thrusting has to be seen in the context of the N-directed movement of the Adriatic indenter displacing the Periadriatic fault from ca. 23–21 Ma onward. This caused substantial N-S shortening in sections across the Tauern window of the Eastern Alps. This N-S shortening is coupled with substantial E-W extension in the context of lateral extrusion of the ALCAPA block (Doglioni and Bosselini, 1987; Ratschbacher et al., 1991a, 1991b; Scharf et al., 2013; Schmid et al., 2013). However, before these Neogene deformations, the South Alpine unit east of the Giudicarie belt was affected by "Dinaric" west-southwest-vergent Paleogene thrusting. Hence, during the Paleogene the eastern Southern Alps hosted the external front of the external Dinarides (Doglioni, 1987). This is compatible with the observation of the main orogenic deformational event affecting the external Dinarides being active during the late Eocene to early Oligocene (Tari, 2002; Mrinjek et al., 2012). Only later, i.e. since Serravallian to Tortonian time (ca. 14–10 Ma) the S- to SSE-directed "Alpine" thrusts accommodating at least 50 km of N-S shortening in the eastern South Alpine unit became active (Venzo, 1940; Castellarin and Cantelli, 2000; Doglioni and Bosselini, 1987; Schönborn, 1999).

Eastwards, i.e. towards and in Slovenia, the frontal thrust of the South Alpine unit of Italy over the undeformed Adriatic continent changes into a S-directed Neogene thrust of the **eastern extension of the South Alpine Unit** over the pre-existent NW-SW striking Dinarides (Fig. 3) that had formed earlier, i.e. during the late Eocene to early Oligocene. The lateral equivalent of the front of the South Alpine unit in Italy is the frontal thrust of the Tolmin nappes of Slovenia over the Trnovo, Hrusica and Snežnik nappes (Placer, 1981, 1999 and 2008) that we mapped as the northwesternmost part of the High Karst unit based on the geological map by Buser (2010) and the relevant 1:100,000 map sheets of former Yugoslavia (Osnovna geološka karta SFRJ). The equivalents of the Italian South Alpine unit are formed by the Tolmin nappes, which host the sediments of the Slovenian basin, a Mesozoic deep-water basin that records geodynamic signals from the opening of both the Alpine Tethys and the Neotethys oceanic domains, and the structurally higher nappes of the Julian Alps characterized by platform carbonates of the "Julian High" (Rozič et al., 2019). The Krn nappe of the Julian Alps carries two small tectonic klippen (Pokljuka nappe and Zlatna klippe) that exhibit stratigraphy and facies of Lower Cretaceous flysch that have a striking similarity with the Bosnian flysch of the inner Dinarides (Goričan et al., 2018). This suggests that these klippen preserve Dinaric thrust contacts at their base that became obliterated by Neogene deformation. This underlines the similarity of the area with the eastern South Alpine unit that also was affected by Dinaric thrusting (see above) and underlines that the present-day boundary between Dinarides and South Alpine unit is a Neogene feature.

East of Ljubljana the frontal thrust of the Tolmin nappes over the

Trnovo nappe (another part of the High Karst unit) is affected by still younger post-nappe emplacement folding (Sava folds of Placer, 2008) that were active in the context of dextral transpression in late Miocene to recent time, kinematically linked with the dextral Idrija, Sava, Šoštanj, and Labot faults (Fodor et al., 1998; Tomljenović and Csontos, 2001; Vrabec and Fodor, 2006; Žibret and Vrabec, 2016). This zone of dextral transpression can be followed eastward into the Pannonian basin (Fig. 3) along the southern side of the Balaton fault (eastern continuation of the Periadriatic fault). The Balaton fault represents the northern WSW-ENE striking boundary of the Mid-Hungarian fault zone defining the boundary between the ALCAPA units in the Transdanubian Range to the north and the elements attributed to the eastern extension of the Southern Alps unit to the south (Fig. 5). The latter are squeezed between Balaton fault and Tisza Mega-Unit, together with the Sava suture zone (Plate 1 and Fig. 5). Southern Alps unit and Sava suture zone run along the southwestern part of the Mid-Hungarian fault zone that formed in the context of lateral extrusion of the ALCAPA mega-unit in the Miocene (e.g. Csontos and Nagymarosy, 1998; Ustaszewski et al., 2008 and references therein). This easternmost extension of the Southern Alpine unit corresponds to the South Karawanken unit and the South Zala unit as defined by Haas et al. (2000) and finds its eastern termination south of Budapest (Plate 1 and Fig. 5).

3.4.2. ALCAPA mega-unit

This mega-unit comprises the Austroalpine unit of the Alps and equivalent units in the Western Carpathians (Fig. 2). Profiles across the ALCAPA mega-unit and underlying tectonic units are presented in Figs. 3 and 4. The far-travelled nappes of the internal Alps are derived from the Adriatic continent, paleogeographically located south of the Alpine Tethys Ocean, and they are referred to as Austroalpine nappes (Schmid et al., 2004; Froitzheim and Schuster, 2008). They find their equivalents in the Central and Inner Western Carpathians (Plašienka, 2008 and 2018). These allochthons constitute, together with tectonically underlying units derived from the Alpine Tethys, what is widely referred to as ALCAPA mega-unit, derived from the names **AL**ps, **CA**rpathians, and **PA**nnonian basin, synonymous with the terms “Northwestern Unit” (Balla, 1987) and “North-Pannonian block” (Csontos et al., 1992) coined by the pioneers of the idea that in the Miocene ALCAPA represented a block that underwent lateral extrusion (Ratschbacher et al., 1991a, 1991b), accompanied by anticlockwise rotation during the Neogene (Márton and Márton, 1996). This block is bounded by subvertical Neogene shear zones: a shear zone constituted by the Pieniny Klippen Belt suture zone in the north and the Mid-Hungarian fault zone including the Periadriatic and Balaton faults in the south. These shear zones form the lateral boundaries of the eastward extruding ALCAPA block (Fig. 2) (Ustaszewski et al., 2008 and references therein).

The individual tectonic units mapped as ALCAPA mega-unit are described in detail by Schmid et al. (2004), Froitzheim and Schuster (2008) and Plašienka (2008 and 2018). The map presented in Plate 1 and Fig. 2 is a simplification in that it minimizes the number of Adria derived allochthons to four; the map is almost identical to the version published by Schmid et al. (2008), except for minor changes.

The tectonically highest unit is the **Upper Austroalpine and Inner Western Carpathian upper plate unit** comprising the Ötztal-Bundschuh and Drauzug-Gurktal nappe systems of the Eastern Alps as defined in Schmid et al. (2004), as well as the Transdanubian Range unit (see profile A in Fig. 3) representing their eastern equivalent in the Inner Western Carpathians (Schmid et al., 2008). These units experienced no or only a low degree of Cretaceous metamorphism. We refer to them as upper plate units in the sense that they overlie an intra-continental Cretaceous eclogitic belt (Eoalpine high-pressure belt). In this sense they represent the frontal part of an initial Cretaceous upper plate overlying the eclogitic subduction channel.

The **Eoalpine high-pressure belt** is represented by the Koralpe-Wölz nappe system (Schuster and Frank, 1999; Schuster et al., 2001, 2004 and 2013; Schmid et al., 2004; Froitzheim and Schuster, 2008). It developed by intra-continental subduction of middle and lower crust that originally had a lower plate position. This high-pressure belt is interpreted in terms of an extrusion wedge exhuming these high-pressure units in the late Cretaceous. Erroneously, this high-pressure belt was interpreted to have formed during the subduction of the western embayment of the Meliata Ocean during Cretaceous (Eoalpine) orogeny by Schmid et al. (2004). We now deviate from this view because (1) the westernmost structures related to the obduction of remnants of the Western Vardar ophiolites are found on top of non- or low-grade metamorphic units of the Northern Calcareous Alps south of Salzburg (Mandl and Ondrejičkova, 1993; Gawlick et al., 2015), i.e. units of the Juvavic nappe system of the Northern Calcareous Alps that escaped Eoalpine high-pressure overprint and which we mapped as parts of the Upper Austroalpine lower plate units (Plate 1 and Fig. 2), and, (2) this high-pressure belt completely lacks remnants of ophiolitic slices and consists exclusively of polymetamorphic basement nappes with a Permian to Triassic HT/LP and an Eoalpine LT/HP metamorphic overprint (Schuster et al., 2001 and 2004). Schuster and Stüwe (2008) and Stüwe and Schuster (2010) showed that it is mechanically plausible that the onset of intra-continental subduction is due to a gravitational instability provided by strength contrasts as a result of a prolonged cooling phase following a Permian thermal event associated with lithospheric thinning. This subduction stated in the Early Cretaceous whereas peak conditions were reached at around 90 Ma (Thöni, 2006). This is a long time after the obduction front of the Meliata ophiolitic mélanges reached the easternmost Eastern Alps (namely the Juvavic nappes; Gawlick et al., 2015) and the Western Carpathians (namely the Gemericum; Kozur, 1991) in the Kimmeridgian (ca. 150 Ma). It also post-dates the onset of the detachment of the Juvavic nappes from their pre-Mesozoic underpinnings in the Valanginian (ca. 135 Ma ago) documented by the deposition of Rossfeld Formation in the Northern Calcareous Alps (Faupl and Wagreich, 2000).

Outcrops of the Eoalpine high-pressure belt of the Eastern Alps can only be followed eastward as far as into the area south of Vienna (Fig. 2) but remnants might also be present in the subsurface of the Pannonian basin. We speculate about the presence of high-pressure rocks along a geophysically defined belt of linear deep-seated faults (Raba and Hurbanovo-Diósjenő faults) mapped in Table 1 and Fig. 2, following Plašienka et al. (1997a, their Fig. 2), as delimiting the northern boundary of the Upper Austroalpine and internal west-Carpathian upper plate units. In any case, the eastern termination of the high-pressure belt is likely to ultimately end along the Hurbanovo−Diösjenő fault (Haas et al., 2012; Klučiar et al., 2016; Koroknai et al., 2001). This fault directly juxtaposes the Transdanubian Range unit (part of the Inner Western Carpathians and hence derived from the upper plate unit) to the south with the Vepor unit (equivalent of the Upper Austroalpine lower plate unit) of the Western Carpathians (Plašienka et al. 1997b; Haas et al. 2012; Klučiar et al., 2016; Koroknai et al., 2001) to the north (Fig. 2).

The Upper Austroalpine nappes to the north and underlying the Eoalpine high-pressure belt, i.e. the **Upper Austroalpine and central Western Carpathians lower plate unit** of Plate 1 and Fig. 2, includes the Silvretta-Seckau nappe system as defined by Schmid et al. (2008). We also included the nappe systems forming the Northern Calcareous Alps with the Greywacke zone at their base (Juvavic, Bajuvaric, Tirolic-Noric and Veitsch-Silbersberg nappe systems), into this unit (see Plate 1) because these nappes were detached from their homeland and transported NE-ward before the Eoalpine high-pressure belt formed behind them. In the case of the

Western Carpathians we consider the Vepor unit (Plašienka, 2008 and 2018) as the equivalent of the Silvretta-Seckau nappe system and the Gemer unit (Plašienka, 2008 and 2018) as the lateral equivalent of the Grauwackenzone (see profile B in Fig. 2). Hence, we map these units as Central Western Carpathian lower plate units. The Western Carpathians also contain a series of detached cover nappes that we also attribute to the Central Western Carpathian lower plate units: the Fatric and Hronic far travelled cover nappes, detached in Early to "Middle" Cretaceous time (110–90 Ma), and the enigmatic Silicic nappes presumably detached earlier (Plašienka, 2018). Furthermore, we also include the Paleozoic of the Szendrö Mountains of northern Hungary, deformed under greenschist facies metamorphism and exhibiting NW-vergent Early Cretaceous-age folding, into the Central West Carpathian lower plate units (Koroknai, 2005). The Szendrö Mountains are tectonically separated from the Bükk Mountains that we consider a displaced part of the Dinarides by the SE-dipping Nekézseny thrust that postdates the deposition of Senonian Gosau-type conglomerates cropping out in the Uppony Hills of northern Hungary (Schréter, 1941; Oravecz et al., 2018). The entire easternmost part of ALCAPA is buried underneath the East Slovakian Neogene basin except for an inselberg near Zemplin (Fig. 2) that exposes a high-grade pre-Triassic basement and its Mesozoic cover (Haas et al., 2012).

The **Lower Austroalpine nappes** of the Austroalpine nappes south of Vienna are mapped together with the **Tatric** and **Infra-Tatric nappes** outcropping in the Male Karpaty near Bratislava and further NE across several windows all the way to the Tatra Mountains at the Slovak-Polish border (Plašienka, 2008 and 2018). These nappes are taken together as they tectonically represent the lowermost elements of the ALCAPA mega-unit, but we are aware of the fact that there are important differences in lithological content and structural style between the Lower Austroalpine and Tatric nappes. Plašienka (2003) documented a succession of pre-orogenic rifting events in the Tatric nappes that are comparable to those described for Lower Austroalpine nappes in the Alps (Bernoulli et al., 1993; Froitzheim and Manatschal, 1996).

The main structures formed during Early and "Middle" Cretaceous orogeny in the Alps and Western Carpathians are sealed by post-tectonic upper Turonian to Coniacian deposits of the Lower Gosau Group in the Alps (Wagreich and Faupl, 1994) as well as in the Western Carpathians (Wagreich and Marschalko, 1995). The Upper Gosau Group (upper Santonian to Eocene) of the Northern Calcareous Alps is characterized by rapid subsidence, and its depositional environment completely differs from that of the Lower Gosau Group. Rapid subsidence was interpreted to be the result of subduction erosion of the Cretaceous (Eoalpine) orogenic wedge, when the thrust front of the Austroalpine nappes migrated towards the Piemont-Liguria Ocean (Wagreich, 1995; Handy et al., 2010). Ortner et al. (1995) describe growth strata within the Upper Gosau Group that document the syn-orogenic nature of this Upper Gosau Group in respect to this later phase of shortening when an active margin formed between the base of the Austroalpine nappe stack and the subducting Piemont-Liguria oceanic series. However, the Gosau Group sediments on top of the Drauzug-Gurktal nappe system further south do not show this same evolution (Froitzheim and Schuster, 2008). In the Western Carpathians Plašienka (2018) provided evidence for a migration of the thrust front towards the Piemont-Liguria Ocean (his Vahic Ocean) during the Campanian.

3.4.3. *Dinarides*

The Dinarides are characterized by top-SW shortening of a thick stack of Mesozoic to Cenozoic sediments, in the internal parts also involving low-grade metamorphic Gondwana-derived Paleozoic rocks. This deformed sequence represents the Adria passive margin facing the northern branch of Neotethys. The external Dinarides are built up of very thick sequences (in places >8 km) of Paleo-Mesozoic platform carbonates of the Adriatic carbonate platform that is, along its SE margin, interrupted by a deep-water Triassic-Cretaceous embayment, namely by the Budva unit in Montenegro (Fig. 13, profile E) (Goričan, 1994; Bernoulli, 2001; Vlahović et al., 2005; Picotti and Cobianchi, 2017). The internal Dinarides comprise a stack of composite nappes, made up of continental units of the more distal Adriatic margin (Schefer et al., 2010; Gawlick et al., 2017a) and carrying obducted Western Vardar ophiolites (Fig. 13). These composite nappes formed by out-of-sequence thrusting with respect to latest Jurassic obduction, mainly during latest Cretaceous to Cenozoic orogeny (Schmid et al., 2008). Our compilation of the Dinarides is partly based on the excellent 1:100,000 map sheets of Osnovna geološka karta SFRJ and own fieldwork. Extensive work by French geologists greatly helped to clarify stratigraphy and tectonic framework, and it introduced a subdivision of the external Dinarides (i.e. into High Karst, Pre-Karst etc.), which is still widely used (Aubouin et al., 1970; Blanchet, 1970a; Blanchet et al., 1970; Cadet, 1970 and 1978; Charvet, 1970 and 1980; Chorowicz, 1975 and 1977; Rampnoux, 1970 and 1974). Overviews over the Dinarides are provided by Dimitrijević (1997 and 2001), Pamić et al. (2002), Pamić et al. (2002) and Karamata (2006). Charvet (2013) presented a historical overview discussing the development of ideas about the geology of the Dinarides.

The **Dalmatian** and **High Karst units** of the external-most Dinarides, characterized by Late Triassic-Cretaceous carbonate platform facies, are separated from each other by the **Budva unit** of Montenegro (Goričan, 1994) in the southeasternmost part of the Dalmatian coast (up to a point ca. 30 km SE of Dubrovnik; Fig. 12). The Budva paleogeographic zone and tectonic unit is known as **Krasta-Cukali unit** in Albania (Robertson and Shallo, 2000) and as **Pindos** unit in Greece (Fleury, 1980; Degnan and Robertson, 1998). The record of sedimentation in the **Budva unit** of Montenegro starts with lowermost Triassic continental deposits, upper Lower Triassic carbonates deposited on a carbonate ramp that drowned in the late Anisian, associated with normal faulting and rifting related sub-volcanic intrusions, volcanic flows, tuffs and radiolarites, ultimately giving way to open marine deposition of mostly deep-water limestones (Cadjenovic et al., 2008; Đaković et al., 2018; van Unen et al., 2019), This sequence is interrupted by a short-lived biocalcification crisis event associated with radiolarite deposition near the Triassic/Jurassic boundary (Crne et al., 2011). Radiolarite deposition also spans the Toarcian to Tithonian time interval and resumed again in the Hauterivian to Turonian (Goričan, 1994). The sediments of the Budva unit differ from those of other Tethyan basins by a lower proportion of carbonate in the Upper Jurassic and Cretaceous sequences (Goričan, 1994). Sedimentation stopped with lower Eocene flysch, which provides a time constraint regarding the age of thrusting of the High Karst unit over the Budva unit. While the Budva unit wedges out somewhere NW of Kotor (Montenegro), the basal thrust of the High Karst unit can be traced easily much farther NW-ward until Split (Plate 1 and Fig. 12). Mapping the boundary between the High Karst and Dalmatian units NW of Split and in Istria is more problematic and not precisely defined. There, it was delineated based on following the prevailing structural trend and by choosing one amongst many thrusts that revealed the largest top-to-the SW displacement (Fig. 12). Given the fact that the amount of shortening within the external Dinarides increases from NW to SE (see discussion below) we regard the wedging out of the Budva unit towards the NW as being a consequence of the paleogeographic configuration; in other words, the two carbonate platforms that presumably existed to both sides of the trough of the Budva unit in Montenegro or in Albania represent one and the same carbonate platform in the NW.

The total amount of predominantly middle Eocene to early Oligocene shortening (Dinaric phase) in the NW part of the external

Dinarides, i.e., in the part of the High Karst and Dalmatian units located to the NW of the Split-Karlovac fault, a dextral fault west of Banja Luka (Plate 1 and Fig. 12), has recently been quantified by Balling et al. (2016 and 2017). A balanced cross section across Velebit Mountain located NW of Zadar and west of a first-order transverse zone in the Dinarides, referred to as Split-Karlovac fault (Chorowicz, 1975 and 1977) reveals some of 50 km shortening. A second transect SE of the Split-Karlovac fault and running across Split reveals a minimum of 130 km shortening. This demonstrates a significant increase in the amount of shortening towards SE. This trend agrees with an estimate of shortening in a transect across the southeastern Dinarides crossing the Budva unit of Montenegro, where balanced cross sections have demonstrated a total amount of Cenozoic shortening in the order of 140–150 km (van Unen et al., 2019). Note, however, that the value calculated by van Unen et al. (2019) may include the 70 km of post-middle Miocene strike-slip and shortening, which according to these authors, occurred after the Dinaric phase and post-dates the deposition of the Dinaric lake sediments (see discussion below). Beside the different amount of shortening to the northwest and southeast of the Split-Karlovac fault, the first-order geometry of thrusting totally changes across this line too. The Velebit structure west of the Split-Karlovac fault represents a major backthrust forming the roof thrust of a classical triangle zone (Balling et al., 2016 and 2017; Tomljenović et al., 2018; Środoń et al., 2018). This contrasts with a second transect with increased amount of shortening located southeast of the Split-Karlovac fault that is dominated by fore-thrusts (Schmid et al., 2008; van Unen et al., 2019; Fig. 13). One of these fore-thrusts transforms the dextral strike slip along the Split-Karlovac fault into a frontal SW-facing thrust that subdivides the High Karst unit into two sub-units (Plate 1 and Fig. 12), although the offset towards the eastern connection with the pre-Karst thrust decreases significantly and transfers to larger backthrusts. A paleogeographical pre-structuration of the Split-Karlovac fault is very likely since the occurrence of Permo-Triassic evaporites is documented only in the area to the ESE of this line, where they locally rise to the surface as diapirs (Kulušić and Borojević Šostaric, 2014). This is in contrast to the area WNW of this line where the transition from Permian to Triassic is characterized by dolomite or clastic facies (e.g. Fio et al., 2010).

Dating the timing of shortening along the coast of Dalmatia has been subject to some controversy since Mikes et al. (2008a) postulated a Miocene age for a substantial part of the flysch of the external Dinarides hitherto considered as Eocene based on nannofossils. These authors claimed that older (Eocene) fossils are all reworked. However, these Miocene ages have since then been put in question and disproved by other experts working in the area (Babić and Zupanić, 2008 and 2012). In northern Dalmatia, Ćorić et al. (2008) found Lutetian and Bartonian nannoplankton assemblages while Miocene forms have not been found. In fact, the main phase of shortening is very well constrained by a stack of middle to upper Eocene or probably lowermost Oligocene Dalmatian "flysch", including foraminiferal marls, overlain by, but occasionally also interfingering with the so-called Promina beds. These Promina beds are a syn-orogenic up to 2 km thick stack of sediments characterized by conglomeratic layers with pebbles predominantly composed of platform carbonate clasts. Exotic clasts of granites and sandstones are found in the Nevesinje basin east of Mostar, suggesting that some of the detritus was shed from the Pre-Karst unit forming the hinterland. These mainly Lutetian to Priabonian Promina beds represent a regressive flexural foreland (molasse) succession that records *syn*-tectonic sedimentation in an evolving thrust wedge-top (piggyback) basin (Dragičević et al., 1992; Babić and Zupanić, 2008 and 2012; Zupanič and Babić, 2011; Mrinjek et al., 2012; Španiček et al., 2017). The Promina beds extend all the way from the area north of Zadar to the area east of Mostar in Herzegovina. They were deposited in piggy-back basins located in both sub-units of the High Karst nappe, across strike all the way from near the frontal thrust of the High Karst unit back to the frontal thrust of the overlying Pre-Karst unit; they are absent in the more external Dalmatian unit. The base of the Promina beds is discordant and found on top of lower Eocene foraminiferal limestones deposited after a phase of non-deposition in the latest Cretaceous to earliest Cenozoic (Španiček et al., 2017), or more rarely, on top of Upper Cretaceous sediments of various age, indicating synorogenic erosion of growing antiforms predating the shedding of the conglomerates. Although the Promina beds are basically synorogenic with respect to the thrusting of more internal units, their stratigraphic base was deformed together with the basal unconformity and underlying strata after their deposition. As a consequence, the duration of the Dinaric phase certainly extended into the early Oligocene.

For the quantification of Miocene and younger (post-Dinaric phase) deformations across the external Dinarides a critical timing indicator is provided by Miocene sediments deposited during the evolution of the Dinarides Lakes System. These lake sediments are dominantly located in the part of the external Dinarides (including the pre-Karst unit described below) located to the E of the Split-Karlovac fault and N the kinematically linked fore-thrust running towards Mostar that subdivides the High-Karst unit into two parts (see Plate 1). The amount of deformation of this endemic and isolated group of intra-montane lakes whose sediments overlie the pre-existing Dinaric orogenic structure and whose sedimentation was enhanced by the Miocene Climatic Optimum at ~17–15 Ma (Harzhauser and Mandic, 2008; Harzhauser et al., 2011; Sant et al., 2018) allows for a rough estimate of the intensity of Miocene and younger deformation. Although a genetic link is not yet clear, the 18 Ma onset of deposition observed in these basins post-dates the ~20 Ma onset of back-arc extension that peaked at around 15–14 Ma in the neighbouring Pannonian basin (de Leeuw et al., 2012; Horváth et al., 2015; Mandic et al., 2012; Pavelić and Kovačić, 2018). In the Dinarides, deposition in smaller basins like the Sinj or Gačko basins terminated at 15 Ma, but was still long active afterwards in many other basins, either in reduced areas such as in the Sarajevo-Zenica basin (until ~ 5.4 Ma) or over larger distances such as in the Livno and Tomislavgrad basins (until ~4.5 Ma), where the timing is based on paleontological correlations with the southern Pannonian basin (Milojević and Sunarić, 1964; Pantić and Bešlagić, 1964; Pantić et al., 1966; de Leeuw et al., 2010; de Leeuw et al., 2011; Mandic et al., 2011; Magyar and Geary, 2012; Sant et al., 2018). Recent analyses of those basins that are located more external in respect to the East Bosnian Durmitor unit and the Bosnian flysch thrusted in its footwall (Andrić et al., 2017; Sant et al., 2018; van Unen et al., 2018 and 2019), show that the onset of sedimentation was always associated with normal faults or systems or normal faults with very variable offsets, from kilometres (such as is the case of Sarajevo-Zenica basin) to tens to hundreds of metres (such as the Kamengrad or Konjic basins) or metres to tens of metres (such as Livno, Bugojno, Glamoc, Drvar or Jajce basins). These analyses have also showed that some of these basins were subsequently truncated or inverted in post- middle Miocene time by NW-SE to N-S oriented strike-slip or E-W oriented high-angle reverse faults and thrusts that have again variable offsets, from metres to 15 km and cumulate in cross-orogenic transects up to 60–70 km (van Unen et al., 2018). The multi-stage deformation is visible, for instance, in the Sarajevo-Zenica intramontane basin (Andrić et al., 2017), which contains upper Oligocene deposits that are synkinematic to a pulse of (renewed) NE-SW oriented contraction along small-offset thrusts. This deformation was followed by a stage of *syn*-depositional early to middle Miocene multi-directional extension and one other stage of post-middle Miocene N-S oriented shortening of the basin.

Most of the post-middle Miocene faults mapped by van Unen et al. (2019) are oblique and crosscut the inherited orogenic structure with notable exceptions in the Bosnian flysch and the Budva unit. The faults indicate roughly a contraction direction that is N-S in the NW and NNE-SSW in the SE, roughly coinciding with the present-day stress field (Bada et al., 2007; Heidbach et al., 2007; Heidbach et al., 2010). This follows the same orientation change in the up to 5 mm/a motions N- to NNE- and NE-ward relative to a stable European framework detected by GPS studies (Grenerczy et al., 2005; Bennett et al., 2008; D'Agostino et al., 2008; Métois et al., 2015). When combined with the high present-day seismic activity (Bada et al., 2007; Herak et al., 2009; Glavatović, 2010; Ustaszewski et al., 2014; Markušić et al., 2015), and geomorphological evidence (Biermanns et al., 2019), this indicates that the presently on-going deformation in the Dinarides to the ESE of the Split-Karlovac fault is continuous since post-middle Miocene time. It is important to emphasize that this post-Middle Miocene shortening does not represent a continuation of shortening during the Dinaric phase but: The two shortening events are interrupted by early to middle Miocene extension that led to the deposition of the lake sediments.

In summary, it is still difficult to know as to how much of the shortening achieved by thrusting and strike slip was exactly distributed during the middle Eocene to early Oligocene Dinaric phase and how much occurred in post-middle Miocene time when contraction in the external Dinarides resumed after a period of extension. Interestingly, they lack any vertical axis rotation expected to result from laterally increasing amounts of shortening during the combined Dinaric and post-middle Miocene phases according to available paleomagnetic investigations. This contrasts with the around 20° counter clockwise deviation of the declination of Paleogene and older sites typical for the northern and central external Dinarides in respect to present-day north (de Leeuw et al., 2012 and references therein) but does not provide a criterion for better quantifying the partitioning between Dinaric phase shortening and late Miocene to recent shortening. It is clear, however, that in the area of the Dalmatian coast the kinematic reconstruction by Ustaszewski et al. (2008) grossly overestimated Miocene shortening and vertical axis rotations since this reconstruction was based on the erroneous age determinations by Mikes et al. (2008a).

The next internal tectonic unit is the **Pre-Karst unit** that thrusted the High-Karst unit (Fig. 12) after the deposition of the Eocene flysch present in the footwall of its basal thrust, i.e. during the middle Eocene to Oligocene Dinaric phase (Aubouin et al., 1970; Blanchet, 1970a; Cadet, 1970 and 1978; Charvet, 1970 and 1980); renewed thrusting occurred again in the late Miocene to recent times (see discussion above). The term "Pre-Karst" denotes a paleogeographic realm thought to be transitional between the carbonate platform environment of the High Karst and Dalmatian units and a second and more internal paleogeographic unit, the so-called "Zone Bosniaque" (Aubouin et al., 1970), characterized by a more distal continental margin facies and a Late Jurassic to Late Cretaceous flysch basin referred to as Bosnian flysch. Large parts of the Bosnian flysch represent the younger stratigraphic cover of the Pre-Karst unit as pointed out by Blanchet (1970b). This flysch is typically found in the more internal parts of the Pre-Karst unit. It is in tectonic contact with the overlying East Bosnian-Durmitor unit, which, however, wedges out northwestward in the area of Banja Luka (Fig. 12).

The pre-Mesozoic basement of the Pre-Karst unit crops out in the Bosnian Schist Mountains and in the Sana-Una Paleozoic series of Bosnia and adjacent Croatia (Hrvatović, 2000a; Hrvatović, 2005; Hrvatović and Pamić, 2005). Radiometric ages from these Paleozoic rocks (Pamić et al., 2004) indicate Cretaceous (121–92 Ma) and Cenozoic (59–35 Ma) low-grade metamorphic overprints, respectively, overprinting older Variscan ages (ca. 335 Ma, Borojević Šoštarić et al., 2012). The external parts of the Jurassic-Cretaceous cover of the Pre-Karst unit are characterized by relatively more condensed series and platform-slope facies, particularly by breccias, which were shed from Middle Jurassic time onwards from the High Karst paleogeographic domain towards the more distal parts of the Adria passive continental margin located further to the NE, drowned during the Early Jurassic. Middle Triassic red nodular Rosso Ammonitico type limestone (the so-called Han Bulog facies, first described by Hauer, 1853) is widespread in these more distal parts of the Adriatic margin and indicates an earlier drowning event (Aubouin et al., 1970). In the more internal parts of the Pre-Karst unit characterized by the Bosnian flysch pelagic sedimentation set in during late Early Jurassic time, possibly due to flexural bending of the Adriatic margin below the advancing ophiolites, followed by the deposition of the Bosnian flysch.

The lower >1 km thick sequence of the **Bosnian flysch** is referred to as Vranduk flysch (Hrvatović, 2000b and 2005) and is of Oxfordian to Valanginian age (Mikes et al., 2008b; Đerić et al., 2010). The flysch reflects trench sedimentation controlled by a dual sediment supply from the advancing Western Vardar ophiolites and from the distal continental margin of the Adriatic plate (Mikes et al., 2008b). The trench formed at the obduction front of the Western Vardar ophiolites. A detailed structural analysis by van Unen et al. (2019) documented polyphase ductile deformation of the Vranduk flysch whereby the first folding phase is associated with NW-SE contraction and is hence likely to be of late *syn*-obduction age. The younger parts of the Bosnian flysch in the Bosna valley section, referred to as Ugar flysch (Hrvatović, 2005), however, represent an entirely different type of flysch sediments that overlie the previously deformed Vranduk flysch above an unconformity. This Turonian to "Senonian" Ugar flysch either covers previously deformed Vranduk flysch, or alternatively, Jurassic to Lower Cretaceous strata of the Pre-Karst unit. Hence it seals pre-Turonian deformations, and in this sense, it represents a kind of "Gosau" deposit. The clasts, including large olistostoliths, consist of carbonate material shed from the carbonate platform of the external Dinarides. The Ugar flysch was deformed by *syn*-depositional NE-SW oriented contraction (van Unen et al., 2019). In western Bosnia and Montenegro, the base of similar flysch deposits is of Late Cretaceous age (Mirković et al., 1972). Sedimentation of the so-called Durmitor flysch in Montenegro (Dimitrijević, 1997), however, set in even later, i.e. in Maastrichtian time and lasted until the late Paleocene (Jolović et al., 2016). The Durmitor flysch formed in connection with Late Cretaceous to early Cenozoic top-SW thrusting of the structurally next higher tectonic unit, the East Bosnian-Durmitor Unit.

The next higher **East Bosnian-Durmitor unit** comprises a tectonic unit with the same name, as defined by Dimitrijević (1997). However, being a composite nappe, it also carries previously obducted Western Vardar ophiolites with the ophiolitic mélange at their base. The Paleozoic (Lim Paleozoic) to Mesozoic series terminate with Middle to Upper Jurassic radiolarites deposited after an early Middle Jurassic drowning event (Radoičić et al., 2009; Vishnevskaya et al., 2009). This composite East Bosnian-Durmitor-ophiolite nappe, emplaced in latest Cretaceous to Paleogene time over the Durmitor flysch, represents a far-travelled (>45 km) nappe according to Dimitrijević (1997) and corresponds to the "Zone Serbe" of Rampnoux (1970) and Aubouin et al. (1970). The NE-SW striking external thrust contact of this unit with the Bosnian flysch is strongly deflected into a N-S orientation in the Sarajevo area (Fig. 12). This deflection is known as the "Sarajevo sigmoid" and involves the underlying and more external Pre-Karst unit that includes the Bosnian flysch as well, in part due to post-middle Miocene dextral strike slip movements (van Unen et al., 2019). NE of Sarajevo, the Paleozoic to Triassic series laterally wedge out (Fig. 12), but the so-called "Radiolarite Formation" (Pamić, 2000a, 2000b), exposed in the Bosna river section stretches NW-ward all the way into the Banja Luka area as the westernmost extension of this unit. This Radiolarite Formation, exclusively consisting of late Middle Jurassic to lowermost Cretaceous radiolarites (Vishnevskaya and Đerić, 2005), is tectonically sandwiched between Vranduk flysch at its base and ophiolitic mélange in its hanging wall. It is important to emphasize that these radiolarites were deposited on the distal Adriatic margin and are hence not part of an ophiolitic unit.

The Drina-Ivanjica thrust sheet is the second and more internal composite tectonic unit carrying obducted ophiolites and underlying mélange (Fig. 12). This thrust sheet was, in contrast to the other tectonic units of the Dinarides, already emplaced in Early Cretaceous time since no rocks younger than Jurassic are found in the footwall of its basal thrust. In the Zlatibor area, the basal thrust of the Drina-Ivanjica thrust sheet front locally ramps up into the previously obducted ophiolites of the underlying East Bosnia-Durmitor nappe (see profile E in Fig. 13) causing a duplication of the ophiolitic units. A recent detailed structural and geochronological study along a transect across the Drina-Ivanjica thrust sheet (Porkolab et al., 2019) detected an early phase of deformation with WNW-facing isoclinal folding that took place under anchizonal metamorphic conditions, dated to have taken place during the 150–135 Ma time interval, i.e. in the Tithonian to Valanginian. Since this is synchronous with and kinematically compatible with the W-directed obduction of the Western Vardar ophiolites, this event is testimony of *syn*-obduction thrusting that probably went together with the thrusting of the entire Drina-Ivanjica thrust sheet over the East Bosnian-Durmitor thrust sheet. Late Cretaceous (Cenomanian to Maastrichtian) sediments, starting with shallow water clastics and rudist limestones and grading into flysch (Kosovska Mitrovica flysch; Dimitrijević and Dimitrijević, 1976 and 1987) unconformably cover the ophiolites of the Drina-Ivanjica thrust sheet. The basal unconformity cuts down, from external to internal, from ultramafics and mélange of the Western Vardar ophiolites to the Mesozoic cover and finally to the Paleozoic strata of the Drina-Ivanjica thrust sheet (see profile E of Fig. 13). This post-tectonic cover proves very substantial exhumation and erosion of the Drina-Ivanjica thrust sheet and overlying ophiolites sometimes during the Early Cretaceous and renewed subsidence in Late Cretaceous time. This renewed subsidence is very likely the result of down bending of the Adriatic continent below the adjacent subduction zones below the Tisza-Dacia mega-units, ending with collision along the Sava suture zone at the end of the Cretaceous. This Upper Cretaceous flysch extends from western Serbia all the way to Kosovo (Dimitrijević, 1997) and further into North Macedonia where it unconformably transgresses the basement of the Pelagonian massif, always forming the immediate footwall of the next higher and yet more internal unit: the Jadar-Kopaonik thrust sheet (Plate 1).

The pre-Mesozoic basement of the Drina-Ivanjica thrust sheet consists of Gondwana-derived Neoproterozoic to Cambrian basement overlain by a Carboniferous clastic sequence (Spahić et al., 2018). Outcrops in the Zlatibor area (near Sirogojno) provide evidence for a late Anisian drowning event after the deposition of a Steinalm-type carbonate ramp, leading to the deposition of red nodular limestone (Han Bulog facies), followed by the deposition of hemipelagic Ladinian limestones and a lower Carnian Wetterstein platform facies (Missoni et al., 2012; Sudar et al., 2013; Gawlick et al., 2017a). In contrast to these workers, we regard the Sirogojno outcrops to represent the immediate cover of the Drina-Ivanjica Paleozoic rather than being a huge gravitationally emplaced block in a mélange (often referred to as “olistoplaka” in the literature following Dimitrijević, 1997). Carnian to Norian Dachstein facies is only locally observed near Sirogojno. Grey cherty and thin-bedded limestones of Late Triassic age (referred to as Grivska Formation by Dimitrijević, 1997) are by far more widespread in other areas of the Drina-Ivanjica thrust sheet (Gawlick et al., 2017a, and references therein). Thick sequences of Jurassic radiolarites are also widespread, indicating pelagic conditions during the Jurassic in this distal part of the Adriatic margin (Rampnoux, 1970; Dimitrijević and Dimitrijević, 1991; Dimitrijević, 1997). These radiolarites are thrust by the ophiolitic tectonic mélange at the base of the Western Vardar ophiolites.

The next higher composite nappe, the **Jadar-Kopaonik thrust sheet**, is named after its two largest occurrences in Serbia and is the innermost composite nappe of the Dinarides. It finds its northern extension in the Medvednica Mountains near Zagreb (Fig. 12) and the Bükk Mountains of northern Hungary (Fig. 5). Southwards it extends as a narrow steep belt of imbricates all the way to northern Greece, located immediately adjacent to the Sava suture zone (Fig. 14). The so-called Jadar “block” consists of non-metamorphic Devonian to lower Carboniferous pelagic carbonates followed by upper Carboniferous shallow water limestones or siliciclastics, covered by Permian Bellerophon limestone (Ebner et al., 2008). The Mesozoic is characterized by a sedimentary succession, which is similar to that of the Drina-Ivanjica thrust sheet (Dimitrijević, 1997). A deformed and metamorphosed sequence exposed in the Kopaonik-Studenica area of southern Serbia (Sudar, 1986; Sudar and Kovács, 2006; Schefer et al., 2010) exhibits Anisian shallow-water carbonates giving way to a drowning event characterized by hemipelagic and distal turbiditic, upper Anisian to Norian cherty meta-limestones (Kopaonik Formation, similar to the Grivska Formation of the Drina-Ivanjica thrust sheet and considered an equivalent of the grey Hallstatt facies of the Eastern Alps, the Western Carpathians, and the Albanides-Hellenides). These are overlain by red hemipelagic limestones and radiolarites of presumably Jurassic age (Schefer et al., 2010). These findings document the palaeographic setting of this tectonic unit at the distal most Adria passive margin facing the northern branch of Neotethys. Filipović et al. (2003) and Sudar and Kovács (2006) recognized very strong similarities of the Paleozoic to Mesozoic sequence in the Kopaonik area with that of the Bükk Mountains of northern Hungary, providing one of the major arguments for regarding the Bükk Mountains a displaced fragment of the internal Dinarides (Kovács et al., 2000 and 2011a; Dimitrijević et al., 2003). In the Medvednica Mountain and neighbouring inselbergs near Zagreb, occupying an intermediate position between the Jadar block and the Bükk Mountains, a lower greenschist facies metamorphic complex is exposed. It partly consists of rocks of Mesozoic age and partly of Paleozoic series. This complex was thrusted by an ophiolitic mélange and deformed during a first deformation event dated as earliest Cretaceous (135–122 Ma) and most probably related to ophiolite obduction (Tomljenović et al., 2008; Pamić and Tomljenović, 1998; Borojević Šoštaric et al., 2012; van Gelder et al., 2015).

Structural studies (Schefer, 2010; Schefer et al., 2011) in the Kopaonik area and westerly adjacent Studenica slice (Dimitrijević, 1997) located in southern Serbia (Fig. 12) revealed polyphase deformation starting with a first ductile event under lower greenschist-facies conditions related to ophiolite obduction, followed a second ductile event of pre-Turonian age. After an erosional event and above an unconformity, the obducted ophiolites and underlying Paleozoic-Mesozoic became buried and covered by another and more internal belt of Upper Cretaceous hemipelagic to flysch-type sediments (with respect to the Upper Cretaceous belt below the basal thrust of the Jadar-Kopaonik unit described above). This second and more internal belt of Upper Cretaceous sediments belt was heavily deformed during end-Cretaceous suturing of the Dinarides with the upper plate Dacia mega-unit along the Sava suture zone. An upright antiform subsequently intruded by the 31.7–30.6 Ma old Koaponik granitoid intrusion (Schefer et al., 2011) documents on going contraction after suturing. This was followed by core complex formation affecting the innermost Dinarides during the Miocene (20–16 Ma; Schefer et al., 2011).

Late Eocene to earliest Miocene magmatic activity is widespread in the inner Dinarides and thought to be related to rollback and possibly slab break-off of the Adriatic plate, contemporaneous with on going contraction that migrated into the external Dinarides (Schefer et al., 2011; Andrić et al., 2018). Miocene core complex formation as observed in the Kopaonik area is very widespread in the rest of the innermost Dinarides and well documented for the northern Bosnia, Cer, Bukulja, Fruška Gora and Jastrebac areas (Ustaszewski et al., 2010; Maţenco and Radivojević, 2012; Stojadinović et al., 2013 and 2017; Toljić et al., 2013; Erak et al., 2017). It is related to roll-back in the Carpathians and/or the Aegean area, inducing local clockwise rotation of the inner Dinarides and causing the sharp bend of the Sava suture zone south of Belgrade in map view (Fig. 12) as documented by paleomagnetic studies (Lesić et al., 2019).

3.4.4. Hellenides

The general strike of the external parts of the Hellenides of mainland Greece differs from that of the Dinarides by some 30°, which suggests a clockwise rotation of the Hellenides with respect to the Dinarides. The change in strike is rather abrupt (Plate 1, and Figs. 12 and 14), occurring across the Skutari-Peć transverse zone located in northern Albania (Cvijić, 1901; Aubouin and Dercourt, 1975; Frashëri et al., 2012; Handy et al., 2019). This abrupt kink coincides with a paleomagnetically well-documented rotation by some 50° clockwise documented for the units in Albania and adjacent Greece; it almost entirely occurred in Neogene time (Speranza et al., 1992 and 1995; Kissel et al., 1995; Kissel and Laj, 1988; van Hinsbergen et al., 2005a). The difference between the paleo-declinations of the Miocene lake sediments with respect to present-day north in the central Dinarides (around 0°; de Leeuw et al., 2012) and the Neogene sediments of similar age in northern Albania (around 35°; Speranza et al., 1995) suggests that this rotation accelerated in latest Miocene to Pliocene times (Speranza et al., 1995; Handy et al., 2019). Rotation and acceleration in the late Miocene were undoubtedly associated with back-arc extension in the Aegean region. As shown by a kinematic restoration (van Hinsbergen and Schmid, 2012) this rotation and back-arc extension were balanced by Miocene-to-recent thrusting at the orogen front.

The SW-NE striking Skutari-Peć transverse zone has a complex history and its location is undoubtedly paleogeographically predetermined (Bernoulli and Laubscher, 1972; Aubouin and Dercourt, 1975). In present day map view, the front of the Western Vardar ophiolites is dextrally offset by some 80 km (Fig. 12). Since the Mirdita ophiolites SE of the Skutari-Peć transverse zone are thrust over considerable distance over the Eocene of the Krasta-Cukali (=Pindos) unit, some of this differential advance of the front of the Mirdita ophiolites, and hence dextral displacement of the Mirdita ophiolites with respect to the Western Vardar ophiolites of the Dinarides probably took place during the Dinaric phase at the end of the Eocene that closed the Krasta-Cukali-Pindos paleogeographic domain. However, the existence of a window exposing structurally deeper Kruja (=Dalmatian) and Ionian units in the Peskopja window, located around the Albanian-North Macedonian border (Fig. 14 and profile F of Fig. 15) and some 90 km to the NE measured from the frontal thrust of the Kruja unit over the Ionian unit, suggests that most of this offset occurred in Miocene to recent time since deformation in the Ionian unit did not start before some 21 Ma ago (Handy et al., 2019; see also discussion below). There is also a possibility of a much older and Late Jurassic to Early Cretaceous (*syn*-obduction) pre-structuration because (1) the at least 10 km thick Mirdita ophiolites, as inferred from geomagnetic anomalies (Kane et al., 2005), located SE of and immediately adjacent to the Skutari-Peć transfer zone are the only Western Vardar ophiolites that escaped post-obduction erosion in the Dinarides-Hellenides, which suggests that they were emplaced into a paleo-topographical depression, and (2), because of the discovery of *syn*-obduction NW-facing folds and thrust faults at the eastern margin of the Mirdita ophiolites in the underlying Mesozoic of the Korab (=Upper Pelagonian) unit also affecting the metamorphic sole and underlying mélange of the ophiolites (Tremblay et al., 2015). Within the wider Skutari-Peć transfer zone the Skutari-Peć normal fault is the youngest and most prominent discrete fault. It discordantly truncates the eastern limit of the flysch in the Cukali half window (Nopcsa, 1911) and delimits the western limit of the Mirdita ophiolites in northern Albania (Xhomo et al., 1999). Near the Albanian-Kosovo border this Skutari-Peć normal fault is well exposed as a NE-dipping Miocene-age low angle normal fault with an estimated vertical throw of some 7 km, with the Miocene fill of the Kosovo basin (Elezaj, 2009) in the hanging wall (Handy et al., 2019).

The thrust sheets of the **Ionian unit,** characterized by a deeper marine basin since the Triassic (Bernoulli, 2001) and by pelagic and hemipelagic sediments in the post-rift Middle Jurassic and Cretaceous strata (Robertson and Shallo, 2000; Durmishi, 2014) stretch all the way from Albania SSE-wards to the northwestern Peloponnesos (Fig. 14). The Ionian facies zone rather suddenly leaves the Dinarides-Hellenides orogen NW of the Skutari-Peć transverse zone in northern Albania, crossing over into the Adriatic foreland and joining the Umbria-Marche basin of the Apennines (Bernoulli, 2001). In the northern and central parts of Albania the frontal and NW parts of the Ionian thrust sheets are covered by the sediments of a thick terrigeneous *syn*-flexural (Oligocene flysch) and synkinematic (Neogene molasse) sedimentary series (Peri-Adriatic depression) whose total thickness of Cenozoic siliciclastic series frequently exceeds 7 km (Roure et al., 2004). The synkinematic Burdigalian to recent sediments allow for dating the onset of deformation in the Ionian unit in Albania at some 21 Ma. A total of about 120 km of Neogene shortening can be inferred from published profiles in central Albania across the thrust sheets of the Ionian unit (Bega and Soto, 2017; their Fig. 4). The amount of Miocene to recent WSW-directed displacement of the front of the next higher **Kruja (=Dalmatian) unit** in central Albania with respect to the undeformed Adria foreland amounts to some 280 km according to the kinematic restoration by van Hinsbergen and Schmid (2012). Much of this displacement is taken up by the subduction of Adria below the Ionian thrust slices (see profile F in Fig. 15). Hence, there is a substantial increase in the amount of Miocene to recent shortening from some 10 km in Montenegro across the Skutari-Peć transverse zone, further increasing across a series of transfer faults near Lezha and Elbasan (Handy et al., 2019), as well as offshore across the Othoni-Dhermi fault between Corfu and the Sazani Peninsula (Jouanne et al., 2012; see also Plate 1). These transverse zones run parallel to the Skutari-Peć transverse zone, as well as to the presently active Kephalonia transfer fault defining the NW termination of the Aegean trench (Kahle et al., 2000; Kokinou et al., 2006). The amount of shortening is expected to further increase southward into mainland Greece and the Peloponnesos (see profile G of Fig. 11) in order to accommodate the estimated 50° Miocene rotation indicated by the paleomagnetic measurement around a Dinarides-Hellenides pivot located at or near the Skutari-Peć transverse zone (van Hinsbergen et al., 2005a and 2006). The Plattenkalk unit exposed in a window of the Mani Peninsula (Peloponnesos), exhibiting high-pressure overprint below a Miocene-age extensional detachment (Jolivet et al., 2003), is attributed to the Ionian unit (Jacobshagen, 1986; Kowalczyk and Dittmar, 1991). It represents deeply underthrusted Ionian units that were buried in the Oligocene in the case of the Plattenkalk unit (van

Hinsbergen et al., 2005a, 2005b and c and references therein).

The Ionian unit underthrust the **Kruja unit** (equivalent of the Dalmatian unit in Montenegro and the **Gavrovo-Tripolitza unit** of Greece), which exposes platform-facies Mesozoic series. The youngest deformed sediments of the Kruja unit in Albania are flysch deposits of early Oligocene age indicating that deformation within the Kruja unit cannot have started before some 28 Ma ago. The end of deformation predates the deposition of Serravallian sediments of the Neogene Tirana basin (Gelati et al., 1997) above an unconformity sealing the thrust imbricates of the Kruja unit by an unknown amount of time (Xhomo et al., 2002; their profile 4). Since the oldest sediments of the Tirana basin are of early Burdigalian age, deformation most probably terminated at around 20 Ma contemporaneous with the onset of deformation in the underlying Ionian unit. Flysch sedimentation in the **Gavrovo-Tripolitza unit** of mainland Greece and the Peloponnesos continued until the late Oligocene in the southwestern Peloponnesos (Faupl et al., 2002 and references therein) suggesting late Oligocene to early Miocene thrusting. Together, the flysch sequences of the Ionian and Gavrovo-Tripolitza units developed in the foreland of the most important thrust of the external Hellenides, namely the far-travelled thrust at the base of the Pindos unit (Konstantopoulos and Zelilidis, 2012; Zelilidis et al., 2015).

The Gavrovo-Tripolitza unit is again exposed as the lowermost unit in the core of the Olympos and Ossa windows (profile G of Fig. 11) located in a very internal position within the Hellenides (Godfriaux, 1970). These lowermost units were thrust by the overlying blueschists of the Pindos unit, after exhumation of the Pindos blueschists, at around 40–36 Ma ago (Schermer et al., 1990; Schermer, 1993). The occurrence of the Gavrovo-Tripolitza unit in this internal tectonic window demonstrates that the Gavrovo-Tripolitza unit extends over a distance of at least 180 km measured from the thrust front to the Olympos window, buried underneath the Pindos unit, Pelagonian massif and Western Vardar ophiolites (Fig. 11). Rocks of the Gavrovo-Tripolitza unit are also found also in the southernmost Peloponnesos and on Crete (Figs. 14 and 16), in two fashions. Non- to low-grade metamorphic carbonates of Triassic to Eocene age with Oligocene flysch are commonly referred to as the Tripolitza unit (e.g., Rahl et al., 2005), whilst HP/LT metamorphosed clastic rocks known as the Phyllite-Quartzite unit with Carboniferous to Triassic ages originally likely formed the stratigraphic underpinnings of the Tripolitza unit (van Hinsbergen et al., 2005b). These overlie the metamorphics of the Ionian unit (Plattenkalk) and are known as Phyllite-Quartzite unit, exposed in the Mani Peninsula (Micheuz et al., 2015) and in Crete (Dornsiepen and Manutsoglu, 1994). Together, these occurrences form a South Aegean window that formed in the Miocene (Jolivet et al., 1996; van Hinsbergen et al., 2005b). This early Miocene HP/LT metamorphism is testimony of very substantial amounts of underthrusting of these most external parts of the Hellenides that took place during formation of the Aegean extensional back-arc.

The substantial underthrusting of the **Pindos unit** beneath the Pelagonian massif by some 180 km in mainland Greece is related to a major event of subduction that took place in the late Eocene to early Oligocene (van Hinsbergen et al., 2005b), post-dating collision along the Sava suture zone. Collision forced the Pindos unit and overlying nappes such as the Pelagonian units together with the previously obducted Western Vardar ophiolitic unit to substantially shorten by top-SW thrusting. At its southwestern front, the basal thrust of the Pindos unit ramps across the Mesozoic into the flysch and carries the Western Vardar ophiolites of central and southern Albania (Devolli and Rehove massifs; Koller et al., 2006) and Greece (Pindos ophiolite; Jones and Robertson et al., 1991; erroneously regarded as issued from a "Pindos Ocean" by some authors) passively all the way to the front of the Pindos unit (Fig. 14 and profile F of Fig. 15). We see no evidence at all for the existence of oceanic relicts in the Pindos unit of mainland Greece whose sediments we regard as having been deposited on continental crust.

In fact, the base of the Mesozoic sediments constituting the Pindos unit of mainland Greece consistently lacks an ophiolitic substrate, all the way back to the Olympos window. The lowermost and oldest sedimentary formation of the Pindos unit is a Triassic clastic formation consisting of sandstones, cherts, marls, overlain by Halobia limestones of Late Triassic age (Fleury, 1980). Occasionally, Middle Triassic basalts with back-arc basin geochemical signatures overlain by Triassic to Jurassic sediments are also found (Pe-Piper and Piper, 1991). These authors postulated their origin in an intra-oceanic back-arc basin, but manifestations of Triassic magmatism are widespread in many Triassic platform sediments of the South Alpine unit, external Dinarides and Hellenides and are simply related to intra-continental rifting (Pamić, 1984). Hemipelagic to pelagic Jurassic to Lower Cretaceous sediments including radiolarites (Fleury, 1980) are overlain by a first Aptian to Turonian turbiditic interval, termed "premier flysch du Pinde" by Aubouin (1957), shedding chrome spinel from an ophiolite source located east of the Pindos basin (Faupl et al., 1998). On going Upper Cretaceous pelagic to hemi-pelagic sediments grade into the flysch of the Pindos unit, dominated by terrigeneous clastic sediments from the upper Paleocene onward (Piper, 2006). Contraction of the flysch deposits and thrusting over the Gavrovo-Tripolitza unit started in the late Eocene. The oldest stratigraphic members of the Meso-Hellenic piggyback basin, contains a total of up to 5 km Cenozoic sediments. It was installed on a thrust sheet overlying the Pindos unit consisting of Beotian flysch and overlying Western Vardar ophiolites. This piggyback basin was interpreted in terms of a forearc basin of Eocene age by Ferrière et al. (2004). This underlines the large amount of underthrusting of the Pindos unit leading to high-pressure metamorphism in the Olympos window. The Meso-Hellenic basin evolved into a piggyback basin in the Oligocene to Miocene and in the course of on going shortening associated with fore- and backthrusts affecting this basin und underlying ophiolites during collision of the Pelagonian block with the Gavrovo-Tripolitza unit (Ferrière et al., 2004). The original width of the Pindos paleogeographical domain in mainland Greece can be estimated to amount to some 300 km at least. Some 180 km are estimated to be overlain by the Pelagonian unit as can be inferred from profile G (Fig. 11), and internal shortening amounts to about 120 km according to Skourlis and Doutsos (2003).

The High Karst unit of Montenegro has hardly any equivalent SE of the Skutari-Peć transverse zone in Albania and northern mainland Greece. Only the **Parnassos unit**, located north of the Gulf of Corinth (Fig. 14), can be considered a lateral equivalent of the High Karst unit of the Dinarides (Aubouin et al., 1970). The overlying **Pre-Karst unit** finds a continuation in the form of a thin band of Vermoshi flysch (Marroni et al., 2009) cropping out in northern Albania. This Vermoshi flysch represents the lateral equivalent of the Vranduk Formation of the Bosnian flysch and terminates southeastward at the Skutari-Peć transverse zone. However, further to the SE in Greece this same flysch characterized by ophiolitic detritus reappears again and is referred to as **Beotian flysch unit**. In Greece it tectonically overlies either the Pindos unit, or where present, the Parnassos unit (Nirta et al., 2018). The next higher East Bosnian-Durmitor unit of the Dinarides has no equivalent at all in the external Hellenides of Albania and Greece.

The Pelagonian "massif" in Albania, North Macedonia and Greece forms a domal structure extending for some 300 km from southern Kosovo to Thessaly in Greece, surrounded by Western Vardar ophiolites in the north, west, and south and unconformably transgressed by a belt of Upper Cretaceous sediments in the east (southern continuation of the Kosovska Mitrovica flysch) (Fig. 14). In North Macedonia, Most (2003) and Kilias et al. (2010) provided convincing evidence for a subdivision into a lower and an Upper

Pelagonian unit that we implemented and extended into Greece in the map of Plate 1 (see also Fig. 14).

The **Lower Pelagonian unit** of North Macedonia consists of an amphibolite facies complex of granitoid gneisses, micaschists and mafic rocks; garnet-staurolite micaschists contain large post-kinematically grown porphyroblasts of kyanite (Dumurdjanov, 1985). In the southerly adjacent Voras Mountains of northernmost Greece, Mposkos and Krohe (2004) determined medium pressure conditions with 11 kbar and 550 °C from the same paragenesis they consider to be Early Cretaceous in age. These amphibolite-facies rocks are overlain by a several km thick sequence of pure calcite and dolomite marbles that were hitherto considered to be of Neoproterozoic to Cambrian in age (Arsovski and Dumurdjanov, 1984). However, $^{87}Sr/^{86}Sr$ isotopy (Most, 2003) and the unlikelihood of occurrence of such voluminous pure carbonate formations of Pre-Mesozoic age suggest a Mesozoic age, and we interpret these marbles as a metamorphosed equivalent of Triassic and Jurassic sediments of the East Bosnian-Durmitor unit. Alpine regional metamorphism of the Lower Pelagonian unit is supported by K–Ar data on biotite and white mica yielding ages between 148 ± 6 Ma and 118 ± 5 Ma, with a clear maximum around 130–140 Ma (Most, 2003). Outcrops at the northern margin of the marbles of the Lower Pelagonian unit and adjacent to a normal fault separating them from the overlying low-metamorphic Upper Pelagonian unit exhibit phengite- and glaucophane-bearing blueschist facies schists (Dumurdjanov and Rieder, 1989; Most, 2003). These yielded a $^{40}Ar/^{39}Ar$ plateau age of 148.7 ± 3.3 Ma (Most, 2003). Zircon fission track data in the basement of the Lower Pelagonian unit are around 86–80 Ma for the easternmost part, decreasing to 71–63 Ma westwards (Most, 2003). All together it appears from the so far available data that the Lower Pelagonian unit was metamorphosed during subduction of the Adriatic margin below the Upper Pelagonian unit and the obducting Western Vardar ophiolites during the earliest Cretaceous, followed by a Barrovian overprint of unknown age and exhumed by normal faulting in Late Cretaceous time from below the Upper Pelagonian unit. The eastern side of the Pelagonian massif of North Macedonia was eroded down to the basement and unconformably covered by Turonian conglomerates grading into the flysch deposits of the southern prolongation of the Kosovska Mitrovica flysch, indicating renewed drowning after substantial pre-Turonian erosion (see profile F in Fig. 15). The southernmost equivalents of the Jadar-Kopaonik unit adjacent to the Sava suture zone and thrusting this Senonian flysch (Prelević et al., 2017) continue from North Macedonia for only a short distance southward across the Voras Mountains of Greece where two slices of continental series are imbricated with Western Vardar (Almopias) ophiolites (Anna and Kakourou-Livadia units of 1: 50′000 map sheet Vitoliste by Galeos, 1998), finally wedging out within the Almopias ophiolite south of the Voras Mountains (Fig. 14).

The Greek part of the Lower Pelagonian unit around the Olympos window and in the area NW of this window (Aliakmon river valley, Vermion Mountains) reveals a similar evolution for latest Jurassic to Early Cretaceous time. Mposkos et al. (2010) demonstrated that two upper blueschist-facies overprints, an Early Cretaceous one and an Eocene one, affected ophiolitic slices derived from the Western Vardar ophiolite complex and interleaved elements of the Pelagonian units (Nance, 1981; Mpsokos and Perraki, 2001) on the northern slopes of the Aliakmon river. This indicates incorporation of thin slices of ophiolites, together with slices of the uppermost parts of the Pelagonian in a subduction channel during obduction of the main body of the Western Vardar ophiolites in the southern Vermion Mountains. The Early Cretaceous age of the first metamorphic event affecting the ophiolitic slices (not necessarily that of the high-pressure peak, which might be older) has been radiometrically dated at around 130–100 Ma by Yarwood and Dixon (1977) for granitoids of the Pelagonian located near the ophiolites and those surrounding the Olympos window (Barton, 1976; Yarwood and Dixon, 1977; Schermer et al., 1990). A late Early Cretaceous age (116 ± 8 Ma) of anatectic melts in a post-kinematic pocket in gneisses of the Pelagonian in the Aliakmon river valley (Schenker et al., 2014) provided evidence that the Early Cretaceous blueschist-facies event was followed by an amphibolite-facies overprint leading to partial melting in the late Early Cretaceous.

Mposkos et al. (2010) consider a second upper blueschist-facies overprint of the ophiolitic slices (see above) to be Paleogene in age in view of ample evidence for a late Paleocene to Eocene high-pressure event affecting the Pelagonian and Pindos units near to and around the Olympos window (Schermer 1990; Schermer et al. 1990 and 1993). $^{40}Ar/^{39}Ar$ spectra from hornblende, white mica, and biotite indicate two Paleogene blueschist-facies events. A first one at 61–53 Ma is related to intense deformation within the Pelagonian unit. A second one at 40–36 Ma is only observed near the basal thrust of Pindos and lowermost Pelagonian units over the Gavrovo-Tripolitza unit and hence related to early stages of thrusting of the Pindos and Pelagonian units over the core of the Olympos window (Schermer et al., 1990; Lips et al., 1998). Zircon fission track data in this southern area located just north of the Olympos window are locally of Miocene age (ca. 24–21 Ma; Schenker et al., 2015) and related to a second event of exhumation by normal faulting in the context of Aegean back-arc extension. Stretching lineations in the Pelagonian and underlying units between Olympos window and South Pelion, are constantly oriented NE-SW (Walcott and White, 1998) regardless of metamorphic grade and with top-NE as well as top-SW senses of shear. This suggests reorientation of older lineations in the late Eocene and during Miocene normal faulting. Kilias et al. (2010) report top-WNW to NW thrusting for the Early Cretaceous event in North Macedonia and adjacent Greek Macedonia.

The **Upper Pelagonian unit** of North Macedonia and Albania (Fig. 14) represents the southern continuation of the Drina-Ivanjica unit of Serbia and northwestern Kosovo (Fig. 12), reappearing in southern Kosovo again from below the obducted Western Vardar ophiolites. This Upper Pelagonian unit is known as the Korab unit in Albania and adjacent North Macedonia. In Albania, it comprises a non-metamorphic to weakly metamorphic (lower greenschist-facies) sedimentary sequence reaching from the Ordovician to the Lower Jurassic (Xhomo et al., 1999; Tremblay et al., 2015). In North Macedonia and Greek Macedonia, the Upper Pelagonian unit also comprises pre-Alpine high-grade rocks characterized by high-T metamorphism (Mposkos and Krohe, 2004). Radiometric ages from the Greek part of the Upper Pelagonian basement document the presence of Neoproterozoic basement complexes amongst the typical Permo-Carboniferous granitoids (Anders et al., 2006). The Mesozoic cover of this basement is non-metamorphic, in contrast to the Mesozoic of the Lower Pelagonian unit.

Metamorphic parts of the Pelagonian units are also found south of the Pelagonian massif s.str., namely on the Pelion Peninsula located SE of the Olympos window, and additionally, in the Sporades. However, we did not map them as “Lower Pelagonian” since they are in a high tectonic position and the age of pressure-dominated metamorphism is not known in this area. These metamorphosed Pelagonian slices are thrust by so-called Eohellenic units (Jacobshagen, 1986), units that are of heterogeneous composition that include serpentinite slices, meta-volcanics, marbles, calcschists, which were emplaced in latest Jurassic to Early Cretaceous time (Jacobshagen and Wallbrecher, 1984; Jacobshagen and Matarangos, 2004). This emplacement was related to obduction of the Western Vardar ophiolites, and the ophiolites are overstepped by Upper Cretaceous “meso-autochthonous” limestones. All these units, including the Upper Cretaceous cover, were metamorphosed during the closing of the Sava Ocean. An upper oceanic complex is found in ophiolite-bearing klippen on Skyros Island and other small islands of belonging to the northern Sporades (Jacobshagen and Wallbrecher, 1984). We attribute these klippen to the Sava suture zone. This indicates that parts of this suture zone have to be looked for immediately offshore north of the Sporades.

Ferrière et al. (2016) recently provided an overview of the non-metamorphic elements of the Pelagonian unit in the Othrys Mountains, Attica, Evvia Island and Argolis Peninsula (Fig. 14). In the case of Othrys and Argolis, these represent the most distal parts of the Adriatic margin adjacent to the western margin of the northern branch of Neotethys (their "Maliac Ocean"). In the Argolis Peninsula, a detailed account of the sedimentological evidence for the evolution of this distal passive margin of Pelagonia, evolving into *syn*-obduction sedimentation and the formation of composite nappes is provided by Baumgartner (1985) and summarized in Ferrière et al. (2016). The Othrys Mountains expose a stack of imbricates above the Pelagonian platform that represents the most distal slices of the Adria continental margin and is overlain by obducted Triassic and Jurassic ophiolitic slices (Ferrière et al., 2012). The syn-obduction stratigraphy of Upper Jurassic to Valanginian sediments as well as the tectonics on Evvia Island are described by Scherreiks et al. (2014).

3.4.5. *Aegean domain*

The Aegean domain is characterized by large-scale displacements and rotations of blocks caused by on-going mantle delamination and back-arc extension overprinting internal nappes (see profile H of Fig. 17) starting in the Eocene in the Rhodopes mega-unit (Brun and Sokoutis, 2007) but accelerating since around 25 Ma and especially since 15 Ma when the rate of extension dramatically increased, perhaps because oceanic lithosphere of the southern branch of Neotethys started to enter the Aegean trench south of Crete (Jolivet and Brun, 2010; van Hinsbergen and Schmid, 2012). In the early Miocene, extensional detachments started to cut a pre-existing nappe stack and exhumed its metamorphosed portions in the Cycladic metamorphic core complexes located north of Crete (van Hinsbergen et al., 2005b). Underthrusting associated with massive trench extension, led to the exhumation of Miocene high-pressure units by return flow and forearc extension during convergence and "synorogenic extension" sensu Jolivet et al. (2003), aided by major trench-parallel extension as argued for by van Hinsbergen and Schmid, 2012) affecting the very external and then still subducting **Gavrovo-Tripolitza** and **Ionian units** in the southern Peloponnese and on Crete (van Hinsbergen and Schmid, 2012). The map view restoration of the Aegean-West Anatolian domain by van Hinsbergen and Schmid (2012) reveals up to 400 km trench-perpendicular extension since 25 Ma and even higher amounts of trench-parallel extension in Crete. The restoration also demonstrates that the Sakarya unit and the Cretaceous ophiolites of Turkey cannot be traced far into the Aegean region and are likely bounded by a pre-35 Ma N-S striking fault zone (see also Okay et al., 2012) that became reactivated since 25 Ma as an extensional detachment located west of Lesbos island (Fig. 16) and the Biga Peninsula (West Biga detachment of van Hinsbergen and Schmid, 2012). Unfortunately, the very important along-strike change between Hellenides and western Anatolia affecting the internal units of the Alpine-Eastern Mediterranean orogen (e.g. Sakarya unit, obducted ophiolites of Jurassic vs. Cretaceous age) across a former pre-25 Ma N-S running fault zone offshore the Biga Peninsula can only be reconstructed very approximately by this restoration. A N-S running lithospheric scale fault nearby probably may also have exerted major control on the localization of slab segmentation presently observed beneath westernmost Anatolia (Berk Biryol et al., 2011; Gessner et al., 2013) and defines the eastern

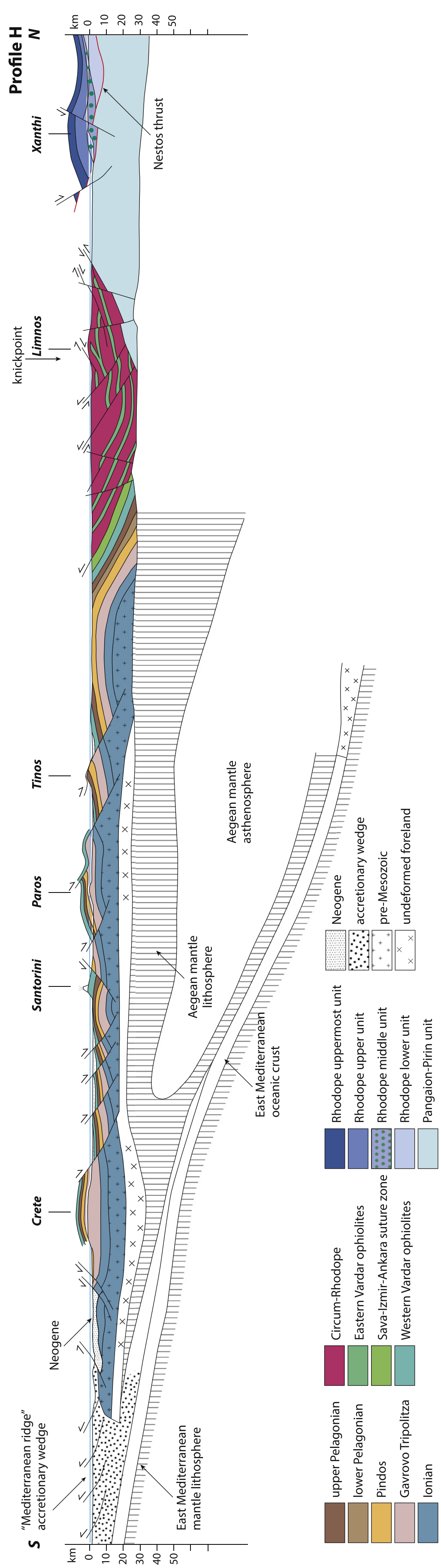

Fig. 17. Crustal-scale profile across the Aegean and into the Rhodopes (profile H). See Figs. 1 and 16 for the trace of profile F. Profile construction is based on the available geological maps of Greece (particularly on Bornovas and Rontogianni-Tsiabaou, 1983) and on Jolivet and Brun (2010). Deep structure and Moho depth after Le Pichon et al. (2002), Li et al. (2003), Sodoudi et al. (2006), Meier et al. (2007), Endrun et al. (2008), Grad et al. (2009), and Bocchini et al. (2018).

termination of the deeply subducted Aegean slab (Spakman et al., 1988; Bijwaard et al., 1998; van Hinsbergen et al., 2005c; Jolivet et al., 2013; van der Meer et al., 2018). These dramatic modifications of the pre-existing crustal structure of the Alpine-Eastern Mediterranean orogen make it challenging to analyse the older substantial along-strike changes between Hellenides and western Anatolia.

The tectonic units in Crete are extremely fragmented by multi-directional post-nappe stacking extension and mapped in Plate 1 and Fig. 16 following a map provided by Martha et al. (2017: their fig. 1). Crete exposes an almost complete section, albeit extremely attenuated and fragmented across the Hellenides with, from bottom to top: (1) Plattenkalk and Trypali units with blueschist-facies overprint mapped as **Ionian unit**, (2) Phyllite-Quartzite sub-unit overprinted by a blueschist-facies overprint (18 kbar/400 °C; Klein et al., 2008; Jolivet et al., 2010), forming the stratigraphic underpinnings of the **Gavrovo-Tripolitza unit** prior to blueschist-facies metamorphic overprint and its subsequent differential exhumation along a normal fault (van Hinsbergen et al., 2005b), (3) non- or anchimetamorphic Tripolitza carbonate unit above an extensional detachment at its base, (4) non-metamorphic Triassic to lower Paleogene sediments of the **Pindos unit**, and (5) a heterogeneous uppermost unit consisting of low-grade slices of sediments, a high grade (Asterousia) crystalline complex and Jurassic as well as Cretaceous ophiolite-bearing mélange (Martha et al., 2017); this uppermost unit was also found on Anafi island north of Crete (Martha et al., 2016). Granodiorite from the Asterousia crystalline Complex yielded zircon populations ranging from 72.5 to 79 Ma in age, suggesting continuous emplacement of granitoids inside of what Martha et al. (2016) interpret as a magmatic arc. A hornblendite from the ophiolitic mélange yielded a U/Pb zircon age of around 163 Ma (Liati et al., 2004), whereas gabbros and dolerites gave K/Ar age ranges of ~160–90 Ma (Koepke et al., 2002). We interpret this uppermost unit as representing small fragments of the **Pelagonian unit** and overlying **Western Vardar** ophiolitic rocks (Fig. 16). This suggests that the geology of the Aegean Islands north of Crete and west of the longitude of the West Biga detachment follow the “Hellenic” logic rather than the “Anatolian” logic.

The Cycladic Islands are located immediately SE of the SE tips of Attica and Evvia and display a similar stack of tectonic units (Fig. 16), later modified by core complex formation. The SE tips of Attica and Evvia, separated from the Pelagonian unit by the steeply NE-dipping Attica-Evvia detachment and transfer zone, better preserve the pre-extension nappe stack (Baziotis and Mposkos, 2011; Xypolias et al., 2012). They expose a trilogy of units in the footwall of the detachment, and hence originally below the massive Pelagonian unit to the NE consisting of, from bottom to top: (1) basal unit with a HP metamorphosed Mesozoic sequence of marbles with schist intercalations and few metabasics at the base covered by a metaflysch we attribute to the Gavrovo-Tripolitza unit, (2) the Cycladic Blueschist unit consisting of HP metamorphosed Permo-Carboniferous to Mesozoic rocks, with rare basic meta-volcanics and meta-ophiolitic lenses that represents the metamorphic equivalent of the Pindos unit, and, (3) an uppermost unit that comprises Permian to Mesozoic sedimentary rocks, ophiolites and Upper Cretaceous greenschist to amphibolite facies rocks that show no evidence of HP metamorphic conditions (Altherr et al., 1994; Papanikolaou, 2009) and that we interpret as relics of the Pelagonian unit and overlying Western Vardar ophiolites. Nappe stacking within individual slices within the Cycladic Blueschist unit is top-ESE and took place during decompression of the blueschist units, associated with top-ESE shearing and pre-dating extension and formation of core complexes (Xypolias et al., 2012).

The extensional core complexes in the Cycladic Islands (Lister et al., 1984; Gautier et al., 1993) exhibit the same original stack of units. The footwall of the detachments exposes metamorphosed equivalents of the **Gavrovo-Tripolitza unit** and/or the **Cycladic Blueschist unit** lateral metamorphic equivalent of the Pindos unit. The non-metamorphic hanging wall, typically preserved near the shoreline of the islands (Grasemann et al., 2018, their Figs. 1 and 12), consists of remnants of the Pelagonian unit (e.g. Andros, Kea, Kithnos) or Western Vardar ophiolites (e.g. Andros, Syros, Tinos, Mykonos, Paros, Naxos). It is not easy in all cases to distinguish post-orogenic stretching lineations and associated senses of shear in the footwall of detachments from stretching lineations related to older shearing during nappe stacking and/or related return flow (so-called “synorogenic extension” of Jolivet et al., 2003). A detailed analysis on Ios Island (Huet et al., 2009) documented opposite senses of shear for the two generations of lineations (top-S for nappe stacking and top-N for unroofing during core complex formation). A compilation of detachment-related senses of shear reveals top-SE senses of shear associated with the West Cycladic detachment, top-NE senses of shear for the North Cycladic detachment and top-N for the Naxos-Paros detachment and the detachments located on islands south of the Mid-Cycladic detachment, such as Ios (Grasemann et al., 2018, their Fig. 1). The compilation of the older “synorogenic” lineations reveals similar orientations of lineations and occasionally opposed senses of shear (Grasemann et al., 2018, their Fig. 8). Interestingly, across the SW-NE striking Mid-Cycladic lineament or detachment (Walcott and White, 1998) the orientation of both generations of lineations changes rather abruptly from dominantly NE-SW striking north of the lineament to dominantly N-S striking south of the lineament. According to the restoration by van Hinsbergen and Schmid (2012) this abrupt change in orientation is caused by trench-parallel extension during a phase of opposite rotation between mainland Greece and westernmost Anatolia since 15 Ma that was accommodated along the Mid-Cycladic detachment. This view was recently substantiated by paleomagnetic and structural analysis of Paros Island where the detachment is exposed (Malandri et al., 2017).

The Cycladic Blueschist unit underwent high-pressure metamorphism at ~55–40 Ma, overprinted by retrograde greenschist-facies metamorphic parageneses between ~30–25 Ma (see reviews in van Hinsbergen et al., 2005b and Philippon et al., 2012). While it appears certain that the Cycladic Blueschist unit is the metamorphosed equivalent of the Pindos unit as exposed in mainland Greece and the Peloponnesos, the question after the existence or non-existence of a “Pindos oceanic domain” remains open. No ophiolite has ever been found in the Pindos unit of mainland Greece or in the Olympos window, unless one considers the ophiolites of the Pindos Mountains to be rooted in a “Pindos Ocean” (see discussion in chapter 3.3.3.). Moreover, the Cycladic Blueschist unit of the Aegean realm typically contains a Permo-Carboniferous metamorphic basement (Keay and Lister, 2002; Tomaschek et al., 2008), locally intruded by Triassic granites (Chatzaras et al., 2013). Bröcker and Pidgeon (2007) also dated Triassic meta-volcanics on Andros, Sifnos and Ios islands, typical rocks found in the thrust sheets derived from the Adriatic continental margin. The stratigraphy of the Mesozoic meta-sediments extends into the Upper Cretaceous (Löwen et al., 2015). Regarding the meta-ophiolite pods occasionally found within the Cycladic Blueschist unit, Bröcker et al. (2014) reported Cretaceous ages of ca. 78 Ma for a meta-gabbro from Samos Island (Selçuk mélange). These ages are in good agreement with similar U–Pb zircon results reported for metagabbros and metaplagiogranites from mélanges on Syros and Tinos (mostly 76 to 81 Ma; Bröcker and Enders, 1999; Bröcker and Keasling, 2006; Bulle et al., 2010; Tomaschek et al., 2003). These ages are younger than those found for the ophiolites of western Turkey and the likelihood that these metagabbros and metaplagiogranites represent parts of an ophiolite sequence is low. Interestingly, the ages coincide with those reported by Martha et al.

(2016 and 2017) for granitoids in the Asterousia Crystalline Complex of Crete and Anafi Island. Should there have been an oceanic basin formed within the Pindos paleogeographical domain, this ocean would have been very narrow since apart from thin ophiolitic slices of uncertain origin on Syros island (Philippon et al., 2011) the Cycladic Blueschist unit is mostly associated with continental crust (Keay and Lister, 2002; Zlatkin et al., 2018). The width of the Pindos paleogeographical realm must have been considerable and is expected to exceed the 300 km width estimated for mainland Greece (see discussion in chapter 3.4.4. Hellenides).

Regarding the lithospheric scale evolution of the Aegean domain, Jolivet and Brun (2010) presented a N-S cross-section running from Crete to the Rhodope massif, parallel to our version of this same section presented in profile H of Fig. 17. They provided a stepwise reconstruction of the tectonic evolution since 65 Ma. Their reconstruction illustrates the progressive migration of the subduction zone during roll back of the delaminated African slab originally located beneath the Hellenides as the African plate rolls back, accelerating at around 23 Ma, when the southern branch of Neotethys entered the trench. They convincingly demonstrated that domains progressively accreted to the orogenic wedge were affected by back-arc extension as the contractional wave migrated southward (see also Tirel et al., 2013). Plate convergence between Africa and Europe since 65 Ma amounts to some 800 km along this profile as calculated by van Hinsbergen et al. (2010a) from the plate circuit of Torsvik et al. (2008). Given the 400 km upper plate extension calculated by van Hinsbergen and Schmid (2012), this means that a total of 1200 km slab length became subducted since 65 Ma. Mantle tomography suggests that the total length of the Aegean (African) slab reaching a depth of at least 1400 km amounts to some 1800 km (van der Meer et al., 2018). Hence an additional 400 km slab must have been subducted before 65 Ma (or more when allowing for slab thickening) as the slab approaches and penetrates the mantle transition zone. Based on this consideration, and the fact that the Circum-Rhodope unit including the Strandja unit were thrust towards north in Early Cretaceous time, a change from top-N to top-S thrusting in "Middle" Cretaceous time at around 110 Ma (Albian) appears to be a reasonable estimate (see discussion in chapter 3.2.5. on the Rhodopes mega-unit).

3.4.6. Western Anatolia

The S- to SE-facing nappe stack exposed in western Turkey located south of the Pontides and the Sava-İzmir-Ankara-Erzincan suture zone, is substantially different from that of the Hellenides in map view (Fig. 16) as well as in profile view (see profile I of Fig. 18). Nevertheless, some similarities do exist and allow for lateral correlations. As mentioned earlier (see chapter 3.3.) there is an important lateral transition regarding the age of the ophiolites that were obducted onto the Adria passive margin - that paleogeographically is contiguous with the Tauride passive margin - towards western Turkey. In western Turkey obduction started in the Late Cretaceous shortly after the initiation of intra-oceanic subduction within the northern branch of Neotethys. The obducted ophiolites found south of the Sava-İzmir-Ankara-Erzincan suture zone are exclusively of Late Cretaceous age and derived from an oceanic plate within the northern branch of Neotethys referred to as Anadolu plate (Gürer et al., 2016). In the south, this lateral change takes place between Crete with its Jurassic ophiolites (Liati et al., 2004) and Karpathos and Rhodos islands with their Cretaceous ophiolites (Koepke et al., 2002). Farther north, the easternmost obducted ophiolite within the Aegean domain offshore Anatolia is found at the westernmost tip of Samos island (Mezger and Okrusch, 1985; Ring et al., 2007a). The non-metamorphic Kallithea ophiolite in Samos is seen to overlie an amphibolite-facies basement, both being separated from the underlying Cycladic Blueschist unit by a late detachment. Since similar high-grade basement is not found in

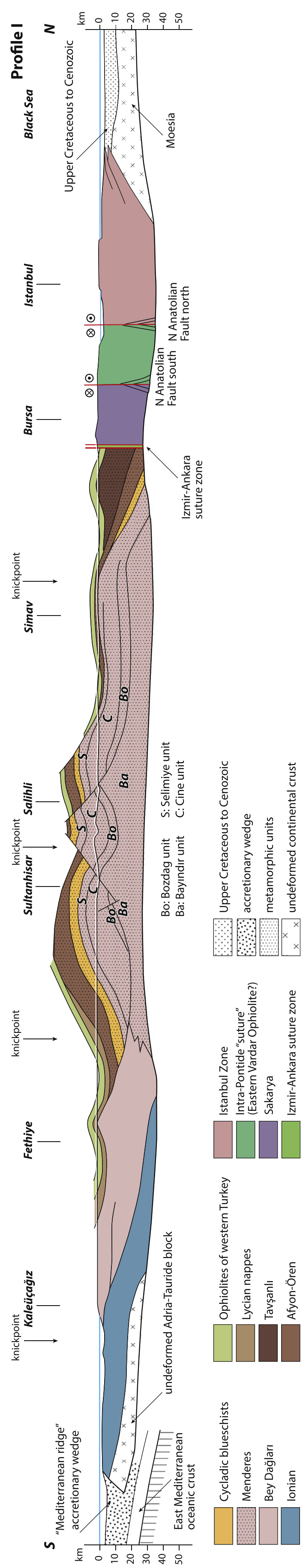

Fig. 18. Crustal-scale profile across Western Turkey (profile I). See Figs. 1 and 16 for the trace of profile I. Profile construction is based on the geological maps of Turkey (Türkecan et al., 2002; Konak et al., 2002a, b; Aksay, 2002; Turhan and Aksay, 2002 and Senel, 2002), Bernoulli et al. (1974), Zitter et al. (2003), ten Veen et al. (2004), van Hinsbergen (2010), Gessner et al. (2011), Munteanu et al. (2011), Aksu et al. (2014). Moho depth after Vanacore et al. (2013).

western Turkey below the obducted ophiolites, we mapped this so far undated ophiolite as a part of the Jurassic ophiolites (Plate 1).

A second and very important difference regards the geodynamic scenario of nappe stacking of units derived from the Adria-Tauride passive margin, which are, from external to internal: Lycian nappes, Afyon-Ören unit, Tavşanlı unit (Fig. 16). The Tavşanlı unit suffered blueschist-facies metamorphism overlapping with or peaking shortly after obduction at ca. 90–80 Ma (Okay et al., 1998; Pourteau et al., 2019; Mulcahy et al., 2014). Imbrication with the Afyon-Ören unit occurred later at around 65 Ma in the process of on going underplating of ophiolites and their metamorphic sole during subduction followed by exhumation and cooling (Plunder et al., 2016). Hence, unlike the situation in the Dinarides-Hellenides, this nappe stacking is not out-of-sequence with respect to an earlier obduction event. Only the most external unit affected by the obduction, the Lycian nappes, escaped high-pressure metamorphism.

A third important factor hampering attempts of lateral correlation is the involvement of westernmost Anatolia in Miocene extensions and rotations in the context of the formation of the Aegean arc that led to the exhumation of the Menderes core complex (van Hinsbergen and Schmid, 2012). Finally, an added complexity is that, in central Anatolia, the Afyon-Ören unit was shown to have been exhumed along detachments within a major latest Cretaceous to Eocene extensional province (Gürer et al., 2018). Whilst structural documentation of the exhumation mechanisms of the Tavşanlı and Afyon-Ören units in western Anatolia is absent, stratigraphic constraints show that these units were also exhumed by Eocene time, i.e. well before Miocene extension of the Menderes massif. Hence, it is feasible that latest Cretaceous to Eocene extension also affected western Turkey.

We used the same colours we used for the Dinarides-Hellenides for mapping the units south of the Sava-İzmir-Ankara-Erzincan suture zone in western Anatolia without implying that they were necessarily contiguous with the composite nappes in the Dinarides-Hellenides (Plate 1). In this case the common colour scheme merely expresses the original position within the Adriatic-Tauride passive margin in a paleogeographical sense, with light brown (external) to dark brown (internal). In terms of terminology and concept, we follow Şengör and Yılmaz (1981) who use the term Anatolides when referring to units that experienced metamorphism related to the closure of the northern branch of Neotethys (e.g. Afyon-Ören unit, Tavşanlı unit) while the term Taurides refers to non- or only weakly metamorphosed domains such as the Lycian nappes. Both are considered to be part of the same Adriatic-Tauride passive margin in a paleogeographical sense ("Anatolide–Tauride Continent" of Şengör and Yılmaz, 1981).

East of Crete, **the Ionian unit** is exposed a last time on Rhodos island, together with slices of the **Gavrovo-Tripolitza and Pindos units** and Cretaceous ophiolites and mélange (Lekkas et al., 2002; Ozsvárt et al., 2017). Offshore the area occupied by the Bey Dağları unit SE of Antalya (Fig. 16), in the area of the submerged Anaximander Mountains Complex located at the junction of the Hellenic and Cyprus Arcs, information is available from Zitter et al. (2003), ten Veen et al. (2004), and Aksu et al. (2014). We correlate the Bey Dağları unit with the Gavrovo-Tripolitza unit (Plate 1) and speculate that the Ionian unit may still be present below a thrust that brings the Jurassic of Bey Dağları unit over Cretaceous strata (Ionian unit) according to the findings in an offshore borehole (Aksu et al., 2014, their fig. 26), as shown in profile I of Fig. 18. To the south, the Anaximander Mountains are adjacent to the Mediterranean ridge, the accretionary prism associated with Oligocene and younger subduction (ten Veen et al., 2004; Zitter et al., 2004) representing the present-day Africa/Europe plate boundary.

The **Bey Dağları unit** builds the calcareous massif of the Bey Dağları (Brunn et al., 1970). The same sequence also occurs in tectonic windows below the pile of the overlying Lycian nappes near the town of Göcek with a platform sequence composed of Cenomanian to lower Burdigalian carbonate rocks, and upper Burdigalian clastics (Bernoulli et al., 1974). These foreland basin clastic sedimentary rocks constrain the timing of thrusting of the Lycian nappes over the Bey Dağları unit (Hayward, 1984; van Hinsbergen et al., 2010b). Langhian (~16–13 Ma) sediments seal the frontal thrust of the Lycian nappes (Flecker et al., 2005; Poisson et al., 2003). On the eastern flank of the Bey Dağları unit, the stratigraphic sequence ends in Paleocene to Eocene time (Ricou et al., 1979; Moix et al., 2013) and is tectonically overlain by the Antalya-Alanya nappes emplaced from the south that will be discussed later. Many correlation schemes of the Bey Dağları unit with respect to the Hellenides have been proposed (Moix et al., 2013 and references therein) but authors generally agree that the platform sedimentary series has similarities with the cover of the metamorphic Menderes massif with which it formed the southern part of a contiguous Menderes-Bey Dağları platform (Collins and Robertson, 1997 and 1999). In view of the observation that large parts of the Menderes massif are overlain by the Pindos unit (Cycladic Blueschist unit, also known as Dilek nappe in western Turkey), we argue for a parallelization of Bey Dağları unit and Menderes massif with the Gavrovo-Tripolitza unit of the Hellenides (Plate 1 and Fig. 16).

To the east, the Bey Dağları unit is thrust by the **Antalya-Alanya nappes**. These consist of a Paleozoic to Paleocene series dominated by carbonate thought to represent the southern passive margin of Bey Dağları unit, and uppermost Cretaceous foreland basin deposits (Robertson and Woodcock, 1981; Angiolini et al., 2007). These are overlain by ophiolites and mélanges that contain ~95–90 Ma old metamorphic sole fragments with ~95–90 Ma $^{40}Ar/^{39}Ar$ cooling ages, comparable with other Anatolian ophiolites (Çelik et al., 2006). Stratigraphic constraints demonstrate that the Antalya-Alanya nappes are a part of a thrust belt that was emplaced from south to north over the Bey Dağları unit (e.g. Okay and Özgül, 1984; Çetinkaplan et al., 2016). The Alanya nappes in the east are outside the area of Plate 1 and comprise obducted ophiolites, as well as a HP-LT metamorphic complex (~85–82 Ma) of continental origin (e.g. Çetinkaplan et al., 2016). The origin of the ophiolites of the Antalya nappes has to be looked for in the southern branch of Neotethys (Fig. 7). Emplacement of the Antalya and Alanya nappes has recently been interpreted to relate to a subduction system that evolved from southeastern Anatolia and radially rolled back westwards, leading to ophiolite emplacement around the Eastern Mediterranean oceanic basin in Arabia, Kyrenia, Cyprus, and the eastern Taurides (Maffione et al., 2017; see also van Hinsbergen et al., 2019). The ophiolite-bearing Alanya nappes located at the southeastern corner of Fig. 16 overlie the Antalya nappes.

Before discussing the Lycian nappes that thrust the Bey Dağları unit towards the SE and the Menderes massif located north of the Bey Dağları unit it is necessary to first discuss the lateral continuation of the high-pressure belt of the **Cycladic Blueschist unit** into Anatolia. This is because this unit characterized by Eocene-age high-pressure metamorphism and parallelized with the Pindos unit of mainland Greece, is a unit that can easily be followed laterally from the Aegean domain into Samos island and westernmost Anatolia. The structural position of the Cycladic Blueschist unit with respect to the Menderes massif, Afyon-Ören unit and Lycian nappes is decisive when discussing the transition from the Aegean domain into western Anatolia. Metamorphism of the Cycladic Blueschist unit in Samos and westernmost Anatolia (Ampelos-Dilek nappe and Selçuk nappe of Ring et al., 2007b) was first described and dated by Oberhänsli et al. (1998) to have occurred at around 40 Ma. This unit can be traced along the northwestern margin of the Menderes massif, which represents a huge composite core complex, all the way to the northern tip of the massif (Ring et al., 2007b; Gessner et al., 2011; Pourteau et al., 2016). This demonstrates that the Menderes massif structurally underlies the Cycladic blueschist unit representing the lateral equivalent of the Pindos unit. The Cycladic Blueschist unit NW of the Menderes massif underwent high-pressure metamorphism starting at about 50 Ma ago. It represents an extrusion wedge formed at around 43–37 Ma and leading to *syn*-convergence ("syn-orogenic") exhumation according to Ring et al. (2007b) that is located between the Menderes massif in the footwall and the Afyon-Ören unit and higher units in the hanging wall. This time interval for extrusion roughly coincides with the age of prograde Barrovian-type metamorphic overprint in the Menderes massif (ca. 43–35 Ma according to Schmidt et al., 2015). Final emplacement of the high-pressure extrusion wedge along the basal Cyclades-Menderes thrust occurred after substantial exhumation at around 39–32 Ma and under lower greenschist-facies conditions (Ring et al., 2007b). The Cycladic Blueschist unit can also be traced for some distance eastward along the southern rim of the Menderes massif (the blueschist facies Kurudere-Nebiler unit of Pourteau et al., 2013). This second belt of the Cycladic Blueschist unit was investigated from a structural point of view by Régnier et al. (2003) and is located south of the Selimiye nappe of the Menderes massif. It formed at ca. 52–45 Ma ago (Pourteau et al., 2013) and separates the Menderes massif lacking blueschist-facies overprint from the southerly adjacent Afyon-Ören that experienced blueschist-facies metamorphism earlier, i.e. at around 70–65 Ma (Pourteau et al., 2013). In this context it has to be mentioned that some authors consider this southern belt of the Cycladic Blueschist unit to represent the cover of the Menderes massif (e.g. Okay, 2008). However, after the discovery of high-pressure metamorphism by Rimmelé et al. (2003), this interpretation appears rather unlikely since the Menderes massif lacks high-pressure overprint elsewhere. However, in some places the exact mapping of the outlines of this belt of the Cycladic Blueschist unit is problematic since unique criteria for its separation from metasediments representing the cover of the Menderes massif are lacking.

As pointed out by Pourteau et al. (2016) the sedimentary realm of the Pindos unit (which they interpreted as oceanic for which we see no conclusive evidence) that became subducted and metamorphosed to issue the Cycladic blueschists was paleogeographically located between the Afyon-Ören unit and the Menderes massif, and it remained open until the Paleocene. These authors point out that the typical Cycladic Blueschist unit, evident in the Aegean and along the rim of the Menderes core complex, does not extend farther into western Anatolia east of the Menderes core complex in map view. Recently, however, deep-marine non-metamorphic sediments accreted in the central Taurides in the ~60–40 Ma time interval were postulated to represent the south-easternmost representatives of the Dilek-Pindos basin (McPhee et al., 2018b). This disappearance of the Dilek nappe (easternmost part of the Cycladic Blueschist unit) towards the east is due to a general axial plunge of the Menderes massif and the Cycladic Blueschist unit towards east that formed in connection with the exhumation of the Menderes core complex. Miocene extension was associated with a vertical axis rotation during the exhumation of the Menderes massif in the Miocene around a pivot point located east of the Menderes massif (van Hinsbergen et al., 2010c). This caused a clockwise rotation difference between the northern and southern Menderes massifs of ~25°–30° in connection with the formation of the Aegean arc (van Hinsbergen and Schmid, 2012; see also Kaymakcı et al., 2018). As a consequence, the continuation of the Eocene thrust located between Menderes massif and Cycladic Blueschist unit is expected to be buried underneath higher structural levels, and in the non-metamorphic Taurides, southeast of the Menderes massif, between the Aladag and Geyikdagi nappes (see McPhee et al., 2018b). Before these severe Miocene modifications, involving rotation and extension in the Menderes massif, the Aegean domain and western Anatolia probably shared a common tectonic evolution as pointed out by Pourteau et al. (2016).

The **Menderes unit** below and thrusted by the Cycladic Blueschist unit is a stack of four nappes, which are, from bottom to top, the Bayındır, Bozdag, Çine and Selimiye units (profile I of Fig. 18). These are built up of a series of low to high-grade Pan-African basement-bearing nappes (Gessner et al., 2013; Oberhänsli et al., 2010, and references therein) intruded by Carboniferous (330–315 Ma) as well as Triassic (250–225 Ma) arc-related granitoids (e.g., Candan et al., 2016; Gessner et al., 2001 and 2013). The pre-Cambrian crystalline basement is overlain by Permo-Carboniferous to Lower Tertiary metasedimentary rocks. The oldest fossiliferous series in the cover sequence are Permo-Carboniferous marble, quartzite and phyllite (Goatee Formation). This is followed by a thick sequence of Mesozoic marbles with emery (meta-bauxite) horizons. The top part of the marble sequence contains shallow marine fossils of Late Cretaceous age and is overlain by red pelagic recrystallized limestones (Okay, 2008). Although there is evidence for earlier Pan-African metamorphism and magmatism in the Çine nappe (Candan et al., 2016; Oberhänsli et al., 2010) garnet Lu/Hf ages of 43–35 Ma indicate Eocene prograde Barrovian metamorphism (Schmidt et al., 2015), indicating that high-grade metamorphic conditions were also reached during the Alpine cycle. The Bayındır nappe is the deepest exposed structural unit of the Menderes massif and penetratively deformed by Alpine deformation while it experienced Alpine greenschist-facies metamorphism (Gessner et al., 2001). It contains uppermost Cretaceous carbonates (Özer, 1998; Özer and Sözbilir, 2003) and yielded a $^{40}Ar/^{39}Ar$ age of 36 ± 2 Ma (Lips et al., 2001). This provides an estimate for the age of accretion of the lowermost Bayındır nappe to the overriding units. Van Hinsbergen (2010, his Fig. 2) proposed a connection of the frontal part of the Bayındır nappe with the Bey Dağları unit in the subsurface, both being considered lateral equivalents of the Gavrovo-Tripolitza unit of the Hellenides. We adopted this interpretation in the profile depicted in Fig. 18 (profile I).

The **Lycian nappes,** including overlying obducted ophiolites, cannot be easily separated from the Central **Taurides** that are also overlain by obducted ophiolites further east and outside the area of Plate 1. The Lycian nappes thrust the Bey Dağları unit towards the SE (Fig. 18, profile I). The window near Göcek exposing Burdigalian sediments (Bernoulli et al., 1974; Hayward, 1984) demonstrates at least 75 km of late-stage SE-ward displacement of the Lycian nappes and adjacent west-Anatolian Taurides relative to the Bey Dağları unit between ~23 and 16 Ma. This was simultaneous with extensional unroofing and exhumation of the Menderes massif to the northwest (Van Hinsbergen, 2010 b and c; van Hinsbergen and Schmid, 2012). In terms of a restoration of the situation before this Miocene thrusting, the Lycian Nappes have to be thought of lying on top of the Menderes unit and the Cycladic Blueschist unit at 25 Ma (Hayward, 1984; van Hinsbergen, 2010). The Menderes massif, representing a huge core complex (see Fig. 18, profile I), was exhumed between 25 Ma and 15 Ma in the north along the Simav detachment (e.g. Ring et al., 2003; Isik and Tekeli, 2001; Isik et al., 2004; Thomson and Ring, 2006; Hetzel et al., 2013; Gessner et al., 2013) and between 15 Ma and 5 Ma along opposite-dipping detachments in the centre (e.g. van Hinsbergen and Schmid, 2012; Gessner et al., 2013). Hence this late-stage thrusting is contemporaneous with N-S extension of the Menderes core complex that is in turn related to extension in the Aegean domain (van Hinsbergen and Schmid, 2012). All this suggests that SE-ward translation of the Lycian nappes over the Bey Dağları unit between ~25 and ~15 Ma was gravity-driven and only affected a few kilometres thick tectonostratigraphy.

The **Lycian nappes**, subdivided into a number of thrust sheets, record continuous sedimentation from the Permian to the Aptian in a passive margin environment south of an oceanic domain (Collins and Robertson, 1999). In the western part of the Lycian nappes clastic sedimentation started during the early Late Cretaceous with flysch-type sedimentation (Bernoulli et al., 1974). In many other places in the Lycian nappes, Upper Cretaceous to upper Eocene foreland basin clastics overstep the non-metamorphic Paleozoic to Paleogene carbonate-dominated sediments (de Graciansky, 1972; Poisson, 1977; Collins and Robertson, 1999). These non-metamorphic thrust sheets are overlain by the frontal-most part of the southward-obducted Cretaceous ophiolites (e.g. Plunder et al., 2016). The ophiolites thrust over the youngest formation of the Lycian nappes, a flysch of probably Maastrichtian age, the Karabörtlen Wildflysch Formation (Bernoulli et al., 1974). During the Cenozoic, the Lycian nappes travelled together with the obducted ophiolites over the Menderes unit and the Cycladic Blueschists unit, and together with the more internal and metamorphic Afyon-Ören and Tavşanlı units that were progressively accreted in the Late Cretaceous and exhumed in the Eocene. The entire length of the accretionary wedge consisting of Lycian nappes and Afyon-Ören and Tavşanlı units thrust over the Cycladic Blueschist unit was schematically restored to the situation in the Paleogene by Plunder et al. (2016, their fig. 6a). Their restoration suggests that some 250 km of Cenozoic shortening was accommodated by the basal thrust that brought Anatolides and Lycian nappes over the Menderes massif and the Cycladic Blueschist unit. Subsequent extension in the Miocene amounts to some 150 km (van Hinsbergen, 2010). Hence the total present-day length of the Late Cretaceous accretionary wedge can be estimated to amount to some 400 km, which is near the present-day distance measured from the front of the Lycian nappes to the Sava-İzmir-Ankara-Erzincan suture zone (see Fig. 16 and profile I of Fig. 18). The eastern boundary of the Lycian nappes with the adjacent non-metamorphic units of the Taurides is not clearly defined and hence not mapped at the southeastern margin of Plate 1. The Lycian nappes are best correlated with the highest thrust sheet of the Taurides, namely the Bozkır unit (Andrew and Robertson, 2002; Özgül, 1976).

The Ören subunit of the **Afyon-Ören** unit is located north of and structurally below the non-metamorphic Lycian nappes in the area south of the Menderes unit. Note that this Ören subunit was formerly, before blueschist facies metamorphic overprint was detected (e.g. Rimmelé et al., 2003; Pourteau et al., 2016), taken as a part of the Lycian nappes. In the northern Menderes massif the Afyon subunit of the Afyon-Ören unit is located structurally below the Tavşanlı unit and in the hangingwall of the Simav low-angle extensional detachment that possibly completely did cut out the rear parts of the Lycian nappe. The Afyon-Ören unit, accreted to the overlying ophiolites, also represents a continent-derived sedimentary sequence. The lithostratigraphy indicates that a lower assemblage of amphibolite-facies schists with mafic and felsic meta-igneous rocks (Pan-African basement) was overlain by a sequence of low-grade metasediments whose dominantly carbonate protoliths were deposited in continental to shallow-water environments (Candan et al., 2005). These were overlain by calciturbidites, siltstone and greywacke with ophiolite and marble olistoliths at the top (Göncüoğlu, 2011). The Afyon-Ören unit was metamorphosed under blueschist-facies conditions reaching 10–12 kbar at about 300 °C (Ören subunit) and 400 °C (Afyon subunit), respectively (Pourteau et al., 2013). The carpholite-bearing assemblages were retrogressed through greenschist-facies conditions at about 67–62 Ma in the Afyon sub-unit, while early retrograde stages in the Ören sub-unit are dated to 63–59 Ma (Pourteau et al., 2013). The metamorphic rocks of the Afyon-Ören unit are unconformably overlain by upper Paleocene to lower Eocene shallow marine sedimentary rocks (Candan et al., 2005).

The highest and most internal structural unit is the composite **Tavşanlı** unit. According to Plunder et al. (2013) this unit consists of, from bottom to top, (1) a metasedimentary unit of continental origin metamorphosed under blueschist-facies conditions (Orhaneli sequence), (2) a Cretaceous ocean-derived accretionary complex, and (3) obducted ophiolite underlain by a metamorphic sole. The blueschist-facies Orhaneli sequence consists of metasedimentary schists and meta-granite (Özbey et al., 2013) transgressed by a metamorphosed Permo-Mesozoic sequence (Okay, 1984 and 1986). This cover sequence starts with graphitic shale overlain by calcite marble grading into pelagic sediments including meta-radiolarite and pelagic shale overlain by an ophiolitic mélange. The rocks of the Orhaneli sequence, which we labelled Tavşanlı unit in Plate 1 and Fig. 16, underwent blueschist-facies metamorphism peaking at around 23 kbar and 430 °C in the case of the Orhaneli sequence (Okay et al., 1998; Okay et al., 2002). From an area outside and east of the area mapped in Plate 1 (Halilbağı Complex east of Ekishehir), Pourteau et al. (2019) reported prograde high-P metamorphism with 21–26 kbar and temperatures between 420–520 °C, dated with Lu–Hf garnet geochronology, to have occurred between 92 and 87 Ma. A garnet amphibolite of the metamorphic sole of the overlying ophiolite was dated at 104.5 ± 3.5 Ma in the same area.

3.4.7. Apennines, Calabro-Peloritani units and Maghrebides

The **Apennines** are a NW-SE striking, NE verging fold-thrust belt, comprising allochthons derived from slices of continental crust scraped off the Adria microcontinent, predominantly composed of sedimentary rocks rarely older than Triassic, thrust upon undeformed continental crust of Adria. These Greater Adria-derived units are structurally overlain by the Ligurian ophiolites and/or ophiolitic melange (see section 3.3.2) that represent the highest tectonic unit (e.g. Dewey et al., 1989; Bernoulli, 2001; Elter et al., 2003; Finetti, 2005; Scrocca et al., 2007; Vezzani et al., 2010; Molli et al., 2010; Calamita et al., 2011). Since the Apennines are not in the focus of our compilation, we collectively mapped the nappes underlying the Ligurian ophiolites as Adria-derived allochthons without further subdivision. A more complete account on these units is given in van Hinsbergen et al. (2019). The connection of the Apennines with the Alps in western Liguria (outside Plate 1) is discussed in Schmid et al. (2017, and references therein). Analysis of existing literature data on the Alps–Apennines transition zone presented in Schmid et al. (2017) reveals that substantial parts of the northern Apennines suffered Alpine-type, i.e. Europe-vergent shortening associated with an E-dipping Alpine subduction zone in pre-Oligocene time and were backthrusted to the NE during Apenninic orogeny that commenced in the Oligocene. Growth of the Apenninic orogenic system, transitioning across the Calabrian nappe stack into the Maghrebides of Sicily and northern Africa by thrusting and nappe stacking in the area of the African and Adriatic continental paleo-margin started in the late Oligocene and reached the external foredeep of the Apennines after the Burdigalian (e.g., Patacca and Scandone, 1990 and 2007; Doglioni, 1991; Doglioni et al., 1999; Mazzoli et al., 2001). Since the Langhian, contraction and crustal thickening in the frontal part of the wedge was accompanied by extension and crustal attenuation in the hinterland, with both compressive and extensional fronts migrating towards the foreland with time (Casero et al., 1988; Patacca and Scandone, 1990 and 2007; Cello and Mazzoli, 1998). On-shore extension in the Apennines was partly associated with off-shore back-arc opening of the Tyrrhenian basin, whose main tectonic phase started in the late Miocene and continued – primarily affecting the southern Apennines – until the early Pleistocene (Malinverno and Ryan, 1986; Royden et al., 1987; Faccenna et al., 1997, 2001; Jolivet and Faccenna, 2000; Rosenbaum et al., 2002, 2008; Nicolosi et al., 2006).

In the northern Apennines the Apenninic orogeny was associated with oroclinal bending in the southernmost Western Alps

(Schmid et al., 2017), the 50° counterclockwise rotation of the Corsica-Sardinia block (Gattacceca et al., 2007; Advokaat et al., 2014) and W-directed subduction and subsequent roll back of the mantle slab below the Apennines (Spakman and Wortel, 2004). This led to the formation of a northern orocline that developed between the Alps-Apennines junction area (Molli et al., 2010; Schmid et al., 2017) located near Genova, the general strike swinging from an E-W-strike near Genova (just west of the area of Plate 1) into the NW-SE strike typical for most of the northern Apennines and even swinging into a N-S-strike along the eastern end of the Umbria-Marche units that is delimited to the east by a N-S-running dextral transfer zone located east of Rome (see Plate 1). This transfer zone, the Olevano-Adrocco strike slip fault zone associated with a lateral ramp (e.g. Calamita et al., 2012) forms the boundary of the Umbria Marche pelagic and hemipelagic series of the N Apennine (e.g. Molli et al., 2010 and literature therein) and the dominantly platform sediments of the adjacent NW-SE-striking units of the central Apennine (e.g. Cosentino et al., 2010). This transverse structure formed during out-of-sequence thrusting in the latest Miocene to Pliocene, i.e. contemporaneous with the final shaping of the arc of the northern Apennines.

The central Apennines continue SE-ward into another tight orocline in Calabria-Sicily discussed below, the Calabrian Arc (e.g. Cifelli et al., 2008; Johnstone and Mazzoli, 2009; Maffione et al., 2013b; Vitale and Ciarcia, 2013). They are constituted by both passive and active margin stratigraphic sequences of the Adria continental margin, composed of carbonate and siliciclastic rocks, overlain by thrusted sedimentary units derived from the oceanic Tethyan domain including ocean-continent transition sequences towards the Alpine Tethys. According to paleogeographic models, going from paleogeographically internal (or SW in present day coordinates) towards external domains (NE), four domains can be recognized in the southern Apennines: the Sicilides, which we mapped together with the melanges below the Ligurian ophiolite unit (see chapter 3.3.2.), the Latium-Abruzzi (in the north) or Apenninic (in the south) platform, the Molise and Sannio (in the north) or Lagonegro (in the south) basins, and the Apulian platform (e.g., Cosentino et al., 2010; Vezzani et al., 2010; Vitale and Ciarcia, 2013).

Much of Calabria is built up of a highest tectonic unit known as the **Calabro-Peloritani** units of Calabria and Sicily. These units are composed of basement-dominated allochthons with a high-grade Variscan overprint and rare occurrences of Mesozoic cover. One of the tectonic units in Calabria, the Aspromonte unit, exposes a section across lower continental crust that is remarkably similar to that exposed in the Ivrea zone of the Alps, both in terms of age and metamorphic grade (Schenk, 1990; Handy et al., 1999). The paleogeographic provenance of the basement nappes of the Calabro-Peloritani units is disputed. Classically they are interpreted to be of "European" origin, i.e. as originally having been connected with Sardinia (e.g. Rosenbaum et al., 2002; Faccenna et al., 2001), in spite of marked differences in the lithologies and grades of metamorphism between the Calabro-Peloritani basement nappes and Variscan Sardinia. According to others they were part of a continental unit representing an extensional allochthon paleogeographically separating two branches of the Alpine Tethys (eastern and western Piemont-Liguria oceanic domains). This continental fragment referred to as ALKAPECA (acronym for Alboran–Kabylia–Peloritani–Calabria; see discussion in Michard et al., 2006 and Handy et al., 2010) also comprises similar lithological units in northern Africa and the Betic cordillera. These highest units were thrust onto the Ligurian ophiolites of southern Basilicata and adjacent Calabria, parts of which reveal a blueschist facies metamorphic overprint, again of disputed age. Rossetti et al. (2001) dated HP/LT metamorphism with peak conditions at 10–12 kbar/330–380 °C at 38–33 Ma by $^{40}Ar/^{39}Ar$ geochronology. This contradicts paleontological evidence indicating that high-pressure metamorphism cannot be older than Aquitanian in the carpholite-bearing Lungro-Verbicario unit (Iannace et al., 2007). The latter data indicate that at least this unit of the Ligurian ophiolites cannot have undergone deformation and metamorphism before the Aquitanian, and hence, that the subducting distal Adria margin did not reach the Calabrian trench before the early Miocene. This open question has bearings on the age of a possible switch in the polarity of subduction of the Piemont-Liguria Ocean from on "Alpine" polarity (subduction below an Adriatic microplate) towards an Apenninic polarity (subduction below a "European" plate also comprising Corsica-Sardinia; see discussion in Schmid et al., 2017). Most authors agree that the rollback of the Calabrian slab (the former dominantly oceanic lithospheric part of the western extension of the Adria microplate, i.e. the Piemont-Liguria Ocean) that started in the Oligocene amounts to well over 1000 km (e.g. Faccenna et al., 2004; van Hinsbergen et al., 2014b).

The **Maghrebides** are an ~E-W trending, S-verging fold-and-thrust belt exposed on the island of Sicily (e.g. Catalano et al., 1996, 2013) extending westwards to Tunisia and Algeria that comprises Africa derived allochthons, overlain by the deep marine Sicilide units derived from an oceanic or ocean-continent transition domain north of the rifted African margin and the Peloritani unit. The base of the orogenic wedge comprises a sequence of Africa-derived carbonate platform and basins, including the Imerese, Panormide, Sicanian, Trapanese-Saccense, and Hyblean domains (e.g., Catalano et al., 1996). Paleogeographically, these continental allochthons derived from the African plate were probably divided from those of the southern Apennines derived from the Adria microplate by the westward extension of the Ionian Ocean, both being part of the southern branch of Neotethys (e.g. Biju-Duval et al., 1977; Handy et al., 2010; Carminati et al., 2012).

4. Concluding remarks

The map compilation and accompanying text presented in this contribution are certainly prone to errors and uncertainties that arise from the data as such, and more importantly, the way they were interpreted and laterally correlated. Concerning the data, sedimentary basins cover large areas, such as the Pannonian, Transylvanian and Thrace basins; hence mapping tectonic boundaries of the units transgressed by these mostly post-tectonic basins is to some extent speculative. More importantly, the map compilation is largely influenced by the concepts used when laterally correlating tectonic units. The authors are aware of the fact that alternative interpretations are possible in many critical areas but hope that the rather long text helps the critical reader to understand the lines of reasoning.

Additional field and laboratory work are clearly needed, for example, for a better understanding of timing and kinematics of deformation in the Rhodopes, only to mention the most problematic area within the area mapped. Our understanding of Paleotethys closure that is somehow linked in time and space to the opening of Neotethys also still remains very limited owing to a scarcity of conclusive geological records of Paleotethys subduction. Any concept about these or similar difficult topics has to stand the test of being kinematically feasible regarding reconstructions of the evolution in space and time (see companion paper by van Hinsbergen et al. (2019) and needs future testing by modelling work.

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Peter Nievergelt is a renowned alpine geologist and petrologist who studied the Penninic-Austroalpine boundary in the Swiss Alps all during his career. Together with Gianreto Manatschal he investigated the Platta Ophiolites and adjacent mid-Jurassic continent-ocean transition, which they reconstructed in great detail despite serious alpine structural overprinting during late Cretaceous nappe stacking. In cooperation with Nikolaus Froitzheim he was one of the first workers to recognize the importance of post nappe low angle detachment fault zones active in the Alps during the early Cenozoic. He followed this normal fault all the way to the eastern margin of the well-dated Bergell granite and its contact metamorphic aureole, which provided one of the best timing constraints regarding the late-stage evolution of the Alps.

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Senecio Schefer (PhD University of Basel, 2010) is a structural geologist who did field work and geochronological investigations in the frame of his PhD in Serbia. Thereby he mainly contributed to a better understanding of the internal Dinarides. He also cooperated in a project with Stefan Schmid and Bernhard Fügenschuh focussing on the Tisza block and its geological neighbourhood. At present he works at the Swiss Geological Survey (swisstopo) being active in the Mont Terri underground rock laboratory in St-Ursanne in the Jura Mountains (Switzerland). He is responsible for geophysical experiments on clay rocks, potentially hosting nuclear waste elsewhere.

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