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NEOPROTEROZOIC TERRANES IN THE MOESIAN BASEMENT AND IN THE ALPINE DANUBIAN NAPPES OF THE SOUTH CARPATHIANS

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(11 figures)

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ABSTRACT. Geological, geochronological and geochemical evidence suggest that several distinct parts of the Neoproterozoic orogen developed at the active margin of Gondwana are preserved in Moesia and South Carpathians. Parts of a Neoproterozoic volcanic arc and foreland basin are preserved in East Moesia while the Alpine Danubian Nappes preserve remnants of a Pan-African (Cadomian) volcanic arc and its marginal backarc basin. The East Moesian basement exposed in Central Dobrogea includes Pre-Cambrian tholeiitic metabasites from Altîn Tepe showing arc/back arc affinities, subjected to a "Cadomian" deformation together with the Neoproterozoic-Early Cambrian turbidites of Histria Formation. In subsurface of South Dobrogea, a Neoproterozoic volcano-sedimentary suite with alkali basalts, related to rifting of the Precambrian Moesian crust, was deformed during latest Neoproterozoic thusting. The Precambrian basement of the Danubian Nappes includes Pan-African island arc metavolcanics (Drăgşan type terranes) with an amphibolite facies metamorphism connected to subduction of the arc complex. Emplacement of late-kinematic, high-K calc-alkaline granitoids records late accretion of oceanic rocks to a continent (Gondwana). Metasedimentary successions (Lainici-Păiuş type terranes) show HT-LP regional metamorphism and pervasive migmatization, prior to the intrusion of Neoproterozoic calc-alkaline and alkali-calcic plutons. Well documented Late Proterozoic ages of the magmatic protoliths in the Lower Danubian basement range between 780 to 570 Ma suggesting Cadomian affinities. The Pan-African tectono-metamorphic evolution can be dated by U/Pb, K/Ar and Ar/Ar isotopic data at 580-560 Ma. Mineralogical and chemical characters of the Tişoviţa-Iuţi mafic-ultramafic complex from the Upper Danubian resemble ophiolite complexes generated in ocean basins associated with arc systems formed at fast- to intermediate-spreading centres.

KEYWORDS: Danubian units, Moesian Platform, Neoproterozoic basement, Pan-African, South Carpathians

1. Introduction

The western margin of the East European Craton (EEC), running NNW-SSE from the Eastern North Sea to the Western Black Sea, is a prominent terrane boundary separating Precambrian Baltica crust from the Phanerozoic orogens of Europe and referred to as the Trans-European Suture Zone (TESZ) (Gee & Zeyen, 1996; Pharaoh, 1999) (Fig. 1).

Surrounded to the north and west by the South Carpathians and to south by the Balkans, Moesia lies at the SE extremity of the TESZ and played an important role inAlpine geodynamics and Cenozoic palaeogeography, when it was part of the European plate (Fig. 1). However, Moesia is one of the poorly known terranes in SE Europe in what concerns the complex Precambrian history, mainly because the Moesian basement is largely concealed by the Palaeozoic to Cenozoic platform cover. Petrographical and geochronological studies of the Moesian basement were published in the sixties and no petrological studies have been undertaken since. Geochemical investigation of the Neoproterozoic basaltic suites started only recently and is still in progress (Crowley *et al*., 2000; Seghedi *et al*., 2000).

The timing of accretion to the EEC, as well as the role of Moesia in the creation of the Carpathian-Balkan double-loop represent still highly debated issues, and most controversial is the palaeocontinental affinity of Moesia.

Late Neoproterozoic metamorphic-granitic terranes were also described in the neighbouring South Carpathians, as Baikalian (Savu, 1970), Cadomian (Berza, 1978) or Pan-African (Liégeois *et al*., 1996). There are authors considering that the Danubian basement represents in fact Moesian basement incorporated in the Danubian Alpine nappes of the South Carpathians (e.g. Săndulescu, 1984), but there is no direct evidence to support this view. The Gondwanan affinity of the Danubian basement from the lowermost South Carpathian nappes is fairly well known and supported by geochronological, geochemical and petrological data (Liégeois *et al*., 1996).

The present paper represents a review, based on published and unpublished data, of the Neoproterozoic rocks from the Moesian Platform and the Danubian units of the South Carpathians, and comments the models for their palaeocontinental affinity.

2. The Phanerozoic situation

The Cretaceous-Tertiary fold and thrust belt of the Central and SE Europe extends from the West and East Carpathians, then to the South Carpathians and further to the Balkans. The double loop, from the N-S structural trend of the East Carpathians to the E-W trend of the South Carpathians, through a tight southward bending to the E-W trend of the Balkans, occurs largely in Romania (Fig. 1). The loop is partly the result of oroclinal bending (Pavelescu & Nitu, 1977; Schmid *et al*., 1998), explained by changes in stress or in transport direction of the Carpathian units around the Moesian platform during Tertiary (Ratchbacher *et al*., 1993; Linzer *et al*., 1998; Mațenco *et al*., 1997; Fügenschuh & Schmid, 2005).

2.1. The Phanerozoic story of the Moesian Platform

North of the Danube River, the Moesian Platform lies on Romanian territory and corresponds geographically to the Danube Plain (Fig. 2). Geophysical and borehole evidence suggest the northward prolongation of the platform underneath the Carpathian belt: east of Intramoesian Fault the platform is underthrusted below the Tertiary nappes of the East Carpathian bend zone, while west of the Intramoesian Fault it is lying beneath the basement-cored Danubian nappe complex of the South Carpathians (Fig. 3). The southern third of the Moesian Platform lies on Bulgarian territory.

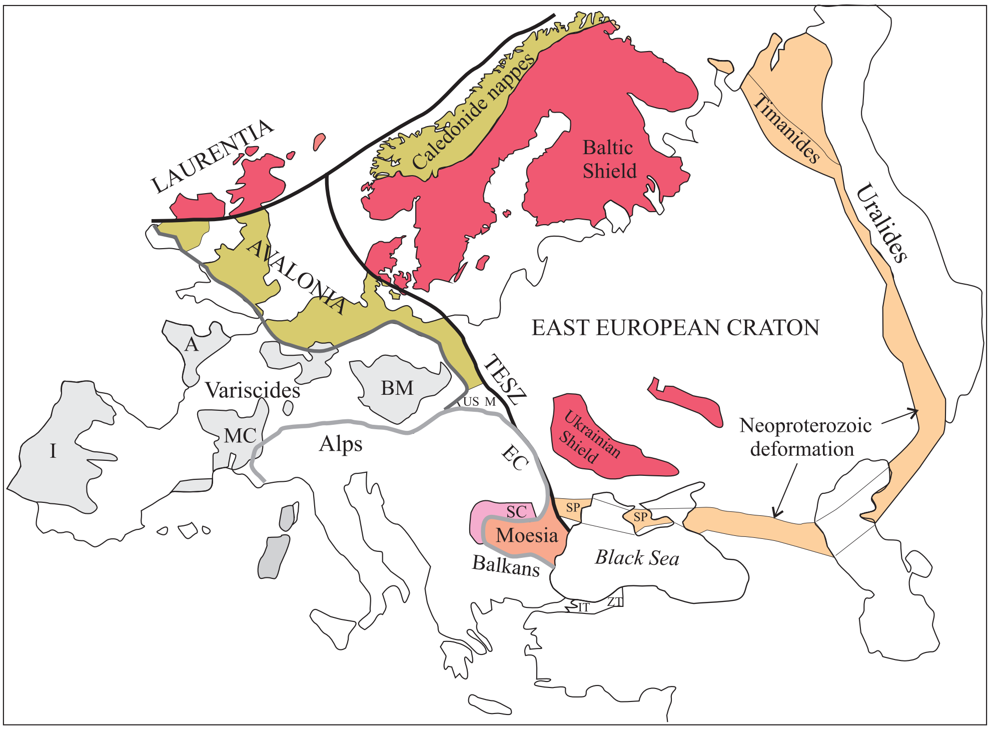

Figure 1. Location of Moesia and South Carpathians (SC) on the map showing the basement structure and Phanerozoic deformation belts in Europe. Key to abbreviations: TESZ – TransEuropean Suture Zone; EC – East Carpathians; M – Malopolska; US – Upper Silesia; SP – Skythian Plate; IT – Istanbul Terrane; ZT – Zonguldak Terrane; Terranes of Armorica Terrane Assemblage (ATA): A – Armorica; MC – Massif Central; I – Iberia; BM – Bohemian Massif.

In Cenozoic palaeogeography, the Moesian Platform represented a part of the European foreland, molded by the Carpathians and the Balkans, with an essential role in the achievement of the Carpathian double-bend (Stille, 1953; Burchfiel, 1980; Pavelescu & Nitu, 1984; Ratchbacher *et al.*, 1993; Berza *et al.*, 1994b; Schmid *et al.*, 1998; Fügenschuch & Schmid, 2005). Clockwise rotation accompanied the molding of the Cretaceous nappes around the western end of the Moesian block, increasing from 20-30º in Serbia to 80º in Banat (Panaiotu *et al.*, 2004). Săndulescu & Visarion (2000) relate the double bend to Cretaceous through Miocene deformation, connected to the interactions of the westward drifting Moesian block and the eastward translated Foreapulian block in conditions of clockwise rotation. The westward drift of Moesia was accommodated by the system of NW trending faults of East Moesia.

The Moesian Platform is separated from the Alpine orogenic belt by major, diachronous thrusts (Fig. 2). The South Carpathian Cretaceous nappe system was overthrusted onto the northern margin of the Moesian Platform in the Miocene (Săndulescu, 1984; Ştefănescu *et al.*, 1988); at depths the platform is in direct tectonic contact with the Danubian units of the South Carpathians (Visarion *et al.*, 1988) (Fig. 3). In Eastern Moesia, thrusting took place in the Sarmatian along the Peri-Carpathian Fault, juxtaposing folded Miocene sediments onto flat-lying, undeformed Sarmatian and older sediments of the platform. The Miocene thrust front is concealed by the Getic Depression, an EW elongated basin that represents the foredeep (Fig. 3). In Late Cretaceous, the southern margin of the platform was

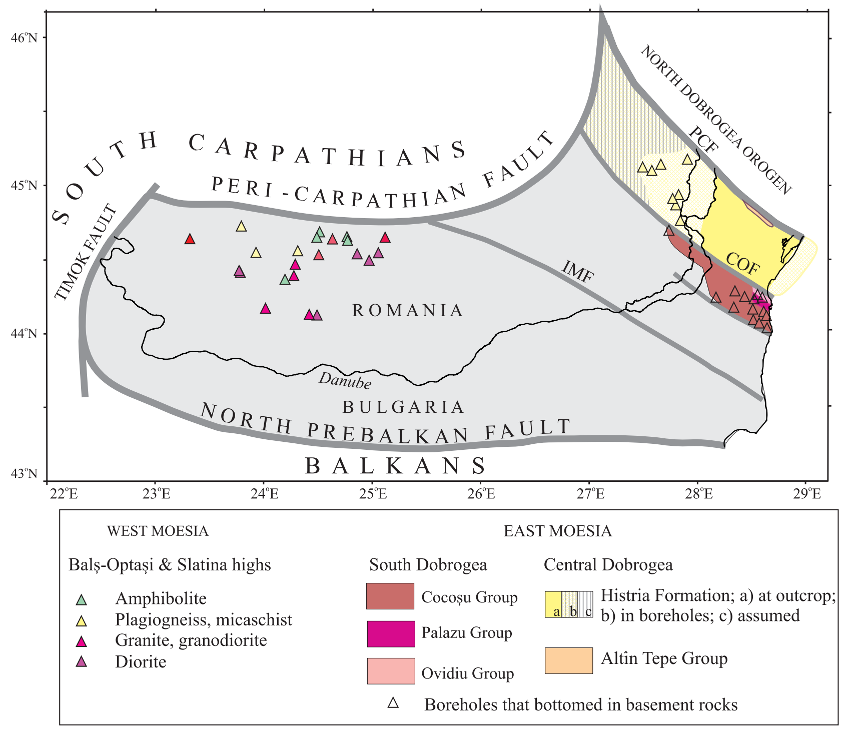

Figure 2. Simplified structural sketch of the Moesian Platform, showing the distribution of basement rocks in East Moesia and locations of boreholes which intercepted basement rocks in West Moesia. Abbreviations: PCF – Peceneaga-Camena Fault; COF – Capidava-Ovidiu Fault; IMF – Intramoesian Fault.

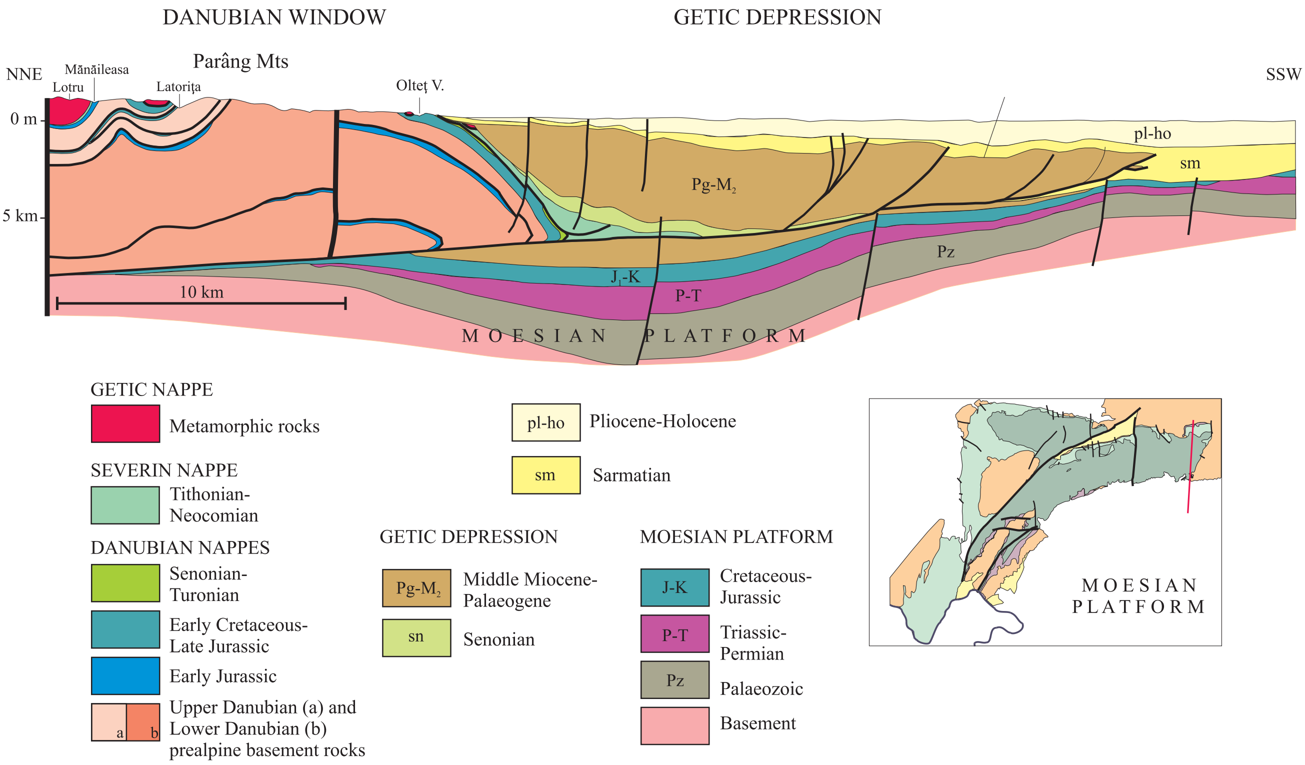

Figure 3. Geological cross-section through the Danubian Window and the northern part of the Moesian Platform, showing the relationships between the Danubian and Moesian basement (modified after Ştefănescu *et al*., 1988). Location of the section is shown in inset map. Legend of the inset map is like in Fig. 5.

underthrusted beneath the Balkans along the North-Prebalkan Fault (Săndulescu, 1984; Visarion *et al*., 1988; Tari *et al*., 1997). Seghedi & Berza (1994) postulated a coeval latest Cretaceous first thrusting of the South Carpathian nappe duplexes along a sole thrust over the Moesian Plaform.

The north-eastern margin of the Moesian Platform is formed by a major crustal fault – the Peceneaga-Camena Fault – developed SE-NW from the Black Sea shore to the Vrancea Zone of the East Carpathians (Fig. 2). The Peceneaga-Camena Fault is a fundamental terrane boundary, separating the Moesian Platform from the North Dobrogea Orogen, a terrane with major Cimmerian deformation.

Running NW from Cape Şabla on the Black Sea shore to NW of Bucharest, the Intramoesian Fault subdivides the Moesian Platform into a smaller north-eastern block (East Moesia) and a larger south-western block (West Moesia), which occupies the Romanian Plain and extends southward of the Danube to the Prebalkans. A major system of E-W trending normal faults is typically developed in West Moesia parallel to the Alpine belt, and is intersected by a second fault system trending about N-S (Barbu & Vasilescu, 1967; Barbu, 1973, 1980; Paraschiv, 1979; Săndulescu, 1984; Săndulescu & Visarion, 1988; Tari *et al*., 1997).

The structure of East Moesia is controlled by a system of NW-SE trending crustal faults (Fig. 2). NE-SW faults in East Moesia represent a secondary system parallel to the Carpathian bend zone. The main faults in East Moesia are the Peceneaga-Camena and Capidava-Ovidiu faults, with a long-lasting history of strike-slip deformation. The Capidava Ovidiu Fault separates the uplifted Central Dobrogea block from the downfaulted South Dobrogea block, with major differences in the constitution of their metamorphic basement and platform cover.

The relief of the basement beneath the Palaeozoic sediments is known largely from correlation of geophysical and borehole data. There are several basement highs or uplifts, controlled by a fault system paralleling the platform margins. From west to east these highs are described as Strehaia, Balş-Optaşi and Bordei Verde (Fig. 4). Olteniţa high represents the continuation north of the Danube of the North Bulgarian high. Below a 2000-4500m thick cover of Mesozoic sediments, basins filled with up to 4000-4500m of Palaeozoic deposits separate the highs.

Pre-Cenozoic rocks are largely concealed in the Moesian Platform, except the areas of Central and South Dobrogea from East Moesia. In East Moesia the Neoproterozoic basement is exposed in the uplifted block of Central Dobrogea. In the southern part of Central Dobrogea, erosional remnants of a Late Jurassic carbonate platform are locally preserved in open synclines on top of the basement. Cretaceous-Miocene sediments are exposed along valleys in South Dobrogea. In West Moesia, the Cretaceous cover is at outcrop south of the Danube, due to northward tilting of the platform.

2.2. The Alpine structure of the South Carpathians

Between the East Carpathians and the Balkans, the South Carpathians form a belt about 500 km long surrounding the northern and western margins of Moesia. The complex nappe structure, achieved during the Middle and Late Cretaceous crustal convergence, consists of two systems of basement-cored nappes with cover nappes sandwiched in between (Murgoci, 1905a, b, 1912; Streckeisen, 1934; Codarcea, 1940; Codarcea *et al*., 1967; Berza *et al*., 1983, 1994b). From top to bottom, the tectonostratigraphy of the Alpine nappe pile in the South Carpathians consists of the Supragetic and Getic nappe systems, the Severin nappe complex and the Danubian nappe systems (Fig. 5).

The Getic-Supragetic units represent slices of the Getic Plate, a continental fragment of the European Margin situated between the Transylvanian Thethys and a Jurassic-Lower Cretaceous Severin ocean (Săndulescu, 1975, 1984, 1994). Both Getic and Supragetic units are made of Variscan basement and Mesozoic ± Palaeozoic cover rocks. The Alpine metamorphism is typically absent in the Mesozoic cover of the Getic-Supragetic nappes.

The thin-skinned Severin and Coşustea nappes represent underthrusted units of the accretionary wedge,

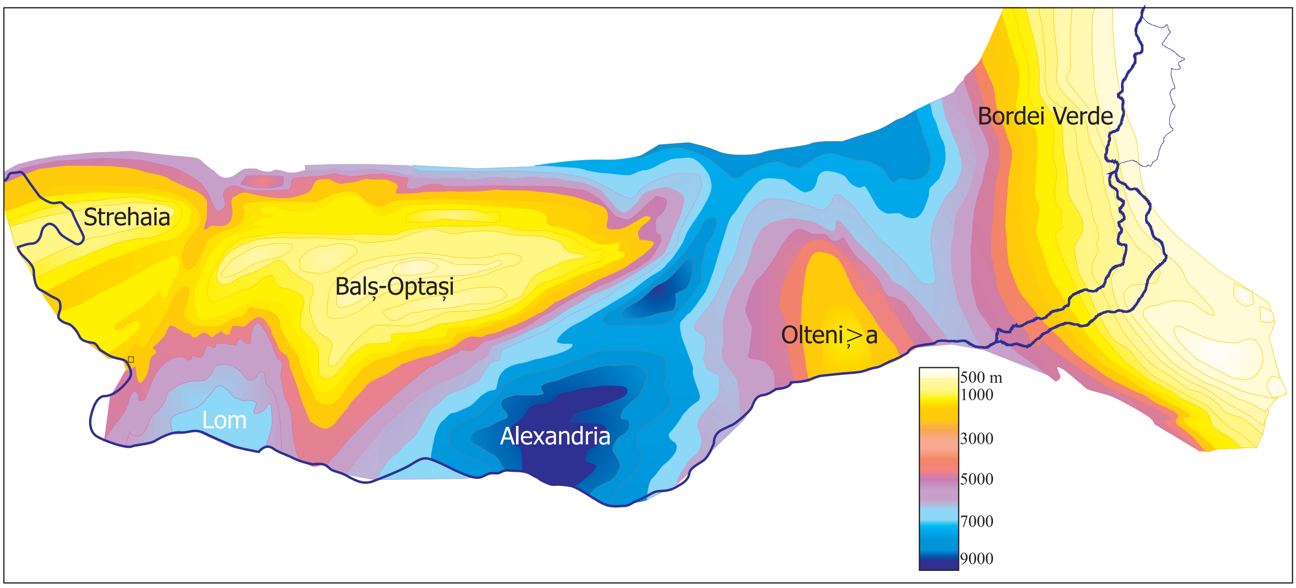

Figure 4. The Precambrian relief of the Moesian basement (modified after Paraschiv, 2001).

emplaced starting with the Early Cretaceous as a consequence of the subduction of the Moesian plate beneath the Getic plate (Seghedi & Oaie, 1997). The age of Severin and Obârşia nappes are considered intra-Aptian underthrusted by analogy with the Ceahlău Nappe from the East Carpathians (Săndulescu, 1984). The nappes have been subsequently deformed during the Late Cretaceous collisional events, responsible for the emplacement of the Getic + Severin complex on top of the Danubian units (Berza *et al.*, 1994b).

In the South Carpathians, the Danubian units are exposed in a tectonic window referred to as the Danubian Window. The Danubian Nappes show the geometry of an antiformal stack eroded in the zone of axial culmination (Seghedi & Berza, 1994), well illustrated in cross-sections (Ştefănescu *et al.*, 1988; Iancu *et al.*, 1998) (Fig. 2). Each Danubian nappe represents a thrust-bounded horse, consisting of a pre-Alpine basement (metamorphic rocks, granitoids and locally thin Palaeozoic formations), overlain by very low to low grade metamorphic Mesozoic successions, starting with the Liassic (Berza *et al.*, 1988a, b). The Triassic is typically absent in the Danubian realm. A major fault of Latest Cretaceous age separates an Upper Danubian nappe stack from a Lower Danubian one, acting as a duplex boundary or as an out of sequence fault. The eastern border of the Danubian Window, including also the Severin nappe, is recognized as a low-angle normal fault marking the Palaeogene detachement of the overlying Supragetic/Getic complex due to orogen-parallel stretching (Schmid *et al.*, 1997; Fügenschuh & Schmid, 2005).

The South Carpathians are obliquely cut by the Cerna-Jiu strike-slip fault, averaging 35 km of dextral translation (Berza & Drăgănescu, 1988). The Cerna-Jiu Fault runs along a NE-SW direction, following the Carpathian bend to Serbia as Porečika-Ravna Fault and merging southward with the Timok–Struma fault system (Kräutner & Krstić, 2003) (Fig. 5). The Tertiary Petroşani Basin formed during Chattian and Badenian activity of the Cerna-Jiu Fault (Berza & Drăgănescu, 1988).

3. The Pre-Neoproterozoic Moesian basement

The areal distribution of the Moesian basement rocks is shown in Figure 2. Schematic logs of the basement lithologies and their Palaeozoic cover rocks are presented in Figure 6 for

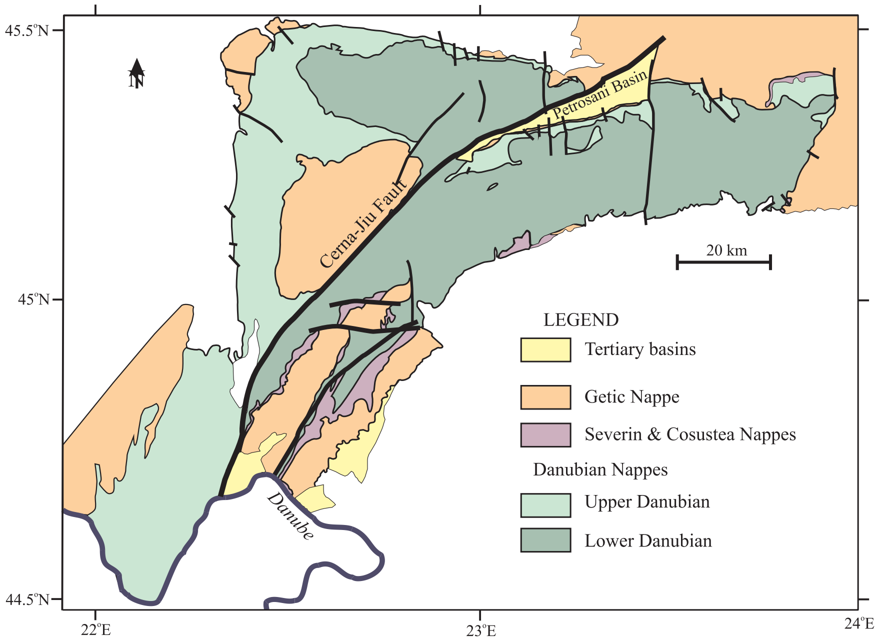

Figure 5. Tectonic map showing the main Alpine units exposed in the Danubian Window (simplified after the map of Berza, Iancu, Seghedi & Drăgănescu, 1994, in Berza *et al.*, 1994b).

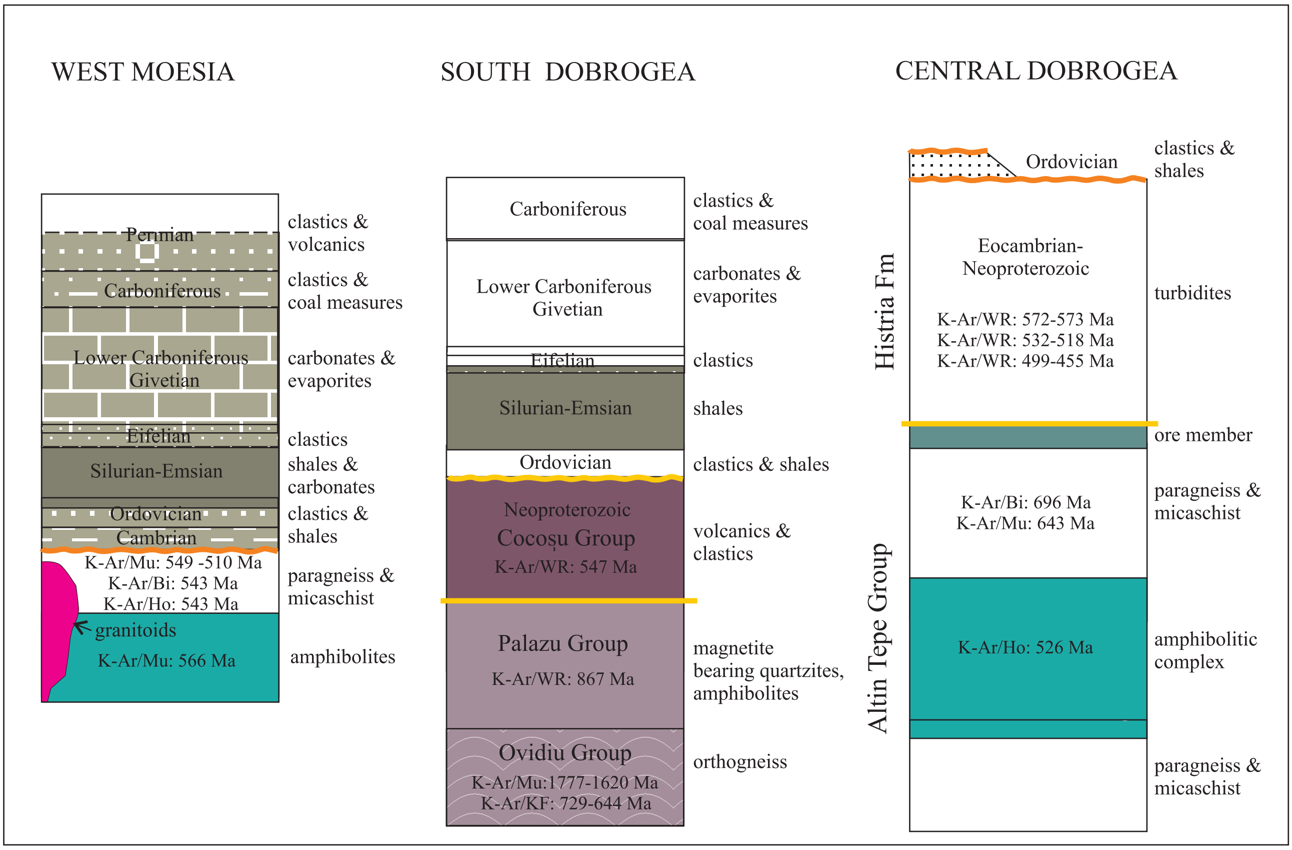

Fig. 6. Schematic lithologic logs of the East and West Moesian basement and their Paleozoic cover (not to scale). Geochronology after Giuşcă *et al.*, 1967, recalculated by Kräutner *et al.*,1988b. Abbreviations: Bi – biotite, Mu – muscovite, Ho – hornblende, KF – K feldspar, WR – whole rock.

West Moesia and South and Central Dobrogea from East Moesia. Except for Central Dobrogea, where the Neoproterzoic basement is exposed, in the rest of the platform the basement is concealed and known only from boreholes. The oldest rocks were found in the South Dobrogea block of East Moesia, where boreholes pierced orthogneisses (Ovidiu Group) and a banded iron formation (BIF) (Palazu Group) (Giuşcă *et al*., 1967, 1976) (Fig. 6). Based on magnetic data (Gavăt *et al*., 1965), a large distribution of this older, “Karelian” basement, was suggested in the central and southern parts of the Moesian Platform, bordered on the northern and southern sides by Neoproterozoic series (Barbu & Vasilescu, 1967; Barbu, 1973). However, this interpretation was not confirmed so far by boreholes drilled in the western part of the platform, which pierced only Neoproterozoic metamorphic rocks.

According to Visarion *et al*. (1979, 1988) and Săndulescu & Visarion (1988), geophysical data indicate that the Palazu-type basement extends both eastward, in the Black Sea offshore, as well as to NW, in the marginal part of the platform, underthrusted beneath the Tertiary nappes in the southern part of the East Carpathians bend zone.

The cratonic basement of South Dobrogea was pierced at depths between 430 and 600 m by boreholes from Palazu Mare – Cocoşu area near Constanţa, SE of Capidava-Ovidiu Fault (Fig. 7B). The Palazu Group consists of two main lithological types, amphibolites and magnetite-bearing quartzitic rocks. They show interbeds of carbonate and quartzitic rocks, as well as of graphitic micaschists, which suggest sedimentary bedding. Detailed mineralogical and chemical studies (Giuşcă *et al*., 1967, 1976; Giuşcă, 1977) revealed that the banding is produced by alternating layers with various amounts of quartz, magnetite, hornblende, cummingtonite, almandine, biotite, dolomite and ankerite. The carbonate beds consist of tremolite, ferrosalite and/or diopside. The protolith of these rocks is a banded chert sequence. This includes cherts, as well as clayey and shally muds. This protolith type represents abyssal oceanic sediments related to sea-floor spreading. The associated hornblende and cummingtonite schists resulted from sideritic and carbonate cherts, respectively; the high initial content of Ni, Cr and V was interpreted to indicate the presence of initial clayey fraction, mixed with the iron-rich sediments (Giuşcă *et al*., 1976). The associated clastic rocks include micaschists, quartzites and microcline gneisses, with frequent graphite and seldom thin amphibolite interlayers. Based on lithological resemblance, the Palazu Group was correlated to the Krivoi Rog series from the Ukrainian shield (Giuşcă *et al*., 1967).

The Ovidiu Group includes an assemblage of migmatic gneisses, granite gneisses and pegmatites. Based on geometric relations in boreholes, where the banded iron formation shows a gradual transition to massive, feldspar rich, and coarse grained gneisses, the Ovidiu gneisses were interpreted as a pre-Karelian (Archaean) basement (Giuşcă *et al*., 1967; Visarion *et al*., 1979).

The oldest K-Ar ages yielded by the Ovidiu orthogneisses range between 1777 Ma (on K feldspar) and 1620 Ma (on muscovite). These ages were interpreted as the overprint of the Karelian amphibolite facies metamorphism of the overlying Palazu BIF, a LP-HT event which produced assemblages with andalusite + sillimanite (600°C and 4-5 kbar) (Giuşcă *et al*., 1976; Giuşcă, 1977). A second age group includes Neoproterozoic ages (867 Ma whole-rock on micaschist, and 729-644 Ma on K-feldspar from granite gneiss) (all ages recalculated by Kräutner *et al*., 1988 b) (Fig. 7). A later, greenschist facies retrogression, was correlated with the very low grade, late Cadomian metamorphism of the overlying Neoproterozoic Cocoşu Formation (Kräutner *et al*., 1988 b).

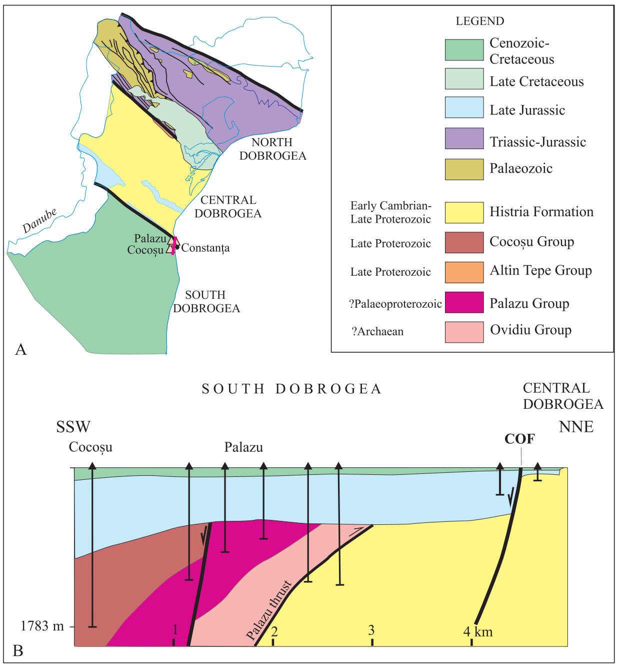

Figure 7. A. Simplified geological map of Dobrogea, showing the location of boreholes and the line of the cross-section. B. Geological cross-section in South Dobrogea south of the Capidava-Ovidiu Fault (COF), showing the relationships between the basement rocks (modified after Kräutner *et al*., 1988b). The Ovidiu Gneisses overthrust the Histria Formation along the Palazu Thrust. The contact of the Cocoşu Group with the Palazu Group, marked by mylonites, is interpreted as a normal fault.

Detailed studies of borehole cores revealed that the Ovidiu gneisses and the Palazu Group are involved in northward directed thrusts (Visarion *et al*., 1979; Kräutner *et al*., 1988 b). Because beneath strongly mylonitic Ovidiu gneisses three boreholes intercepted pelitic-psammitic rocks ascribed to the Vendian Histria Formation, this thrust was interpreted as Late Proterozoic (the Palazu Thrust, Visarion *et al*., 1979) (Fig. 7B). The Neoproterozoic Cocoşu Formation, lying on top of the Palazu Series, shows a brittle tectonic contact with its basement rocks.

3.1. The Neoproterozoic basement of East Moesia

3.1.1. The Neoproterozoic basement of South Dobrogea: the Cocoşu Group

Several boreholes emplaced in the area of the magnetic high from the north-eastern part of South Dobrogea, south of the Capidava-Ovidiu Fault, intercepted the Cocoşu Group below Late Jurassic carbonate platform limestones (Visarion *et al*., 1979) (fig. 7B). Geophysical data indicate that the basement of South Dobrogea is downfaulted to the south, where Cambrian and Ordovician overstep sequences were pierced by boreholes below the Late Jurassic limestones.

The lithological succession of the Cocoşu Group comprises two formations, well illustrated in borehole 5051 Cocoşu (Fig. 8). The lower, volcano-sedimentary formation, consists of basalt flows separated by coarse to fine-grained volcaniclastics interbedded with epiclastic rocks. The mafic rocks are massive to porphyritic basalts and dolerites, with well preserved primary magmatic minerals and petrographic features indicating submarine volcanism. Abundance of large limestone clasts in the epiclastic beds suggests a shallow-marine depositional environment. Geochemical studies indicate that the basaltic rocks are basanites and trachybasalts. They are products of an alkaline mafic volcanism with intraplate affinities, interpreted to suggest a transtensional rift setting (Seghedi *et al*., 2000).

The upper, siliciclastic member, is dominated in the lower part by red and green clays, sometimes thinly laminated with green siltstone layers. A coarser facies of conglomerates and sandstones occurs to the top of the succession. Clast petrography of the conglomerate beds indicate a source area consisting of quartz, granites, gneisses and rhyolites.

A penetrative, steeply dipping slaty cleavage (75-90^{o}) develops in the siliciclastic succession, penetrative in the lower, shally part. The low-angle of bedding/cleavage intersection suggests that the borehole was drilled on the limb of a steep fold. A tectonic contact between the upper and lower formations is indicated by brittle deformation affecting the slates and development of a mylonitic foliation in the volcano-sedimentary succession.

The age of the Cocoşu Group is pre-Ordovician, as indicated by geological evidence, and is ascribed to the Neoproterozoic (Vendian), coeval with the Histria Formation (Giuşcă *et al*., 1967; Kräutner *et al*., 1988b). However, an Ediacaran age cannot be precluded in the absence of reliable geochronological evidence. Red slates from the upper member of the Cocoşu Group yielded a whole rock K-Ar age of 547 Ma, interpreted as a slightly disturbed Cadomian/Baikalian age of their very low grade, subgreenschist facies metamorphism (Giuşcă *et al*., 1967; Kräutner *et al*., 1988b).

3.1.2. The Neoproterozoic basement of Central Dobrogea: Altîn Tepe Group and the Histria Formation

In Central Dobrogea the basement is largely exposed and consists of Neoproterozoic metamorphic rocks (Altîn Tepe Formation) and a thick Neoproterozoic-Early Cambrian (Vendian) turbidite succession (Histria Formation) (Fig. 6).

The Altîn Tepe Group crops out south of the Peceneaga-Camena Fault, in the core of an antiformal fold, which trends NW and plunges to SE beneath the Histria Formation (Fig. 7A). It consists of polymetamorphic rocks ascribed to the Upper or Middle Proterozoic (Codarcea-Dessila *et al*., 1966; Mirăuţă 1969; Kräutner & Savu, 1978, respectively). Mureşan (1972) separated three formations in the Altîn Tepe Group (Fig. 6):

- a lower terrigeneous complex (gneiss-micaschist formation), including an alternation of micaschists, paragneisses and quartzites;
- a middle, basic tuffaceous complex (amphibolite formation), consisting by banded amphibolites with thin interlayers of micaschists and metagabbro sills;
- an upper terrigenous complex (quartzite-micaschist formation) is made of biotite and biotite-muscovite quartzites, micaschists and paragneisses. Geochemical features suggest that the Altîn Tepe metabasites are tholeiitic, showing arc/back arc affinities (Crowley *et al*., 2000).

An early, amphibolite facies (staurolite zone) metamorphism is suggested by the mineral assemblage (Kräutner *et al*., 1988b). Based on K-Ar geochronology, the Altîn Tepe Group was ascribed to the Late Proterozoic (Codarcea-Dessila *et al*., 1966; Giuşcă *et al*., 1967; Semenenko *et al*., 1969, in Kräutner *et al*., 1988a; Mirăuţă, 1969), or to the Middle Proterozoic (Kräutner & Savu, 1978; Kräutner *et al*., 1988b). K-Ar ages range from 696-643 Ma (yielded by biotite from micaschists, Giuşcă *et al*., 1967) to 526 Ma (on hornblende from amphibolites, Semenenko *et al*., 1969) (all ages recalculated by Kräutner *et al*., 1988b). These ages were interpreted as the Late Neoproterozoic age of the amphibolite facies metamorphism (Giuşcă *et al*., 1967). Kräutner *et al*. (1988b) ascribed the Altîn Tepe Group to the Mesoproterozoic, interpreting the Neoproterozoic ages as connected to a late Proterozoic regional retrogression, due to partial Ar loss during the Cadomian metamorphism of the Histria Formation. The Altîn-Tepe Group was traditionally regarded as the basement of the overlying Neoproterozoic-Eocambrian turbiditic succession of East Moesia (Ianovici & Giuşcă, 1961; Giuşcă *et al*., 1967). The top of the metamorphic

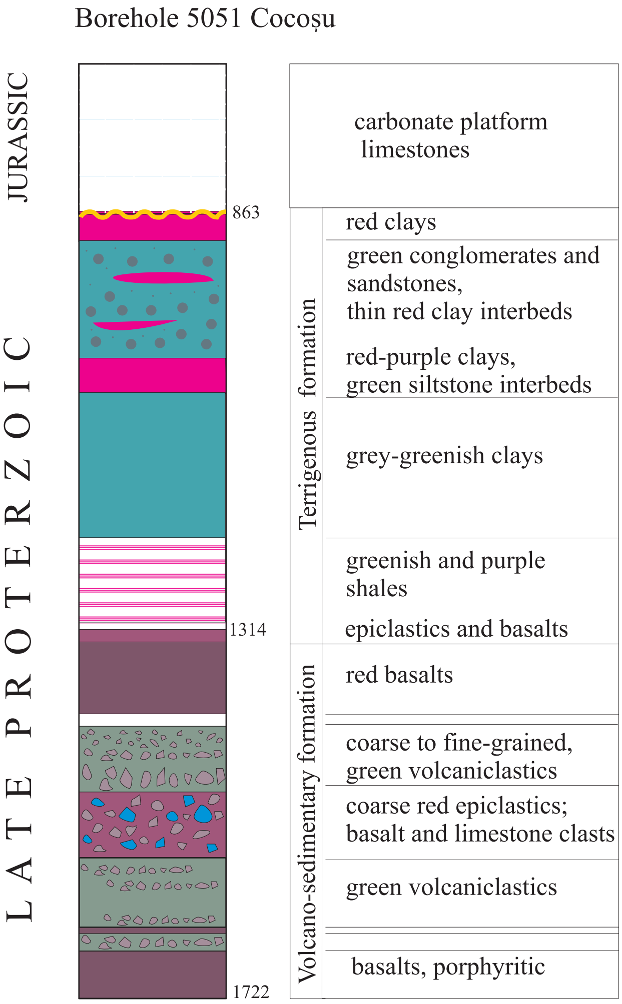

Fig. 8. Lithologic log of the Cocoşu Group in borehole 5051 Cocoşu. Cores from the lowermost interval of the borehole (1722-1783m) are not preserved.

suite shows a ductile mylonitic zone, formed in lower greenschist facies along the contact with the Histria Formation (Mureşan, 1971, 1972; Kräutner *et al.*, 1988b; Seghedi & Oaie, 1994). The geometry of the contact is that of a tectonic window, and was interpreted as "the Altîn-Tepe window" below the nappe of the Histria Formation (Mureşan, 1971). The geometry and the ductile nature of the contact, the brittle deformation in the overlying, younger turbiditic successions, as well as kinematic indicators, all suggest that this contact is rather a shallow extensional detachment, like that shown in metamorphic core complexes (Seghedi *et al.*, 1999). However, detailed structural analysis is required to document the structure of this area.

The Histria Formation is exposed over the entire area of Central Dobrogea, overlain by small, scattered remnants of a largely eroded Late Jurassic carbonate platform succession. West of the Danube, several boreholes drilled in the westward prolongation of the outcrop area from Central Dobrogea bottomed in the Histria Formation after piercing the Palaeozoic cover of this unit (Fig. 2). In boreholes this succession is overstepped by Ordovician quartzitic sandstones and green shales dated on graptolites (Murgeanu & Spassov, 1968).

The Histria Formation consists of a turbiditic succession about 5000 m thick, as suggested by geological and seismological data (Mirăuţă, 1969; Visarion *et al.*, 1988). The turbidites represent submarine fan deposits, forming a northward prograding sequence accumulated in a deep basin floored by continental crust (Seghedi & Oaie, 1995; Oaie, 1999). Geophysical data suggest that the basin floor consisted partly of Altîn Tepe type metamorphic rocks, as well as gneisses, probably like those from South Dobrogea (Visarion *et al.*, 1988). The shape of the magnetic anomalies in the Palazu area suggests that the Palazu iron-bearing rocks might continue at depth north-east of the Capidava-Ovidiu Fault in the basement of the Histria Formation (Visarion *et al.*, 1988; Stănică & Stănică, 1989).

The Histria Formation includes a lower and upper member of coarse, sandstone dominated, channelized midfan turbidites (Seghedi & Oaie, 1995; Oaie, 1999). Between the coarse members a thin member of distal turbidites was maped, dominated by lower fan and abyssal plain turbidites. Based on sedimentological data, a foreland basin setting is supposed for the Histria Formation. A detailed presentation of the sedimentological, petrographic and geochemical features of the Histria Formation is given by Oaie *et al.* (this volume).

The age of the Histria Formation was ascribed to the Late Neoproterozoic-Early Cambrian, based on palynological assemblages (Iliescu & Mutihac, 1965), and on the medusoid *Nemiana simplex* Palij, identified in fine-grained turbidites (Oaie, 1992; 1999). The deformation of the turbidites by open folds in very low-grade metamorphic conditions took place during the Baikalian/Cadomian events (Giuşcă *et al.*, 1967; Mirăuţă, 1969). This deformation predated the deposition of the unconformable Ordovician sediments, and occurred at the end of Neoproterozoic according to K-Ar data (Giuşcă *et al.*, 1967; Kräutner *et al.*, 1988b).

Mineralogical studies indicate that the turbidites were sourced by an active continental margin and a volcanic arc (Seghedi & Oaie, 1995; Oaie, 1999; Oaie *et al.*, this volume). U-Pb SHRIMP zircon data indicate provenance from Archaean and Grenvillian metamorphic sources and Neoproterozoic magmatic and metamorphic sources (Żelaźniewicz *et al.*, 2001).

3.2. West Moesia

The Balş-Optaşi basement uplift from the central-western part of West Moesia is still poorly known, as it was intercepted by only a few drillings (Paraschiv, 1974) (Figs 2 & 4). Several boreholes bottomed in amphibolite facies metamorphic rocks, locally showing strong retrogression or mylonitization in greenschist facies conditions (Paraschiv, 1974). Amphibolites represent the dominant lithology, but mylonitic micaschists with garnet porphyroblasts, as well as paragneisses and quartzite-muscovite schists have been found in three boreholes. The Late Neoproterozoic age of the metamorphic rocks is indicated by stratigraphic data and K-Ar geochronology. The oldest Palaeozoic sediments overlying the basement are Cambrian in age (Paraschiv, 1974). K-Ar ages yielded by metamorphic samples range between 566 ± 11 Ma (muscovite from amphibolite schists), 549 ± 16 Ma (muscovite from micaschist), 543 ± 17 Ma (biotite and amphiboles from micaschist), 510 ± 16 Ma (muscovite in paragneiss)(Paraschiv *et al.*, 1982, 1983; Paraschiv, 1986a). They were interpreted to reflect the late Neoproterozoic greenschist facies mylonitization of rocks (Paraschiv *et al.*, 1982, 1983).

A series of boreholes intercepted magmatic rocks in the Balş-Optaşi and Strehaia uplifts. Permo-Triassic or Jurassic sediments directly overlay most of these rocks.The Balş-Optaşi uplift includes two types of older magmatic rocks, with granodiorites and diorites-gabbros prevailing to the west, and granites dominating to the east (Barbu & Dăneţ, 1970) (Fig. 2). There is very little information regarding the petrographical features of the intrusive rocks. Opinions regarding the age of the intrusives range from entirely

Precambrian (Mutihac & Ionesi, 1975; Pătruţ, in Paraschiv *et al*., 1982), Precambrian and Palaeozoic (Barbu & Dăneţ, 1970), to entirely Palaeozoic, connected to Caledonian or Variscan events (Paraschiv 1974, 1979; Paraschiv *et al*, 1982, 1983). The oldest K-Ar age is 371 ± 11 Ma, the other ages ranging between 350 ± 11 and 281 ± 10 Ma (Paraschiv *et al*., 1982, 1983; Paraschiv, 1986a, b).

4. The Danubian basement in the South Carpathians

The pre-Alpine basement of the Lower Danubian Nappes includes late Neoproterozoic plutons and high-grade metamorphic groups underlying low-grade metamorphosed Ordovician to Lower Carboniferous formations, the latter presented in detail by Iancu *et al*. (this volume). Two types of pre-Ordovician metamorphic groups, with contrasting protoliths, metamorphism and associated magmatism are involved in a pre-Permian nappe structure (Fig. 9): the Lainici-Păiuş Group, dominated by HT-LP metasediments, and the Drăgşan Group, dominated by MT-MP metabasites (Pavelescu, 1958; Berza, 1978). Together with the plutons and dyke swarms crossing the metamorphic rocks and with their unconformable Ordovician to Lower Carboniferous covers, these two groups constitute two distinct Variscan units. These units are in contact along a thrust bringing Drăgşan on top of Lainici-Păiuş and sealed by Permian-Mesozoic sedimentary sucessions (Berza & Seghedi, 1983; Berza & Iancu, 1994; Berza *et al*., 1994b).

In the Upper Danubian Nappes, the pre-Alpine basement also exposes this Variscan nappe structure involving Drăgşan and Lainici-Păiuş type metamorphic sequences. However, there are some particular features found both north of Danube, in Banat (Romania), and south of it, in Miroć (Serbia), where large pre-Ordovician mafic-ultramafic massifs and (possibly) Variscan granitoid plutons occur (Stan, 1996). Both in Romania and Serbia, the mafic-ultramafic exposures are tectonically bordered and no isotopic dating is yet available. Gabbros of the Tişoviţa-Iuţi complex from Southern Banat (Romania) are reworked in Ordovician conglomerates (Iancu *et al*., 1990), along with serpentinite, flasser gabbro, amphibolite and granite boulders, suggesting pre-Ordovician emplacement in a Drăgşan-type environment.

4.1. Lainici-Păiuş metasedimentary Group and associated magmatism

The Lainici-Păiuş Group (Manolescu, 1937) is outcropping on large areas in the pre-Alpine basement of Lower Danubian Schela-Petreanu and Lainici Nappes, but similar metasedimentary sequences also exist in the basement of some of the Upper Danubian Nappes. Berza (1978) has formalized two lithostratigraphic units: a lower "Carbonate-Graphitic Formation", consisting of marble, graphite mica gneiss, amphibolite and calc-silicate gneiss, and an upper "Quartzitic and Biotite Gneiss Formation", where these two main lithologies contain minor marble, graphite mica gneiss, amphibolite and calc-silicate gneiss (Fig. 10). Graphite is a typical mineral for most Lainici-Păiuş rocks, being a major constituent in mica gneisses, locally attaining up to 60% in case of some ore deposits. Both formations show identical metamorphic evolution, consisting in high temperature, medium pressure assemblages (sillimanite-andalusite-cordierite-almandine in metapelites; forsterite-diopside-phlogopite-pargasite in marbles and calc-silicate rocks), partly overprinted by low-temperature minerals, due to Variscan and Alpine tectonics.

Typical for Lainici-Păiuş Group outcrops is the presence of heterogranular leucogranite with black K-feldspar, muscovite and garnet, forming a dense, meter-sized network which crosses all lithologies, including marble and pure quartzite and sometimes accumulating as bodies up to 1 km large. Zircons from such a leucogranite have been analysed by Grünenfelder *et al*. (1983), and their data recalculated by Liégeois *et al*. (1996) give an upper intercept of discordia at 582 ± 7 Ma. This metamorphic-leucogranitic assemblage is cut by elongated plutons (up to 100 km long and 15 km wide) covering almost half of the exposure area (Fig. 9). The plutons were ascribed to two distinct suites (Manolescu, 1937; Berza, 1978): the Şuşiţa-type, medium-K calc-alkaline mostly granodioritic-tonalitic (Savu, 1970; Savu *et al*., 1971) and the Tismana-type, very high-K calc-alkaline, mostly granitic but including kilometre-size pyroxene diorite schlieren and ultramafic cumulate pods (Berza, 1978; Duchesne *et al*., 1998). There is no direct evidence for a time sequence of these two contrasting suites, or reliable isotopic dating for Şuşiţa-type plutons, but it is

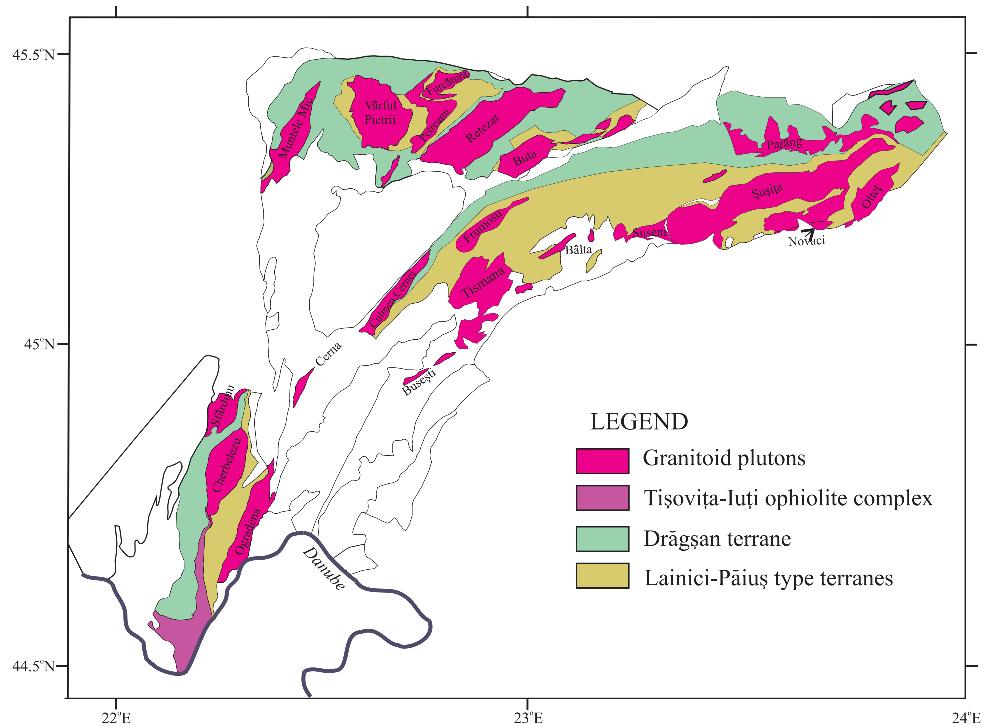

Fig. 9. Distribution of the Danubian basement rocks in the Danubian Window. Symplified after Berza *et al*. (1994b).

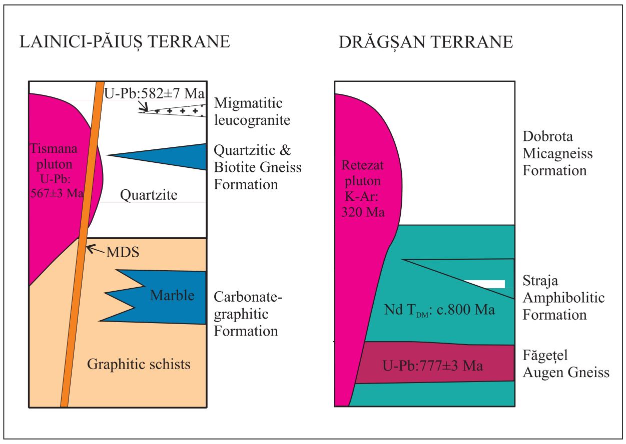

Fig. 10. Schematic logs showing the main lithologies in the Drăgşan- and Lainici-Păiuş-type terranes (not to scale). Geochronology after Liégeois *et al*. (1996). MDS – Motru dyke swarm.

considered that the latter are the first emplaced plutons. The Vârful Pietrii pluton is a special case, consisting of a hundred of metre-size marginal hornblende biotite diorite facies and of a central muscovite garnet leucogranitic core, exposed on a circular area with 15 km in diameter. The pluton is chemically alike the leucogranitic core (Duchesne *et al.*, 1998; Gandrabura & Kasper, 2001), but with different isotopic signature (juvenile versus crustal, Duchesne *et al.*, 1998).

Tismana-type intrusions were dated at Tismana with an upper discordia intercept at 567 ± 3 Ma (Liégeois *et al.*, 1996) and at Novaci by Grünenfelder *et al*. (1983), whose data recalculated by Liégeois *et a*l. (1996) give an upper discordia intercept at 588 ± 5 Ma. Together with biotite K-Ar ages in the 600-550 Ma range (Grünenfelder *et al.*, 1983) and with hornblende and muscovite Ar-Ar ages of 596-560 Ma (Dallmeyer *et al.*, 1996, 1998), isotopic dating certifies that high-temperature low-pressure metamorphism, leucogranite injection and both Şuşiţa-type and Tismana-type plutonism occured during late Neoproterozoic, covering an age range typical for high-K calc-alkaline intrusions from the Pan-African belt of Sahara (Liégeois *et al.*, 1996).

A dyke swarm of porphyric rocks was described by Berza & Seghedi (1975a) as pre-Silurian lamprophyres, microdiorites, microgranodiorites and micrograntes and recognized as a younger magmatic phase, cross-cutting Lainici-Păiuş crystalline schists with their associated leucogranites and the Şuşiţa- and Tismana-type plutons, but not intruding the overlying Ordovician-Silurian sediments. Féménias (2003) has described these dykes as basalts, andesites, dacites and rhyolites forming the Motru Dyke Swarm, a medium-K calc-alkaline suite emplaced in a high level sub-volcanic environment. Using the magnetic fabric of the dykes, Féménias *et al*. (2004) have found that the large (1m to 30 m thick) dykes are strongly asymmetrical, presumably indicating a regional field of sinistral transcurrent shearing during the time as they acted as feeders to volcanic processes. The age of this transcurrent faulting and coeval calc-alkaline volcanism is not yet known, but it postdates the 570 Ma plutons and predates the overstepping Upper Ordovician-Silurian cover.

4.2. Drăgşan metavolcanic Group and associated magmatism

The Drăgşan Group (Pavelescu, 1953) outcrops in the pre-alpine basement of the Lainici Lower Danubian Nappe, but similar sequences are also exposed in the basement of Upper Danubian nappes from the Romanian and Serbian South Carpathians (Berza *et al.*, 1994b; Kräutner & Krstić, 2003). The Drăgşan Group (Fig. 10) includes the lower Făgeţel Augen gneiss Formation, the dominant Straja Amphibolitic Formation including kilometre-size ultramafic cumulates and the upper Dobrota Micagneiss Formation (Berza & Seghedi, 1983; Kräutner *et al.*, 1988a). For all these lithological units, high-grade minerals are hornblende-andesine-garnet in amphibolites, kyanite-staurolite-garnet in mica gneisses and microcline-andesine-biotite in augen gneisses, pointing to medium pressure/medium temperature metamorphic conditions (Berza & Seghedi, 1975b; Berza, 1978; Solomon, 1985). Extensive Variscan and Alpine dynamic retrogressions had induced a greenschists overprint over large areas, especially in the Upper Danubian basement.

Major and trace elements of both light-(tonalitic) and dark-coloured (dioritic-gabbroic) samples from the banded amphibolites display an island arc signature, with three differentiation trends, evolving from an early tholeiitic to a more differentiated low-K calc-alkaline trend (similar to the early Pan-African juvenile terranes of Sahara, Liégeois *et al.*, 1996). The late Neoproterozoic age of the Drăgşan Group is well documented with isotopic ages by Liégeois *et al.* (1996). Augen gneiss protoliths (granite or rhyolite) were emplaced at 777±3 Ma (zircon U-Pb data), which is a minimum age of emplacement and close to the T_{DM} model ages around 800 Ma of the amphibolites (Sm-Nd isochron age of 835 ± 200 Ma). An oceanic origin of both black and white layers of the banded amphibolites, with no evidence for continental crust contamination, is indicated by very low Sr_i ratios (0.7007-0.7019) and strong positive ε_{Nd} (8.24-9.79). Later acretion of these oceanic rocks to a continent is marked by the emplacement of the augen gneiss protolith as high-K calc-alkaline granitoids, in the genesis of which continental crust participated ($Sr_i = 0.7045$, $\varepsilon_{Nd} = -5.39$).

The age of the regional kyanite-staurolite zone metamorphism and of late to post-kinematic granitoids emplaced in Drăgşan Group are not precisely known. An Rb-Sr errorchron for 8 banded amphibolite samples at 434 ± 130 Ma points to partial resetting of the system during Variscan tectonics (Liégeois *et al.*, 1996). Plateau 300 Ma hornblende and 296 Ma muscovite Ar-Ar ages (Dallmeyer *et al.*, 1998), as well as K-Ar ages in the same range (Grünenfelder *et al.*, 1983), confirm late Carboniferous tectonothermal activity. A Pan-African metamorphism for the Drăgşan Group is not yet demonstrated, but could be represented by the coronitic garnet-bearing amphibolites, suggesting an eclogite facies stage.

Post-kinematic granitoid plutons emplaced in the Drăgşan Group comprises two contrasting types represented by the Retezat and Parâng plutons. The Retezat pluton contains mostly leucocratic granodiorite, tonalite and granite, with up to 3% biotite and 1% epidote, but usually with 2-5% muscovite; it is unique in the South Carpathian basement for its primary epidote. The strip of hornblende biotite diorite developed along its western border represents a distinct intrusion. The Retezat geochemistry (Berza *et al.*, 1994a; Berza *et al.*, 1997; Duchesne, 1997; Berza & Tatu, 2002) points to a medium- to high-K series, with enrichment in light REE, absence of Eu anomaly and negative anomalies for Nb, P and Ti. The Parâng pluton contains mesocratic hornblende biotite granodiorite and biotite granite, similar to several other plutons intruding the Drăgşan Group. The lithological types define a high-K series, with enriched LREE and negative Eu anomaly, as well as negative Ba, Nb, P and Ti anomalies (Savu *et al.*, 1973a, b, 1976; Duchesne, 1997; Berza & Tatu, 2002).

Both Retezat pluton and Parâng-type plutons represent late-stage intrusions in respect to the medium-grade metamorphism of their host Drăgşan Group rocks. The contrast between the epidote-bearing granitoids (Retezat) and the hornblende-bearing granitoids (Parâng) suggests an important difference in the emplacement level of the two plutons, deeper for Retezat and shallower for Parâng. As K-Ar ages for these plutons range between 320 and 150 Ma (Grünenfelder *et al*., 1983), show partial Alpine resetting, and, in the absence of other isotopic data, it is not possible currently to precise the age of emplacement of the plutons intruding the Drăgşan Group. If the Ar-Ar plateau ages around 300 Ma of these amphibolites and mica gneisses (Dallmeyer *et al*., 1998) date the kyanite-staurolite regional metamorphism of the Drăgşan Group, as both Retezat and Parâng-type plutons postdate it, a post-Carboniferous emplacement is required. The Retezat pluton is overlain by Permian conglomerates (Pavelescu, 1953), so the time bracket for their emplacement could be narrow.

Drăgşan Group is also cross-cut by dykes of porphyritic rocks, but much less abundant and more acid (only rhyolite) than the Motru Dyke Swarm intruding the Lainici-Păiuş Group. No age data are available for these rhyolites, but they postdate the Retezat and Parâng-type plutons and do not cross-cut the Permo-Mesozoic cover.

4.3. The Tişoviţa-Iuţi mafic-ultramafic complex

Four major occurrences of pre-Alpine ophiolites outcrop in the Alpine belt along the western margin of Moesia, including the Tişoviţa-Iuţi complex (Romania) and Deli Jovan (Serbia) in the South Carpathians and two massifs in the Balkans (or Stara Planina): Zaglavac and Tcherni Vrah (Bulgaria). Geological and structural evidence indicates that both Tişoviţa-Iuţi and Deli Iovan complexes were parts of the same massif, located in the basement of an Upper Danubian Nappe and dismembered by Oligocene dextral translations along the Cerna-Timok Fault. Because it is uncertain if the continuation of the Danubian and Getic-Supragetic Nappes is developed in the western Stara Planina from the Bulgarian Balkans (Berza, 2000), it is not yet proven that all four occurrences belong to the same ophiolitic massif, as previously suggested (Haydoutov, 1989; Haydoutov & Yanev, 1997; Savov *et al*., 2001). If the metamorphic basement in the Stara Planina represents the continuation of the Danubian units from the South Carpathians (e.g. Săndulescu, 1984; Kräutner, 1996; Kräutner & Krstic, 2003), then the Zaglavac and Tcherni Vrah occurrences belong to the same ophiolitic massif as Tişoviţa-Iuţi and Deli Jovan. Alternatively, they might represent ophiolitic successions located within the Getic-Supragetic nappe complex from the Balkans. In the Tcherny Vrach massif cumulate, sheeted dykes and pillow lava units were recognized (Haydoutov 1989; Haydoutov & Yanev 1997) and the U-Pb zircon age of the Tcherny Vrach gabbro is 563±5 Ma (von Quadt *et al*., 1997).

Field relations prove that the Iuţi gabbro is older than Late Carboniferous, being overlain by Westphalian to Early Permian continental successions (Stănoiu & Stan, 1986). The Baicu conglomerate in the Upper Danubian basement reworks gabbroic clasts identical to Iuţi gabbro, along with foliated gabbros as those from Plavişeviţa, serpentinites, amphibolites and granites (Iancu *et al*., 1990). This indicates that conglomerates were sourced by the Tişoviţa-Iuţi complex. However, because the evidence for dating is not conclusive, the conglomerates are ascribed to the Ordovician or the Carboniferous (see Iancu *et al*., 2005). Questionable evidence for a latest Neoproterozoic age comes from the Deli Jovan Massif in Serbia, which southward is in tectonic contact with the Duboćane Formation (Kräutner & Krstic, 2003) supposed to overstep the massif and containing marbles in which Lower Cambrian Archaeocyathids were found (Kalenić, 1966). Similar marbles are known also eastward of Tişoviţa-Iuţi complex in the controversial Corbu tectonic zone (Mărunţiu & Seghedi, 1983a, b; Stan, 1984, 1985; Dinică, 1989; Kräutner & Krstic, 2003), but no fossils were found there.

The Tişoviţa-Iuţi (-Glavica, across the Danube, in Serbia) mafic-ultramafic complex represents a plutonic sequence, well preserved in the Upper Danubian basement, tectonically sandwiched and partly dismembered in a pre-Late Carboniferous nappe complex (Mărunţiu, 1984, 1987, Mărunţiu *et al*., 1997; Iancu *et al*, 2005). Various parts of the ophiolite occur in a low-angle tectonic pile, as well as along a N-S trending vertical shear zone related to dextral strike-slip (Mărunţiu *et al*., 1997), where the ophiolite is tectonically juxtaposed onto highly mylonitised medium-grade metasediments ascribed to the Corbu or Neamţu sequences (Gunnesch & Gunnesch, 1978; Mărunţiu & Seghedi, 1983a, b; Stan, 1984, 1985; Dinică, 1989; Kräutner & Krstic, 2003), resembling the Lainici-Păiuş Group. Both the ophiolite and the overthrusted gneissic unit are covered by sedimentary successions of the Variscan molasse (starting with Upper Carboniferous, Westphalian-Stephanian clastics with coal measures and followed by Permian red terrigeneous clastics and rhyolitic volcanics and volcaniclastic rocks; Stănoiu & Stan, 1986).

A reconstructed schematic lithologic log of the ophiolite is illustrated in Fig. 11 (Mărunţiu *et a*l., 1997). The Tişoviţa-Iuţi complex includes two main units which show the ophiolite igneous stratigraphy: a lower unit with upper mantle lithologies and an upper association of plutonic cumulates. Plutonic and effusive rocks in the eastern, Plavişeviţa shear zone, are interpreted as the upper part of the ophiolite sequence.

The mantle peridotite unit, forming the lower part of the ophiolite igneous stratigraphy, consists mainly of harzburgite with tectonite structure and subordinate dunite, hosting small podiform chromitites. The cumulate sequence includes an ultramafic unit, a transitional zone and a layered mafic unit (Mărunţiu *et al*., 1997). Ultramafic cumulates are dominated by layered dunites, which show lens-shaped bodies of plagioclase-bearing dunite, troctolite, olivine gabbro and gabbro (Fig. 11). The transitional zone (maximum 500 m thick) consists of alternating cumulates, mafic (troctolite, olivine gabbro, gabbro) and ultramafic (dunite, plagioclase dunite, wehrlite, clinopyroxenite). At the top of the cumulate unit, mafic cumulates include gabbro, olivine gabbro and troctolite, forming a rhythmic layered sequence which is cross-cut by isotropic olivine gabbro (Fig. 11). The

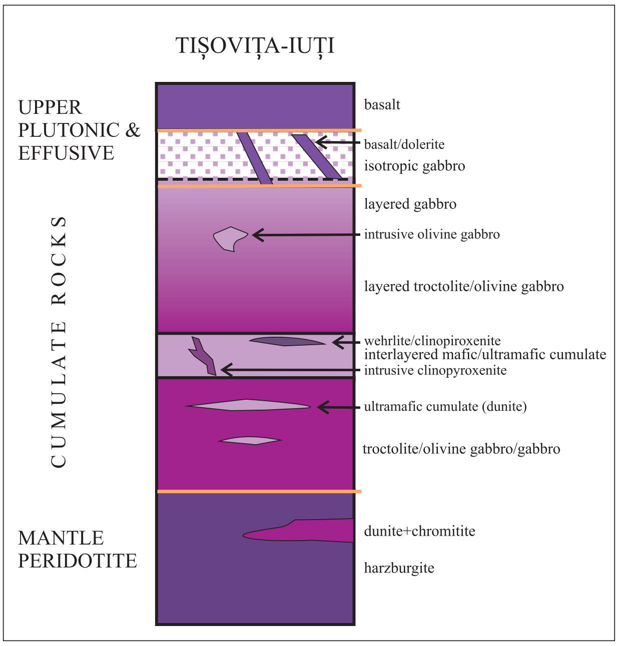

Fig. 11. Schematic lithologic log in the Tişoviţa-Iuţi ophiolite (not to scale). After Mărunţiu *et al*. (1997).

entire cumulate pile is characterised by rhythmical layering, marked by alternation of isomodal, modally graded and grain-size graded layers. The adcumulus texture is most widespread, with mesocumulus texture shown only by some mafic cumulates.

The mylonitic gabbro in the eastern shear zone (Plavişeviţa zone) represents a highly tectonized mixture of various lithologies (Bercia & Bercia, 1975; Mărunţiu, 1984, 1987). This mylonite zone is characterized by association of cumulate and isotrope gabbro, cross-cut by high level intrusives (dykes of dolerite and porphyritic basalt with preserved chilled margins, along with dykes of plagiogranite). The association suggests remnants of the upper part of the plutonic complex. Large tectonic slabs of homogeneous, green mylonite of basaltic composition suggest that relics of the upper effusive sequence may also exist.

Mineral and whole-rock geochemistry indicate the residual features of the mantle peridotite, suggesting derivation by at least 20% of partial melting from a slightly depleted lherzolite source. The composition of cumulates indicates a high-Ti ophiolite of MORB or transitional-type. The mineralogical and chemical features of the Tişoviţa-Iuţi complex are consistent with those of ophiolite complexes formed in ocean basins associated with arc systems (spreading back-arc basins). The great total thickness of the cumulate pile, the cyclic organization of cumulates, the occurrence of multiple intrusions cutting the layered sequence, limited cryptic variation of cumulus phases and the dominant adcumulus texture indicative a high heat flow, are all consistent with the formation of the Tişoviţa-Iuţi complex at fast- to intermediate-spreading centres (Mărunţiu *et al.*, 1997). The ophiolitic complex, along with discontinuous oceanic-type metasediments, is interpreted to mark a late Neoproterozoic or Early Palaeozoic plate boundary (Iancu *et al.*, 1997, 2005).

5. Discussion and conclusions

The largest part of the European continent south-west of the TESZ consists of peri-Gondwanan terranes which, depending on the time of their rifting from Gondwana, are known as Avalonia and Armorica Terrane Assemblage (ATA) (Fig. 1; Franke, 2000; Tait *et al.*, 2000; Winchester *et al.*, 2002; Winchester, 2003). Along the western margin of the East European Craton, there is a narrow zone of displaced terranes, amalgamated to the Baltica margin and interpreted either as Gondwana- or as Baltica-derived (Cocks *et al.*, 1997; Cocks & Fortey, 1998; Pharaoh, 1999; Unrug *et al.*, 1999; Belka *et al.*, 2000; Żelaźniewicz *et al.*, 2001; Winchester *et al.*, 2002, 2004, 2005; Winchester, 2003; Oczlon *et al.*, 2005).

Avalonia-type terranes, originating from ca. 1.3-1.0 Ga juvenile crust within the Panthalassa-type ocean surrounding Rodinia, were accreted to northern Gondwana by 650 Ma (Murphy & Nance, 1991; Nance *et al.*, 1991; Nance & Murphy, 1996). Their distinctive feature is the presence of a "Rondonian" event (about 1.5 Ga), characteristic for the Ganderian basement of Avalonia, and the occurrence of the distinctive middle and late Ordovician "Celtic" faunas (Cocks *et al.*, 1997; Cocks & Fortey, 1998). Following accretion, Avalonian terranes were subsequently detached and eventually docked to the western margin of Baltica by latest Ordovician. Due to oblique collision with Avalonia, a zone of dextral shear/transform developed along the western Baltica margin, and proximal Baltica terranes were detached and displaced along the TESZ (Winchester *et al.*, 2005). This model is consistent with the identification of Avalonian terranes in the East Carpathians (Munteanu & Tatu, 2003).

The ATA is characterized by distinctive Cadomian basement, formed along the West African margin by recycling Palaeoproterozoic and Archaean West African crust (2-3 Ga) (Nance *et al.*, 2004), mixed Silurian-Devonian fauna, presence of Vendian and late Ordovician glaciogenic sediments (Cocks *et al.*, 1997; Cocks & Fortey, 1998) and late Carboniferous accretion to Laurussia.

However, there is evidence that the Late Neoproterozoic records active margins and accretion not only in peri-Gondwanan environments, like Avalonia (Murphy & Nance, 1991), but also along the Timanian (Roberts & Siedlecka, 2002; Scarrow *et al.*, 2001), Uralian (Puchkov, 1997) and Scythian margins of Baltica (Muratov *et al.*, 1968; Plakhotnyy, 1969; Neaga & Moroz, 1987; Khain, 1994; Seghedi *et al.*, 2003). Consequently, the Neoproterozoic development is no longer critical for evaluating Gondwanan terrane provenance (Winchester *et al.*, 2004, 2005; Oczlon *et al.*, 2005).

Geological, geochronological and geochemical evidence indicate that parts of the Neoproterozoic orogen developed at the northern active margin of Gondwana are preserved in the Alpine belt of the South Carpathians and Balkans. They were interpreted to represent an island arc accreted to the continental margin following subduction, and a back-arc basin which was opened on the continental margin above the subduction zone. There is enough geological and geochemical evidence that the Danubian terranes show Pan-African affinities with the Tuareg shield of West Africa, and on this basis a north Gondwanan provenance is likely (Liégeois *et al.*, 1996). Their lithological, geochemical and geochronological features suggest a tectonic affinity with the Altîn Tepe terrane from East Moesia.

For Central Dobrogea, K-Ar ages and petrological data (Giuşcă *et al.*, 1967; Crowley *et al.*, 2000) suggest a Neoproterozoic volcanic arc affinity. Geochronological data of 696-643 Ma can be interpreted as the age of accretion, coeval with the accretion of Avalonian terranes to Gondwana margin. ***Evidence for subduction in the form of calc-alkalic plutonic and volcanic rocks is missing in the*** Altîn Tepe terrane, but this shows a very limited outcrop area. The thick, deep sea turbidite fan of Histria Formation has accumulated in a foreland basin setting, in a basement floored by continental crust, part of which was the juvenile Altîn Tepe arc. Detritus into the basin was supplied by a source of Grenvillian and Archaean ages, which could suggest an Avalonia-type source, especially in the absence of Paleoproterozoic detrital zircons.

Neoproterozoic deformation and metamorphism affected the Archean and Palaeoproterozoic crust of South Dobrogean, involved in a Neoproterozoic thrust system as indicated by younger K-Ar ages (Giuşcă *et al.*, 1967). This thrust system appears to involve the volcano-sedimentary pile accumulated in a Neoproterozoic rifted basin (Cocoşu Group). Available geochronological data suggest that rifting occurred prior to the latest Vendian-Early Cambrian, when the basaltic suite was deformed in very low-grade metamorphic conditions (Giuşcă *et al.*, 1967). Although well constrained stratigraphic ages are lacking, one possible interpretation is that this rift might be part of/or coeval with the Volhyn rift system within Baltica, in connection to its separation from Rodinia.

The older basement, ascribed to the Archean and Palaeoproterozoic, is the most enigmatic. It is known only on a restricted area in the NE corner of South Dobrogea and was pierced by four close boreholes. The magnetic anomaly corresponding to the magnetite–bearing quartzites dies out to the NE and SW, while still continuing with a smaller, en-echelon anomaly, to the NE of the Capidava-Ovidiu Fault (Romanescu *et al.*, 1972; Visarion *et al.*, 1988). If the oldest Moesian crust correlates to the Ukrainian shield, as suggested by lithology (Giuşcă *et al.*, 1967, 1976; Kräutner *et al.*, 1988b), then it might represent a small sliver of Baltica, detached and displaced on strike-slip faults along the TESZ (Oczlon *et al.*, 2005). However, clear evidence for provenance of this older Proterozoic basement is still missing and needs to be solved by accurate geochronology.

A comparison of the metamorphic basement from West Moesia, Central Dobrogea and the Danubian Nappes does not provide sufficient evidence to answer the question of terrane provenance, due to very limited information on the former. Therefore the question of the relation of Neoproterozoic metamorphic rocks from West Moesia with the margin of Gondwana or Baltica is unresolved.

A possible correlation is suggested by the comparison of the Palaeozoic formations ovelying Danubian and West Moesian basement. For details regarding the Palaeozoic formations an overview is given by Iancu *et al.* (this volume) for the South Carpathians, and Seghedi *et al.* (this volume) for the Moesian Platform. The Cambrian is not known in the Danubian basement of the South Carpathians. Archaeocyathid-bearing limestones were described west of the Deli Iovan massif in Serbia (Kalenić, 1966), indicating that this area belonged to the north Gondwanan margin. However, these limestones seem to overly Getic type basement (Berza, 2000). During the Palaeozoic and Mesozoic, the external Danubian, represented by the Lower Danubian nappes, was the closest area to the West Moesian realm (Săndulescu, 1984). The Danubian Valea Izvorului Formation (Upper Ordovician-Lower Silurian) shows a clastic facies correlatable with that of West Moesia, and the Lower Carboniferous Ideg limestones from the internal Danubian (upper Danubian nappes), more distal from West Moesian margin, suggest that the carbonate platform extended over a larger area, including the Moesian and the Danubian realms, as already mentioned by Patrulius & Neagu (1963). The paralic clastic facies and coal measures of the Late Carbonifeorus are again typical for both Moesian and Danubian realms, as well as for the Getic-Supragetic domain. This facies is also widely developed on the Scythian margin of the East European Platform (Neaga & Moroz, 1987; Seghedi *et al.*, 2003) and in the Zonguldak terrane of the Istanbul Zone (Kaya, 1980; Sengör, 1995; Kozur & Stampfli, 2000).

The similarity of the Palaeozoic facies suggests that the Danubian terrane was possibly accreted to Moesia at least before the Ordovician, as proposed in the palaeogeographic model of Winchester *et al.* (2005). The model is based on the Pan-African affinity of the Danubian basement and absence of diamictites in the overlying Ordovician sediments, suggesting that the Danubian Terranes might belong to Avalonia, similar to the Istanbul terrane from the Western Pontides. This model explains the detachment of the tip of Avalonia upon accretion to the Bruno-Silesian Promontory, represented on the western margin of Baltica by terranes now forming Upper Silesia, Malopolska and Moesia. These terranes are supposed to have been already accreted to Baltica at the end of the Cambrian. After detachment, the tips of Avalonia were displaced along strike-slip faults and, in the case of the Danubian terranes, accreted along the southern margin of the new continent of Laurussia (the result of collision between Baltica and Laurentia).

Various lines of evidence indicate that parts of the Neoproterozoic orogen are preserved in the basement of the South Carpathian Danubian Nappes and of the Moesian Platform, although there is no unequivocal proof that they were part of the same belt. There is reliable evidence that the Danubian basement includes an island arc accreted to the continental margin following subduction, and a back-arc basin which was opened on the continental margin above the subduction zone developed at the northern active margin of Gondwana. As for the Neoproterozoic Moesian basement, further geochronological evidence is required to establish the affinity of the volcanic arc and the foreland basin from Central Dobrogea, although a Grenvillian source for the latter suggests Avalonian affinity.

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