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The Alpine-Carpathian-Dinaridic orogenic system: correlation and evolution of tectonic units

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ABSTRACT

A correlation of tectonic units of the Alpine-Carpathian-Dinaridic system of orogens, including the substrate of the Pannonian and Transylvanian basins, is presented in the form of a map. Combined with a series of crustal-scale cross sections this correlation of tectonic units yields a clearer picture of the three-dimensional architecture of this system of orogens that owes its considerable complexity to multiple overprinting of earlier by younger deformations.

The synthesis advanced here indicates that none of the branches of the Alpine Tethys and Neotethys extended eastward into the Dobrogea Orogen. Instead, the main branch of the Alpine Tethys linked up with the Meliata-Maliac-Vardar branch of the Neotethys into the area of the present-day Inner Dinarides. More easterly and subsidiary branches of the Alpine Tethys separated Tisza completely, and Dacia partially, from the European continent. Remnants of the Triassic parts of Neotethys (Meliata-Maliac) are preserved only as ophiolitic mélanges present below obducted Jurassic Neotethyan (Vardar) ophiolites. The opening of the Alpine Tethys was largely contemporaneous with the Latest Jurassic to Early Cretaceous obduction of parts of the Jurassic Vardar ophiolites. Closure of the Meliata-Maliac Ocean in the Alps and West Carpathians led to Cretaceous-age orogeny associated with an eclogitic overprint of the adjacent continental margin. The Triassic Meliata-Maliac and Jurassic Western and Eastern Vardar ophiolites were derived from one single branch of Neotethys: the Meliata-Maliac-Vardar Ocean. Complex geometries resulting from out-of-sequence thrusting during Cretaceous and Cenozoic orogenic phases underlay a variety of multi-ocean hypotheses, that were advanced in the literature and that we regard as incompatible with the field evidence.

The present-day configuration of tectonic units suggests that a former connection between ophiolitic units in West Carpathians and Dinarides was disrupted by substantial Miocene-age dislocations along the Mid-Hungarian Fault Zone, hiding a former lateral change in subduction polarity between West Carpathians and Dinarides. The SW-facing Dinaridic Orogen, mainly structured in Cretaceous and Palaeogene times, was juxtaposed with the Tisza and Dacia Mega-Units along a NW-dipping suture (Sava Zone) in latest Cretaceous to Palaeogene times. The Dacia Mega-Unit (East and South Carpathian Orogen, including the Carpatho-Balkan Orogen and the Biharia nappe system of the Apuseni Mountains), was essentially consolidated by E-facing nappe stacking during an Early Cretaceous orogeny, while the adjacent Tisza Mega-Unit formed by NW-directed thrusting (in present-day coordinates) in Late Cretaceous times. The polyphase and multi-directional Cretaceous to Neogene deformation history of the Dinarides was preceded by the obduction of Vardar ophiolites onto to the Adriatic margin (Western Vardar Ophiolitic Unit) and parts of the European margin (Eastern Vardar Ophiolitic Unit) during Late Jurassic to Early Cretaceous times.

1. Introduction

Our analysis of the Alpine-Carpathian-Dinaridic system of orogens including the Pannonian and Transylvanian basins, and of its complex temporal and spatial evolution, is based on a tectonic map that includes the entire system and crosses many national boundaries. This map, presented in Plate 1, was constructed by compiling of available geological maps and subsurface information for those parts of the system covered by very thick Mio-Pliocene (in case of the Pannonian basin) or mid-Cretaceous to Late Miocene deposits (in case of the Transylvanian basin).

The map leads to a better understanding of the mobile belts formed during Late Jurassic, Cretaceous and Cenozoic times that are characterized by extreme changes along strike, including changes in subduction polarity (Alpine-Carpathian polarity vs. Dinaridic polarity; e.g. Laubscher 1971; Schmid et al. 2004b; Kissling et al. 2006). In addition it serves as a base map for

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future palinspastic reconstructions that are required to arrive at realistic paleogeographic and paleotectonic reconstructions. The first obvious step towards this goal consists in establishing the Early Miocene geometry of the various tectonic units of the system. A retro-deformation of the very substantial Miocene rotations and translations was first sketched by the pioneering work of Balla (1987). Subsequent attempts included kinematic inversion of pre-Miocene rotations and translations in order to establish the motion and deformation of the different tectono-stratigraphic units involved in the system during the earlier Palaeogene and/or Cretaceous orogenic phases (i.e. Royden & Baldi 1988; Csontos et al. 1992; Csontos 1995; Fodor et al. 1999; Csontos & Vörös 2004). Moreover, numerous contrasting reconstructions have been advanced for the opening and closure of the different oceanic domains that form part of Neo-Tethys, including its Alpine branch (e.g. Săndulescu 1980, 1988; Golonka 2004; Haas & Pero 2004; Stampfli & Borel 2004).

All attempts at retro-deformation for Palaeogene and/or Mesozoic times must consider the present-day complex three-dimensional configuration of the entire Alpine-Carpathian-Dinaridic system. Our maps and profiles attempt to provide correlations on the basis of which the paleogeographic and paleotectonic evolution can be deduced. The data presented here are derived from literature studies and extensive own fieldwork. Naturally the divisions shown on the map (Plate 1) build on existing compilations (e.g. Săndulescu 1975; Channell & Horvath 1976; Royden & Horvath 1988; Csontos & Vörös 2004; Kovács et al. 2004), but use only primary data and own observations for defining the individual units.

Starting with a brief overview of the first order tectonic elements, we proceed with detailed descriptions of the individual tectonic units. In support of these descriptions, a series of crustal-scale profiles will be presented. These provide a three-dimensional picture of this complex system of orogens, which formed by a long-lasting evolution, that started in Late Jurassic times and is still going on today (Weber et al. 2005).

2. Method of map compilation

The tectonic map of the Carpathian-Balkan Mountain Systems edited by Mahel (1973) served as base map for the entire area covered by our compilation (Plate 1), apart from the Alps, for which we used a simplified version of the tectonic map published by Schmid et al. (2004a). The tectonic map by Mahel (1973) was progressively updated by new data. The most important sources of information are referenced. Longitudes and latitudes are marked at the margins of Plate 1. A version of this map giving 1° longitude and latitude grid points can be obtained from the first author. Figure 1 gives geographic and geological names mentioned in the text.

The Pannonian and Transylvanian basin fills, covering large parts of the Alpine-Carpathian-Dinaridic tectonic units, must be removed to understand the correlations and relationships between the different units. The sedimentary fill of these basins is regarded as "post-tectonic" as it rests unconformably on parts of the Alpine, Carpathian and Dinaridic orogens; however, these sediments were later deformed in many places by deformation associated with basin formation and inversion. In Plate 1, white lines give the outlines of the Pannonian and Transylvanian Basin, and of numerous "inselbergs" in which units of the basin floor are exposed. Buried tectonic boundaries were drawn by projecting their suspected or known position to the surface. However, in many places the location of tectonic boundaries below the basin fill remains rather uncertain.

In the Pannonian basin, the post-tectonic fill conceals many crucial contacts between Alpine, Dinaride and Carpathian tectonic units. The basin fill mostly consists of Miocene sediments up to 6 km thick (see overview given in Haas 2001). The sediments of the post-tectonic cover of the Pannonian basin typically start with either Late Cretaceous or Palaeogene strata.

Note that we assigned the Cretaceous to Eocene Szolnok Flysch (Haas 2001) to the "bedrock"-units rather than to the post-tectonic fill of the Pannonian basin, because the evolution of this flysch basin clearly differs from that of the rest of the Pannonian Basin (e.g. Baldi & Baldi-Becke 1985). In our compilation the Szolnok Flysch is correlated with similar units occurring in the "Pienides" of Northern Romania (Săndulescu et al. 1981a; Tischler 2005; Tischler et al. 2007) and in the "Pennine" units known from the subsurface of the East Slovak Basin (Iňačovce-Krichevo Unit; i.e. Soták et al. 1999). The subsurface information for this basin largely stems from drill holes (e.g. Fülop & Dank 1987) and from seismic data (e.g. Tari et al. 1999).

The Late Cretaceous to Miocene fill of the Transylvanian Basin (e.g. Huismans et al. 1997; de Broucker et al. 1998) covers many of the more internal units of the East Carpathians and Apuseni Mountains. Only recently, a wealth of information from the floor of this basin became available from hydrocarbon exploration (e.g. Kreszek & Bally 2006). Much of this as yet largely unpublished information, consisting of seismic reflection data partly calibrated by bore holes was used during compilation of Plate 1.

The crustal-scale cross-sections given in Plates 2 & 3 were constructed during the compilation the map (Plate 1). Profile construction led to additional insights that significantly improved parts of the map. Note however, that the deeper portions of the crustal-scale profiles in Plates 2 & 3 are not constrained by geophysical or borehole data.

3. Overview of the major groups of tectono-stratigraphic units

The tectono-stratigraphic units described in detail below define eight groups (Plate 1). Before going into greater details regarding map compilation and individual tectonic units, these groups are briefly introduced.

3.1. Undeformed foreland

The northern and eastern flexural foredeep of the Alps and Carpathians ("External Foredeep") is variably floored by Va-

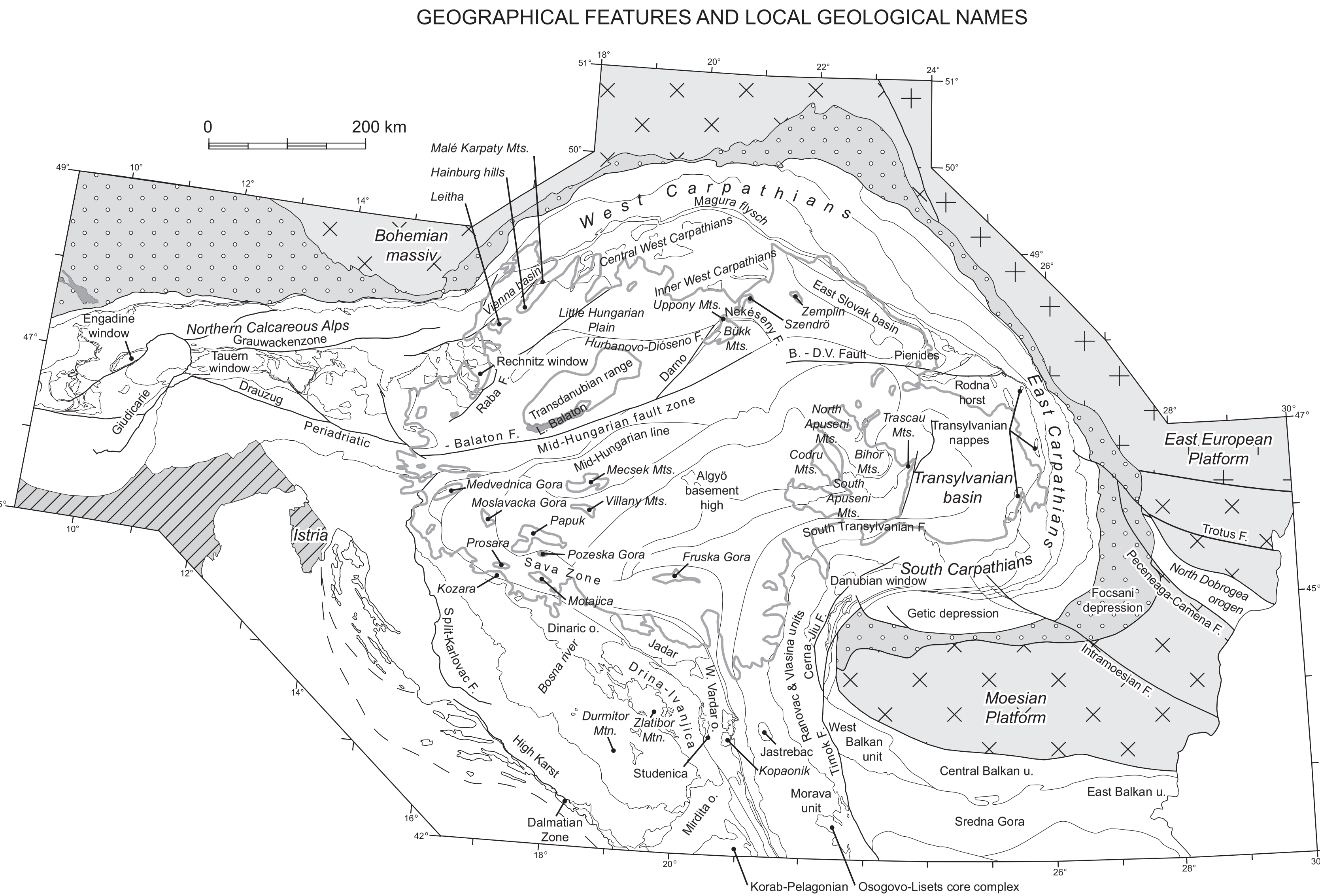

Fig. 1. Index map of geographic and geological names used in the text (see also Plate 1).

riscan, Caledonian or Precambrian basement that is overlain by little or undeformed Mesozoic to Cenozoic sediments. The East European and Scythian platforms were essentially consolidated during Precambrian times. More recently, the Scythian Platform was re-interpreted as the passive margin of the East European craton that was strongly involved in latest Precambrian to Early Paleozoic (pre-Variscan) tectonic events and that was reactivated during younger and less important events (Stephenson et al. 2004; Saintot et al. 2006). The SW boundary of the East European Platform ("Precambrian platform" in Plate 1) coincides with the NW–SE striking Teisseyre-Tornquist Zone that was repeatedly reactivated during Late Paleozoic, Mesozoic and early Cenozoic times (Ziegler 1990). The Pre-Mesozoic basement of areas located to the SW of the Tornquist-Teisseyre Zone consists of an array of essentially Gondwana-derived terranes that were accreted to the margin of the East European craton during the Caledonian and Variscan orogenies (Ziegler 1981, 1990; Pharaoh et al. 2006). The Moesian Platform, for example, underwent significant Variscan deformation and was accreted to the Scythian Platform during the Late Carboniferous (Seghedi 2001). In the process of this, the latter also became overprinted by Variscan deformation (e.g. Zonenshain et al. 1990).

The most external units of the allochthonous East Carpathian Miocene flysch belt partly override the Teisseyre-Tornquist Zone. Following an Early Triassic rifting event, the Moesian Block was probably separated from the Scythian Platform along the SE-most segment of the Teisseyre-Tornquist Zone, only to be re-accreted to it during the Jurassic Cimmerian North Dobrogean orogeny (e.g. Murgoci 1929; Seghedi 2001) that occupies a special position within the Alpine-Carpathian-Dinaridic system (Plate 1). Along the SE-most segment of the Teisseyre-Tornquist Line the Moesian Platform and the North Dobrogea Orogen were welded along the Peceneaga-Camena Fault Zone before the end of the Early Cretaceous (e.g. Murgoci 1915; Hippolyte 2002) when intense tectonic activity along it ceased. Both units are separated by the pre-Neogene Trotus fault from the Scythian Platform (Săndulescu & Visarion 1988).

With respect to the post-Early Cretaceous tectonic activity, the North Dobrogea Orogen, and the Moesian and Scythian platforms are considered as "undeformed foreland". Only minor reactivation along faults shown by thick black lines in Plate 1 occurred in Miocene-Quaternary times (Tărăpoancă et al. 2003; Leever et al. 2006). The northern fault segment near Bacau (Plate 1), however, is not part of the pre-Neogene Trotus fault, but a Quaternary northern splay of the latter (see Matenco et al. 2007).

The term "Adriatic plate" often refers to the undeformed lithospheric plate or "subplate" (Channell & Horvath 1976) that includes the present day undeformed areas of Istria and the Apulian carbonate platform and is framed by the external thrust belts of Southern Alps, Dinarides and Apennines. This part of the Adriatic plate acted as a rigid indenter during its collisional interaction with the Alpine and Dinaridic orogens to the north and east (Fig. 2c), respectively (e.g. Channell et al. 1979; Schmid & Kissling 2000; Pinter et al. 2005). However, note that the term "Adria" is also used for denoting the paleogeographical affiliation of structural entities that had originally formed part of a larger Adriatic (or "African/Adriatic"; Channell & Horvath 1976) promontory or micro-continent but later became involved in its fringing folded belts (Dercourt et al. 1986). In this contribution the term "Adria" and "Adriatic" will therefore be used for the broader paleogeographical realm that flanked the Alpine Tethys to the south ("Apulia" in the sense of Schmid et al. 2004a), rather than just for the present-day rigid Adriatic plate. Structural entities that later became incorporated into the Alpine-Carpathian-Dinaridic system will be referred to as "Adria-derived".

3.2. Miocene external thrust belt:

This thrust belt is the only structural element that extends continuously along the margin of the Alpine-Carpathian orogenic system from the Western Alps through the Western and Eastern Carpathians into the South Carpathians where it ends (Plate 1). In the Carpathians this foreland fold-and-thrust-belt (i.e. Săndulescu et al. 1981a,b; Morley 1996; Matenco & Bertotti 2000; Krzywiec 2001; Oszczypko 2006) formed during the Neogene when the ALCAPA (**Al**ps-**Ca**rpathians-**Pa**nnonia) and Tisza-Dacia Mega-Units moved to the northeast and east into an eastward concave embayment in the European foreland. Their soft collision with the European foreland (e.g. Balla 1987) was triggered by a combination of lateral extrusion (e.g. Ratschbacher et al. 1991a,b) and, more importantly, by the retreat (roll-back) of a subducting eastern European oceanic lithospheric slab (in the sense of Royden 1988, 1993). Calc-alkaline and alkaline magmatism within the Carpathian arc was closely related to subduction, slab-rollback, slab-detachment and extension in the overriding plate (Nemčok et al. 1998; Wortel & Spakman 2000; Seghedi et al. 2004). In the South Carpathians, where the Miocene fold-and-thrust-belt ends, the inner units of the South Carpathians were juxtaposed with the Moesian Platform during Palaeogene-Miocene time in response to a combination of strike-slip movements along curved fault systems (Plate 1) and by oblique thrusting (Ratschbacher et al. 1993; Răbăgia & Matenco 1999; Fügenschuh & Schmid 2005; Răbăgia et al. 2007).

3.3. Europe-derived units in the Alps and in the "Dacia" Mega-Unit

This rather heterogeneous group of tectonic units denotes allochthonous units, commonly interpreted to have been derived from the European continent. In the Alps these units comprise the Helvetic, Ultrahelvetic and Subpenninic nappes, as well as nappes derived from the continental Briançonnais fragment that broke off Europe (Fig. 2c) in the Early Cretaceous in conjunction with opening of a partly oceanic scar (Valais ocean, e.g. Schmid et al. 2004a). In the Carpathians analogous allochthonous units, but not considered as the direct lateral continuation

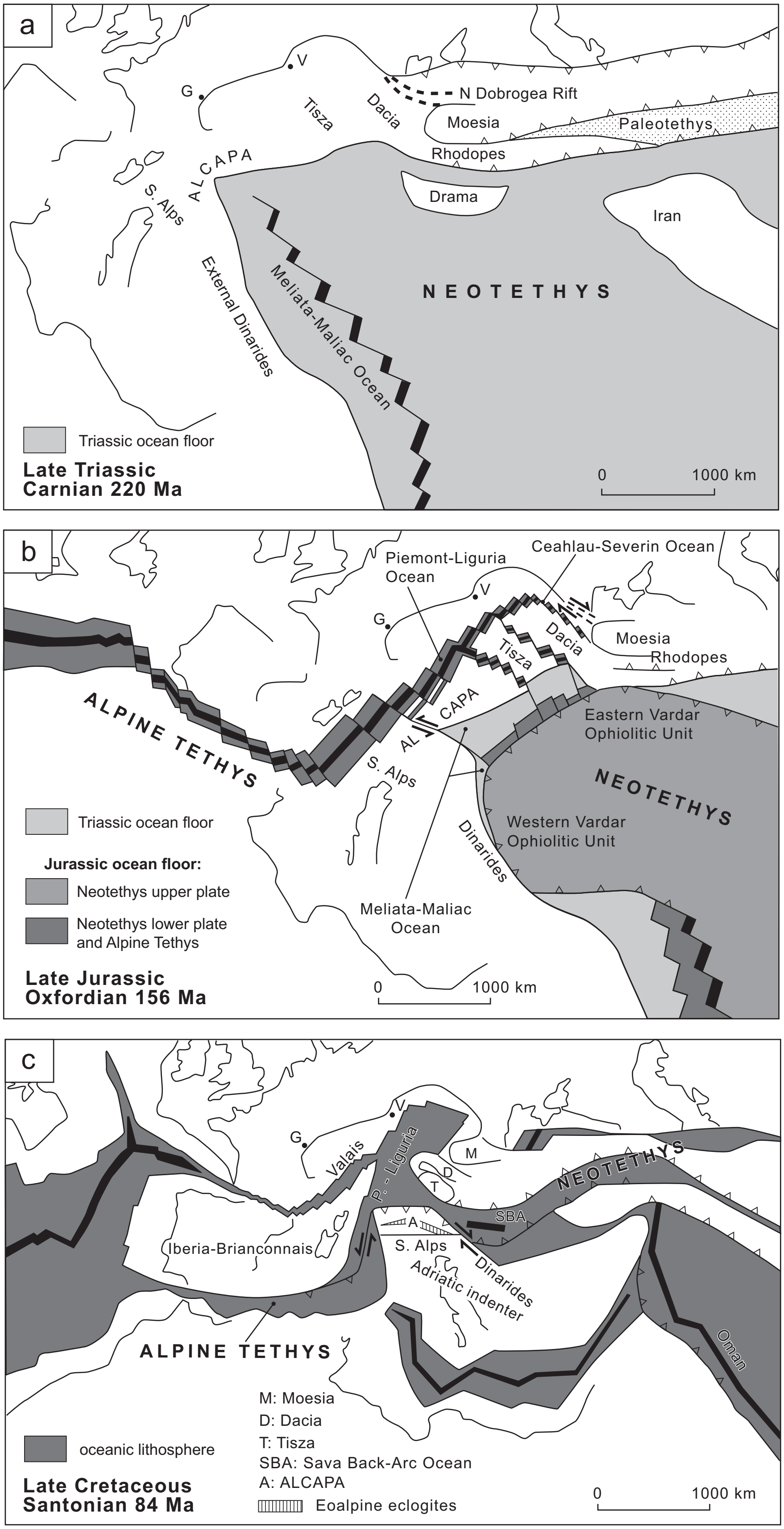

Fig. 2. Schematic palinspastic sketches visualizing concepts presented in the text and incorporating ideas mainly taken from Frisch (1979), Frank (1987), Dercourt et al. (1993), Stampfli (1993), Săndulescu (1994), Stampfli et al. (2001), Marroni et al. (2002) & Schmid et al. (2004a). Positions and projections of the outlines of the European and African continents are those given by Stampfli et al. (2001), except for the coastlines of the Adriatic Sea that we assume to remain firmly attached to the African continent until Cretaceous times. G: Geneva; V: Vienna.

a) Opening of the Neotethyan Meliata-Maliac Ocean in the Triassic (Carnian, 220 Ma).

b) Opening of the Piemont-Liguria Ocean (Alpine Tethys), leading to the separation of the Adria Plate (including the future Southern Alps and ALCAPA), as well as the Tisza and Dacia Mega-Units from the European continent in Oxfordian time (156 Ma). Simultaneously, the Triassic Meliata-Maliac Ocean is being subducted south-eastward below the upper plate Jurassic parts of Neotethys (future Western and Eastern Vardar Ophiolitic Units). The latter are about to become obducted onto the Adria margin, i.e. the future Dinarides, as well as onto the Dacia Mega-Unit.

c) In Santonian times (84 Ma) the two branches of the Alpine Tethys (Valais and Piemont-Liguria Oceans) connect via the Carpathian embayment with younger, i.e. Cretaceous-age branches of Neotethys such as the back-arc ocean that will form the future Sava Zone. Note the onset of northward movement of the Adriatic indenter, contemporaneous with the opening of the East Mediterranean Ocean and the northward drift of Africa.

of their Alpine counterparts, define what is commonly referred to as the Dacia (Fig. 2) Mega-Unit or terrane in the Hungarian literature (e.g. Csontos & Vörös 2004). Parts of this Mega-Unit consist of an assemblage of far-travelled nappes (Infrabucovinian-Getic-Kraishte-Sredna Gora and Serbo-Macedonian-Supragetic-Subbucovinian-Bucovinian-Biharia units; also referred to as Median Dacides by Săndulescu 1994) that were derived from blocks, which were detached from the European margin during Jurassic rifting (Fig. 2b). Whilst oceanic lithosphere of the Ceahlau-Severin Ocean separated parts of these units from Europe, it is unlikely that oceanic lithosphere had separated other units from the European margin. It appears that some nappes, such as the Danubian nappes of the South Carpathians (also referred to as Marginal Dacides; Săndulescu 1994; Kräutner 1996), and units such as the Central Balkan and Prebalkan units of Bulgaria (Georgiev et al. 2001) were directly scraped off the European margin (Moesian Platform; Fig. 1) in response to strong collisional coupling of the orogenic wedge and the foreland (Ziegler et al. 1995).

3.4. Inner Balkanides

While the External Balkanides (Prebalkan, Central Balkan and Sredna Gora units; e.g. Ivanov 1988) can be correlated with other units that form part of Dacia, the Inner Balkanides ("inner" with respect to the Moesian foreland of the Balkanides) form a distinct group of tectonic elements whose origin remains controversial. Two of these units, that are not the focus of our analysis, reach the southern margin of the map of Plate 1. These are: (i) the Strandja Unit (e.g. Georgiev et al. 2001; Okay et al. 2001) that was strongly deformed, metamorphosed and thrust northward during Late Jurassic to Early Cretaceous time and represents a fragment of an older ("Cimmerian") orogen, located near the Paleotethys-Eurasia-Gondwana triple point, that was later incorporated into the N-vergent Balkan Orogen; (ii) The Rhodope core complex delimited to the north and west by Cenozoic normal and strike-slip faults (e.g. Sokoutis et al. 1993; Kilias et al. 1999; Brun & Sokoutis 2007). Its bounding Cenozoic faults post-date Early Jurassic to Cretaceous top-S (SW) nappe stacking and high- to ultrahigh-pressure metamorphism within the gneissic units of the Rhodope Mountains, and they obscure the original relationships of the Rhodope core complex with the neighbouring units (Burg et al. 1996; Turpaud 2006; Perraki et al. 2006; Mposkos & Krohe 2006). The structurally lower unit within the Rhodope core complex possibly represents a terrane (Drama terrane in Fig. 2a) that collided with the rest of the Rhodopian units in Jurassic times along the eclogite-bearing Nestos suture (Turpaud 2006; Bauer et al. 2007).

3.5. Units with mixed European and Adriatic affinities: Tisza

The crustal fragment, which presently constitutes the nappe sequence of the Tisza Mega-Unit, was separated from Europe during the Middle Jurassic (Fig. 2b), presumably in conjunction with the opening of the eastern parts of the Alpine Tethys (or Piemont-Liguria Ocean; Haas & Pero 2004). With the opening of an ocean between Europe and Tisza, the latter moved into a paleogeographic position comparable to that of the Austroalpine or Southalpine realm. The facies of the post-rift sediments within Tisza, such as radiolarites and pelagic Maiolica-type limestones, exhibits Adriatic affinities; the faunal provinces are of the "Mediterranean" type (e.g. Vörös 1977, 1993; Lupu 1984; Haas & Pero 2004). Much of the tectonic units of Tisza (Mecsek, Bihor and Codru nappe systems) are only exposed in isolated and rather small inselbergs within the Pannonian plain (i.e. Mecsek nappe system in the Mecsek Mountains in Hungary; Haas 2001). A coherent nappe sequence is only present in the North Apuseni Mountains of Romania (Bihor and Codru nappe systems; Balintoni 1994). The tectonically highest and most internal nappe system of the North Apuseni Mountains is the Biharia nappe system (Balintoni 1994), traditionally considered as a constituent of Tisza (Csontos & Vörös 2004), i.e. the Internal Dacides (Săndulescu 1984; Balintoni 1994). However, we correlate this Biharia nappe system with the Bucovinian nappes (Plate 1) and hence attribute it to Dacia (i.e. the Median Dacides in the sense of Săndulescu 1984) for reasons discussed later.

3.6. Adria-derived far-travelled nappes of the internal Alps and the West Carpathians (ALCAPA Mega-Unit):

This group of tectonic elements is referred to as the Austroalpine nappes in the Alps (e.g. Schmid et al. 2004a) and as the Central and Inner West Carpathians in the Carpathians (e.g. Plašienka et al. 1997a,b). In the Alpine literature these elements are often referred to as being derived from Adria (or "Apulia" in the sense of Schmid et al. 2004a). In the Alps this implies that their initial paleogeographical position was to the south of the Piemont-Liguria Ocean, the main branch of the Alpine Tethys (Fig. 2b). Presently, these elements represent far-travelled thin crustal slices found in an upper plate position above the Upper Penninic (or Vahic in case of the West Carpathians) suture zone and the underlying Europe-derived elements of the Alps and West Carpathians.

Note, however, that the notion of Adria as a single paleogeographic entity becomes problematic in the easternmost Alps and the adjacent Carpathians and Dinarides owing to the occurrence of branches of a second group of oceanic realms that formed part of the so-called "Neotethys" (e.g. Haas 2001). Neotethys formed an eastward opening oceanic embayment in the Adria paleogeographical realm that had opened during Triassic times (Fig. 2a) and is referred to as Meliata Ocean in the Alps and Carpathians (i.e. Channell & Kozur 1997), but as Maliac Ocean in the Hellenides (e.g. Stampfli & Borel 2004). Hence the Austroalpine nappes and their extension into the West Carpathians include tectonic elements that were positioned north, west and south of the Meliata-Maliac Ocean (Fig. 2; Schmid et al. 2004a). Sea-floor spreading continued in Neotethys during the Jurassic, but the connections between the various oceanic realms formed in Jurassic times with those of the Alpine Tethys are not yet properly understood; Figure 2 merely represents an attempt to satisfy the data compiled in this work.

3.7. Adria-derived thrust sheets: Southern Alps and Dinarides

The units that comprise parts of the Adria paleogeographic realm presently located south of the Periadriatic Line and its eastern continuation (Balaton Line), i.e. south of the ALCAPA Mega-Unit, constitute the Southern Alps and all the non-ophiolitic tectonic units of the Dinarides, including a fragment dislocated along the Mid-Hungarian Fault Zone (Bükk Mountains of Northern Hungary). During the Triassic these units were also located south (or rather southwest, Fig. 2a) of the Meliata Ocean, and thus they were derived from the northern passive margin of Adria that faced the Meliata-Maliac-Vardar branch of Neotethys. In our view the Drina-Ivanjica, Korab-Pelagonian, Bükk, Jadar and Kopaonik "terranes" or "blocks" (e.g. Dimitrijević 2001; Karamata 2006) all represent units that structurally underlie remnants of what we refer to as Vardar Ocean in this contribution, i.e. Neotethyan ophiolites of Jurassic age that were obducted onto the Adria margin during the Late Jurassic (Fig. 2b). Presently, the most external northern Dinaridic units are separated from the Southern Alps by the eastward continuation of a south verging dextrally transpressive thrust front, which formed at a late stage (Mio-Pliocene) in the tectonic history of the Alpine-Carpathian-Dinaridic system. From northeastern Italy and Slovenia we trace this transpressive thrust front into Hungary as far to the east as south of lake Balaton (Plate 1).

In summary, this group of units was derived from continental crustal domains located far to the south of the Alpine Tethys, and even south (or southwest in case of the Southern Alps) of the Neotethys (Fig. 2). Note that many of the detached units of this group overlie an autochthonous basement that is presently still associated with its lithospheric underpinnings (Schmid et al. 2004b); they thus represent deformed parts of the present-day Adriatic plate.

3.8. Ophiolites and accretionary prisms

This group of tectonic units is characterized by ophiolitic and/or flysch-type rock associations and is extremely heterogeneous. It comprises tectonic elements, some of which can be traced along strike over long distances; they often define important sutures and/or important mobile zones between and occasionally also within the above described groups of tectonic elements.

Here we will use the term "ophiolite" in a wider sense, that is we do not restrict the term to rock associations formed at mid-ocean ridges. We also include other types of magmatic and sedimentary rock associations that are indicative of the presence of former oceanic lithosphere. Hence we include, for example, supra-subduction ophiolites and/or subduction-related volcanic arc rocks that developed within pre-existing oceanic crust (see discussion on ophiolite models in Robertson 2002). Moreover, we use the term "ocean" as denoting paleogeographic domains we suspect to have been floored by oceanic rather than continental lithosphere.

It is important to realize, however, that in some cases the ophiolites were obducted as thrust sheets and hence do not mark the location of a deeper suture zone (e.g. Western Vardar Ophiolitic Unit of Plate 1). Others (e.g. the remnants of the Meliata-Maliac Ocean) are only found as blocks within mélange formations beneath coherent ophiolitic thrust sheets such as the Western Vardar Ophiolitic Unit that were obducted during the Late Jurassic (Fig. 2b). Some ophiolitic mélange and obducted ophiolites became subsequently involved in the development of composite thrust sheets or nappes, consisting of continental units and previously obducted ophiolites that formed by out-of-sequence thrusting during Cretaceous and Cenozoic orogenic cycles.

The interrelationships between individual elements of this group of tectonic units still remain uncertain in many cases. This is due to a poor understanding on how Miocene deformations and even more so Palaeogene and Cretaceous deformations and translations ought to be retro-deformed. In this sense, all existing paleogeographic reconstructions for the Alpine-Carpathian-Dinaridic area for Triassic, Jurassic or Cretaceous times, of course including those presented in Fig. 2, must be regarded as purely speculative.

These uncertainties necessitate the continued use of many regional or local names for these ophiolitic and/or accretionary wedge units (see discussion by Zacher & Lupu 1999). Nevertheless, we find it convenient to use the terms Alpine Tethys and Neotethys in order to denote two groups of oceans that began to open in Permian (?) to Mesozoic times in conjunction with the break-up of Pangaea (e.g. Stampfli & Borel 2004).

We use the term "Alpine Tethys" collectively for all oceanic realms in the Alpine-Carpathian domain, the opening of which was kinematically directly linked to sea-floor spreading in the Central Atlantic (Figs. 2b and 2c) that was initiated in late Early Jurassic times (Favre & Stampfli 1992). The term "Neotethys" is used for all oceanic realms located in an area southeast of the Alpine Tethys and the future Western Alps that opened during and after the Permian to Triassic closure of Paleotethys, which at the end of the Variscan orogeny had still separated Gondwana and Laurussia (e.g. Stampfli & Borel 2004). Opening of the oceanic Neotethys basins was neither temporally nor kinematically linked to the opening of the Alpine Tethys. In contrast to the Alpine Tethys, parts of Neotethys were consumed in the context of obduction during Late Jurassic times (Fig. 2b), with a kinematic link to the opening of the Central Atlantic (e.g. Dinaridic ophiolites; Laubscher 1971; Pamić et al. 2002a), while others opened later during the Cretaceous (Sava Back-Arc Ophiolites of Fig. 2c). Yet it is debatable whether during the Jurassic the oceanic realms of the Alpine Tethys and Neotethys were completely separated or linked somewhere in the Alpine-Carpathian-Dinaridic realm (as proposed in Fig. 2b). However, we will present arguments to propose that Cretaceous-age oceanic branches of Neotethys (such as the Sava Back-Arc Ocean,

Fig. 2c) were directly linked with the Alpine Tethys along the present-day Sava-Zone mapped in Plate 1.

We attribute the following oceanic basins to the Alpine Tethys: (i) The Valais, Rhenodanubian and Magura flysch units forming a northern branch (Schnabel 1992; Plašienka 2003; Schmid et al. 2004a), and (ii) the Piemont-Liguria-Vahicum-Pieniny Klippen Belt (Birkenmajer 1986; Plašienka 1995a) units defining a southern branch. Both branches were at least partly separated by the continental Briançonnais fragment of the Western Alps (Frisch 1979; Stampfli 1993) and possibly by smaller analogous units found as relics in the Pieniny Klippen Belt of the West Carpathians (Birkenmajer 1986; Trümpy 1988) that probably were not directly linked with the Briançonnais.

According to our interpretation the Magura Flysch and the Piemont-Liguria-elements of the Alpine Tethys can be traced SE-ward via the Iňačovce-Kriscevo Unit of Eastern Slovakia and the Ukraine (Soták et al. 1993, 1999) into the flysch units referred to as the Pienides of northern Romania (Săndulescu et al. 1981a). In map view, the latter form a tight E-verging arc (Plate 1) and were thrust onto the Tisza-Dacia Mega-units during Miocene times (Fig. 3; Tischler et al. 2007). We propose to connect the Pienides with the Szolnok Flysch in the subsurface of the Pannonian Basin. The ophiolitic Szolnok Flysch belt ("ophiolite-bearing Intrapannonian belt" of Channell et al. 1979) can be traced westward along the Mid-Hungarian Fault Zone towards Zagreb where it links up with the ophiolite-bearing Sava Zone (Plate 1). This zone consists of a belt of ophiolitic (relics of the Sava Back-Arc Ocean, Fig. 2c), magmatic and metamorphic rocks that extends SE-ward from Zagreb to Belgrade ("North-western Vardar Zone" of Pamić 1993; "Sava-Vardar Zone" of Pamić 2002). The Sava Zone defines the Late Cretaceous to Palaeogene suture between Tisza and the Dinarides. Between Zagreb and Belgrade this Sava Zone connects the SE branch of the Alpine Tethys with the Cretaceous-age branches of Neotethys further to the southeast (Fig. 2c). At present the Sava Zone, located between the Western and Eastern Vardar Ophiolitic Units (see Plate 1), represents the suture between the Dinarides and the Tisza-Dacia Mega-Units. The latter are derived from older branches of Neotethys.

The narrow, possibly only partly oceanic Ceahlau-Severin rift opened during Middle to Late Jurassic times (e.g. Ştefănescu 1995; Fig. 2b). Hence it is considered as the easternmost branch of the Alpine Tethys rather than a branch of Neotethys. Note, however, that in present-day map view (Plate 1) it does not directly connect with the easternmost branch of the Alpine Tethys of the Western Carpathians in Northern Romania and adjacent Ukraine, nor does it link up with the Transylvanian branch of the Eastern Vardar Ophiolitic Unit. Westward, units attributed to the narrow Ceahlau-Severin Ocean join the Magura Flysch Unit (Plate 1) that in Miocene times was thrust towards the SE (Fig. 3; Tischler et al. 2007) and that connects with the Mid-Hungarian Fault Zone (see above). Southward, the Ceahlau-Severin Ophiolitic Unit, which is well exposed in the South Carpathians, eventually appears to wedge out in western Bulgaria and eastern Serbia; no equivalents of this narrow ocean have been ascertained in the Balkan Orogen. Figure 2b depicts two additional eastern branches of the Alpine Tethys, which are proposed to kinematically connect with the Triassic Neotethys Ocean and whose opening led to the separation of the Dacia and Tisza blocks from Europe. The relics of the branch between the Dacia and Tisza Mega-Units (Fig. 2b) will form the suture between the Dacia and Tisza Mega-Units in Early Cretaceous times, only preserved in the subsurface (see Plate 3, Profile 3). The branch between the Tisza and ALCAPA Mega-Units (Fig. 2b) forms part of the Mid-Hungarian Fault Zone (Fig. 1, Plate 1), i.e. the "ophiolite-bearing Intrapannonian belt" of Channell et al. (1979).

In the Alpine-Carpathian-Dinaridic system of orogens the westernmost branch of Neotethys (Fig. 2a) is known as the Meliata Ocean (Kozur 1991) that had opened earlier, i.e. in Triassic times (e.g. Velledits 2006). However, in our area, the ophiolitic remnants of the Meliata-Maliac Ocean do not form coherent thrust sheets, but are only preserved as blocks contained in Mid to Late Jurassic-age ophiolitic mélange formations. Such mélange formations occur in the Eastern Alps and West Carpathians (Kozur & Mostler 1991). Remnants of this same Meliata-Maliac Ocean are also found as blocks in Jurassic mélange formations that immediately underlie the obducted Western Vardar ophiolites in the Bükk Mountains (Monosbel nappe of Csontos 1999, 2000) and in the Dinarides (ophiolitic mélange formations; e.g. Babić et al. 2002). In the Dinarides these mélange formations are also referred to as "Diabas-Hornstein" (Kossmat 1924), "Diabase-Radiolarite" Formation (Ćirić & Karamata 1960) or "wildflysch with ophiolitic detritus" (Laubscher & Bernoulli 1977). Note that all obducted ophiolitic thrust sheets, which overlie these ophiolitic mélange formations, as seen in the Bükk Mountains (Darno-Szavarskö ophiolites; Csontos 2000) and Dinarides (Western Vardar Ophiolitic Unit; e.g. Pamić et al. 1998; Dimitrijević 2001; Karamata 2006), consist of Jurassic age ophiolites. We suggest that these Jurassic Western Vardar ophiolites, together with the Triassic Meliata-Maliac ophiolites, once formed part of one and the same Neotethyan oceanic branch before obduction began (Fig. 2b), i.e. before Mid-Jurassic times. During the latest Jurassic to Early Cretaceous the young (Early to Mid-Jurassic) Western Vardar Ophiolitic Unit was obducted onto the passive continental margin of Adria, the ophiolite mélange formation defining the tectonic contact zone.

The Eastern Vardar Ophiolitic Unit extends into the South Apuseni and Transylvanian ophiolite belt and represents another part of the Meliata-Maliac-Vardar Neotethys (Săndulescu 1984) that was probably obducted onto the European Margin (Dacia Mega-Unit, Fig. 2b). Its original location with respect to that of the Western Vardar ophiolites is still enigmatic. We emphasize that the Eastern Vardar ophiolitic belt does not extend along the Sava Zone towards Zagreb. Hence, the Eastern Vardar Ophiolitic Unit does not form part of the Dinarides. Instead, it is now the most internal, structurally highest tectonic element on top of the Dacia Mega-Unit. In turn this implies that the Eastern Vardar ophiolites cannot be linked with the

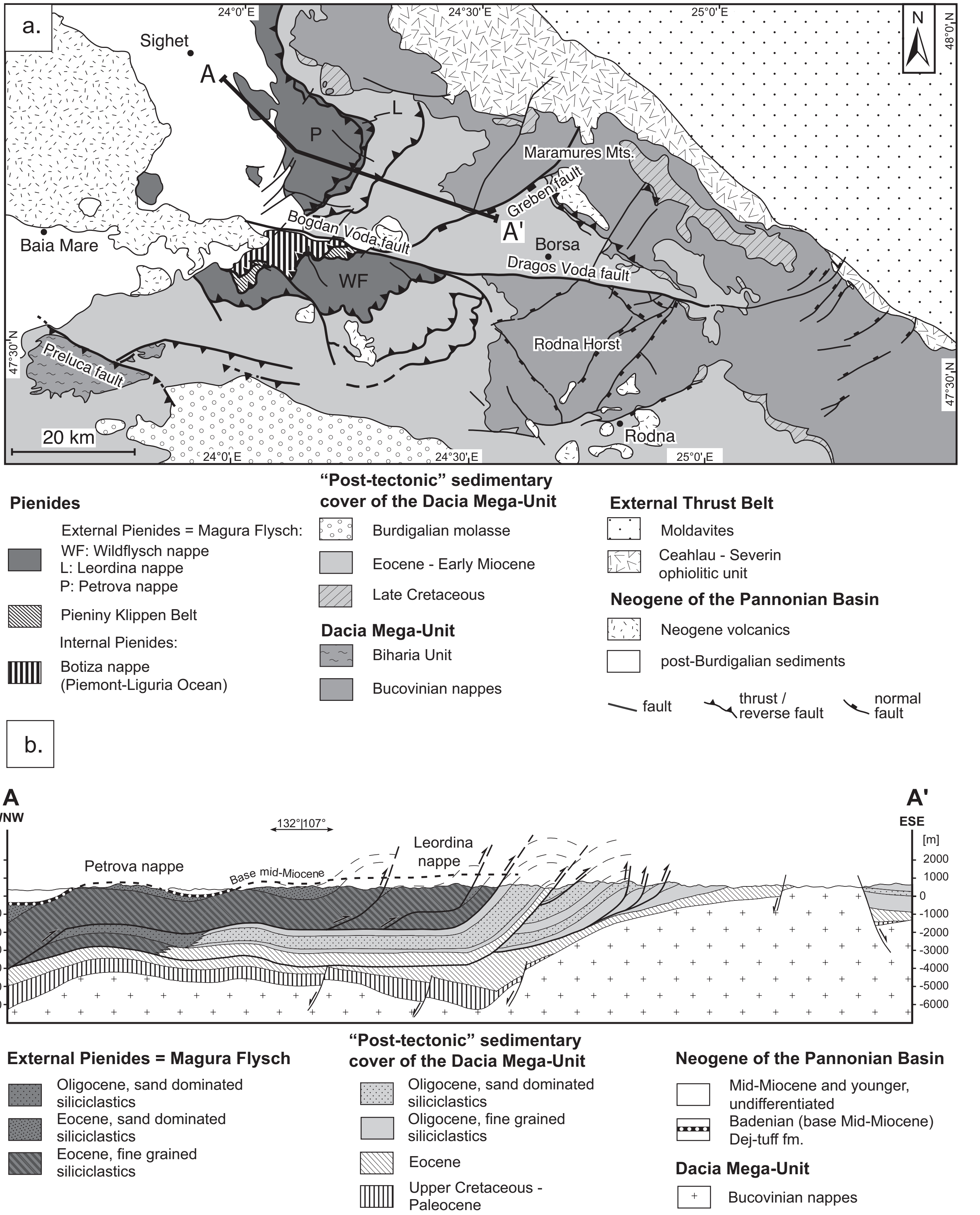

Fig. 3. Map (a) and cross-section (b) through the contact area between the ALCAPA and the Tisza-Dacia Mega-Units in Northern Romania (Maramures) after Tischler (2005) and Tischler et al. (2007). The Pienides of the Maramures area in Northern Romania form a tight arc in map view and were thrust onto the Tisza-Dacia Mega-units during Miocene times.

Alpine Tethys around the western margin of Tisza, nor do they extend E-ward into the North Dobrogea Orogen. In the south, i.e. between Beograd and Skopje, the Neotethyan Western and Eastern Vardar Ophiolitic Units parallel each other but are separated by a narrow band of the Sava Zone that, in our view, represents a suture zone between the Dinarides and the Tisza-Dacia Mega-Units which, in the Late Cretaceous, included their earlier obducted ophiolite sheets (Plate 1).

We conclude (1) that none of the branches of the Alpine Tethys and of Neotethys extends eastward into the North Dobrogea Orogen, although such connections have been proposed by many paleogeographic reconstructions (e.g. Stampfli & Borel 2004). The Triassic North Dobrogea rift (i.e. the Niculitel Zone; Savu et al. 1977; Seghedi 2001), which formed during a Permo-Triassic rifting event that affected also central Moesia, dies out rather rapidly northwestward (see review by Tari et al. 1997). We propose (2) that the ophiolitic remnants of Neotethys (the Triassic-age Meliata-Maliac ophiolites and the Jurassic-age Vardar ophiolites) occurring in the area of consideration formed part of one and the same Meliata-Maliac-Vardar oceanic basin. We will show later that there is no evidence for the existence of a "Pindos Ocean" and thereby accept the "one-ocean hypothesis" first formulated by Bernoulli & Laubscher (1972). We are convinced (3), that the different branches of Tethys found in the Alpine-Carpathian-Dinaridic system can only be followed eastward via the Dinarides and Hellenides into Turkey, but not into the Cimmerian system of the Black Sea to the north.

4. Detailed description of individual tectonic units of the Alps, Carpathians and Dinarides

This section provides the most important sources used for defining tectonic units and constructing the map (Plate 1). Moreover, a more detailed overview of all the tectonic units is presented and supported by a series of crustal-scale cross-sections (Plates 2 & 3). Names of tectonic units listed in the legend of Plate 1 are given in bold letters. The Alps will not be treated in detail (see Schmid et al. 2004a for a comprehensive review). We provide a relatively more comprehensive overview of the Dinarides since they are the least known part of the system.

4.1. The Miocene fold-and-thrust belt of the Alps and Carpathians

The Cretaceous and Cenozoic flysch units of the external Carpathian fold-and-thrust belt underwent most, though probably not all their total shortening during Miocene times. Deformation progressively migrated towards the foreland (Săndulescu et al. 1981a,b; Roca et al. 1995; Morley 1996; Zweigel et al. 1998; Matenco & Bertotti 2000).

The Neogene evolution of this flysch belt was mainly driven by the retreat of the Alpine Tethys subduction slab into the Carpathian embayment (Royden 1988). Prior to the Late Cretaceous onset of subduction, this embayment was occupied by Late Jurassic to Early Cretaceous oceanic lithosphere of the Alpine Tethys that was attached to the European continent (Balla 1987; Fig. 2c). Subduction of this oceanic lithosphere started from the Late Aptian to Albian onwards. It's covering Late Jurassic – Early Cretaceous sediments were scraped off, forming the Ceahlau/Severin nappe that was emplaced about 75 Ma ago ("Laramide" emplacement of the Ceahlau-Severin Ophiolitic Unit; e.g. Săndulescu et al. 1981a,b). There are several popular models such as slab detachment (e.g. Wortel & Spakman 2000; Sperner et al. 2005; Weidle et al. 2005), slab delamination (e.g. Girbacea & Frisch 1998; Gvirtzman 2002; Knapp et al. 2005), or thermal re-equilibration of the lithosphere (e.g. Cloetingh et al. 2004) that have been proposed to explain the evolution of this subduction and its present-day expression as a near vertical high velocity mantle anisotropy beneath Vrancia (e.g. Martin et al. 2006). Whatever model is adopted, all of them invoke a Miocene retreat of the subducted oceanic slab as the principal driving force for the final emplacement of the continental ALCAPA and Tisza-Dacia Mega-Units that occupy the internal parts of the Carpathian loop.

The Carpathian thrust belt consists mainly of flysch units in its inner part and grades outward into progressively more shallow-water strata, the most frontal thrusts involving the inner parts of the Carpathian foredeep. The most internal thrusts were emplaced in Late Cretaceous time whereas the external thrusts are of Neogene age. The age of the youngest thrust deformation at the front of this fold-and-thrust belt decreases from the West Carpathians (ca. 18 Ma) towards the Polish and Ukrainian Carpathians where the latest movements are dated as Late Miocene (ca. 12 Ma; Krzywiec 2001). Further to the SE and S along the entire East Carpathians to the bending zone north of Bucharest the age of the youngest thrusts remains constant at around 11 Ma, though the amount of shortening accomplished across them increases significantly in the Romanian segment (Roure et al. 1993; Dicea 1995; Matenco & Bertotti 2000; Krzwiec 2001). This was partly accommodated in the internal parts of the Carpathians by coeval late-stage sinistral strike-slip motion along the contact between ALCAPA and Tisza-Dacia (e.g. Tischler et al. 2007). The external flysch belt in the East Carpathians is completely allochthonous and contains sediments formerly deposited on continental or oceanic crust of the Carpathian embayment. However, these sediments, which are incorporated in the Miocene-age nappe sequence, do not contain any detritus from an oceanic basement (e.g. Săndulescu 1988). Hence the hypothesis that parts of the Carpathian embayment were underlain by oceanic crust (e.g. Balla 1987) remains speculative.

The external Carpathian fold-and-thrust belt consists of three nappes or thrust systems, some of which can be traced all the way along strike. These nappe systems are best developed in the East Carpathians where they are referred to as the Moldavides (Săndulescu 1994).

The external-most tectonic unit consisting of Middle Eocene to Late Miocene sediments, the ***thrusted internal foredeep***, is called Subcarpathian nappe in the East Carpathians (Săndulescu 1981a,b; Dicea 1995). This unit, which is thrust over the undeformed sediments of the Carpathians foreland, i.e. the external foredeep (Plate 3, Profile 3), can be traced north along strike into Ukraine where it wedges out between external foredeep and more internal thrust sheets (e.g. Kováč et al. 1999; Oszczypko 2006). In the foreland of the Romanian Carpathian bend zone and the South Carpathians this unit is partly buried beneath the latest Miocene-Pliocene (post-11 Ma), essentially undeformed sediments of the Dacic Basin (e.g. Jipa 2006) (Plate 2, Profile 4). The depocenter of these sediments is located near the town of Focsani ("Focsani Depression"; Tărăpoancă et al. 2003).

The Cenozoic Dacic basin forms the western branch of the larger and endemic Eastern Paratethys basin that was isolated from the Mediterranean and World Ocean since the beginning of the Oligocene in response to orogenic uplift of the Alpine-Carpathian-Tauride Caucasus domains (e.g. Rögl 1999). Parts of the Dacic basin, located between the Intramoesian and Trotus/Peceneaga-Camena Faults, were inverted during the Quaternary by folds and reverse faults involving the lower plate of the Eastern Carpathian thrust system; this led to the development of the synclinal configuration of the Focsani Depression (Plate 3, Profile 3; see also Leever et al. 2006).

In the South Carpathians foreland, the most internal and deformed part of the foredeep is referred to as Getic Depression. Although its latest Cretaceous to Late Miocene sediments are buried beneath the post-tectonic cover of the Dacic basin, subsurface data show that these were thrust over the Moesian foreland (Motaş & Tomescu 1983; Săndulescu 1988) (Plate 2, Profile 4). During the progressive N-, NE-, E- and ESE-ward transport of the Tisza-Dacia Mega-Units around Moesia, the Getic Depression was dominated by dextral strike-slip displacements. However, the total amount of Cenozoic shortening in the East Carpathians (<160 km; Ellouz et al. 1994) was largely accommodated within the South Carpathians (Fügenschuh & Schmid 2005; Răbăgia et al. 2007) rather than by shortening within the sediments of the Getic Depression. Palaeogene orogen-parallel extension within the South Carpathians (Schmid et al. 1998) was followed by Palaeogene to Early Miocene dextral movements and transtension along curved faults systems such as the Cerna-Jiu and Timok faults (Berza & Drăgănescu 1988; Ratschbacher et al. 1993; Kräutner & Krstić 2002, 2006). For this reason it is suspected that the predominantly Miocene-age shortening in the East Carpathians may have already started in Palaeogene times and that Middle to Late Miocene transpression and thrusting of the internal foredeep over Moesia is only of secondary importance (Răbăgia & Matenco 1999).

The ***Marginal Folds*** and ***Tarcau*** thrust system makes up the middle of three major Neogene thrust systems of the East Carpathians and involves Early Cretaceous to Early Miocene flysch series. Towards the SW the amount of shortening across this thrust system decreases and is gradually transformed into limited transpressional deformations where it merges into the South Carpathians (Plate 1). To the west of the bending area this thrust system wedges out completely and is missing in the South Carpathians where transpression was absorbed in the Subcarpathian nappe (Ştefănescu et al. 2000; Matenco et al. 2003). Northwards, these two units are traced into the ***Skole*** Unit of Poland (e.g. Oszczypko 2006) where this intermediate thrust system wedges out to the west.

Only the oldest and most internal thrust system of the Carpathian Miocene fold-and-thrust belt can be followed all along the East and West Carpathians into the Alps. In the East Carpathians it consists of, from external to internal, of the ***Audia, Macla*** and ***Convolute Flysch*** units that contain Late Cretaceous to Palaeogene strata (Săndulescu 1981a,b). These units gradually wedge out westward in the South Carpathians where they record SE-directed thrusting. Their internal geometry indicates associated strike-slip movements (Matenco & Bertotti 2000). Northwards, these units can be correlated with the external ***Subsilesian,*** intermediate ***Silesian*** and internal ***Dukla*** units of the Polish West Carpathians (Morley 1996; Kováč et al. 1999) that comprise Late Jurassic to Miocene sediments (Oszczypko 2006). Further west, in Slovakia, where the Dukla Unit is tectonically covered by the Magura Flysch Unit (Plate 2, Profile 2), the Silesian and Subsilesian Units can be traced westward into the Czech Republic (Zdanice Unit; Picha & Stranik 1999) and finally into the Waschbergzone of the easternmost Alps ("Molasse Zone" of Wessely 1987 in Profile 1 of Plate 2). Further westwards this unit wedges out. The Subalpine Molasse thrust slices of the Alps in western Austria and Switzerland were also mapped as part of this thrust system. Note, however, that in terms of basin evolution and tectonic position these Molasse sediments have little in common with their counterparts in the Eastern Alps and Carpathians.

4.2. Alps and West Carpathians (ALCAPA Mega-Unit)

4.2.1. What is ALCAPA?

The term ALCAPA (Alps-Carpathians-Pannonia) as defined by Csontos & Vörös (2004) stands for a tectonic mega-unit that includes the Eastern Alps, the West Carpathians and the Transdanubian ranges north of Lake Balaton. The Miocene fill of the Pannonian Basin separates these three major areas of older rock exposures from one another. We mapped the exact boundaries of ALCAPA as follows on Plate 1:

The contact between the Rhenodanubian and Magura flysch units along with the Miocene flysch fold-and-thrust belt delimits the northern boundary of this Mega-Unit, which underwent a combination of lateral escape, thinning and subsidence during the Miocene evolution of the Pannonian Basin in Miocene times (Ratschbacher et al. 1991a,b; Horvath et al. 2006). The Periadriatic Line and its eastern extension, the Balaton Line (Fodor et al. 1998), that were active during this escape, define the southern boundary of ALCAPA all the way to Lake Balaton. The Balaton Line was delineated according to the compilation of subsurface data provided by Haas et al. (2000). Further east, we define the southern boundary of ALCAPA according to subsurface data provided by Csontos & Nagymarosy (1998) and Fülöp & Dank (1987). However, and in contrast to the us-age of the term ALCAPA by previous authors, we exclude the Bükk Mountains from this Mega-Unit. The principal reason is that the Bükk Mountains can nowadays be considered as a piece of the Dinarides (Kovács et al. 2000, 2004; Dimitrijević et al. 2003; Velledits 2006) rather than part of the Alpine-West Carpathian chain. The Bükk Mountains were displaced within the Mid-Hungarian Fault Zone (Tischler et al. 2007), a broader fault zone that is delimited to the south by the Mid-Hungarian Line (Csontos & Nagymarosy 1998), or "Zagreb-Zemplin Line" (Haas et al. 2000), and to the north by the eastern continuation of the Balaton Line. Consequently, we let the southern boundary of ALCAPA coincide with the southern part of the Darno Line and further eastwards with the Nekéseny Fault (Haas 2001; see Fig. 1). These fault zones separate the Bükk Mountains from two inselbergs that we consider as part of the Inner West Carpathians (Uppony and Szendrö Mountains of NW Hungary; see Haas 2001, his Figs. 78 & 106). The eastern tip of ALCAPA extends into Northern Romania where it was identified from subsurface and field data (Săndulescu et al. 1978, 1993; Kováč et al. 1995; Tischler et al. 2007).

4.2.2. Europe-derived units within ALCAPA

In Plate 1, and regarding the Alps, we grouped the Helvetic and Ultrahelvetic cover nappes including the External Massifs (***Helvetic***) together with the ***Subpenninic*** nappes. The term "Subpenninic" denotes pre-Mesozoic basement units, on which the sediments now exposed in the Helvetic and Ultrahelvetic nappes were originally deposited, as well as more distal parts of the European upper crust, and its metamorphic cover, as exposed in the Tauern window (Schmid et al. 2004a; see Fig. 1). The most internal Europe-derived nappes are paleogeographically assigned to the ***Briançonnais***, a crustal block that was detached from the European passive margin during the Early Cretaceous opening of the Valais branch of the Alpine Tethys (Fig. 2c; Frisch 1979; Stampfli 1993). The easternmost Briançonnais nappes are exposed in the Engadine window (Fig. 1).

However, according to our interpretation Subpenninic units extend in the subsurface much further to the east. For instance Subpenninic units, referred to as "Penninic" in the crustal-scale profile across the easternmost Alps by Tari (1996), are needed to fill the space between the base of ophiolitic nappes attributed to the Alpine Tethys (as exposed nearby in the Rechnitz window) and the Moho (Plate 2, Profile 1). However, we suspect the presence of such Subpenninic units in the subsurface also further to the east, namely in the West Carpathians (Plate 2, Profiles 1 & 2). Information on the deep structure of the West Carpathians is provided by a seismic transect presented by Tomek (1993) who assigned a large rock volume below the Alpine Tethys suture (Vahicum; Plašienka 1995a,b) to the Briançonnais. Yet, as the Briançonnais paleogeographic domain proper terminates west of the Tauern window (Schmid et al. 2004a), except for some small analogous crustal slivers found in the Pieniny Klippen Belt of the West Carpathians (Birkenmajer 1986; Trümpy 1988), we prefer to attribute this rock volume to the distal European margin, that is, to the Subpenninic units.

A direct comparison of the profiles across the Eastern Alps and West Carpathians (Plate 2, Profiles 1 & 2) highlights an important difference pertaining to the northern boundary of ALCAPA, which is only evident in cross sections. In the profile across the easternmost Alps (Plate 2, Profile 1) the basement of the undeformed European foreland is traced along a gently inclined thrust at the base of the Alpine accretionary wedge far to the south into the area of the Hungarian Plain (Tari 1996). By contrast, the profile through the West Carpathians (Plate 2, Profile 2) depicts a steeply dipping dextral strike-slip zone in the depth projection of the Pieniny Klippen Belt according to our interpretation. This crustal-scale sinistral strike-slip fault zone was active during the Miocene and formed the northern boundary of the ALCAPA Mega-Unit (Central and Inner West Carpathians) during its eastward escape (i.e. Nemčok 1993; Sperner et al. 2002). It thus post-dates nappe emplacement in the Inner West Carpathians. The northern parts of Profile 2 (Plate 2) suggest that the European foreland crust was imbricated (modified after Roca et al. 1995) during the transpressional emplacement of ALCAPA and the development of the Miocene West Carpathian fold-and-thrust belt. In this process the Outer West Carpathian Magura Flysch Unit was imbricated and accreted to ALCAPA whilst severe late-stage Miocene NNE-SSW-directed back-thrusts affected the more internal units in the High Tatra Mountains (Plate 2, Profile 2) (Sperner et al. 2002).

4.2.3. Remnants of the Alpine Tethys in ALCAPA

Most authors accept the existence of two branches of the Alpine Tethys, the ***Piemont-Liguria*** and ***Valais*** Oceans, respectively. However, this division can only be made where remnants of the intervening Briançonnais micro-continent are present. This is no longer possible east of the Engadine window. Hence, at first sight a paleogeographic separation of the Alpine Tethys into two separate branches (Froitzheim et al. 1996; Schmid et al. 2004a) becomes redundant in the Eastern Alps (see discussion in Kurz 2005 and Schmid et al. 2005) where these branches apparently merged into one single oceanic basin. Nevertheless, we make a distinction between the eastern equivalents of these two branches since an evaluation of the timing of accretion of these oceanic units (Late Cretaceous vs. Cenozoic) to the Austroalpine, respectively to the West Carpathian Tatric-Veporic-Gemeric upper plate, is an important criterion for distinguishing between the equivalents of the two branches of the Alpine Tethys east of the Engadine window (for additional criteria see discussion in Schmid et al. 2005). Consequently, we correlate the Valais Ocean with the ***Rhenodanubian Flysch*** since accretion of both units the to the Alpine nappe sequence occurred in Eocene times. The lateral equivalent even further east and in the West Carpathians is the ***Magura Flysch*** accreted to the Tatric-Veporic-Gemeric orogenic wedge during Late Oligocene to Miocene times.

The ***Pieniny Klippen Belt*** (Birkenmajer 1986) was mapped separately (Plate 1) because it represents a rather peculiar, subvertical and narrow structural unit at the surface (Plate 2, Profile 2). It forms the boundary between Outer and Central West Carpathians for some 700 km from the St. Veit Klippen near Vienna to northern Romania. This belt typically consists of relatively erosion-resistant non-metamorphic blocks of Mesozoic strata of many different facies surrounded by less competent Late Cretaceous to Palaeogene marlstones and flysch. Only small parts of this belt, that were derived from the pelagic Kysuca-Pieniny Basin, hint to the former presence of a now completely subducted eastern continuation of the Piemont-Liguria Ocean, referred to in the West Carpathians as the ***Vahic*** Ocean (Plašienka 1995a, 2003). Southward subduction of the Vahic oceanic crust started in the Senonian (Plašienka 1995a,b). This ocean was located south of sedimentary rocks assigned to the Czorsztyn Ridge, the second major constituent of the Pieniny Klippen Belt that probably represents an eastern equivalent of the Briançonnais continental block (Birkenmajer 1986). It is important to emphasize that pebbles of Jurassic-age blueschists, thought to be derived from the Pieniny Klippen Belt (Dal Piaz et al. 1995), were not encountered in primary deposits, but were found as recycled components in the Gosau-type sediments of the Klape Unit ("Periklippen Zone"; Ivan et al. 2006). This unit is not considered as forming part of the Pieniny Klippen Belt s.str., but as a displaced fragment of the Central West Carpathians (Plašienka 1995b). Hence these blueschist pebbles more likely originated from the Meliata-Maliac oceanic domain which crops out further to the south in the Central and Inner West Carpathians (see below) and which is not part of the Alpine Tethys.

In contrast to the Alps, in which ophiolitic nappes derived from the Piemont-Liguria Ocean are well exposed (Schmid et al. 2004a), remnants of the Vahicum that represents the eastern extension of the latter do not outcrop, except for very small tectonic windows some 10 km south of the Pieniny Klippen Belt (Belice Unit exposed in the Považský Inovec Mountains; see Plašienka 1995a,b for details). Based on seismic data (Tomek 1993), we suspect, however, that in the Central West Carpathians remnants of the Vahicum separate the Tatricum from suspected Subpenninic elements at depth (Plate 2, Profile 2). In Eastern Slovakia and Ukraine boreholes have reached, in the subsurface of the East Slovak basin, the ***Iňačovce-Kriscevo*** Unit that includes some serpentinite bodies (Soták et al. l993, l994, 2000) (Fig. 1). These rocks are along strike with the Vahicum and hence we regard them as equivalents of the Piemont-Liguria Ocean.

The easternmost occurrences of the Pieniny Klippen Belt (Kysuca-Pieniny-type sediments) form part of the so-called Pienides of the Maramures area in Northern Romania (Poiana Botizei locality; Săndulescu et al. 1979/1980; Bombită & Savu 1986). In map view they form a tight arc (Plate 1, Fig. 3). These easternmost Pieniny Klippen Belt lithologies are a key-element for understanding the relationship between West and East Carpathians (Săndulescu et al. 1981a, 1993; Tischler at al. 2007). This is because they separate, within the Pienides, the more internal Botiza nappe, a lateral equivalent of the Iňačovce-Kriscevo Unit of Eastern Slovakia and Ukraine that is attributed to the Piemont-Liguria Ocean, from the easternmost equivalents of the Magura Flysch (Wildflysch nappe of the External Pienides, see Fig. 3). The Pienides, including both these oceanic units and the Pieniny Klippen Belt in between them, were thrust SE-ward during the Early Burdigalian over the Dacia Mega-Unit, including its "post-tectonic" Late Cretaceous to Cenozoic cover. Later they were dissected by the Bogdan-Dragos-Voda fault system (Tischler et al. 2007). As such, the Pienides define the easternmost tip of ALCAPA that is thrust over the Tisza-Dacia Mega-Units (Fig. 3; see Tischler 2005; Tischler et al. 2007; Márton et al. 2007). This thrusting was kinematically linked to movements along the Mid-Hungarian fault zone during the lateral extrusion of ALCAPA. The geometry of the tight Pienides arc in northern Romania led us to correlate the Botiza nappe, which we consider as the lateral equivalent of the Piemont-Liguria branch of the Alpine Tethys, with the Szolnok Flysch that occurs in the subsurface of the Pannonian Basin in eastern Hungary near Debrecen (Plate 1). Furthermore, we propose to connect this Szolnok flysch belt, which forms part of the Mid-Hungarian Fault Zone (or the ophiolitic "Intrapannonian Belt" in the sense of Channell et al. 1979) with the westernmost parts of the ophiolite-bearing Sava Zone of the Dinarides near Zagreb. According to our interpretation, Miocene dextral movements within the Mid-Hungarian Fault Zone (Zagreb-Zemplin line; Haas et al. 2000) disrupted the original connection between the Sava Zone of Croatia and northern Bosnia and the Pienides of northern Romania.

4.2.4. Adria-derived nappes in ALCAPA and remnants of the Triassic Meliata-Maliac Ocean

The compilation of Adria-derived units, known as the Austroalpine nappes in the Alps, follows the scheme proposed by Schmid et al. (2004a), except for minor modifications in the easternmost Alps that are discussed below. Correlation of these Alpine units across the Vienna Basin and the Little Hungarian Plain with the equivalent ones in the West Carpathians is based on compilations of subsurface (seismic and/or borehole) data provided by Fülöp & Dank (1967), Fülöp et al. (1987), Tari (1994, 1996), Plašienka et al. (1997a), Tari et al. (1999), Haas et al. (2000), and Haas (2001).

The ***Lower Austroalpine*** nappes representing the structurally lowermost units of the Austroalpine nappe sequence were derived from the northern passive margin of Adria that faced the Piemont-Liguria Ocean. At the eastern margin of the Alps, the subdivision proposed by Schmid et al. (2004a) was slightly modified: the Semmering nappe, without the overlying nappes that are built up by the so-called Grobgneiss Unit and the Stralegg Complex (Schuster et al. 2001, 2004), was included into the Lower Austroalpine nappe pile. This permits a correlation of the Lower Austroalpine units with an equivalent nappe sequence of the West Carpathians known as the ***Tatricum*** (i.e.

Plašienka et al. 1997a). Very probably the rocks in the Leitha Mountains of the easternmost Alps SE of Vienna extend eastward into the Hainburg Hills and the Malé Karpaty Mts. of the West Carpathians near Bratislava (Plašienka et al. 1991). Structurally, the Tatricum represents the lowermost Adria-derived nappe that consists of Variscan basement and its sedimentary cover and that was derived from a position adjacent to the Vahic (= Piemont-Liguria) Ocean (Dumont et al. 1996; Plašienka et al., 1997a). As revealed by deep seismic transects, the Tatricum forms a tabular, upwards slightly convex and more than 10 km thick thrust sheet (Plate 2, Profile 2) that extends into the lower crust below the Veporic units (Tomek 1993; Bielik et al. 2004).

Subdivision of the Upper Austroalpine nappe system and its West Carpathian equivalents into ***northern margin of Meliata, Eoalpine high-pressure belt and southern margin of Meliata units***, as proposed in our compilation (Plate 1), follows essentially the new subdivision of the Upper Austroalpine units of the Alps as proposed by Schuster et al. (2004) and Schmid et al. (2004a). Note, however, that this concept, as briefly outlined below, applies strictly only to the West Carpathians and the eastern parts of the Eastern Alps since the oceanic Meliata-Maliac embayment did not extend into the Austroalpine domain of the more western areas. This applies also to the related Eoalpine high-pressure belt, the westernmost parts of which are present in the southern Ötztal basement (Schmid et al. 2004a).

This new subdivision is supported by the following data and/or concepts:

(1) The sedimentary nappes of the Northern Calcareous Alps were derived from the northern passive margin of the Triassic ***Meliata Ocean*** that formed a branch of the Neotethys. The most distal parts of this margin are characterized by the classical Hallstatt facies (e.g. Mandl 2000; Gawlick & Frisch 2003). We do not share the view of Neubauer et al. (2000) who proposed that within the domain of the present-day Northern Calcareous Alps a Meliata "suture" separates two conjugate passive margins. Instead, we interpret the stratigraphic evidence for Late Jurassic compressional tectonics in the Northern Calcareous Alps (Gawlick & Frisch 2003; Gawlick & Schlagintweit 2006) to be related to the development of an accretionary wedge at the front of an obducted Jurassic-age ophiolite body (Vardar Ocean). This obduction event followed an earlier pulse of intra-oceanic subduction (see chapter Dinarides) during which the Triassic parts of the Neotethyan oceanic lithosphere, namely the Meliata Ocean, were consumed.

(2) After this obduction event, major deformations accompanied by the stacking of the Austroalpine nappe units started during the Late Valanginian (ca. 135 Ma). This is indicated by the development of the syn-orogenic Rossfeld sedimentary basin in the Northern Calcareous Alps (Faupl & Wagreich 2000) in the Northern Calcareous Alps. It is this shortening, referred to as the Eo-Alpine tectono-metamorphic event, that we relate to the final closure of the westernmost parts of the Neotethys oceanic realm. A SE- to E-dipping subduction zone developed within and/or near the western termination of the Neotethys oceanic embayment into Adria (Fig. 2), possibly along a Jurassic strike-slip fault (see discussions in Schuster & Frank 1999; Frank & Schlager 2006). In the Alps and along this intra-continental subduction zone located at the western end of the Meliata Ocean, crustal rocks were transported to great depths some 90 Ma ago (Thöni 2006).

(3) The intra-continental, partly eclogitic, former subduction zone, referred to as ***Eoalpine high-pressure belt*** in Plate 1, subdivides the Upper Austroalpine basement nappe sequence into two parts (Schmid et al. 2004a): the Silvretta-Seckau nappe system in its footwall (***northern margin of Meliata*** in Plate 1, together with the Northern Calcareous Alps and the Grauwackenzone) and the Ötztal-Bundschuh and Drauzug-Gurktal nappe systems in its hanging-wall (part of the ***southern margin of Meliata*** in Plate 1, together with the Southern Alps).

Correlation of the Upper Austroalpine units of the Alps with the major tectonic units of the West Carpathians (Plate 1) has to contend with considerable difficulties, since major changes occur along strike in the architecture of the Cretaceous-age (Eoalpine) orogen. Hence, individual Upper Austroalpine units cannot be cylindrically traced eastwards. The absence of exposed Eoalpine high-pressure rocks east of the Alps, possibly due to their burial beneath the fill of the Pannonian Basin, presents an additional difficulty.

We correlate the Drauzug-Gurktal nappe system with the Transdanubian ranges since we interpret both of these units to structurally represent the hanging-wall with respect to the Eoalpine high-pressure belt. Therefore Drauzug-Gurktal nappe system and Transdanubian ranges paleogeographically represent realms along the southern margin of the Meliata-Maliac Ocean. The Transdanubian ranges are made up of weakly metamorphosed Variscan basement that is overlain by a non-metamorphic Permo-Mesozoic cover. This cover exhibits close similarities to the Drauzug and the Southern Alps (e.g. Kazmer & Kovács 1985; Haas et al. 1995). This correlation is also warranted on structural grounds; the Transdanubian ranges are located immediately N of the Balaton Line, the eastern extension of the Periadriatic Line, along which the Austroalpine units (Drauzug and Transdanubian ranges) escaped to the east (Kazmer & Kovács 1985). Hence, the Drauzug-Gurktal nappe system and the Transdanubian Ranges are interpreted as allochthonous tectonic units that structurally overlie the postulated eastern extension of the Eoalpine high-pressure units (Plate 2, Profile 1).

Conversely we correlated all units of the Central and Inner West Carpathians with the Silvretta-Seckau nappe system, the Northern Calcareous Alps and the Grauwackenzone. We propose that the easternmost parts of these units were derived from the northern margin of the Meliata-Maliac Ocean. These units form the lower plate with respect to the Eoalpine high-pressure belt of the Alps and include the thick-skinned

Veporicum and Gemericum thrust sheets, as well as a series of detached sedimentary nappes. From external to internal, the latter are referred to as Fatricum, Hronicum and Silicicum units (Plašienka 1997a; see Plate 2, Profile 2). According to our interpretation this group also includes rock associations that crop out in isolated inselbergs, the Uppony, Szendrö and Zemplin mountains (Fig. 1), whose assignment to specific units is much debated. These inselbergs are located to the northwest and north of the southern boundary of ALCAPA, marked by the Darno and Nekészeny Faults (Fig. 1).

The structural and paleogeographic relationships between the most internal and structurally highest system of cover nappes (Silicicum in Slovakia or Aggtelek-Rudabánya Unit in Hungary; Haas 2001) and the Meliaticum are still enigmatic. One school of thought considers the Meliaticum as a suture zone, referred to as Rožňava suture (i.e. Plašienka et al. 1997a,b). In this case, however, the Silicicum, together with a weakly metamorphic cover slice at its base referred to as Tornaicum (Martonyi in Hungary; see Fodor & Koroknai 2000; Less 2000) or as South Rudabányaicum (Kozur & Mock 1997), would represent the upper plate with respect to this suture as they overlie the Meliaticum. Hence, these units ("Inner West Carpathians" of Plašienka 1997a) would, contrary to our interpretation (Plate 1 and Plate 2, Profile 2), represent the southern margin of the Meliata-Maliac Ocean.

We chose another option. According to our interpretation the Silicicum-Aggtelek Unit was derived from the northern passive continental margin of the Triassic Meliata-Maliac Ocean, with the Bodva sub-unit in northern Hungary (e.g. Kovács 1992; Kovács et al. 1997) representing its southernmost and most distal parts (Fig. 4a). The paleogeographic position of the Silicicum-Aggtelek Unit is very reminiscent of that of the Juvavic nappes of the Northern Calcareous Alps (Fig. 4a). Due to the development of a triangle structure formed during Late Jurassic obduction of the West Vardar Ophiolitic Unit, a part of the northern distal passive margin of the Meliata-Maliac Ocean (Silicicum-Aggtelek unit) now tectonically overlies ophiolitic mélange with remnants of the Meliata-Maliac Ocean (Fig. 4b, right). Hence, for the Western Carpathians we propose that S-directed back thrusting, associated with the formation of this triangle structure, is responsible for the present-day tectonic position of the Silica-Aggtelek cover nappes in the hanging-wall of obducted remnants of the Triassic Meliata-Maliac Ocean (Plate 2, Profile 2; Fig. 4b, right). Our field observations support S-directed Jurassic thrusting for the southernmost exposures of the West Carpathians in Northern Hungary. Subsequent Cretaceous deformation was N-directed (Balla 1987), with in-sequence thrusting gradually propagating northward (Fig. 4c, right; Plašienka 1997a,b).

Consequently, we place the suture of the Meliata-Maliac Ocean further to the south, namely between the southernmost exposures of the Central and Inner West Carpathians of the Uppony Mountains in the north and the Bükk Mountains to the south. The latter represent a displaced part of the Dinarides. This suture formed during the Early Cretaceous and was severely overprinted by Cenozoic strike-slip faulting along the Darno Line during the lateral eastward extrusion of the Alps and Dinaridic fragments (Plate 1 and Plate 2, Profile 2).

A geometrically similar situation is found in the easternmost Alps (Mandl & Ondrejickova 1991; Kozur & Mostler 1991; Neubauer et al. 2000; Frank & Schlager 2006). In spite of the great similarities between the nappe units of the West Carpathians and the Eastern Alps (Fig. 4), we prefer a different scenario for the Northern Calcareous Alps, largely following Mandl (2000) and Frank & Schlager (2006). In this interpretation Late Jurassic/Early Cretaceous emplacement of the most distal elements, the so-called "Tiefjuvavikum" including Meliata-type ophiolitic mélanges (Fig. 4b, left), was followed by out-of-sequence Cretaceous imbrication of the "Hochjuvavikum" (Fig. 4c, left).

The ***Meliata*** Unit of the West Carpathians and easternmost Eastern Alps includes different elements that do not represent ophiolitic bodies in a strict sense. The Meliaticum of the West Carpathians consists of two units: (1) the metamorphosed Bôrka Unit, that was derived from near the ocean-continent transition of the northern margin of the Meliata-Maliac Ocean and which underwent low-temperature, high-pressure (12 kbar) metamorphism (Faryad 1995a & b, 1997; Mello et al. 1998), and (2) the Meliata Unit s.str. (Mock et al. 1998), a non-metamorphic ophiolite-bearing mélange which forms part of a Jurassic accretionary flysch complex containing radiolarites, olistostromes, mélanges and ophiolitic bodies (Kozur & Mock 1997). Glaucophane-bearing basalts from the base of the Meliata accretionary complex (Bôrka Unit) overriding the southernmost Gemericum yield Middle to Late Jurassic ages (150–160 Ma) for their HP-LT metamorphism (Maluski et al. 1993; Dallmeyer et al. 1996; Faryad & Henjes-Kunst 1997). This demonstrates that Mid-Jurassic intra-oceanic subduction processes preceded latest Jurassic obduction onto the distal northern continental margin of the Meliata-Maliac Ocean that was associated with the development of an accretionary wedge. This subduction clearly pre-dated, and is hence not related to the Late Valanginian onset of nappe stacking in the Eastern Alps (Gawlick & Frisch 2003) and West Carpathians.

We can only speculate that the Eoalpine high-pressure belt of the Eastern Alps, which represents a Cretaceous-age partly intra-continental suture that post-dates obduction of the Meliaticum onto the Juvavic and Gemeric units of the Alps and West Carpathians, respectively, might also be present in the subsurface of the Pannonian plain. We suggest its presence along a geophysically defined belt of linear deep-seated faults consisting of the Raba and Hurbanovo-Diósjenő faults (Fig. 1), based on data from Plašienka et al. (1997a; their Fig. 2). This Cretaceous-age suture (Eoalpine high-pressure belt of Plate 1) would coincide with the northern limit of the Transdanubian Range Unit along the Diósjenő Fault, as proposed by Haas (2001). Borehole data analysed by Koroknai et al. (2001), revealed immediately north of the Diósjenő Line the occurrence of basement that is characterized by Eo-Alpine amphibolite-grade metamorphism, typical for the Veporic Unit of the West

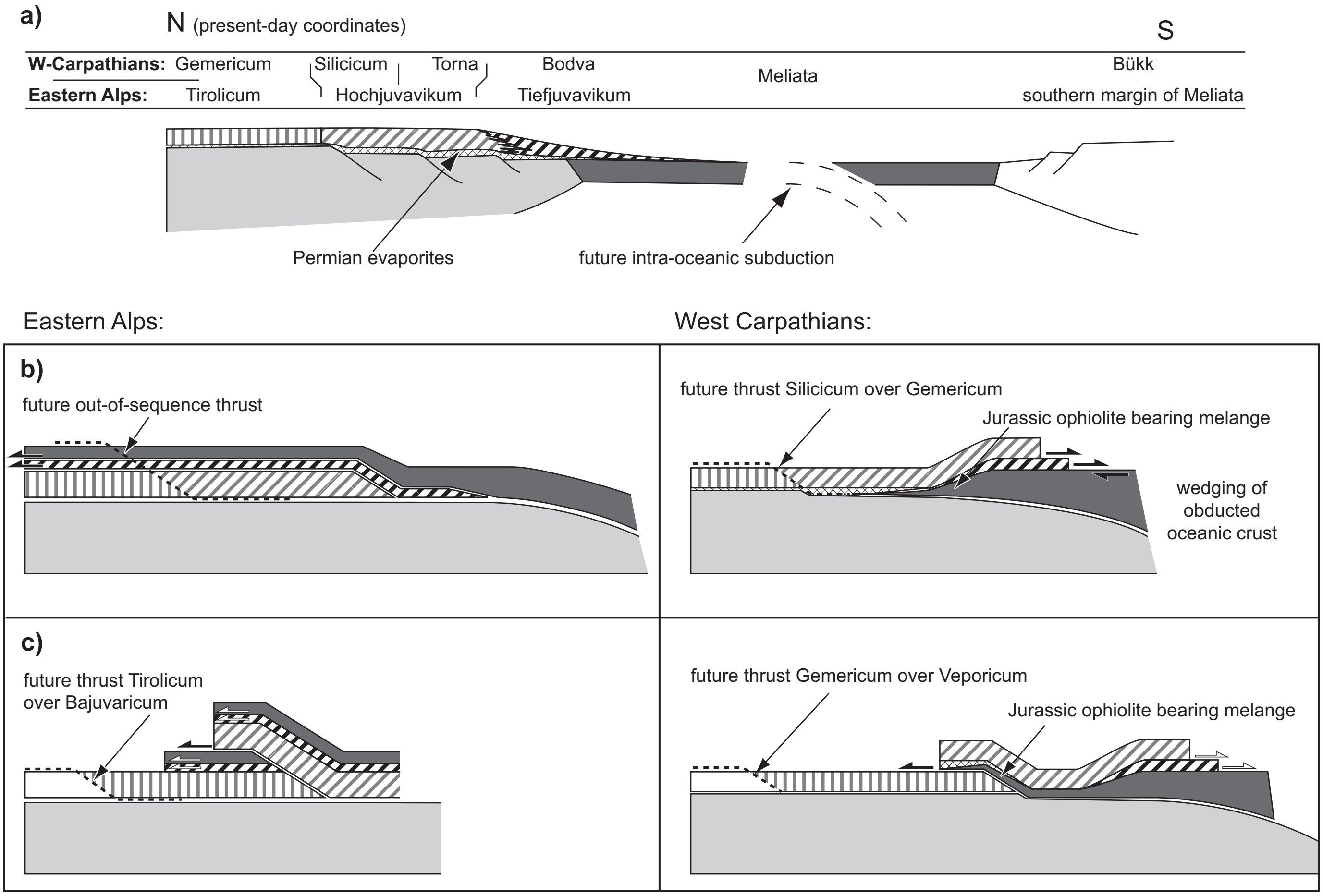

Fig. 4. Conceptual model for Late Jurassic and Cretaceous thrusting affecting the former passive continental margin adjacent to the Meliata Ocean in the Eastern Alps (Northern Calcareous Alps) and in the West Carpathians. a) Former passive margin of Neotethys and terminology of the various paleogeographic domains and tectonic units as used in the West Carpathians (top line) and the Eastern Alps (bottom line). b) Late Jurassic thrusting in the Alps (left) and in the West Carpathians (right: triangle structure). c) Overprinting during Early Cretaceous orogeny in the Alps (left) and in the West Carpathians (right).

Carpathians. Moreover, these data indicate that this high-pressure belt probably wedges out eastward (Plate 1). In addition, these findings provide further evidence for the existence of an important tectonic boundary that separates the non-metamorphic Transdanubian Ranges, derived from south of the Meliata-Maliac Ocean, from the Veporic Unit of the Central West Carpathians that we place to the north of the Meliata oceanic embayment.

4.3. East Carpathians, South Carpathians, Transylvanian Basin and Carpatho- Balkanides (Danubian nappes and "Dacia")

4.3.1. Overview

The internal tectonic units of the East and South Carpathians, which are fringed to the E and S by the external Miocene thrust belt, were emplaced in Cretaceous time (i.e. Săndulescu 1984, 1994; Kräutner et al. 1988; Kräutner 1996). Săndulescu (1984, 1994) referred to these units as the Marginal, Outer and Median Dacides. In the Hungarian literature, the Outer and Median Dacides are referred to as the Dacia Mega-Unit or terrane (i.e. Csontos & Vörös 2004), whilst Burchfiel (1980) refers to them as the "Rhodopian fragment".

The Danubian nappes, also referred to as Marginal Dacides, however, were originally part of the Moesian foreland. The Outer and Median Dacides of the Dacia Mega-Unit, together with more internal units located below the Transylvanian Basin and cropping out in the North Apuseni Mountains (Tisza Mega-Unit; e.g. Haas & Pero 2004, referred to as Internal Dacides by Săndulescu 1984), moved into the Carpathian embayment during Cenozoic times (i.e. Royden 1988; Fodor et al. 1999; Márton 2000; Wortel & Spakman 2000; Csontos & Vörös 2004; Horváth et al. 2006). These internal units were finally docked against the European foreland during the Miocene, controlling the evolution of the earlier described external fold-and-thrust belt that involves Cretaceous to Tertiary-age flysch units.

Along the western margin of the Moesian Platform the South Carpathian units can be traced further southward into eastern Serbia and western Bulgaria. There they are referred to as the Carpatho-Balkanides. They also include, what Serbian authors refer to as the Serbo-Macedonian Massif (e.g. Dimitrijević 1997). Some of these units can then be traced eastward into the Balkan Orogen whilst others follow the eastern rim of the Eastern Vardar Ophiolitic Unit and extend southward into Northern Greece (Plate 1). Two curved strike-slip faults, the Cerna-Jiu Fault (Berza & Drăgănescu 1988) and the Timok Fault (Visarion et al. 1988; Moser 2001; Kräutner & Krstić 2002, 2006), displace the Cretaceous nappe sequence along the contact between the Dacia Mega-Unit and the Moesian Platform. These faults accommodated Early Oligocene to Early Miocene dextral strike-slip motion with total displacements of up to 100 km (Moser 2001; Fügenschuh & Schmid 2005) as Dacia was mouldered around the western tip of the Moesian Platform (Ratschbacher et al. 1993). The Timok Fault can be traced southward and appears to be kinematically linked to extensional deformation in the Osogovo Mountains (Kounov et al. 2004) and along the Strymon fault system of Northern Greece (Dinter 1998; Kilias et al. 1999).

4.3.2. Danubian nappes

The ***Danubian nappes*** consist of a Neoproterozoic basement (e.g. Liégeois et al. 1996; Seghedi et al. 2005), Paleozoic rocks that were deformed during the Variscan cycle (e.g. Iancu et al. 2005) and Carboniferous to Late Cretaceous cover sequences (Berza et al. 1983, 1994). This imbricated nappe sequence developed by eastward thrusting during the latest Cretaceous (Late Campanian – Maastrichtian; often referred to as the "Laramide" phase in the Romanian literature; e.g. Kräutner 1993; Berza et al. 1994) under low-grade (sub-greenschist to lowermost greenschist facies) metamorphic conditions (Berza & Iancu 1994). It is best exposed in a tectonic window in the South Carpathians referred to as the Danubian window (e.g. Murgoci 1905). After Late Cretaceous nappe emplacement this unit was exhumed during Late Eocene to Oligocene time (Schmid et al. 1998; Fügenschuh & Schmid 2005). The rocks in the Danubian nappes have a Neoproterozoic ("Panafrican") tectono-metamorphic evolution similar to that of the basement in the Moesian Platform (e.g. Liégeois et al. 1996; Seghedi et al. 2005), a unit with Northern Gondwana paleogeographic affinities (Vaida et al. 2005). The Mesozoic cover of the Danubian nappes is characterized by Early Jurassic Gresten facies (e.g. Năstăseanu et al. 1981), followed by Late Jurassic to Early Cretaceous platform carbonates, Albian to Turonian deeper marine pelagic limestones and marls that give way to a Turonian and Senonian flysch sequence. Most authors regard the Danubian nappes as having been detached from the Moesian Platform, as shown in Profile 4 (Plate 2).

Following Visarion et al. (1978) and Ştefănescu et al. (1988) we suggest that the Danubian nappes continue northward in the subsurface below the East Carpathians (see Plate 3, Profile 3) as far north as the pre-Neogene Trotus Fault (Fig. 1). Southwards the Danubian nappe sequence extends into eastern Serbia and western Bulgaria (see Plate 1 & Plate 3, Profile 5) as indicated by the 1 : 100'000 sheets of Geological Maps of former Yugoslavia (Osnovna Geološka Karta SFRJ) and data provided by Kräutner & Krstić (2002, 2006) and Cheshitev et al. (1989). In Serbia (Dimitrijević 1997) the Danubian nappes are locally known as Miroč Unit (west of the Timok fault) and Vrška Čuka Unit (east of the Timok fault), or as Vrška Čuka-Miroč terrane (Karamata 2006). The Vrška Čuka Unit extends into western Bulgaria, where it is known as the West Balkan Unit (Kounov 2002). However, in the Balkan Mountains north-directed Eocene age thrusting overprinted many of the older Cretaceous structures (Boyanov et al. 1989).

4.3.3. The Ceahlau-Severin Ocean

The ***Ceahlau nappe*** (including the Black Flysch and Baraolt thrust sheets) of the East Carpathians and the equivalent ophiolite-bearing ***Severin nappe*** of the South Carpathians form a wedge that was accreted already in mid-Cretaceous (Aptian) time to the continental units of the overlying Bucovinian-Getic nappes (Săndulescu 1984). Ceahlau and Severin Ophiolitic Units represent the relics of what we refer to as the Ceahlau-Severin Ocean.

The Ceahlau nappe of the East Carpathians is the largest and main unit derived from this ocean. The Black Flysch and Baraolt thrust sheets occupy a higher tectonic position but were mapped as part of the Ceahlau-Severin Ophiolitic Unit in Plate 1. The Black Flysch nappe (Bleahu 1962; Săndulescu 1975) exposes a basement that consists of massive basaltic flows and dykes; younger basalts also intrude into the overlying Kimmeridgian-Aptian sediments. Since the mafic complex displays intra-plate geochemical characteristics it is not part of an ophiolitic sequence in the strict sense (Săndulescu et al. 1981a). These rocks were possibly located at the NW margin (in present day coordinates) of the Carpathian oceanic embayment and perhaps formed along large transtensional faults (Badescu 1997). Near the southern termination of the East Carpathians, the equivalent of the Black Flysch nappe is the Baraolt nappe (Ştefănescu 1970). This nappe is exclusively made up of Berriasian to Aptian sandy-calcareous turbidites. The large Ceahlau nappe consists of basal Late Jurassic radiolarites, with basic igneous rocks and other deep-water deposits (Azuga Facies), which are overlain by mostly shaly and calcareous Tithonian to Neocomian (Sinaia beds) and Barremian to Aptian proximal turbidites (Comarnic Beds). Late Aptian to Albian massive sandstones and conglomerates (Bucegi conglomerate) unconformably cover the tectonic contact between Ceahlau nappe and overlying Getic nappe (e.g. Patrulius 1969; Ştefănescu 1976).

In the South Carpathians the Danubian nappes are overlain by Senonian "wildflysch", which constitutes a tectonic mélange complex (Cosustea mélange; Seghedi & Oaie 1997). This complex, which is characterized by a block-in-sheared-matrix structure, formed during early stages of tectonic accretion of these sediments to the overlying ophiolitic Severin nappe during the latest Cretaceous (presumably Maastrichtian). The basement of the overlying Ceahlau-Severin Ocean crops out in the Severin nappe and consists of strongly dismembered ophiolitic lithologies (Savu et al. 1985; Maruntiu 1987) such as harzburgitic ultramafics, gabbros and pillow basalts. The basalts show ocean-floor tholeiitic affinities (Cioflica et al. 1981). The overlying sediments are much thinner compared with units in the East Carpathians with which they are correlated and include Late Jurassic radiolarites (Azuga beds) followed by Early Cretaceous terrigenous turbidites (Sinaia and Comarnic Flysch) (see Codarcea 1940; Pop et al. 1997).

The tectonic units derived from the Ceahlau-Severin Ocean were thrust eastward in two stages (Ştefănescu 1976; Săndulescu et al. 1981a,b). During the earlier Aptian to Albian event (referred to as "Austrian" in the Romanian literature) they were accreted to the overlying Getic nappe related to west directed (in the East Carpathians and in present-day coordinates) subduction. The Baraolt and Black Flysch nappes of the Eastern Carpathians were emplaced over the more external Ceahlau nappe during the first stage, as dated by their Late Albian to Cenomanian post-tectonic cover (Săndulescu 1984). The second ("Laramide") event occurred in the latest Cretaceous, when the Ceahlau-Severin Ophiolitic Units were thrust above more external units: the most internal flysch units of the Moldavides (Convolute Flysch nappe) in the East Carpathians and the Danubian nappes in the South Carpathians (Plate 3, Profile 3). The age of this second event is constrained by a Late Campanian to Maastrichtian post-tectonic cover (e.g. Săndulescu 1984; Melinte & Jipa 2005).

The Ceahlau-Severin Ophiolitic Unit can be followed southwards into the boundary region between eastern Serbia and western Bulgaria based on the compilation by Kräutner & Krstić (2002, 2006). However, no traces of this ophiolitic unit (including the associated Sinaia-type turbidites) are present further south in the Carpatho-Balkan Orogen or further east in the Balkan Orogen. Hence, the oceanic Ceahlau-Severin rift, which represents a branch of the Alpine Tethys according to our interpretation, ended eastward within continental units of the European foreland and was not connected to any of the branches of Neotethys. This Ceahlau-Severin oceanic rift split off a narrow continental ribbon (Dacia), which later formed the Cretaceous-age Getic to Supragetic (or Bucovinian) nappe pile of the Carpathians described below (Fig. 2b). Because the Ceahlau-Severin Ocean ended to the east, the equivalents of this nappe pile in the Balkan Orogen (Central Balkan and Sredna Gora units; Ivanov 1988) are thrust above the Prebalkan Unit and the Moesian Platform without an apparent intermediate oceanic suture. Moreover main thrust contacts formed during the Cretaceous were reworked during Eocene times in the Balkan Orogen.

4.3.4. Getic-Supragetic (Bucovinian) nappe sequence

The east-facing Bucovinian nappe sequence of the East Carpathians developed during Early to mid-Cretaceous times (e.g. Popescu-Voitesti 1929; Kräutner 1938; Kräutner 1980) (see Plate 3, Profile 3). It represents the lateral equivalent of the Getic-Supragetic nappe sequence of the South Carpathians (Săndulescu 1984, 1994) (see Plate 2, Profile 4). This nappe sequence has the general geometry of a large antiform and consists of low- to medium-grade metamorphic rocks of Late Precambrian to Cambrian age (e.g. Balintoni & Gheuca 1974) separated from each other either by pre-Alpine thrusts or by Triassic to Lower Cretaceous sedimentary series (Balintoni 1981). The uppermost Alpine tectonic unit is the ***Bucovinian nappe*** whose Mesozoic cover series grade upward into Early Cretaceous wildflysch that lies structurally below the overriding ophiolite-bearing Transylvanian nappes (Patrulius et al. 1969; Ştefănescu 1976; Săndulescu 1984). A similar Paleozoic basement composition and Permian to Early Cretaceous sedimentary cover characterize the underlying ***Sub-Bucovinian*** **nappe**. The ***Infra-Bucovinian nappe***, the lowermost unit of the thrust succession, contains medium-grade metamorphic basement overlain by a Permian to Jurassic sedimentary cover, which was locally metamorphosed in Cretaceous times (Kräutner 1980; Gröger 2006; Dallmeyer et al. in press). This lowermost nappe crops out only in small isolated tectonic windows. Following Săndulescu (1984, 1994) we correlated the Infra-Bucovinian nappe of the East Carpathians with the ***Getic nappe*** of the South Carpathians, whereas the two higher units of the Bucovinian nappe sequence are correlated with the ***Supragetic*** nappes of the South Carpathians.

The Bucovinian nappe sequence of the East Carpathians and its lateral equivalent, the Getic and Supragetic nappes of the South Carpathians (Murgoci 1905; Streckeisen 1932), consist of Europe-derived continental crustal material (Dacia) that was separated from the European foreland along the Ceahlau-Severin oceanic rift. To the south, in Serbia and western Bulgaria, we also included the structurally highest unit, referred to as ***Serbo-Macedonian "Massif"*** (e.g. Dimitrijević 1957, 1997), into this nappe sequence. However we do not imply that this also applies to a unit that carries the same name in Greece and which experienced a severe Alpine metamorphic overprint (Kilias et al. 1999).

The Getic-Supragetic nappe sequence (see Plate 2, Profile 4) involves a medium- to high-grade metamorphic Neoproterozoic to Early Paleozoic gneissic basement and sub-greenschist to epidote-amphibolite grade Paleozoic successions. These are unconformably overlain by Late Carboniferous to Permian continental clastics and Mesozoic strata (Iancu et al. 2005). The Mesozoic rocks contain Middle Triassic carbonate platform deposits followed by detrital Early Jurassic strata (Gresten facies). Locally sedimentation ended with Middle Jurassic radiolarites, but was elsewhere followed by Late Jurassic to Early Cretaceous pelagic series. Post-tectonic strata begin with Albian to Cenomanian Molasse-type deposits, proving that this nappe sequence essentially formed during the mid-Cretaceous ("Austrian") orogenic pulse. However, Late Cretaceous series were locally also affected by latest Cretaceous "Laramide" deformations (Săndulescu 1984, 1994).

We mapped the Getic nappe across the Danube into Serbia mainly based on the 1 : 100'000 Geological Map of former Yugoslavia (Osnovna Geološka Karta SFRJ) and the work of Kräutner & Krstić (2002, 2006). In Serbia the Getic nappe, referred to as Kučaj-Ljubaš Zone by Kräutner & Krstić (2002, 2006), is locally also known by many different names (Dimitrijević 1997; Karamata 2006). We also include a structurally higher and more internal nappe digitation (Saska-Gornjak Unit) as part of the Getic nappe in Serbia (see Plate 3, Profile 5).

Further south, in the border area between Serbia and Bulgaria, additional and tectonically higher nappes, referred to as ***Kraishte units*** by Kräutner & Krstić (2002, 2006) have been mapped as part of what appears to be a Getic nappe "system" in Serbia and Bulgaria, rather than only one single Getic nappe. These Getic nappes and the overlying Struma unit now form part of the Osogovo-Lisets metamorphic core complex and were exhumed in Late Eocene to Oligocene times below a SW-dipping detachment below a Supragetic unit, locally known as Morava Unit (Kounov et al. 2004).

In Bulgaria the Kraishte units overlie the more external ***Sredna Gora Unit***, also included as part of the Getic nappe system in Plate 1, together with the more external Luda Kamčija Zone of Ivanov (1988), identical with the East Balkan Unit of Georgiev et al. (2001). The Sredna Gora Unit extends eastward to the Black Sea and is well known for a superimposed thick Late Cretaceous volcano-sedimentary succession. This same succession and its associated magmatic rocks also extend westwards into eastern Serbia (Timok eruptive area of Dimitrijević 1997; Timok-Sofia Basin of Kräutner & Krstić 2002, 2006). This Late Cretaceous basin and its associated Turonian to Campanian (92–78 Ma) magmatic activity ("banatites") evolved in a back-arc setting with respect to the N- to NE-dipping Neotethys Ocean subduction zone (Fig. 2c) that lay farther to the south and southwest (i.e. Berza et al. 1998; Georgiev et al. 2001; Heinrich & Neubauer 2002; Neubauer & Heinrich 2003; von Quadt et al. 2005). Development of this continental back-arc basin is post-tectonic with respect to the Lower to mid-Cretaceous main deformation phase ("Austrian") of the Getic-Supragetic nappe succession (Kräutner & Krstić 2002, 2006; Osnovna Geološka Karta SFRJ). Post-tectonic sedimentation commenced in Bulgaria during the Cenomanian (Georgiev et al. 2001) and in eastern Serbia during the Albian (Dimitrijević 1997).

In the South Carpathians the Supragetic nappes consist of two alpine aged thrust sheets, that is, from bottom to top, the Locva and Bocşa units (Năstăseanu et al. 1991). These Supragetic nappes were traced across the Danube into Serbia based on Kräutner & Krstić (2002 & 2006) and the 1 : 100'000 Geological Map of former Yugoslavia. There, these units are known as, from bottom to top, Ranovac and Vlasina units. They consist predominantly of low to intermediate-grade Proterozoic and Paleozoic rocks (Ranovac-Vlasina terrane of Karamata 2006). These units extend into Western Bulgaria, where they are known as the "Morava Unit" and form the upper plate of the Osogovo-Lisets core complex (Kounov et al. 2004), that we assigned to the Getic nappe sequence. While some authors (e.g. Dimitrijević 1997) regard these Supragetic units as part of the Serbo-Macedonian Massif, others (i.e. Kräutner & Krstić 2002, 2006) restrict the term Serbo-Macedonian Massif to a strip of high-grade units, which follows the Eastern Vardar Ophiolitic Unit along the western edge of Dacia. We interpret the contact between the high-grade Serbo-Macedonian Massif and the overlying low-grade units ("high-grade basement" and "low-grade Paleozoic" in Plate 3, Profile 5) to be of pre-Mesozoic age. Hence, we consider both high- and low-grade parts of the Serbo-Macedonian Massif in Serbia as a part of the Supragetic nappe sequence. In Serbia, Triassic sediments (Malešević et al. 1978) in sub-greenschist facies locally cover the high-grade part of the Serbo-Macedonian Massif. This shows that high-grade metamorphism of the westernmost part of the Serbo-Macedonian Massif of Serbia is pre-Mesozoic. Note that, by contrast, the Serbo-Macedonian Massif of Greece exhibits Cretaceous-age amphibolite-grade metamorphism (i.e. Kilias et al. 1999).

We interpret the entire Getic-Supragetic nappe sequence, formed in mid-Cretaceous times, to assume an upper plate position during the latest Cretaceous and Early Cenozoic suturing with the Dinarides. The latter, including the W-Vardar and Sava Ophiolitic Units, essentially occupy a lower plate position (see Plate 3, Profile 5). Our reconnaissance fieldwork, using the map by Malešević et al. (1978), showed that the Eastern Vardar ophiolites are located in a structurally higher position with respect to the Serbo-Macedonian Massif. Hence, in contrast to the generally accepted opinion, this easternmost branch of the "Vardar ophiolites" that in map view is spatially linked to the South Apuseni and Transylvanian ophiolites (Plate 1), tectonically overlies the Getic-Supragetic nappe sequence. This, together with findings presented below, led to the assignment of the Biharia nappe system of the Apuseni Mountains to the Getic-Supragetic nappe sequence, as shown in Plate 1.

The ***Biharia*** nappes of the Apuseni Mountains occupy the highest structural position in the N- to NW-facing North Apuseni Orogen (Bleahu et al. 1981; Balintoni 1994). Consequently, they are commonly regarded as an integral part of Tisza (i.e. Csontos & Vörös 2004). However, when combining surface mapping in the Apuseni Mountains (i.e. Balintoni 1994) with our own interpretation of subsurface data from the Transylvanian Basin (see also Krézsek & Bally 2006) it becomes clear that both the Biharia nappe system as well as the Bucovinian (= Getic-Supragetic) nappe sequence structurally underlie the ophiolitic units of the South Apuseni ("Metaliferous") Mountains and the Transylvanian ophiolitic units. Both nappe sequences are parts of the same structural unit, the structurally highest unit of the East Carpathians (e.g. Săndulescu 1994). Hence we consider the Biharia nappe system as a part of the Dacia Mega-Unit.

Information available on the age of the tectonic contact between the South Apuseni ophiolites and the underlying Biharia nappe system is at first sight contradicting. In the Trascau Mountains the contact between island-arc-type volcanics that are part of the South Apuseni ophiolites and the underlying Biharia unit appears to be of Jurassic age. According to Săsăran (2006) and our reconnaissance work, the contact between these volcanics and the continental Biharia basement is sealed by Late Jurassic (Late Oxfordian-Early Kimmeridgian) to Early Cretaceous platform limestones and, thus, must have formed prior to their deposition. As these platform carbonates are not metamorphosed, the low-grade metamorphism of the Biharia nappes system must be Jurassic rather than Cretaceous in age, at least locally in the Trascau Mountains. In this context it is interesting to note that Jurassic metamorphism, pre-dating the deposition of Late Jurassic platform carbonates that have not been metamorphosed, is also known from the ophiolite-bearing Circum-Rhodope Belt of Northern Greece (Michard et al. 1994) that we correlate with the Eastern Vardar Ophiolitic Unit as mapped in Plate 1. We interpret this Late Jurassic deformation and metamorphism as related to the obduction of the Eastern Vardar Ophiolitic Unit onto parts of the Dacia Mega-Unit in latest Jurassic times (see Fig. 2 and later discussion). The final east-directed emplacement of the Transylvanian ophiolites (the eastern continuation of the South Apuseni ophiolites) over the Bucovinian nappes (the eastern equivalents of the Biharia nappe system), however, is younger and did not occur before mid-Cretaceous times.

Radiometric dating by Dallmeyer et al. (1999) and fission track studies by Schuller (2004) indicate a poly-metamorphic history for the Biharia nappe system as a whole, with ages falling into two groups, namely Jurassic (186–156 Ma) and Early to middle Cretaceous (124–111 Ma). Thus, a Middle to Late Jurassic tectonic and metamorphic event, followed by Early Cretaceous top-E nappe stacking within the Dacia Mega-Unit, need to be distinguished. The final emplacement of the nappes in the North Apuseni Mountains involving top-W to NW superposition of the Biharia, Codru and Bihor nappe systems (see Plate 3, Profile 3), did not occur before Turonian time. This is documented by the Late Turonian "Gosau" unconformity (e.g. Balintoni 1994; Schuller 2004). This youngest tectonic event was preceded by two earlier thrusting events of latest Jurassic and Early Cretaceous ages.

The W-ward continuation of the Biharia nappe system beneath the subsurface of the Pannonian Basin as shown in Plate 1 is based on subsurface data compiled by Bleahu et al. (1994), Tari et al. (1999) and Lelkes-Felvari et al. (1996, 2003, 2005). We excluded, however, the Upper Codru nappes from the Biharia nappe system since they form an integral part of the Tisza Mega-Unit. On the other hand, we included parts of the Algyö basement high into the Biharia nappe system. This is based on the finding of Lelkes-Felvari et al. (2005) that in boreholes drilled on the Algyö basement high of southern Hungary, units characterized by a high-grade Cretaceous metamorphism (Dorozsma Complex) occur in the highest tectonic position.

4.3.5. Eastern Vardar Ophiolitic Unit (including South Apuseni Ophiolites and Transylvanian nappes)

The ***Eastern Vardar Ophiolitic Unit*** is part of the Carpatho-Balkan Orogen (or the Rhodopian fragment; Burchfiel 1980). The Sava Zone separates this ophiolitic unit from the Western Vardar ophiolites of the Dinarides (Plate 1). Hence the unspecified term "Vardar" ought to be abandoned. In Plate 1 we trace the Eastern Vardar ophiolites of Macedonia and eastern Serbia in the subsurface of the southernmost Pannonian Basin across Vojvodina into the ophiolites of the South Apuseni Mountains (Kemenci & Čanović 1997; Čanović & Kemenci 1999). As mentioned above, the Eastern Vardar ophiolites, and their equivalents in the Guevgeli ophiolites and the Circum-Rhodope Belt of northern Greece (Kockel et al. 1971, 1977; Michard et al. 1994) were metamorphosed and tectonically emplaced eastward on the Serbo-Macedonian Massif during the Late Jurassic, probably by obduction. However, final emplacement of the Eastern Vardar Ophioltic Unit (including the South Apuseni and Transylvanian ophiolites) occurred during Early Cretaceous east-facing nappe stacking that primarily shaped the Dacia Mega-Unit (Romanian Carpathians and their continuation into the Carpatho-Balkan Orogen).

The Eastern Vardar Ophioltic Unit in Serbia ("Central Vardar Subzone" of Dimitrijević 1997, "Main Vardar Zone" of Karamata 2006) represents a piece of MORB-type oceanic lithosphere. Following their emplacement on the Serbo-Macedonian Massif these ophiolites were overstepped by Late Jurassic reef limestones (Karamata 2006) that are succeeded by a Cretaceous flysch ("Paraflysch"; Dimitrijević 1997). The Late Jurassic emplacement of the Eastern Vardar Ophiolitic Unit onto at least parts of the Biharia nappe system and its lateral equivalent, the Serbo-Macedonian Massif, is probably related to the obduction of the Eastern Vardar ophiolites onto parts of the Dacia Mega-Unit, i.e. the European margin. However, no metamorphic sole has yet been found at the base of the East Vardar ophiolites. The age of the Eastern Vardar ophiolites is constrained by modern radiometric dating in northern Greece only, where igneous rocks with ages between 164 Ma and 155 Ma are present (Guevgeli ophiolites; Anders et al. 2005). These ages are very similar to those reported from the Dinaridic and Western Vardar ophiolites (see later discussion concerning the Dinarides). However, the ages stem from granitoid igneous rocks that intrude the Guevgeli ophiolites and hence, are slightly younger. The presence of granitoids within the Guevgeli ophiolites indicates that the latter formed in an island arc or back-arc setting (Brown & Robertson 2004).

The ophiolite-bearing units of the ***South Apuseni*** Mountains are grouped together with ophiolitic nappes that occur beneath the Neogene sediments of the Transylvanian Basin and those that are exposed in the Transylvanian nappes of the East Carpathians. This group of ophiolitic nappes is also known as "Transylvanides" (Săndulescu 1984). Neither a metamorphic sole, nor associated Jurassic ophiolitic mélanges have so far been reported from this group of ophiolite-bearing nappes. Only parts of the South Apuseni ophiolites represent MORB-type oceanic lithosphere formed during the Middle Jurassic. Other parts represent Late Jurassic intra-oceanic island-arc volcanics (Savu et al. 1992; Bortolotti et al. 2002a; Nicolae & Saccani 2003; Bortolotti et al. 2004a). The latter include basalts, andesites, rhyolites, as well as some granites. Similar to the Guevgeli ophiolites in Northern Greece these igneous rocks are only slightly younger than the surrounding MORB-type ophiolites, which they intrude. They are particularly widespread in the Trascău Mountains south of Cluj. Similar to the Eastern Vardar Ophiolitic Unit of Serbia, the South Apuseni ophiolites are also unconformably overlain by Late Jurassic to Early Cretaceous reef limestones (Bortolotti et al. 2002a; Săsăran 2006). As mentioned above, the Late Oxfordian to Early Kimmeridgian base of these reef limestones also oversteps the contact between the Apuseni ophiolites and the underlying basement of the Baia de Aries nappe, which forms part of the Biharia nappe system (Săsăran 2006 and our own observations). This suggests that the South Apuseni ophiolites were obducted over the Biharia nappe system before Late Oxfordian times.

The Late Jurassic to Early Cretaceous platform carbonates give way to Barremian to Aptian flysch deposits (Feneş Formation) and Aptian to Albian wildflysch (Meteş Formation; Bleahu et al. 1981; Săndulescu 1984; Suciu-Krausz et al. 2006). The wildflysch of the Meteş Formation formed during the second and main (pre-Cenomanian or "Austrian") deformation event that affected the South Apuseni Mountains. It is during this event that the South Apuseni ophiolites, including their prolongation in the subsurface of the Transylvanian Basin (Plate 3, Profile 3), were finally thrust over the Bucovinian nappe sequence of the Dacia Mega-Unit. Late Campanian to Maastrichtian ("Laramide") top-ESE thrusting (high-angle pop-up in the Trascau Mountains; see Profile 3, Plate 3) reflects a third compressional event that affected the South Apuseni and Transylvanian ophiolites including their Jurassic to Late Cretaceous sedimentary cover. "Laramide" thrusting is also reported from the Mures valley area where the most frontal tectonic slices of the South Apuseni units were imbricated towards the N to NW (Cris and Grosi nappes; Balintoni 1994) and thrust onto the Gosau-type sedimentary cover of the eastern parts of the Biharia nappe system (see Profile 4, Plate 2).

According to the interpretation given in Profile 3 of Plate 3 the ***Transylvanian Ophiolites***, which are preserved beneath the Late Cretaceous to Cenozoic sedimentary fill of the Transylvanian Basin, represent the direct continuation of the South Apuseni ophiolites. This is based on the interpretation of numerous borehole, gravity, magnetic and seismic data (e.g. Săndulescu & Visarion 1977; de Broucker et al. 1998), that document their presence as well as that of the overlying latest Jurassic to Early Cretaceous platform carbonates beneath the western parts of the Transylvanian Basin. In the western and central-northern part of this basin, our own assessment of borehole data suggests a similar situation as in the South Apuseni Mountains, namely that Late Jurassic to Early Cretaceous reef limestones rest directly on the ophiolites. According to our interpretation (Profile 3, Plate 3) these ophiolites were eroded in the eastern parts of the Transylvanian Basin during an exhumation event that pre-dates the deposition of the Late Cretaceous post-tectonic cover, which in many parts of the westernmost Transylvanian Basin rests directly on the basement of the Bucovinian nappe. This Late Cretaceous erosional event is genetically coupled with Late Cretaceous normal faulting (see normal fault east of the Tarnave Basin depocenter in Profile 3, Plate 3). Such normal faulting led to substantial exhumation of the previously metamorphosed Bucovinian nappe sequence (Dallmeyer et al. in press) along the northern margin of the Transylvanian Basin (Rodna horst; Gröger 2006; Gröger et al. in press). There, zircon fission track data yield evidence for rapid and substantial cooling from temperatures in excess of 300 °C during Coniacian to Campanian times.

These data support the interpretation that the Bucovinian nappe stack was originally covered by a substantial overburden and was exhumed in the footwall of the Tarnave normal fault. Seismic interpretation and wells in the hanging-wall of this large west-dipping normal fault demonstrate the presence of (half-) grabens. These grabens contain Late Cretaceous sediments, referred to as the "Uppermost Cretaceous rift mega-sequence" by Krézsek and Bally (2006). Our own reflection-seismic interpretations indicate that these series clearly pre-date syn-tectonic deposits that are associated with subsequent "Laramide" thrusting. Further to the west, development of these extensional structures is reflected in the South Apuseni Mountains by the gradual subsidence and deepening of a sedimentary basin in which the Bozes and Remeti flysch accumulated during the Late Cretaceous (Balintoni et al. 1984; Schuller 2004), that is, between the Austrian and Laramide tectonic events. Overall, the onset of extension-induced subsidence becomes gradually younger westward and highlights the importance of this Late Cretaceous (post-"Austrian", pre-"Laramide") extensional event. Note that juxtaposition of the South Apuseni ophiolites and the Getic-Supragetic nappe sequence across the South Transylvanian fault in the Mures Valley (Plate 2, Profile 4) largely results from the offsets across this large E–W oriented (in present-day coordinates) transfer fault produced during the Late Cretaceous extensional event, as evident in its eastern segment that is buried beneath the younger sedimentary fill of the Transylvanian Basin. During the Late Campanian to Maastrichtian "Laramide" event, the Tarnave and South Transylvanian faults were inverted, involving top-ESE thrusting and transpression, respectively.

The Transylvanian ophiolites crop out again as isolated tectonic klippen assigned to the so-called Transylvanian nappes that form the structurally highest unit of the East Carpathians. These nappes consist of Triassic to Albian sedimentary rocks, that were partly derived from continental margin units on the one hand, and from truly ophiolitic units on the other hand (Săndulescu 1974; Săndulescu & Russo-Săndulescu 1979; Săndulescu et al. 1981a; Săndulescu 1984). Modern geochemical work on these ophiolites by Hoeck & Ionescu (2006) evidence the occurrence of different types of magmatic rocks, namely true MORB and back-arc type ophiolites, but also a few andesites believed to have formed in a continental magmatic arc setting. The age of these ophiolites is still uncertain, although the above cited authors favour a Triassic age. The facies of the Triassic sediments occurring in some of the Transylvanian nappes (i.e. siliceous limestones, Hallstatt-type limestones; Săndulescu 1975) suggests that the Transylvanian nappes as a whole were presumably derived from a continent-ocean transition of a Triassic-age Meliata-Maliac-type ocean. Note, however, that the structural position of the Transylvanian nappes, which according to our interpretation is similar to that of the Jurassic-age ophiolites occurring beneath the Transylvanian Basin, indicates that both Triassic and Jurassic ophiolites form part of one and the same oceanic Meliata-Maliac-Vardar Neotethys that is referred to by Săndulescu (1994) as the "Main Tethyan Suture Zone".

Suturing of the Transylvanian nappes to Dacia (Bucovinian nappes) occurred during mid-Cretaceous times and is documented by the sealing of the nappe contacts by an Albian- to Cenomanian-age post tectonic cover. However, top-E compression had commenced already during the Late Barremian and was accompanied by the deposition of wildflysch, now occurring beneath the Transylvanian nappes (Săndulescu 1975). As shown in Profile 3 of Plate 3, we interpret the North Apuseni units (Tisza) to represent the upper plate with respect to this mid-Cretaceous suture. Note however, that the Transylvanian and South Apuseni ophiolites, that tectonically overlie the North Apuseni units, appear to be rootless in a present-day cross section view. This is the reason for depicting a large backfold and -thrust, respectively, shown schematically and conceptually in Profile 3 of Plate 3, very similar to an interpretation that was advanced much earlier by Săndulescu & Visarion (1977). This structure developed presumably during a Turonian compressional event that post-dated the top-E mid-Cretaceous emplacement of the Transylvanian ophiolites but which is not documented at all within Dacia. During this Turonian event the NW-facing North-Apuseni nappe stack developed, that is so characteristic for the Tisza Mega-Unit, as will be described in the next chapter.

4.4. Tisza Mega-Unit of the southern Pannonian Basin and the N-Apuseni Mountains

4.4.1. Overview

The Tisza Mega-Unit (Haas & Pero 2004; Csontos & Vörös 2004) comprises a sequence of nappes consisting of basement and its Mesozoic cover with some common characteristics. According to our interpretation this Mega-Unit is surrounded by mobile zones, most if not all of them probably representing oceanic sutures. It was recognized early on that the faunal assemblages ("European" faunal province) contained in the Triassic and Early Jurassic sediments of units exposed in the Mecsek Mountains, located near the northern rim of this Mega-Unit, strongly contrast with those found in the northerly adjacent series that are exposed in the Transdanubian ranges ("Mediterranean" faunal province) which form part of ALCAPA (Vörös 1977, 1993). Moreover, paleomagnetic evidence indicates contrasting rotations and translations for Tisza and ALCAPA. Rotations during Cretaceous to Miocene times are generally counter clockwise in ALCAPA but clockwise in the eastern parts of Tisza and adjacent Dacia (e.g. Patrascu et al. 1994; Panaiotu 1999; Márton 2000, 2001). The western parts of Tisza, however, were highly mobile, particularly during Tertiary times, and show small block rotations in opposing directions (Márton 2000).

The pre-Triassic basement of the Tisza Mega-Unit consists of various Variscan high-grade metamorphic series including anatectic granites and migmatites (Kovács et al. 2000). Recently Klötzli et al. (2004) provided evidence that the Late Paleozoic granitoids of the Mecsek Mountains likely formed at a location S or SSW of the Rastenberg granodiorite of the Bohemian Massif. This basement is therefore considered to represent a former part of the Moldanubian Zone.

The nappe succession that forms the Tisza Mega-Unit comprises, from external (NW) to internal (SE) the Mecsek, Bihor and Codru nappe systems (see Plate 2, Profile 2 & Plate 3, Profile 3). In general, thrusting occurred during the Turonian; it locally affects Early Turonian sediments and pre-dates the latest Turonian onset of deposition of the post-tectonic Late Cretaceous Gosau sediments (Balintoni et al. 1984; Schuller 2004). The Triassic cover sequences show substantial facies variations (Burchfiel & Bleahu 1976; Bleahu et al. 1981, 1994; Kovács et al. 2000; Haas & Pero 2004). A Germanic Muschelkalk-type facies of the most external units grades into more massive Mid-Triassic carbonate build-ups that are overlain by a Late Triassic Keuper facies (Haas & Pero 2004), and finally into the Schreyeralm-Hallstatt-type facies of the most internal and structurally highest units such as the Vascău nappe of the Codru nappe system (Bleahu et al. 1994). Hence, as pointed out by Burchfiel & Bleahu (1976), Triassic facies changes and nappe emplacement in the North Apuseni Mountains are reminiscent of what is known from the Central and Inner West Carpathians. A dramatic change occurred during the Bathonian, as documented in the Mecsek Unit by an increase in depositional water depths that can be related to the separation of the Tisza from the European continent, presumably in connection with the opening of the Alpine Tethys. Oxfordian radiolarites, followed by Rosso Ammonitico and Calpionella Limestone, combined with the faunal characteristics, indicate that Tisza had now become part of the Adriatic paleogeographic realm. Paleomagnetic data (Márton 2000) indicate, however, that rotations and translations of the Tisza micro-continent did not substantially depart from those of the European continent until about 130 Ma ago. This coincides with the Valanginian to Barremian volcanic activity of the Mecsek Mountains (Császár 1998). Interestingly, this event also approximately coincides with the opening of the Valais-Magura Ocean (e.g. Frisch 1979) and the onset of Early to mid-Cretaceous ("Austrian") crustal shortening in the neighbouring Dacia and ALCAPA Mega-Units, as described earlier.

4.4.2. Tectonic contacts between Tisza and neighbouring units, timing of nappe stacking within the Tisza Mega-Unit

The ***Mecsek*** nappe system consists, according to Haas & Pero (2004), of the Mecsek and Szolnok Units. Since exposures are restricted to the Mecsek Mountains, the internal structure of these units and their contacts with surrounding ones are poorly defined. The northern tectonic contact between the Mecsek nappe system and the Szolnok-Sava ophiolite-bearing belt of the Mid-Hungarian Fault Zone was mapped (Plate 1) according to Haas (2001) in the western sector, whilst in the area of Debrecen, we substantially departed from previous published compilations (e.g. Csontos & Vörös 2004). In this area the contact between the Mecsek nappe system and the Szolnok-Sava belt, as shown in Profile 2 (Plate 2), is based on a combination of borehole (Sáránd 1) and seismic data (Horvath & Rumpler 1984; Windhoffer et al. 2005). A Palaeogene thrust, overprinted by Miocene normal faulting, defines the contact between the previously stacked Mecsek and overlying Codru nappe systems with the Late Cretaceous Szolnok Flysch. The latter is considered as part of the Szolnok-Sava Zone ("ophiolite-bearing Intrapannonian Belt" of Channell et al. 1979) rather than simply the youngest cover of the Mecsek nappe system.

The ***Bihor nappe*** system (Villany-Bihor Zone or unit of Bleahu et al.1994 and Haas & Pero 2004, respectively) crops out in the Villany Mountains of southern Hungary, as well as in the North Apuseni Mountains of Romania. We also included units that crop out in the Papuk inselberg of Croatia (Pamić et al. 1996) as part of the Bihor nappe system. We mapped its northern contact with the Mecsek nappe system after Csontos & Vörös (2004). However, we excluded the basement that crops out in the Moslavačka Gora inselberg in Croatia, previously considered as the westernmost tip of Tisza, from the Mecsek nappe system, and attributed it to the Sava Zone owing to the occurrence of mid-Cretaceous gabbros (109 ± 8 Ma; Balen et al. 2003) and Late Cretaceous-age high-temperature metamorphism and magmatism (Starijaš et al. 2006). The southern contact between the Bihor and the Codru nappe system is rather poorly constrained by subsurface data, except for the area south of Debrecen, where S-dipping reflectors indicate a shear zone (Posgay et al. 2006, their Fig. 2) that we interpret to mark the tectonic boundary between these two nappe systems. In the North Apuseni Mountains, however, the tectonic contact between Bihor "Autochthon" (Bleahu et al. 1981) and the Codru nappe system is well exposed and closely dated. The Codru nappe system (Bleahu et al. 1981) was thrust NW-ward over Early Turonian, the youngest sediments of the Bihor nappe system, whilst Late Turonian to Early Coniacian Gosau beds seal the tectonic contact between these two units (Schuller 2004).

The ***Codru*** nappe system, as mapped in Plate 1, corresponds to the Békés-Lower Codru nappe system of Bleahu et al. (1994) and Haas & Pero (2004), but it includes also the Upper Codru nappes that these authors considered as part of the Biharia nappe system. The Codru nappe system mostly consists of a succession of exclusively sedimentary nappes, mostly preserved in the western parts of the North Apuseni Mountains (Codru Mountains). Only the structurally lowest unit, the Finis-Girda Unit (Profile 3, Plate 3), also contains pre-Triassic basement. This basement can be traced eastwards around a structural dome of the Bihor nappe system exposed in the Bihor Mountains. In contrast to the non-metamorphic Codru sediments, the pre-Triassic basement of the Girda Unit in the Bihor Mountains was mylonitized under greenschist facies conditions during Turonian nappe emplacement (Balintoni pers. comm. and our own observations). Stretching lineations and associated shear sense criteria in these mylonites indicate top-WNW nappe transport; kinematic indicators observed further W and within the sedimentary Codru nappe system indicate top-NW nappe transport.

Turonian-age top-NW thrusting also involved the structurally highest nappe system, the earlier described Biharia nappe system that we consider to form part of Dacia (see above). According to our interpretation (Profile 3, Plate 3) the Biharia nappe system, which was previously affected by mid-Cretaceous top-E nappe transport, was back-thrust towards the NW during the Turonian. Hence, it now forms the highest structural unit of the North Apuseni Mountains. This back-thrust is interpreted to be responsible for a substantial offset of the "Main Tethyan Suture Zone" (Săndulescu 1994), the suture zone between Tisza and Dacia that formed during the mid-Cretaceous orogeny. This interpretation is supported by the radiometrically constrained Jurassic (186–156 Ma) and Early Cretaceous (124–111 Ma) ages of metamorphism within the Biharia nappe system (Dallmeyer et al. 1999), which clearly pre-date Turonian thrusting.

In summary, and according to the interpretation presented above, the northern margin of Tisza, characterized by Germanic Muschelkalk-type facies, is preserved in the Mecsek Mountains and faced, during the Jurassic, a branch of the Alpine Tethys Ocean. The southern margin of the Tisza-block, however, faced the Triassic Meliata-Maliac Ocean that formed part of the Neotethys (Fig. 2b). Remnants of this Triassic passive continental margin are preserved in the highest sedimentary nappes (Vascau and Coltesti nappes) of the Codru nappe system and possibly in the Transylvanian nappes.

4.5. Dinarides

4.5.1. Overview

The tectonic units of the Dinarides were primarily compiled on the basis of the excellent geological 1 : 100'000 maps of former Yugoslavia (Osnovna Geološka Karta SFRJ). In some, but not all aspects we depart from schemes proposed by previous authors (i.e. Kossmat 1924; Petković 1961; Aubouin et al. 1970; Aubouin 1973; Dimitrijević 1982, 1997, 2001; Pamić et al. 2000; Hrvatović & Pamić 2005; Karamata 2006). The main difference with respect to previous compilations pertains to the distinction between a more external and a more internal ophiolitic belt, corresponding to the "Dinaridic and Mirdita-Pindos Oceanic Basin" and the "Vardar Zone Western Oceanic Basin" of Karamata (2006), respectively, as well as to their tectonic relationship with the continental Drina-Ivanjica and Jadar "blocks". Almost all previous workers considered these two ophiolite belts as representing two distinct oceanic realms, while the intervening continental blocks were regarded as terranes that separated them (e.g. Karamata 2006).

According to our interpretation, however, we have to deal with one single Jurassic ophiolite sheet in the Dinarides, namely the Western Vardar Ophiolitic Unit that was obducted onto the passive margin of Adria during latest Jurassic times (see Figs. 2 & 5). In this interpretation the Drina-Ivanjica and Jadar blocks simply represent tectonic windows below a single ophiolitic thrust sheet in which the most distal paleogeographic domains of Adria are exposed. Furthermore, we suggest that in map view the present-day separations between ophiolitic and continent-derived tectonic units resulted from Cretaceous to Cenozoic out-of-sequence thrusting, severely modifying the originally rather simple geometry that resulted from obduction (Plate 3, Profile 5). This view is not new and was already proposed by Bernoulli & Laubscher (1972) and Baumgartner (1985) for the southward adjacent Hellenides. Similarly, Smith & Spray (1984), Bortolotti et al. (2004b) and Bortolotti & Principi (2005) more recently emphasised the similarities between the ophiolitic belts in the Dinarides and Hellenides.

4.5.2. External Dinaridic Platform: Dalmatian Zone, Budva-Cukali Zone and High Karst Unit

The proximal parts of the Adriatic margin are characterized from the Triassic to Cenozoic times by carbonate platforms that are interspersed by intervening narrow deep-water basins (e.g. Bernoulli et al. 1990; Bernoulli 2001). One of these deep-water basins that are floored by crust of uncertain composition (thinned continental or oceanic) is the Pindos-Olonos Zone of Greece (e.g. Aubouin 1959) that is referred to as the Krasta-Cukali Zone in Albania (Aubouin & Ndojaj 1964; Robertson & Shallo 2000) and as the Budva Zone in Montenegro (Goričan 1994; Dimitrijević 1997).

The ***Budva-Cukali Zone*** of the Dinarides (Nopcsa 1921; Dimitrijević 1997), the sedimentary record of which starts with Triassic deep-water facies and ends with Cenozoic flysch, separates the relatively more external carbonate platform of the ***Dalmatian Zone*** from that of the more internal ***High Karst Unit*** (Aubouin et al. 1970). However, the Budva-Cukali Zone wedges out to the north between Kotor and Dubrovnik (Plate 1). Further north the Dalmatian Zone and High Karst Unit can still be separated by a thrust between the Dalmatian Zone and the High Karst Unit (Cadet 1970) that can be traced northwestward into the area of Split (Blanchet 1970) where it runs offshore, interfering with the N–S striking dextral Split-Karlovac transpressive zone (Chorowicz 1970, 1975, Fig. 1). According to the 1 : 100'000 Geological Map of former Yugoslavia (Osnovna Geološka Karta SFRJ) thrusting occurred exclusively during the Late Eocene to Early Oligocene (so-called Dinaridic phase of the Southern Alps), as documented by the accumulation of flysch-type sediments in an evolving flexural foreland basin from the Middle to the Late Eocene (e.g. Tari 2002). However, De Capoa et al. (1995), also found Miocene faunas along this flysch belt and hence provided evidence for ongoing Late Miocene thrusting of the High Karst Unit onto the Dalmatian Zone. Also the Split-Karlovac transpressive zone was active later since it affected Miocene-age strata in intramontane basins. Merlini et al. (2002) report ongoing thrusting in the subsurface of the foreland of the Dinarides until Quaternary times for the Trieste area. All this indicates that shortening continued during Late Miocene to recent (?) times, as clearly documented in Albania (Carminati et al. 2004), and off-shore Montenegro within the Dalmatian Zone (e.g. Picha 2002).

Although many authors claim that the Pindos Trough of the Pindos-Olonos Zone was underlain by oceanic lithosphere (e.g. Stampfli & Borel 2004) no ophiolitic basement flooring this deep-water through is exposed anywhere. Moreover, it is clear that the so-called Pindos Ophiolite Group of the Pindos Mountains in Greece, including its metamorphic sole and underlying, presumably Jurassic-age mélange formation (identical with the typical "Diabase-Radiolarite Formation" of the Dinarides), tectonically overlie the Cenozoic Pindos-Flysch, that is, the youngest sediments of the Pindos Zone (Aubouin 1959, Bornovas & Rondogianni-Tsiambaou 1983). This, and the fact that the Budva Zone has no northward continuation in Dalmatia, strongly argues in favour of the "one-ocean thesis" that was first formulated by Bernoulli & Laubscher (1972). We propose that the so-called, and in our opinion misnamed "Pindos ophiolites", form indeed the southern continuation of the Dinaridic-Mirdita ophiolite belt, but that these far-travelled ophiolite sheets have a more internal paleogeographic origin with respect to the Pindos Zone. Thus, the term "Mirdita-Pindos Ocean" is totally misleading and ought to be abandoned (see also Burchfiel 1980 who reached the same conclusion).

In Albania, the thin-skinned Miocene thrust belt of the Ionian Zone (Frasheri et al. 1996; Robertson & Shallo 2000; Carminati et al. 2004) is characterized by deeper marine slope and basin facies (Bernoulli 2001). It is external with respect to the Dalmatian Zone but does not reach the area of our map and strikes north-westward into the Adriatic Sea. The carbonate platform of the Dalmatian Zone is the northern equivalent of the more internal Kruja Zone of Albania (Aubouin & Ndojaj 1964; Robertson & Shallo 2000) and the Gavrovo-Tripolitza Zone of Greece (Jacobshagen 1986), which also crops out in the Olympos window of the Pelagonian Zone (Aubouin 1973). Moreover, in northeastern Albania, the Kruja (= Dalmatian) and Cukali (= Budva) zones, including Eocene-Oligocene sediments, are exposed in a window below the Korab (= Pelagonian) Massif and overlying Western Vardar ophiolites south of the area mapped in Plate 1 (Bortolotti et al. 2005). The carbonate platform of the High Karst Unit of Dalmatia, on the other hand, probably wedges out in Albania. In Greece, we interpret platform carbonates that are exposed in the Parnassos window of the Pelagonian Zone (Aubouin 1973; Jacobshagen 1986) as probable southern equivalents of the High Karst Unit. The existence of all these windows was the rational for extrapolating at depth the external Dinaridic platform units over considerable distance (about 120 km) eastward beneath a Cenozoic thrust at the base of the more internal Dinaridic units (Pre-Karst and Bosnian Flysch Unit and East Bosnian – Durmitor thrust sheet) and the Drina-Ivanjica (= Pelagonian) thrust sheet, as shown in Profile 5 of Plate 3.

4.5.3. Internal Dinaridic Platform: Pre-Karst and Bosnian Flysch Unit

The term ***Pre-Karst Unit***, introduced by Aubouin et al. (1970) and co-workers (Blanchet 1970; Cadet 1970; Charvet 1970), refers to a paleogeographic realm that was thought to be transitional between the carbonate platforms of the High-Karst Unit and a second and more internal tectono-stratigraphic unit, the so-called "Zone Bosniaque" of Aubouin et al. (1970), that is characterized by Late Jurassic to Cretaceous flysch. This flysch zone we refer to as ***Bosnian Flysch*** ("Flysch Bosniaque") largely corresponds to the so-called "Sarajevo sigmoid" of Dimitrijević (1997). According to our own observations, however, the Bosnian Flysch does not represent a different first-order tectonic entity, and hence we interpret the Pre-Karst and Bosnian Flysch units as part of a single unit, the ***Pre-Karst and Bosnian Flysch Unit***. In western Bosnia, large parts of the Bosnian Flysch simply represent the latest Jurassic and younger sedimentary sequences of the Pre-Karst Unit, typically found in the more internal parts of one and the same tectonic unit. The Bosnian Flysch is, however, in tectonic contact with the overlying units, mostly the East Bosnian-Durmitor thrust sheet except in the NE where the latter wedges out (see Plate 1).

The Paleozoic of the Pre-Karst Unit, which exhibits no or only low-grade Variscan metamorphism, crops out in the Bosnian Schist Mountains and in the Sana-Una Paleozoic series of Bosnia and adjacent Croatia (Hrvatović 2000a; Hrvatović 2005; Hrvatović & Pamić 2005). Radiometric ages indicate Cretaceous (121–92 Ma) and Cenozoic (59–35 Ma) low-grade metamorphic overprints, respectively (Pamić et al. 2004). The external parts of the Pre-Karst Unit are characterized by Jurassic-Cretaceous transitional platform-slope facies. From Mid-Jurassic times onward, breccias were shed from the High Karst paleogeographic domain NE-ward towards the more distal parts of the Adria passive continental margin that was drowned during the Early Jurassic. Departures from the facies of the external Dinaridic platform occurred, however, locally already during Mid-Triassic times, as evidenced by the deposition of red nodular Rosso Ammonitico type limestones (the so-called "Han-Bulog" facies, first described by Hauer 1853). This facies is very widespread in the more distal parts of the Adriatic margin (Aubouin et al. 1970). In the more internal parts of the Pre-Karst and Bosnian Flysch Unit, pelagic sedimentation began during Early Jurassic times and well before the deposition of the Bosnian Flysch. The boundary between the Pre-Karst and Bosnian Flysch Unit and the High Karst Unit, as shown in Plate 1, in the SE follows essentially the one drawn by Aubouin et al. (1970) and Dimitrijević (1997). Further to the NE we based the mapping of the thrust contact between the Pre-Karst and Bosnian Flysch and the High Karst units in Plate 1 on the 1 : 100'000 Geological Map of former Yugoslavia (Osnovna Geološka Karta SFRJ), this thrust being mostly accompanied by a narrow stripe of Cenozoic strata. Hence, we also included, for example, units such as the Mid-Bosnian Schist Mountains and the Raduša Unit (Hrvatović 2000a; Hrvatović 2005; Hrvatović & Pamić 2005), or the Sana-Una Paleozoic and its Mesozoic cover (Blanchet 1970; Pamić & Jurković 2002), in our combined Pre-Karst and Bosnian Flysch Unit.

The Bosnian Flysch s. str. ("Flysch Bosniaque" of Blanchet 1966 and Aubouin et al. 1970) comprises latest Jurassic (Tithonian) to Cenozoic flysch-type deposits that vary in depositional age, paleotectonic environment and source area both along and across strike. In the most internal zones of the Pre-Karst and Bosnian Flysch Unit, deposition of flysch started during the Late Jurassic but becomes progressively younger in the more external parts where it commenced variably during the Senonian, Maastrichtian or even the Palaeogene. Aubouin et al. (1970) recognised that the basal parts of the Bosnian Flysch represented synorogenic deposits with respect to what these authors referred to as the development of the Paleo-Dinarides. Other authors invoked, however, a passive margin scenario (distal Adriatic continental margin) to explain these flysch deposits (Pamić et al. 2000). According to our observations the latter model cannot be supported, at least for those parts of the Bosnian Flysch, which contain abundant ophiolitic detritus, such as the Late Jurassic to Berriasian age Vranduk Flysch that is exposed along the river Bosna north of Sarajevo (Blanchet 1966; Olujić 1978; Hrvatović 2000b). This flysch, which contains also radiolaritic pelagic intervals, formed at the leading edge of the Western Vardar Ophiolitic Unit during its obduction onto the Adria passive margin, the most distal parts of which were obviously converted into an active margin during the Late Jurassic.

The younger parts of the Bosnian Flysch in the Bosna Valley section, referred to as Ugar Flysch (Hrvatović 2000b) represent, however, a younger and entirely different type of flysch basin. According to our observations this Turonian to Senonian flysch rests unconformably either on previously deformed Vranduk Flysch or on Jurassic to Early Cretaceous strata of the Pre-Karst Unit. Hence it formed in response to a pre-Turonian orogenic event and, in this sense, represents a kind of Gosau Basin. Clasts contained in this flysch include large olistostoliths consisting of carbonates derived from the External Dinaridic platform. In Western Bosnia and Montenegro the base of similar flysch deposits is also Senonian in age (Mirković et al. 1972). Deposition of the so-called Durmitor Flysch in Montenegro (Dimitrijević 1997), however, commenced even later during Maastrichtian time and lasted into the Palaeogene. Accumulation of this flysch can be related to Late Cretaceous to Early Cenozoic top-SW thrusting and emplacement of the structurally next higher tectonic unit, the East Bosnian-Durmitor thrust sheet.

4.5.4. East Bosnian – Durmitor thrust sheet

The ***East Bosnian-Durmitor*** thrust sheet, as shown in Plate 1, comprises the unit of the same name defined by Dimitrijević

(1997), but includes also more internal tectonic units such as the Lim Paleozoic and overlying Triassic-Jurassic strata. The Triassic-Early Jurassic carbonate platform, which constitutes a large part of this unit, was drowned in Early Jurassic time and sunk below the CCD in the Middle Jurassic (Bajocian; Đerić pers. comm.) when the deposition of radiolarites started (e.g. Rampnoux 1970). The radiolarites give upward way to a Late Jurassic ophiolitic tectonic mélange ("Diabase-Radiolarite Formation") that was deposited during the obduction of the directly overlying West Vardar Ophiolite Units. Hence the East-Bosnian-Durmitor thrust sheet is a composite tectonic unit; it consists in its basal part of Paleozoic and Mesozoic formations detached from the Adriatic passive margin in Cenozoic times, and in its upper part of the Western Vardar Ocean ophiolites that were obducted during the Late Jurassic to earliest Cretaceous. An out-of-sequence thrust, that post-dates the Late Jurassic obduction of the Western Vardar Ophiolitic Unit over the East-Bosnian-Durmitor Unit, is also found in the hanging wall of the East Bosnian-Durmitor composite thrust sheet. It juxtaposes the more external East Bosnian-Durmitor composite thrust sheet against a more internal structural unit, the Drina-Ivanjica Unit, which is also a composite tectonic unit (Plate 3, Profile 5). According to Dimitrijević (1997) the East Bosnian-Durmitor thrust sheet represents a far-travelled (>45 km) thrust sheet that corresponds to the "Zone Serbe" of Rampnoux (1970) and Aubouin et al. (1970). Note that on Plate 1 only the Paleozoic-Jurassic series, which underlie the ophiolitic mélange, were mapped as the East Bosnian-Durmitor thrust sheet, while the previously obducted ophiolites and associated mélanges were mapped as part of the Western Vardar Ophiolitic Unit; the same holds for the more internal Drina-Ivanjica and Jadar-Kopaonik composite thrust sheets, respectively.

The NW-SE striking external thrust contact between the East Bosnian-Durmitor thrust sheet and the Bosnian Flysch is strongly deflected towards N–S in the Sarajevo area (hence the term "Sarajevo sigmoid" for the underlying and more external Pre-Karst and Bosnian Flysch Unit). N of Sarajevo, the Paleozoic to Triassic formations of the East Bosnian-Durmitor thrust sheet laterally wedge out towards the NW. We regard, however, the so-called "Radiolarite Formation" (Pamić 2000), that is exposed in the Bosna River section and extends northwestward into the Banja Luka area (Plate 1), as forming part of the East Bosnian-Durmitor thrust sheet as well. In the Banja Luka area this Radiolarite Formation consists exclusively of late Middle Jurassic to earliest Cretaceous radiolarites (Vishnevskaya & Đerić 2005; Đerić pers. comm.) that are tectonically sandwiched between Vranduk Flysch in the footwall and an ophiolitic mélange at the base of the Western Vardar Ophiolitic Unit in the hanging wall.

The East Bosnian Durmitor thrust sheet can locally be subdivided into second-order tectonic units, such as the more external Durmitor sub-unit of Montenegro and the more internal Lim sub-unit of adjacent southern Serbia. These are involved in a number of domal anticlines that are upheld by low-grade Paleozoic sediments (i.e. Foča and Lim Paleozoic). The Triassic (i.e. Rampnoux 1970) is either characterized by thick platform carbonates, such as those found in the Durmitor sub-unit, or by more distal slope or basinal facies typical for the more internal Lim sub-unit, which are occasionally characterized by siliceous thin-bedded limestones (Ladinian-Carnian age Grivska Formation defined by Dimitrijević & Dimitrijević 1991). This slope facies is very reminiscent of the so-called Pötschenkalk-facies of the Eastern Alps (e.g. Gawlick 2000) or a coeval deep-water limestone in Central Greece Adhami Limestone; Baumgartner 1985) and is also found in the more internal Drina-Ivanjica and Jadar-Kopaonik thrust sheets. Mid-Triassic rift-related volcanism is widespread throughout this unit (Pamić 1984). Very thick (>100 m) Jurassic radiolarite successions (Rampnoux 1970) are typical for the more internal Lim sub-unit in southern Serbia (Zlatar area) and resemble those of the late Middle Jurassic to earliest Cretaceous (Vishnevskaya & Đerić 2005) "Radiolarite Formation" of the Bosna Valley section (Pamić 2000) much further to the NW, which are also part of the East Bosnian-Durmitor thrust sheet and were deposited prior to the Late Jurassic to Early Cretaceous obduction of the Western Vardar ophiolites onto the distal Adriatic margin. Note that most previous authors (e.g. Dimitrijević 1997; Pamić 2000) erroneously interpreted these radiolarites, which according to our observations tectonically underlie the obducted Western Vardar Ophiolitic Unit, as part of the ophiolitic succession.

4.5.5. Drina-Ivanjica thrust sheet

The ***Drina-Ivanjica*** thrust sheet contains even more distal parts of the Adriatic passive margin. It was probably emplaced in Early to mid-Cretaceous times on top of the East Bosnian – Durmitor thrust sheet and, similar to the East Bosnian-Durmitor composite thrust sheet, also carried passively the previously obducted Western Vardar ophiolites (i.e. the Zlatibor ophiolites; see Plate 3, Profile 5). The front of this thrust sheet was mapped by compiling the 1 : 100'000 sheets of the Geological Map of former Yugoslavia (Osnovna Geološka Karta SFRJ). NE of Sarajevo (Devetak area), Triassic to Middle Jurassic strata form mainly the frontal parts of this thrust sheet. SW-wards, in the Zlatibor area, the sole of the thrust front locally ramps up into the previously obducted ophiolites. This caused a duplication of the ophiolitic units since the footwall of the "youngest" rocks of the East Bosnian – Durmitor nappe also consists of ophiolites (see Plate 3, Profile 5).

The more internal parts of the Drina-Ivanjica thrust sheet consist largely of low-grade metamorphic Paleozoic formations (Milovanović 1984) that form the basement of the Drina-Ivanjica Mesozoic. The facies of the Mesozoic strata shares many similarities with that of the East-Bosnian-Durmitor thrust sheet (Dimitrijević & Dimitrijević 1991). Local occurrences of red nodular Late Anisian limestones (Han-Bulog facies; Sudar 1986), siliceous thin-bedded Ladinian to Carnian limestones (Grivska Formation) indicate that these

Mesozoic rocks also were deposited on a relatively distal part of the Adriatic passive margin, which started to form in Triassic times (Rampnoux 1970; Dimitrijević & Dimitrijević 1991). Thick sequences of Jurassic radiolarites are also typical for this thrust sheet. In places, the drowning below the CCD started rather abruptly in Aalenian times, radiolarite deposition persisting until Callovian times (Đerić et al. 2007) and being followed by ophiolite emplacement. Above a major unconformity, Cenomanian to Maastrichtian sediments commence with shallow water clastics and rudist limestones, which grade upward into flysch (Kosovska Mitrovica Flysch; Dimitrijević & Dimitrijević 1976, 1987). These series progressively unconformably cover, from west to east, ultramafics and mélange of the Western Vardar Ophiolitic Unit, Mesozoic and finally Paleozoic strata of the Drina-Ivanjica thrust sheet (Plate 3, Profile 5). This sedimentary cover is post-tectonic and post-metamorphic with respect to the Late Jurassic to earliest Cretaceous obduction of the Western Vardar Ophiolitic Unit and to Early to mid-Cretaceous deformations.

We correlate the base of the Late Cretaceous sedimentary cover of the Drina-Ivanjica thrust sheet with the base of the Ugar Flysch that occurs in the Bosna Valley section and that is also post-tectonic with respect to Cretaceous compressional deformation affecting the Vranduk Flysch. Since neither Late Cretaceous nor Cenozoic flysch is present below the frontal thrust of the Drina-Ivanjica thrust sheet, we propose that thrusting of this unit also occurred during the Early to mid-Cretaceous deformation of the internal Dinarides. It is this thrust event, combined with subsequent erosional denudation, which we hold responsible for the development of the unconformity at the base of this more than 200 km long strip of Late Cretaceous Kosovska Mitrovica Flysch (Plate 3, Profile 5). This flysch extends from western Serbia southward into the Kosovo (Dimitrijević 1997) in the immediate footwall of the next higher and yet more internal unit: the Jadar-Kopaonik thrust sheet.

In summary, this interpretation strongly deviates from the view expressed by practically all previous authors who regarded the Drina-Ivanjica Unit as a continental terrane that was originally located between two separate oceanic basins (e.g. Dimitrijević & Dimitrijević 1973; Robertson & Karamata 1994; Dimitrijević 2001; Karamata 2006) or who postulated that this element was derived by out-of-sequence thrusting from the European margin (Pamić et al. 1998; Hrvatović & Pamić 2005). However, as the tectonic position of the Drina-Ivanjica thrust sheet is identical to that of the ***Korab*** element in Albania, referred to as ***Pelagonides*** in Macedonia and Greece, our re-interpretation has consequences for the entire Dinaridic-Hellenic Orogen. Based on our interpretation of the cross section through the Dinarides we challenge the postulate of most authors (i.e. Robertson & Karamata 1994; Stampfli et al. 2004) according to which in the area of the future Hellenides the Pelagonian micro-continent was originally located between a more external "Pindos" or "Sub-Pelagonian" Ocean and a more internal "Maliac" or "Vardar" Ocean.

4.5.6. Jadar-Kopaonik thrust sheet

The present-day tectonic contact between the Drina-Ivanjica thrust sheet and the Jadar-Kopaonik thrust sheet, derived from the most distal Adriatic passive margin, is very steep (Plate 3, Profile 5) and has a strong dextral strike-slip component (Gerzina & Csontos 2003). In the literature this contact is referred to as "Zvornik suture" (Dimitrijević 1997) that is supposed to mark the ophiolitic suture between the continental Drina-Ivanjica and Jadar block terranes (Karamata 2006). In our view also this innermost thrust sheet passively carries previously obducted ophiolites of the Western Vardar Ocean. Hence, according to our interpretation this Zvornik "suture" simply represents the northwestern continuation of the long belt of Senonian flysch, that marks the tectonic boundary between the Drina-Ivanjica and Jadar-Kopaonik thrust sheets. The Jadar-Kopaonik thrust sheet represents the third and innermost thrust sheet that was activated after the Late Jurassic obduction of the Western Vardar Ophiolitic Unit. Also elsewhere Late Cretaceous ("Senonian") flysch unconformably overlies the Jadar-Kopaonik thrust sheet, the largest occurrence being present SSW of Belgrade in the Ljig area.

The so-called ***Jadar*** "block" consists of a non-metamorphic Paleozoic basement that is covered by Permian Bellerophon limestones, which are followed by a Triassic succession that is similar to the one of the Drina-Ivanjica thrust sheet (Dimitrijević 1997). Moreover, Filipović et al. (2003) recognized very strong similarities between these rocks and those in the ***Bükk*** Unit of Northern Hungary. This correlation provides one of the major arguments to consider the Bükk Mountains as a displaced fragment of the internal Dinarides (Kovács et al. 2000, 2004; Dimitrijević et al. 2003). Geographically, the Medvednica Mountain and neighbouring inselbergs near Zagreb occupy an intermediate position. Tomljenović (2000, 2002) showed that a low-grade metamorphic complex, partly consisting of Mesozoic rocks and partly of Paleozoic series, was covered by an ophiolitic mélange during a first deformation event that possibly equates with the Late Jurassic ophiolite obduction of the Western Vardar Ophiolitic Unit (see also Pamić & Tomljenović 1998; Pamić et al. 2002a). The low-grade metamorphism of the Paleozoic and Mesozoic rocks is of Aptian (115–123 Ma) age (Belak et al. 1995); moreover, a blueschist event of unknown age (Belak & Tibljaš 1998) has also been reported from this area.

In southern Serbia two other complexes crop out, which according to our interpretation belong to the most distal Adriatic margin, and are surrounded by the Western Vardar Ophiolitic Unit: a window in the ***Kopaonik*** area (Fig. 1) and a smaller westerly adjacent window below the Western Vardar ophiolites referred to as Studenica slice (Dimitrijević 1997). According to our investigations both these windows expose low-grade metamorphosed Paleozoic to Mesozoic rocks. Karamata (2006) considered the Kopaonik block as yet another continental terrane, which he traced northward into the Belgrade area. We regard, however, also the rocks in the Kopaonik and Studenica window, together with the Jadar block and more southerly occurrences found in the Kosovo, as having been derived from the distal Adriatic passive margin. Southwest of Belgrade, the tectonic contact between the Drina-Ivanjica and Jadar-Kopaonik thrust sheets turns in map view NW-ward, owing to some differential rotation between the southeastern and northwestern Dinarides; this is one of the reasons for not extending the Kopaonik ridge into the Belgrade area in Plate 1.

According to a cross-section published by Grubić et al. (1995) and our observations, the Kopaonik window is a late-stage antiform structure that folded the previously obducted Western Vardar Ophiolitic Unit preserved on both sides of this structure (see also Sudar & Kovács 2006). On the eastern flank of this antiform, which was later intruded by an Oligocene granodiorite, occurs a strip of Senonian flysch. This flysch rests unconformably either on the Western Vardar ophiolites or on the underlying continental units of the Kopaonik window and is over-thrust by the Eastern Vardar Ophiolitic Unit that was earlier obducted onto the Serbo-Macedonian Massif (Profile 5, Plate 3). We regard the corresponding E-dipping thrust fault as a first order tectonic contact that marks the suture between the Dinarides in a lower-plate position and the Carpatho-Balkan orogen in an upper-plate position. Towards the hangingwall this Senonian flysch contains huge olistoliths, as well as abundant detrital material that was eroded from the overriding Eastern Vardar Ophiolitic Unit, thus dating the closure of this flysch basin as Late Senonian. We interpret this strip of Senonian flysch, which here marks a suture, as the southerly extension of the Sava Zone (Plate 3, Profile 5).

The continental units underlying the Western Vardar ophiolites in the Kopaonik area, as well as in the W-ward adjacent Studenica area, largely consist of Triassic carbonates, often transformed into marbles, and associated meta-basites (Zelić et al. 2005). Sudar (1986) determined Carnian and Norian ages for some of the carbonates occurring in the Kopaonik area and in other areas of the adjacent Kosovo (metamorphic Trepča series), proving their Mesozoic age. They stratigraphically rest on Early Triassic Werfen beds and Paleozoic schists. Sudar & Kovács (2006) estimated the temperature of this metamorphism to be slightly higher than 400 °C based on the colour alteration index established for conodonts and on microstructural observations. The exact age of this metamorphism, which pre-dates the unconformity at the base of the Senonian flysch, is still unknown (Jurassic or Early to mid Cretaceous?).

4.5.7. Ophiolites obducted onto the Adria margin: the Western Vardar Ophiolitic Unit

In former Yugoslavia the mapping of the Western Vardar ophiolites shown on Plate 1 is based on the 1 : 100'000 geological maps of former Yugoslavia (Osnovna Geološka Karta SFRJ) whilst for Albania we used the 1 : 200'000 map of Albania (Geological Map of Albania 2002) and an overview presented by Robertson & Shallo (2000). We mapped all ophiolitic occurrences west of the Sava belt as the same unit that is here collectively referred to as Western Vardar Ophiolitic Unit. Also in Albania (Mirdita ophiolites) a traditional distinction between different ophiolitic belts cannot be made on structural grounds, although geochemical and petrological differences have been documented (see below). A high-grade metamorphic sole typically underlies the obducted ultramafic massifs. We also included the ophiolitic mélange underneath the metamorphic sole, as well as the post-obduction overlapping younger strata into this same tectonic unit on Plate 1.

In the territory of former Yugoslavia two belts of largely isolated ultramafic massifs forming klippen overlying the ophiolitic mélange (Diabase-Radiolarite Formation) are evident in map view (Plate 1). These belts are separated by a structural culmination of the continental Drina-Ivanjica thrust sheet. The more external belt, referred to by most authors as the Dinaridic ophiolites (or Dinaridic ophiolite belt; Pamić et al. 2002a; Karamata 2006), is also known as the Central Dinaridic ophiolite belt (Lugović et al. 1991). Most of the ophiolitic klippen of this external belt overly the external parts of Drina-Ivanjica thrust sheet and a few the internal parts of the East-Bosnian-Durmitor thrust sheet in Serbia and Montenegro. The more internal belt is referred to as Western Vardar ophiolites by Karamata (2006), but also referred to under a variety of names such as Inner Dinaridic ophiolite belt (Lugović et al. 1991), External Vardar Subzone (Dimitrijević 1997, 2001) or simply Vardar Zone (Pamić et al. 2002a). We propose that the collective term "Vardar" should not be used any longer, as the Eastern Vardar Ophiolitic Unit is part of the Carpatho-Balkan Orogen (or the Rhodopian fragment; Burchfiel 1980) and structurally separated from the Western Vardar Ophiolitic Unit by the Sava Zone (see below). Note that we excluded ophiolitic occurrences in the Fruška Gora Mountain NW of Beograd (Dimitrijević 1997) and in the northern parts of the Kozara Mountains in Bosnia (Karamata et al. 2000a) from the Western Vardar Ophiolitic Unit as we consider them as forming part of the more internal Sava Zone (Plate 1).

The SW ophiolite belt ("Dinaridic" in the sense of Karamata 2006) of the Western Vardar Ophiolitic Unit is dominated by fertile mantle rocks (lherzolites), though harzburgites occur locally as well. Extrusive rocks are relatively rare (Lugović et al. 1991; Trubelja et al. 1995). Depleted mantle rocks (harzburgites) predominate in the NE ophiolite belt ("Western Vardar" in the sense of Karamata 2006) of the Western Vardar Ophiolitic Unit that contains abundant extrusive rocks with a supra-subduction geochemical signature (i.e. Karamata et al. 1980; Spray et al. 1984; Lugović et al. 1991). Most authors concluded that the ophiolites of the western belt are predominantly of the MORB-type, whilst those of the eastern belt represent supra-subduction oceanic island-arc type ophiolites.

In northern Albania both belts merge into one larger thrust sheet, the ***Mirdita*** ophiolitic thrust sheet (e.g. Bortolotti et al. 1996, 2005; Gawlick et al. 2007) that was thrusted over the Budva-Cukali Zone into a relatively more external structural position southwest of an important transverse zone known as Skutari-Pec line (Plate 1). This ophiolite unit, and other smaller klippen in Central Albania, show a general trend from dominantly lherzolitic types in the west towards harzburgitic types in the east (e.g. Bortolotti et al. 2002b). Most authors working in Albania, however, express doubts that these two types of ophiolites represent different oceanic basins that were separated by a micro-continent. Some claim along-strike variations (e.g. Hoeck et al. 2002), whereas others infer that these ophiolites represent a single piece of oceanic lithosphere formed at a slow spreading ridge with variable amounts of amagmatic (west) vs. magmatic (east) spreading (Nicolas et al. 1999). Bortolotti et al. (2004b, 2005) also postulated a single oceanic basin, but invoked mature intra-oceanic NE-directed subduction to explain the supra-subduction-type ophiolites found in the eastern parts of the Albanian and Greek ophiolites. From a purely structural point of view (one sheet of obducted ophiolites), and owing to the widespread occurrence of metamorphic soles both in the western and eastern belt (see below), we subscribe to the latter interpretation.

The age of the ophiolitic ultramafic and mafic plutonic rocks of both belts is Middle to early Late Jurassic (163–148 Ma; Callovian to Early Tithonian), based on radiometric age determinations (Spray et al. 1984; Bazylev et al. 2006; Ustaszewski et al. submitted) and palaeontological data from the oldest overlying sediments (Late Bajocian to Early Oxfordian radiolarites, preserved in Albania only; Prela 1994; Chiari et al. 1994; Marcucci et al. 1994; Marcucci & Prela 1996). Interestingly, the range of ages obtained for of the metamorphic sole at the base of the obducted ophiolites appears to be slightly older (174–157 Ma, Aalenian-Oxfordian; Lanphere et al. 1975; Okrusch et al. 1978; Dimo-Lahitte et al. 2001). In Bosnia, Serbia and Albania a metamorphic sole is present almost everywhere at the base of the obducted ophiolites. Metamorphic conditions span a large range of temperatures from >300 to 850 °C while the range of the reported pressures (4–10 kbar) is relatively small (Carosi et al. 1996; Karamata et al. 2000b; Operta et al. 2003). Mafics (amphibolites) predominate, though interlayered metasediments, probably derived from the sedimentary cover of the lower plate during intra-oceanic subduction, occur as well (Karamata et al. 2000b; Bortolotti et al. 2005). Most authors agree that, given the inverted metamorphic field gradient and the large range of temperatures over only a few hundreds of meters, residual heat from the overlying young and still hot oceanic lithosphere represents the heat source for metamorphism affecting the protoliths of the lower plate rocks during the intra-oceanic obduction stage (Michard et al. 1991) of the future Western Vardar Ophiolitic Belt (Bortolotti et al. 2005). Shear sense indicators (Simpson & Schmid 1983) are preserved in the mylonitic metamorphic sole at Zlatibor Mountain (Dinaridic ophiolites of Southern Serbia, our observations), around the Bresovica locality in Kosovo (Karamata et al. 2000b and our observations), as well as in the Mirdita ophiolites of Albania (Carosi et al. 1996). The senses of shear consistently vary between top-W to top-NW. This indicates that intra-oceanic obduction was WNW-directed, that is oblique to nearly parallel to the present-day strike of the Dinaridic chain.

Given the Aalenian to Oxfordian ages for the formation of the metamorphic sole at a depth of 35 km (corresponding to 10 kbar), intra-oceanic subduction in the Meliata-Maliac-Vardar Neotethys Ocean must have started no later than at least some 5 Ma earlier (assuming a subduction rate of 1 cm per year and a 45° dipping subduction zone), namely at around 179 Ma ago (Toarcian). This suggests that older oceanic lithosphere (Early Jurassic and Triassic, see below) must have been present in this subducting lower plate (see Fig. 5).

The ophiolitic mélange, which occurs below the metamorphic sole flooring the obducted ophiolites (Diabase-Radiolarite Formation of former Yugoslavia, referred to as Rubik complex in Albania; e.g. Bortolotti et al. 2005), typically contains a mixture of (1) rock types derived from the lower plate, mechanically scraped off and accreted to the upper plate, and (2) gravitationally emplaced olistoliths derived from the upper plate. The blocks derived from the lower plate consist, amongst other lithologies, of Triassic ultramafics and mafics (MORB-type ophiolitic blocks up to several km in diameter) that were derived from the Meliata-Maliac-Vardar Ocean, the age of which was inferred from preserved stratigraphic contacts with Triassic radiolarites. Such Triassic ophiolites are found, for example, in the Darno-Complex adjacent to the Bükk Mountains (Dimitrijević et al. 2003), in the area around Zagreb (i.e. Halamić & Goričan 1995), in Serbia (Ovcar-Kablar gorge; N. Đerić pers. comm.) and in Albania (e.g. Kodra et al. 1993; Bortolotti et al. 1996; Gawlick et al. 2007). Amongst the blocks derived from the lower plate, Triassic strata derived from the adjacent Adria passive margin mafics predominate, however, over ophiolites. These Triassic strata consist of platform carbonates, slope to basinal facies such as Hallstatt limestone, cherty limestone, thin-bedded radiolarite – pelagic limestone successions or radiolarites that are of Late Anisian to Norian age (e.g. Chiari et al. 1996; Dimitrijević et al. 2003; Goričan et al. 1999, 2005; Bortolotti et al. 2005; Gawlick et al. 2007). The composition of the rocks derived from the lower plate documents that the oceanic Triassic and Early Jurassic crust of the Meliata-Maliac-Vardar Neotethys Ocean, which was attached to the Adriatic passive margin was completely subducted, except for the blocks preserved in the Diabase-Radiolarite Formation. Hence we interpret the ophiolitic mélange as an accretionary prism that formed during the late emplacement stages of the obducted upper plate ophiolites (Western Vardar Ophiolite Unit) that are Jurassic in age. The gravitationally emplaced olistoliths mostly comprise blocks derived from the metamorphic sole and the Western Vardar ophiolites of the upper plate, including their overlying Jurassic radiolarites (e.g. Bortolotti et al. 2005).

The matrix of the ophiolitic mélange contains palynomorphs that yields ages ranging from the Hettangian to the Late Bajocian (Babić et al. 2002) in the area near Zagreb; hence mélange formation must post-date the Late Bajocian. The final obduction stages of ophiolites and associated ophiolitic mélange are dated by strata, which overstep the previously emplaced Western Vardar Ophiolitic Unit. After a period of erosion, during which parts of the obducted ophiolites were removed, sedimen-

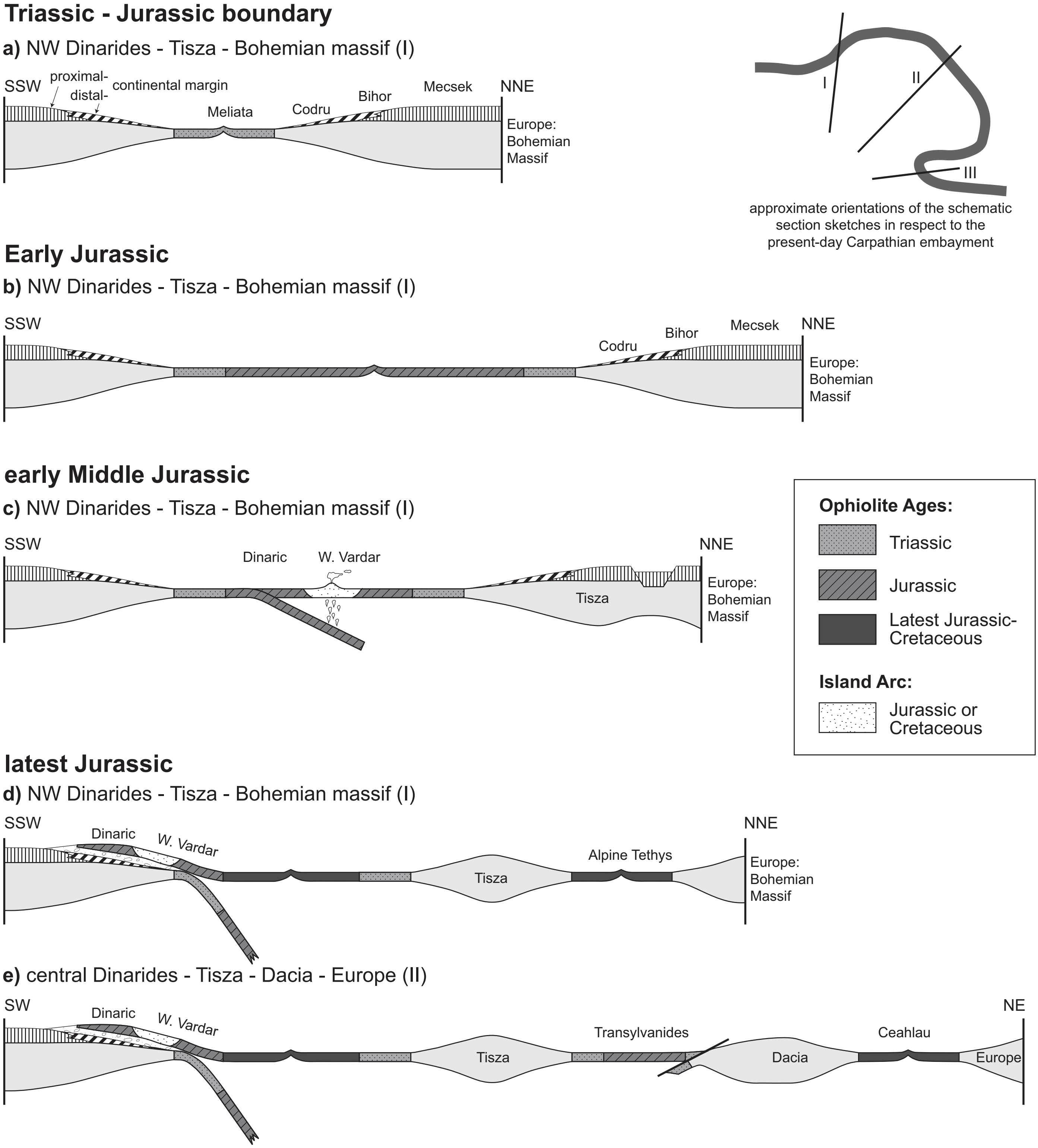

Fig. 5. Schematic serial sections depicting the plate tectonic evolution of the Alps-Carpathians-Dinarides system in pre-Cenozoic times. Given the complexity of the structures depicted in Plate 1, the sections were drawn in 3 different directions (see inset for location of the sections in present-day map view): a, b, c, d along section I; e, f, h along section II; g, i along section III. Note that important out-of-section translations of some of the units must have occurred (see Fig. 2). Furthermore, associated vertical axis rotations cannot be appropriately depicted in the sections alone. See text for detailed discussion.

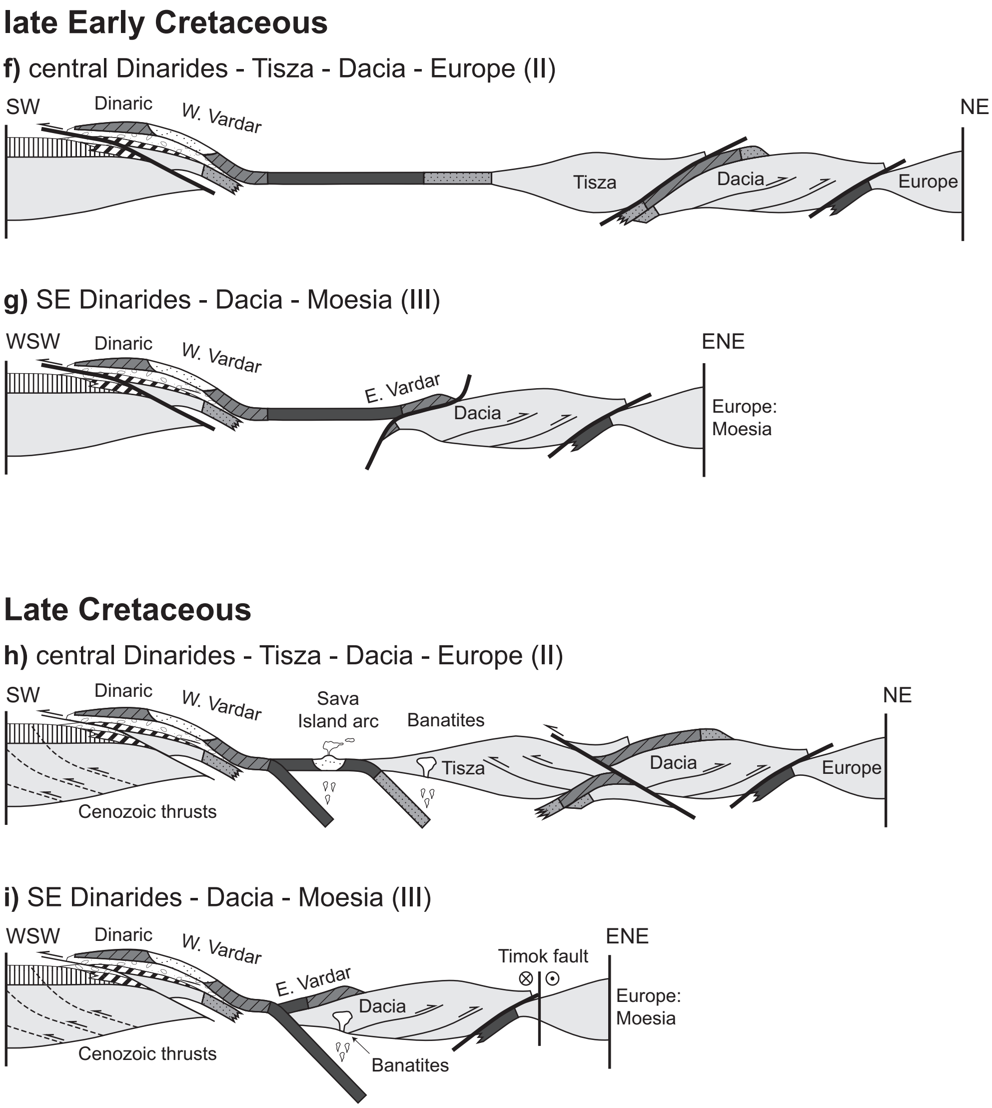

Fig. 5. (Continued).

tation locally resumed (e.g. Bosna River section) with fluvial conglomerates and sandstones (Pogari Formation; Blanchet et al. 1970; Pamić & Hrvatović 2000; Neubauer et al. 2003). These Tithonian to Berriasian conglomerates contain reworked ophiolitic as well as continental crustal material, including Late Permian granites (Neubauer et al. 2003). A similar over-step sequence that consists of mixed continental and ophiolitic detritus and ranges in age from Tithonian to Valanginian is known from Albania (Simoni mélange; Bortolotti et al. 2005). In former Yugoslavia platform carbonates and reefal build-ups of Late Jurassic (Kimmeridgian-Tithonian) and Early Cretaceous age are more widespread. These overlie erosionally truncated ophiolites and interfinger laterally with the Pogari conglomerates (Pamić & Hrvatović 2000). These overstepping formations indicate that the final stages of obduction occurred during Late Jurassic times. Hence, final obduction onto the Adriatic margin closely followed, within some 10 Ma, intra-oceanic subduction. However, since the age range of radiolaria in the underlying radiolarites of the continental East Bosnian Durmitor thrust sheet extends into the Berriasian and Valanginian (Đerić & Vishnevskaya 2006; Đerić pers. comm.) this obduction possibly did not come to a halt before Early Cretaceous times. Our kinematic analysis in the Zlatibor area (Rudine locality), performed on the weakly metamorphosed matrix of ophiolitic mélange immediately underlying the metamorphic sole, indicated top-WNW thrusting. Thus, also the final stages of ophiolite obduction involved transport oblique to sub-parallel to the present-day strike of the Dinarides.

After the Late Jurassic to Early Cretaceous period of alluvial to neritic sedimentation, the overstep basin was involved in Early to mid-Cretaceous folding and thrusting, subjected to erosion and subsequently unconformably covered by the Senonian series which locally rest on the non-ophiolitic parts of the Drina-Ivanjica and Jadar-Kopaonik thrust sheets. These series grade upward into latest Cretaceous flysch that immediately predates the Palaeogene development of the major tectonic contacts between the different units of the Inner Dinarides.

4.5.8. Sava Zone: the Cenozoic suture between Dinarides and Tisza-Dacia

According to our interpretation, the ***Sava Zone*** marks the position of a suture between the upper plate Tisza and the Dacia Mega-Units to the NE and SE, respectively, and the lower plate internal Dinarides (Fig. 5 and Plate 3, Profile 5). Originally this belt of ophiolitic, magmatic and metamorphic rocks that extends from Zagreb to Belgrade was referred to as "North-western Vardar Zone" (Pamić 1993) or "Sava-Vardar Zone" (Pamić 2002a). It was interpreted as a Late Cretaceous to Early Palaeogene volcanic (back-) arc basin that had remained open until the Mid-Eocene collision between the Dinarides and the Tisza Block by Pamić et al. (2002a). Since the Western and Eastern Vardar Ophiolitic Units were closed by obduction in Late Jurassic times (see above) we prefer not to use the term "Vardar" for the Sava Zone, which contains relics of a Cretaceous age back-arc ocean (Fig. 2c) and presently represents a Palaeogene-age suture zone. Southward from Belgrade we extend the Sava Zone as a narrow belt of Late Cretaceous ophiolite-bearing flysch (Senonian Flysch of Dimitrijević 1997). There the Sava Zone separates the Dinarides (including the Western Vardar Ophiolitic Unit) from the Carpatho-Balkan orogen (including the Eastern Vardar Ophiolitic Unit) (Plate 1; Plate 3, Profile 5).

Unfortunately, near Zagreb there are no exposures of the Sava Zone in the area of the junction of the Mid-Hungarian Fault Zone that includes the ophiolitic "Intra-Pannonian Belt" (Channell et al. 1979) and the westernmost parts of the ophiolite-bearing Sava belt of the Dinarides. Furthermore, the tectonic position of the Moslavačka Gora inselberg, characterized by Cretaceous-age gabbros (109 ± 8 Ma, Balen et al. 2003) and Late Cretaceous metamorphism and magmatism (Starijaš et al. 2006), is not very clear. Although traditionally attributed to the Tisza Mega-Unit, we interpreted the Moslavačka Gora inselberg as part of the Sava Zone, guided by the fact that areas with a strong Mesozoic metamorphic overprint are unknown from the adjacent internal Dinarides and the Tisza Unit.

The best exposures of the Sava Zone are present in northern Bosnia and Eastern Croatia (Kozara, Prosara, Motajica and Požeška inselbergs; Pamić 2002). In the northern part of the Kozara Mountains the southernmost and structurally lowest part of the Sava Zone is thrust SW-ward over the Western Vardar Ophiolitic Unit and its Cretaceous to Palaeogene post-obduction sedimentary cover. The structurally lowermost unit of the Sava Zone consists of Late Cretaceous ophiolites with bimodal volcanic suites (Karamata et al. 2000a; Ustaszewski et al. submitted). These younger ophiolites indicate that oceanic island arc-type crust (Sava Back-Arc Ocean in Fig. 2c) was generated during Late Cretaceous time within a remnant of the once much larger Neotethys Ocean. This confirms the view of Karamata et al. (2000a) that a remnant of the Vardar Ocean stayed open until Campanian times. Northward these Late Cretaceous ophiolites are covered by Maastrichtian to Eocene siliciclastic flysch (Cretaceous to Early Palaeogene flysch of Pamić et al. 2002a). Still further north, this flysch becomes progressively metamorphic (Pamić et al. 1992). This latest Cretaceous to Early Palaeogene metamorphism reaches lower amphibolite facies (Ustaszewski et al. 2007) and is followed during the Late Oligocene by the intrusion of S-type granites (Ustaszewski et al. 2006). This is in line with Pamić (1993, 2002) who provided evidence that the Sava Zone formed during the Palaeogene final collision of the Internal Dinarides with the Tisza Mega-Unit. The next and more easterly located inselberg occurs west of Belgrade in the Fruška Gora and exposes a very heterogeneous suite of blueschist metamorphic and non-metamorphic ophiolitic and non-ophiolitic rocks (Milovanović et al. 1995; Dimitrijević 1997). According to Milovanović et al. (1995) this metamorphism is Barremian in age (123 ± 5 Ma) and thus, documents an earlier stage of ongoing subduction of the Vardar Ocean.

South of Belgrade, a narrow strip of Senonian flysch represents the suture between the Dinarides and the Eastern Vardar Ophiolitic Unit that we consider as an integral part of the Carpatho-Balkan Orogen, and which, together with the Serbo-Macedonian "Massif", forms here the upper plate during end-Cretaceous-Palaeogene suturing (Plate 3, Profile 5). Further to the east, a window through the Serbo-Macedonian Massif, the Jastrebac window (Fig. 1; Grubić 1999; Kräutner & Krstić 2002), exposes low-grade metamorphosed Late Cretaceous to Palaeogene flysch that is overlain by greenschists, marbles and meta-pelites. These rocks, which we attribute to the Sava Zone (Plate 3, Profile 5), clearly demonstrate that the previously E-facing thrust succession of the Carpatho-Balkan Orogen formed the upper plate with respect to the Dinarides during their Palaeogene final suturing. Furthermore, mantle xenoliths occurring within Palaeogene magmatic suites in the Carpatho-Balkan Orogen of eastern Serbia are evidence for deep reaching subduction of the Dinarides lower plate, causing magmatism in the upper plate according to Cvetković et al. (2004)

5. Summary and outlook

The bewildering geometric complexity of the Alpine-Carpathian-Dinaridic orogenic system largely results from its long lasting deformation history and local changes in polarity of subduction and thrusting. This led to multiple overprinting of older deformations by younger ones. In Figure 6 we attempted to group the age of tectonic contacts within this orogenic system into six time slices. We are aware that such a subdivision is rather artificial, particularly for those parts of the system that underwent a progressive and rather continuous history of deformation. In contrast, in other parts of this orogenic system, the kinematics of deformation did change rather abruptly between the different deformation episodes. A second difficulty encountered during the construction of Figure 6 was that when along a given contact zone repeated tectonic activity had occurred we had to decide which deformation episode was the most important.

Late Jurassic deformations (170–150 Ma), associated with the obduction of parts of the Vardar Ocean, affected mainly the area of the future Dinarides (Figs. 5d,e), whereas the Eastern Alps were only marginally affected by this episode of deformation. The occurrences of ophiolitic mélange formations (Meliata unit) in the Western Carpathians show that the Jurassic Vardar Ocean was not only obducted onto the Adriatic southern margin of the Meliata-Maliac-Vardar Neotethys embayment in the domain of the Dinarides, but also less dramatically over its northern margin in the domain of the Western Carpathians. Note that in the Dinaridic domain this obduction occurred top-WNW with low-angle obliquity to the strike of the future orogen. On the other hand, the exact nature of the Late Jurassic, still rather enigmatic contact between the Eastern Vardar-South Apuseni and Transylvanian ophiolites with the Europe-derived Dacia units (Fig. 5e) is difficult to assess owing to a strong overprint by Early Cretaceous deformations.

Early Cretaceous orogenic processes strongly affected the Eastern Alps and the Dacia Mega-unit, but not the Tisza Mega-Unit. In the Eastern Alps, the Early Cretaceous orogeny was accompanied by the subduction of large volumes of continental crust along the western end of the Meliata embayment. This led to the development of an eclogitic subduction channel, which was subsequently exhumed. The onset of this orogeny around 135 Ma (Valanginian) is documented by the deposition of the Rossfeld Formation in the Northern Calcareous Alps (Faupl & Wagreich 2000). In the Europe-derived Dacia Mega-unit, this deformation episode controlled the formation of the presently E-facing succession of nappes, closure of the Ceahlau-Severin Ocean and their suturing to the European foreland (Fig. 5f). At the same time, the Eastern Vardar Ophiolitic Unit overrode the internal, western margin of the Dacia Mega-Unit. In this area, the Early Cretaceous orogeny ("Austrian" phase) commenced during Late Barremian time (130 Ma), as evidenced by the onset of wildflysch accumulation, and ended prior to the deposition of the post-tectonic cover, between Aptian and earliest Cenomanian (125–100 Ma) times, depending on the location. Shortening was immediately followed by extensional collapse of the orogenic edifice. The Tisza Mega-Unit was, however, not affected by Early Cretaceous orogenic activity. This, and the fact that the kinematics of deformation of Turonian-age thrusting and nappe formation in the Tisza Mega-Unit are drastically different, i.e. top NW rather than top-E (see Plate 3, Profile 3), led us separate this Early Cretaceous episode from a younger one that started in early Late Cretaceous times.

The second, early Late Cretaceous, orogenic episode peaked during the Turonian. Typically, a Late Turonian to Coniacian-age unconformity truncates structures that resulted from this orogenic episode and forms the base of the overstepping Gosau-type basins. In the area of the Tisza-Dacia Mega-Units, this second Cretaceous deformation episode can be rather clearly separated from the first one. Such a distinction is more problematic in the ALCAPA Mega-Unit of the Eastern Alps and the Western Carpathians. In these areas pre-Gosau deformation appears to have been rather continuous during the Cretaceous with the thrust front progressively migrating towards the European foreland. However, in both the Tisza-Dacia and ALCAPA Mega-Units Cretaceous-age metamorphism was followed by rapid exhumation (e.g. Thöni 2006), and the related Late Cretaceous extensional collapse that appears to have affected also the Dacia Mega-Unit. This indicates that in all areas considered the Early and early Late Cretaceous orogenic pulses were clearly separated from the Maastrichtian to Cenozoic orogenic phases.

In the Dinarides, Cretaceous orogenic events are currently not well dated. The most prominent angular unconformity occurs below the basal strata of the Turonian to "Senonian" overstep basins. For this reason we tentatively assigned the basal thrust of the Drina-Ivanjica thrust sheet to the early Late Cretaceous (100–85 Ma, Late Albian-Coniacian) compressional event (Figs. 5h,i). Rare radiometric ages so far available for the low-grade or blueschist metamorphism in the Dinarides available so far (Milovanović 1984; Belak et al. 1995; Milovanović et al. 1995; Pamić et al. 2004; Ilić et al. 2005) span, however, the entire 130–92 Ma time interval (Barremian-Cenomanian), suggesting that also in the Dinarides Cretaceous orogenic activity may have been rather continuous.

Compression resumed during the latest Cretaceous to Cenozoic orogenic cycle. A very short-lived orogeny affected during the latest Santonian and the Maastrichtian ("Laramide" phase) the East and South Carpathians, including the Transylvanian Basin. However, no Palaeogene compressional deformations have been recorded in the Tisza and Dacia Mega-Units. Indeed, some areas, such as the South Carpathians, underwent even extension during this period of time.

By contrast, the ALCAPA Mega-Unit and the Dinarides are characterized by a completely different Maastrichtian to Palaeogene deformation history. Palaeogene orogenic activity essentially shaped the present-day Alps, Western Carpathians and Dinarides and is thus recognized as the dominant deformation episode in this part of the Alpine-Carpathian-Dinaridic-orogenic system. It was associated with a large magnitude of N–S convergence between Adria and Europe, estimated to be about 600 km in case of the Alps by Schmid et al. (1996). Contemporaneous shortening of this magnitude did not occur in the Tisza and Dacia Mega-Units during this time period. Hence, it is obvious that crustal shortening in the Alps, which was associated with a differential N-ward displacement of the Adria plate, must have been accommodated in the Dinarides by very substantial dextral strike-slip movements. The steepness of the more internal thrusts of the Dinarides indeed is suggestive of a transpressive tectonic setting during their Palaeogene evolution that essentially terminated at the Eocene-Oligocene transition when the subducted Adriatic slab was detached from

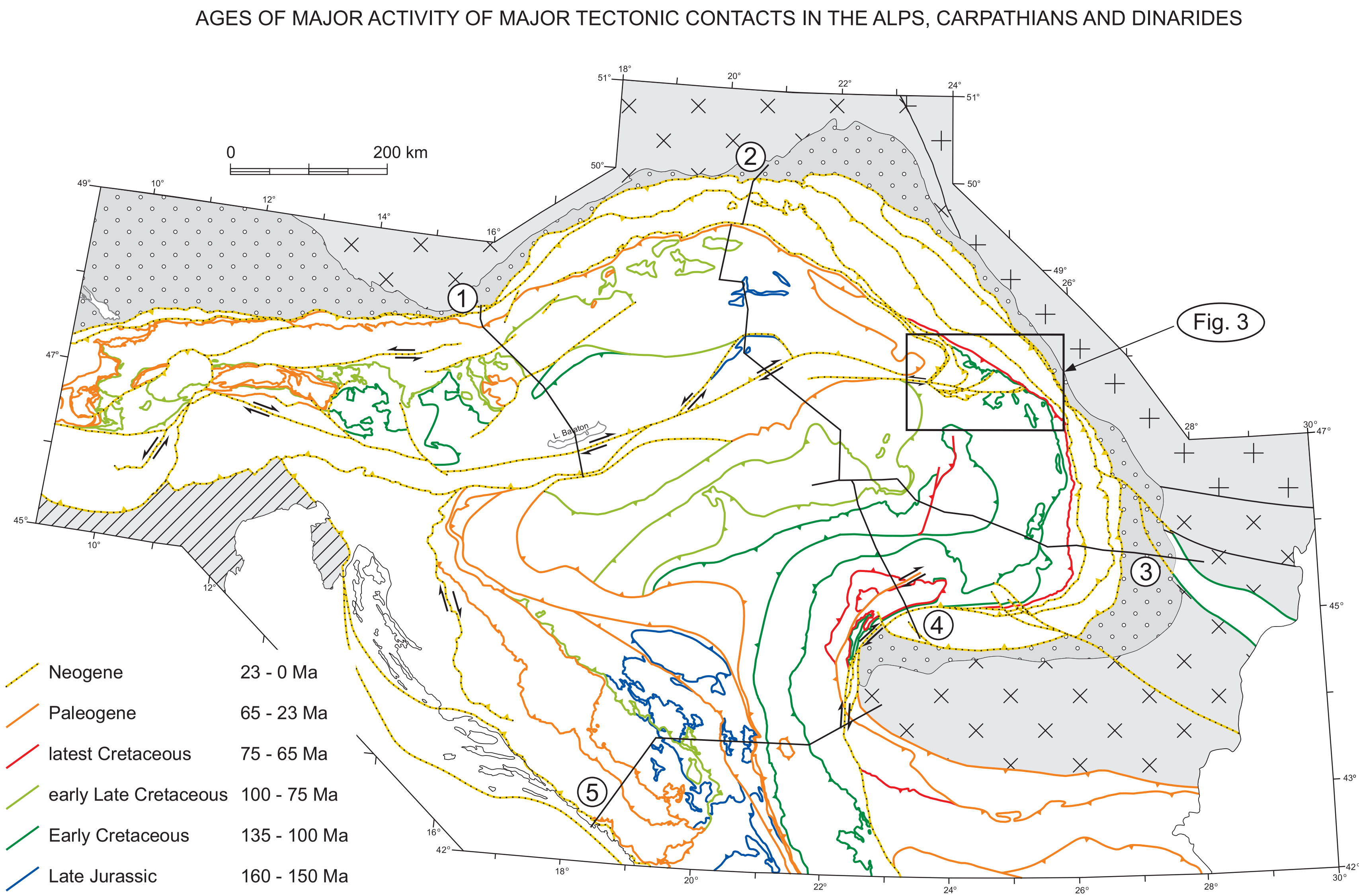

Fig. 6. Ages of main activity along major tectonic contacts in the Alps, Carpathians and Dinarides are colour-coded. Six time slices are depicted. Although some contacts were repeatedly active, only the age of the main deformation is shown. This figure also shows the locations of the traces of cross-sections given in Plates 2 and 3 and the area covered by Figure 3.

the lithosphere, giving rise to an Oligocene to Early Miocene high-K calc-alkaline and shoshonitic magmatism (Pamić et al. 2002b; Harangi et al., 2006). As already pointed out by Laubscher (1971) in his pioneering paper that addressed the Alps-Dinarides connection, a change in subduction polarity takes place between Alps and Dinarides. This change presumably occurs somewhere in the area now occupied by the Mid-Hungarian Fault Zone, i.e. the eastern continuation of the Periadriatic Line. However, this change in subduction polarity cannot be understood before the rather dramatic effects of Neogene deformation have been unravelled and palinspastically restored.

The Neogene tectonic setting of the Alpine-Carpathian-Dinaridic orogenic system is rather closely constrained. Continued crustal shortening in the Alps was accompanied by N-ward movement of the Adriatic plate, dextral wrench movements along the Periadriatic-Sava Zone of the Dinarides and E-ward lateral extrusion of the ALCAPA Mega-Unit along the Alpine Periadriatic-Mid-Hungarian Fault Zone. Retreat of the subducted European lithospheric slab beneath the inner Carpathians facilitated the advance of the ALCAPA and Tisza-Dacia Mega-Units into the Pannonian embayment and controlled the development of the back-arc-type Pannonian Basin (i.e. Horvath et al. 2006; Cloetingh et al. 2006). Simultaneously, the ALCAPA and Tisza Dacia Mega-Units invaded the still partly oceanic Carpathian embayment, involving substantial amounts of strike-slip faulting that interact with extension. Furthermore, the formation of this basin was kinematically linked to shortening in the external Miocene thrust belt of the Carpathians (Royden 1988).

Conclusions drawn from our correlations have serious paleogeographic implications. Alpine Tethys and Neotethys denote two separate groups of oceanic basins that opened during the break-up of Pangea (Fig. 2). The Neotethys opened during Triassic and Early to Mid-Jurassic times whilst the Alpine Tethys started to open during the Mid-Jurassic (Fig. 2; Figs. 5a-d). Remnants of the Triassic parts of the Neotethys, referred to as the Meliata Ocean, are preserved only in ophiolitic mélanges. Jurassic opening of the Alpine Tethys was largely contemporaneous with partial closure of the Neotethys oceanic lithosphere and the obduction of its Jurassic parts represented by the Eastern and Western Vardar Ophiolitic Units (Fig. 5d). Both, Triassic and Jurassic ophiolites formed part of one and the same branch of Neotethys, referred to as the Meliata-Maliac-Vardar Ocean (Fig. 2; Fig. 5b). By rooting the Transylvanian and South Apuseni ophiolites beneath the Tisza Mega-Unit, we postulate that a branch of the Jurassic Neotethys had separated the Tisza and Dacia Mega-Units. Erroneous interpretation of the complex geometries that resulted from out-of-sequence thrusting during the Cretaceous and Palaeogene deformations underlies a variety of multi-ocean concepts that were advanced in the literature. We propose that such models are incompatible with field evidence we have gathered and/or synthesized from the literature.

Moreover, different branches of the Alpine Tethys must have opened throughout the complex Alpine-Carpathian-Dinaride orogenic system. For instance, the oceanic Valais and Piemont-Liguria units, including their eastern counterparts, were apparently only partly separated from each other by such ribbon continents as the Iberia Briançonnais micro-continent (Fig. 2c) and narrow elongated continental fragments found within parts of the Pieniny Klippen Belt (too small to be depicted in Fig. 2c). In present-day map view, the strike of the Piemont-Liguria main branch of the Alpine Tethys changed sharply in the Maramures area (area of the Pienides, see Fig. 1). From there, after turning south and southwest, it followed the Mid-Hungarian fault zone to finally link up with the Sava Zone, the ophiolitic constituents of which represent a back-arc basin of the Meliata-Vardar Ocean. The Ceahlau-Severin Ocean of the Eastern Carpathians represents an eastern branch of the Alpine Tethys, which in contrast, terminated eastwards and did not reach the present-day Balkan Mountains of Bulgaria (Fig. 2b).

Therefore we conclude that (1) none of the branches of the Alpine Tethys and Neotethys can be followed E-ward into the North Dobrogean Orogen, although such connections have been proposed in the past (e.g. Stampfli & Borel 2004). Instead, in the area of the Sava Belt the main branch of the Alpine Tethys was connected with the Neotethyan Meliata-Maliac-Vardar Ocean, representing the only oceanic realm that can be traced via the Dinarides and Hellenides eastward into Turkey (Fig. 2c). (2) We propose that all ophiolitic remnants of Neotethys found in the area under consideration formed part of one and the same oceanic basin that started to open in Triassic times and continued to open during the Jurassic, though its closure commenced during the Middle Jurassic at the same time as the Alpine Tethys began to open.

It is hoped that the correlation of tectonic units presented here will provide a solid basis for future research into the kinematics and dynamics of the Alpine-Carpathian-Dinaridic system and to better-constrained sequential palinspastic reconstructions of the various deformation episodes. Such research and testing of our interpretations will finally lead to a better understanding specifically of the kinematics and dynamics of pre-Neogene orogenies, orogenies that, in contrast to the Neogene deformations, still remain rather enigmatic.

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MAJOR TECTONIC UNITS OF THE ALPS, CARPATHIANS AND DINARIDES

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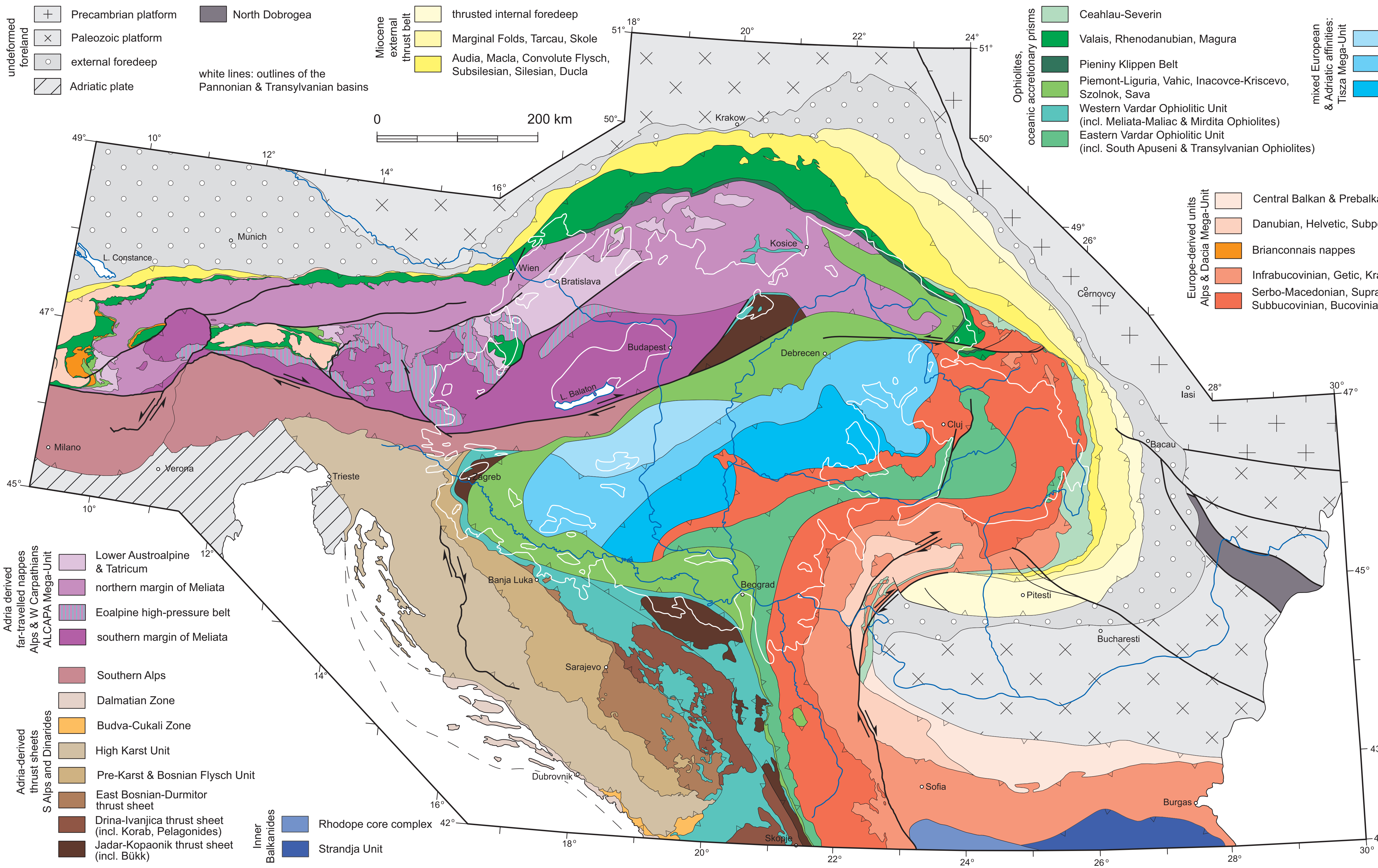


Plate 1: Major tectonic units of the Alps, Carpathians and Dinarides 1 : 5'000'000. Note that the locations of the cross-sections given in Fig. 3 and Plates 2 & 3 are found in Fig. 6.

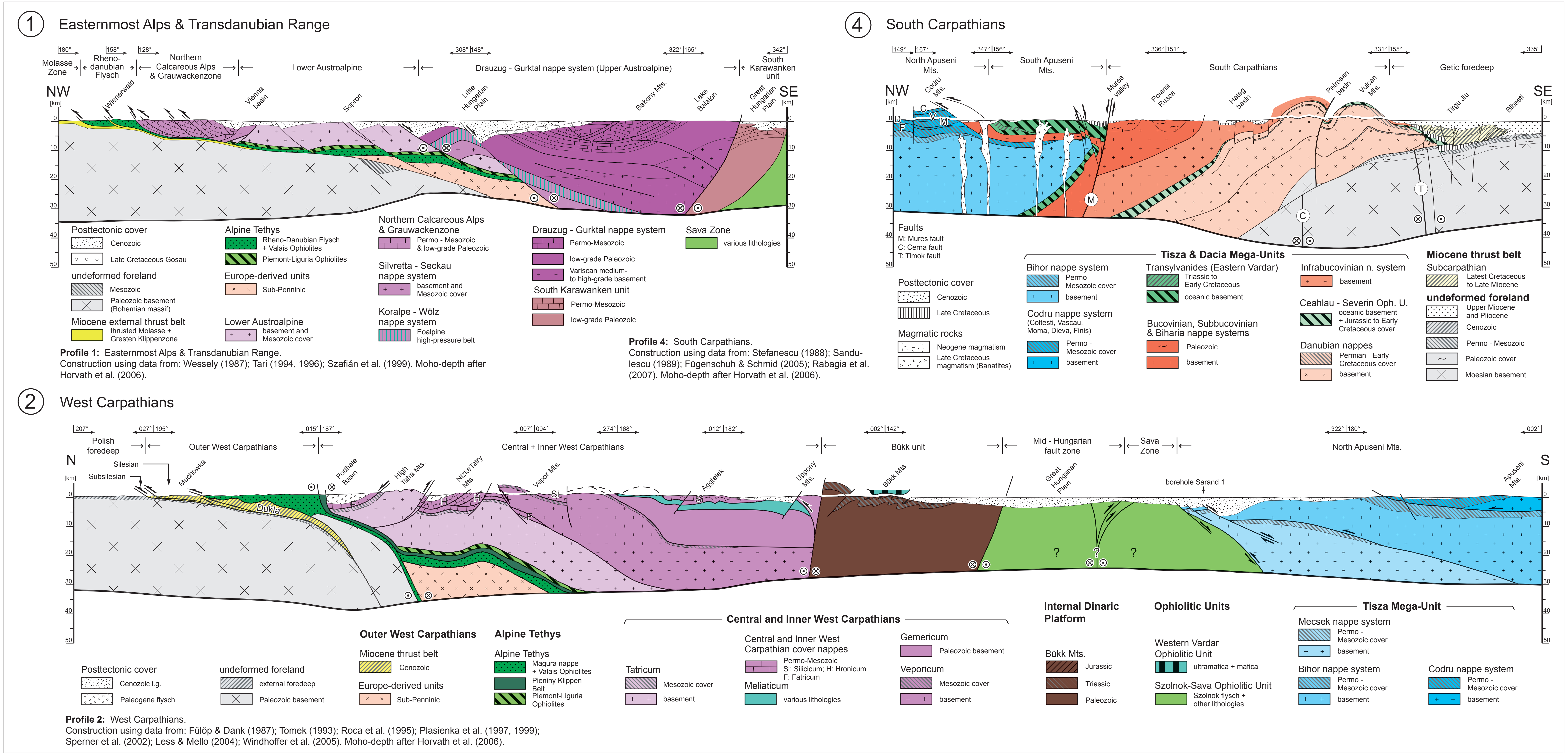
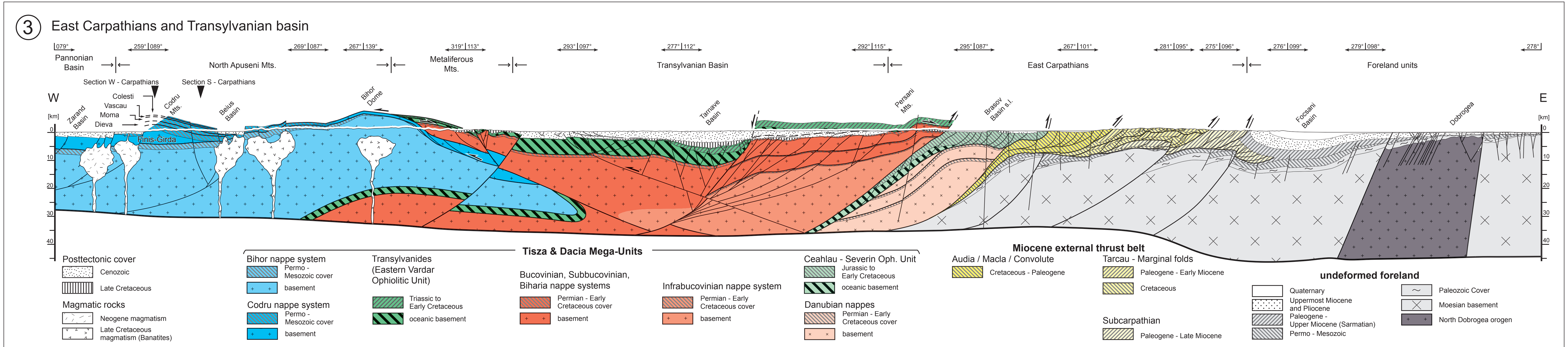
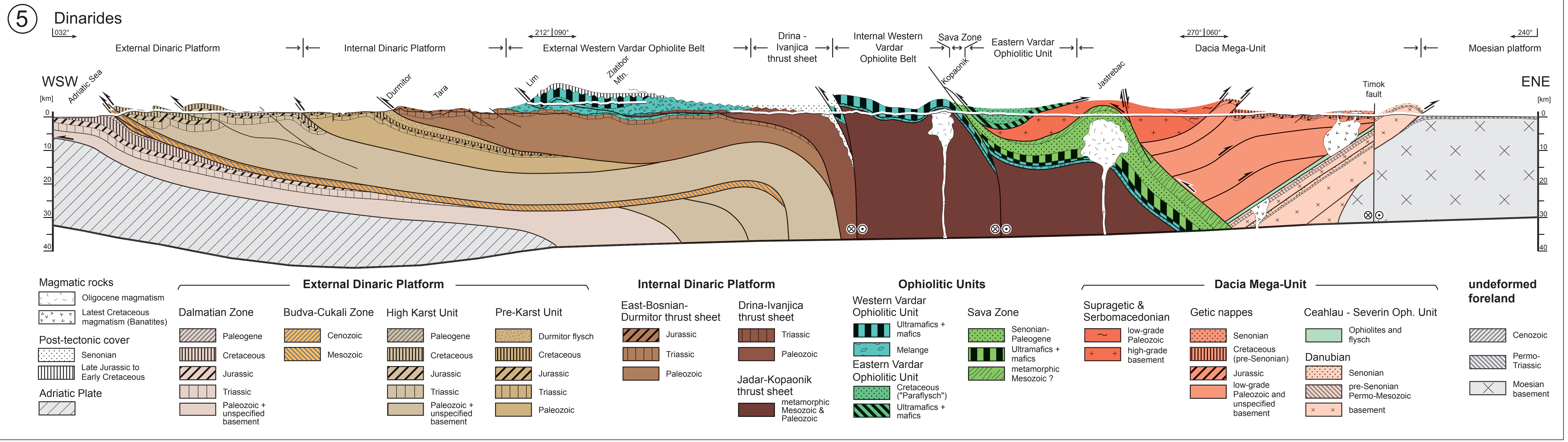


Plate 2: Crustal-scale cross-sections through the Alps, Carpathians and Dinarides (for location see Fig. 6).



Profile 3: East Carpathians and Transylvanian Basin.
Miocene flysch belt and its foreland after Matenco & Bertotti (2000), Bocin et al. (2005), Leever et al. (2006), Matenco et al. (2007a); Persani Mountains, Brasov Basin and Ceahlau nappe modified after Stefanescu et al. (1988) and Visarion (1988); Transylvanian Basin after Matenco et al. (2007b); Apuseni Mountains according to reconnaissance field work of the first author and on the basis of 1:50'000 and 1:200'000 map sheets of the Geological Institute of Romania. Moho depth after Martin et al. (2005) and Hauser et al. (2007).



Profile 5: Dinarides.
Profile construction based on a compilation of all the 1:100'000 geological maps of former Yugoslavia (Osnovna Geoloska Karta SFRJ) along and near the trace of the profile. Moho depth after Marovic et al. (2002) and Dimitrijevic (2002).

Plate 3: Crustal-scale cross sections through the Alps, Carpathians and Dinarides (for location see Fig. 6).