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GEOLOGY AND GEOMORPHOLOGY DOBROGEA AND DANUBE DELTA

FIELD TRIP GUIDEBOOK

Editura GeoEcoMar
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DOBROGEA AND DANUBE DELTA: GEOLOGY AND GEOMORPHOLOGY

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1. GEOLOGY OF THE DOBROGEA HIGHLAND AREA

The Dobrogea Region represents the south-easternmost part of the Romanian Carpathian foreland, bordered by the Danube towards W and N, and by the Black Sea to the E. This highland region displays three distinct geotectonic units (Northern, Central and Southern Dobrogea), which are separated by major crustal faults, which are from N to S, Peceneaga-Camena and Capidava-Ovidiu (Fig. 1). North Dobrogea is the exposed part of the North Dobrogea orogen, while Central and South Dobrogea are tectonic blocks of the Moesian Platform (Săndulescu *et al*., 1978).

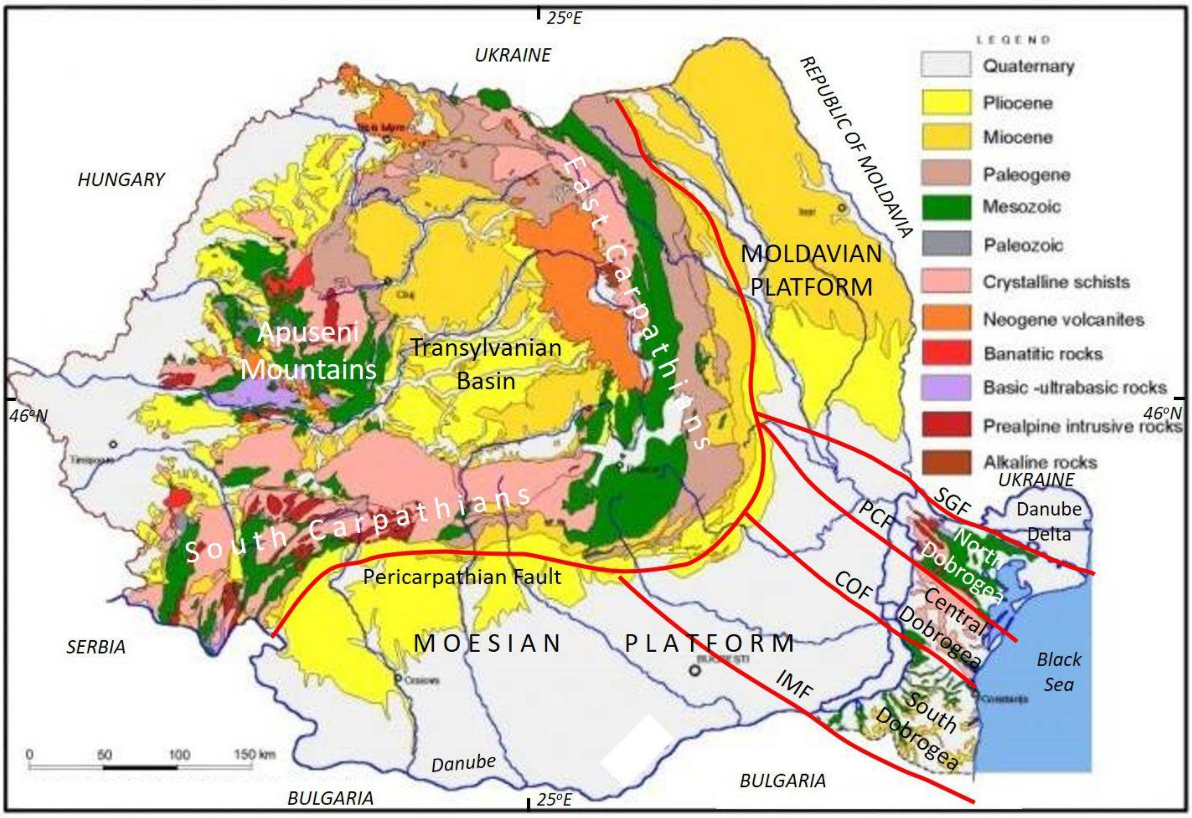


Fig. 1. Location of Dobrogea in the Carpathian foreland, on the geological map of Romania (modified after a symplified version of the geological map of Romania, scale 1:1.000.000 of Săndulescu *et al*., 1978). SGF, Sfântu Gheorghe Fault; PCF, Peceneaga-Camena Fault; COF, Capidava-Ovidiu Fault; IMF, Intramoesian Fault.

North Dobrogea

The Cimmerian (Early Alpine) orogenic belt of North Dobrogea is bound by two crustal faults, the northern Sfântu Gheorghe Fault and the southern Peceneaga-Camena Fault (Dumitrescu & Săndulescu, 1970; Săndulescu, 1984; Săndulescu & Visarion, 2000) (Fig. 1). The pre-Triassic basement of North Dobrogea, a remnant of the European Hercynian Chain, contains Paleozoic slate belts, calc-alkaline volcaniclastics and granitoids (Seghedi, 2001) (Fig. 2). Early Triassic-middle Jurassic sedimentary deposits include continental clastics, limestone turbidites, limestones and marls and terrigenous turbidites, followed by late Jurassic carbonate platform sediments (Mutihac, 1964; Mirăuţă *et al*., 1968; Patrulius *et al*., 1974; Grădinaru, 1981, 1984, 1988; Baltres, 1993). A 700-800 m thick pile of upper Cretaceous shelf carbonate and carbonate-detrital succession represents the post-tectonic cover of the Cimmerian belt (Mirăuţă & Mirăuţă, 1964; Patrulius *et al*., 1974).

Two types of Hercynian Paleozoic formations can be distinguished in North Dobrogea (Seghedi, in Baltres *et al*., 1992), in tectonic contact with each other as indicated by geophysical evidence (Fig. 2). This contact, supposed to represent a strike-slip fault, is concealed by the overlying Triassic deposits. Correlation of outcrop evidence with deep and shallow borehole information shows that the Paleozoic formations of the two basement types are different in what concerns their lithology, facies association and fauna type, suggesting that they have formed in distinct tectonic settings and on different type of crust (Seghedi, 1999).

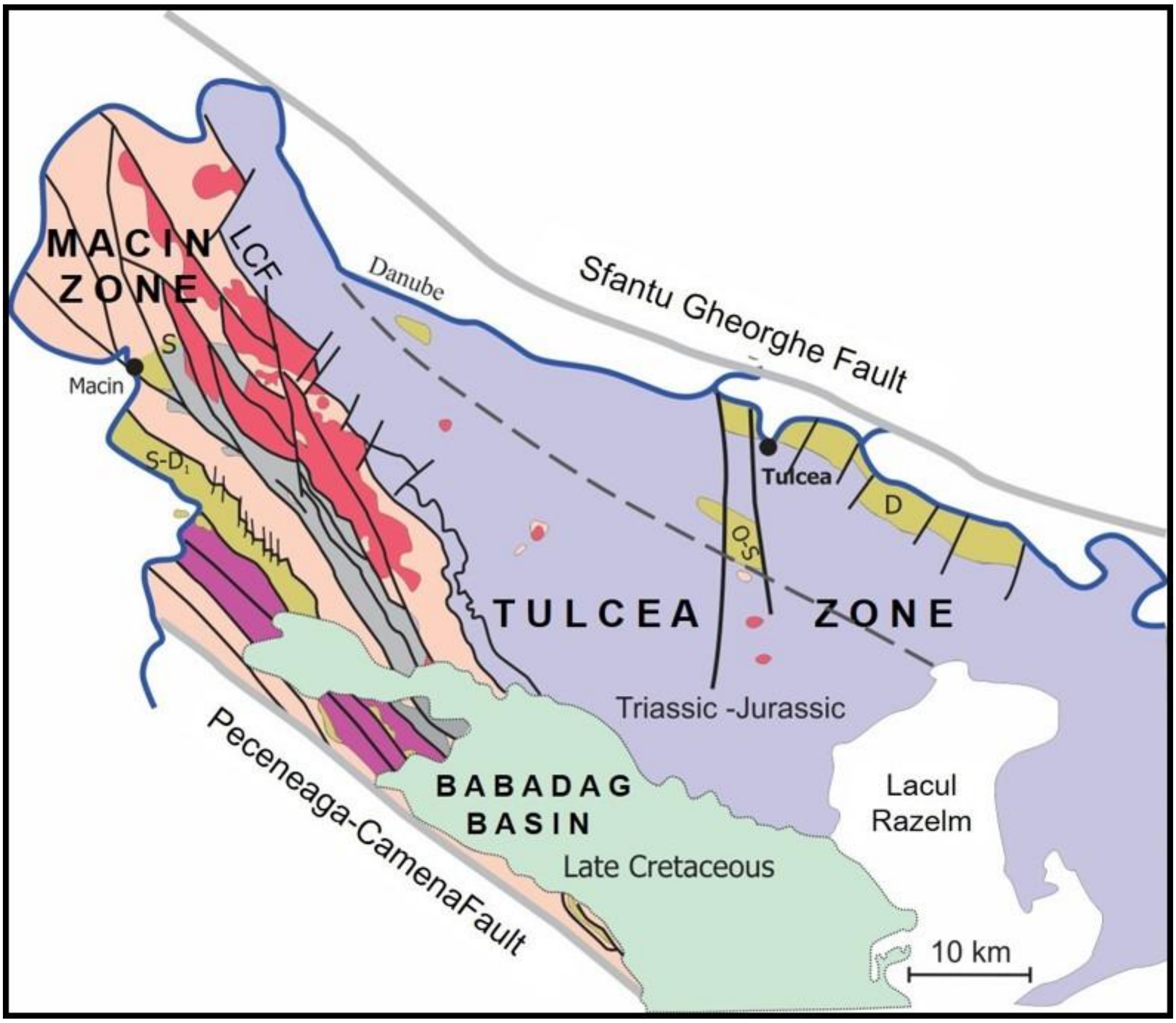


Fig. 2. Quaternary subcrop map of North Dobrogea, showing the development of the Paleozoic formations (modified after Seghedi, 1999). Triassic-Jurassic formations not separated. Red, Hercynian granitoids; pink, metamorphic rocks (amphibolite and greenschist facies); grey, late Carboniferous-early Permian Continental deposits (Carapelit Formation); purple, late Permian alkaline granitoids and rhyolites; S, Silurian; S-D_1, Silurian-lower Devonian; O-S, Ordovician-Silurian; D, Devonian. Dashed line in Tulcea zone represents the contact between the two types of Paleozoic successions.

The lithology of the succession accumulated on sialic basement, exposed largely in the western part of North Dobrogea (Măcin type), shows an upward shallowing trend, from deeper marine in Silurian to shallow marine, neritic in the lower Devonian, to reefal in the middle Devonian and finally to continental in the late Paleozoic (Seghedi, 1985; Seghedi & Oaie, 1995) (Fig. 3). The transition from marine to continental sedimentation is marked by a stratigraphic gap corresponding at least to the lower Carboniferous (Seghedi & Oaie, 1986). Both the Paleozoic deposits and the metamorphic basement rocks are intruded by Paleozoic magmatic suites. The Paleozoic successions from the northern part of North Dobrogea (Tulcea type) reflect deep marine sedimentation on oceanic crust and possibly in a trench, from Ordovician to lower Carboniferous.

The early Alpine geology of the Northern Dobrogea starts in the upper Permian, when alkaline magmatic rocks were emplaced as dyke swarms and subvolcanic bodies into the Hercynian basement. Consequently, thin subvertical dykes of the basalt-trachyte association occur all along the northern border of the North Dobrogea (Seghedi *et al*., 1994; Seghedi, 2001). During the Triassic (the Spathian-Anisian interval), massive and pillow basalts, basaltic volcaniclastics, and rhyolithic volcaniclastics and lava flows were emplaced (Savu *et al*., 1988) (Fig. 4).

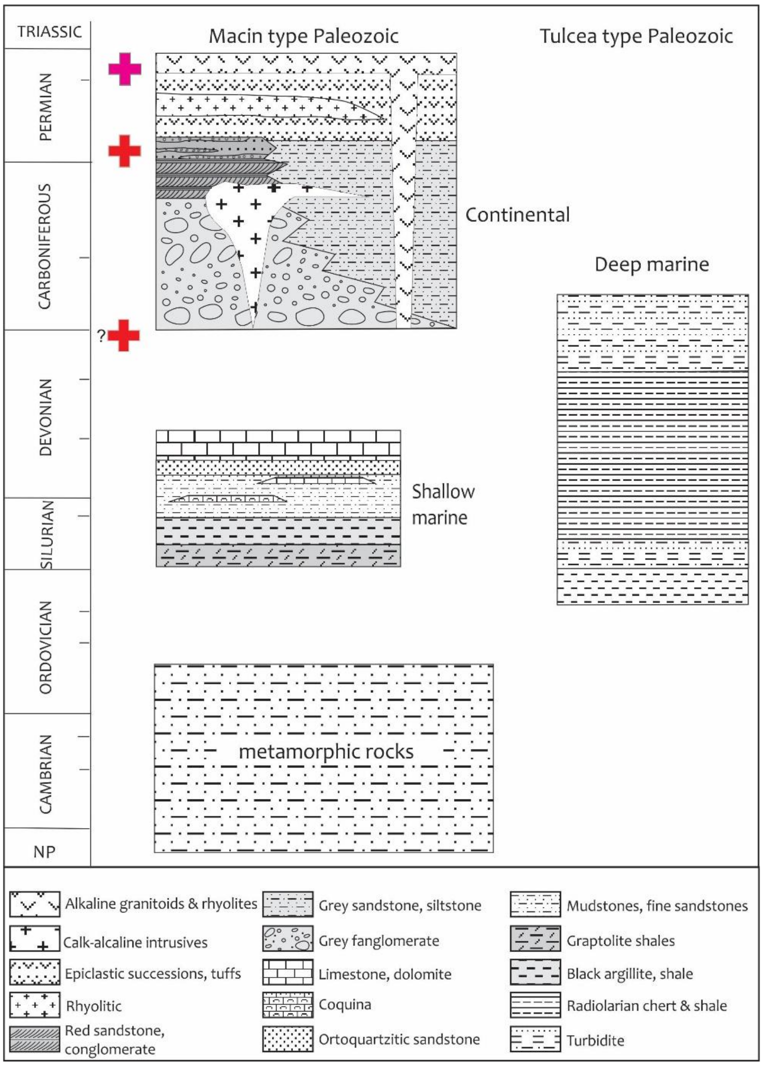


Fig. 3. Lithological chart for the Paleozoic deposits of North Dobrogea (modified after Seghedi, 2012).

The Mesozoic formations are exposed mainly in the Tulcea zone, the area east of the Luncaviţa-Consul Fault. Two main successions have been recognized, basinal and shallow marine (Grădinaru, 1984; Baltres, 1992) (Fig. 5). The main Triassic facies includes continental clastics in the lowermost Triassic, grading upward to marine deposits. The lower Triassic (Spathian) consists of limestone turbidites (Somova Formation), the middle Triassic (Anisian-Ladinian) is made of cherty and nodular limestones, interbedded with basalt flows, the Carnian (late Triassic) is marly and the Norian to Bathonian (late Triassic - middle Jurassic) is turbiditic (Baltres, 1992). Based on ammonoid zones, a continuous deposition across the Triassic-Jurassic boundary, at least in some places, is indicated (Grădinaru, 1995).

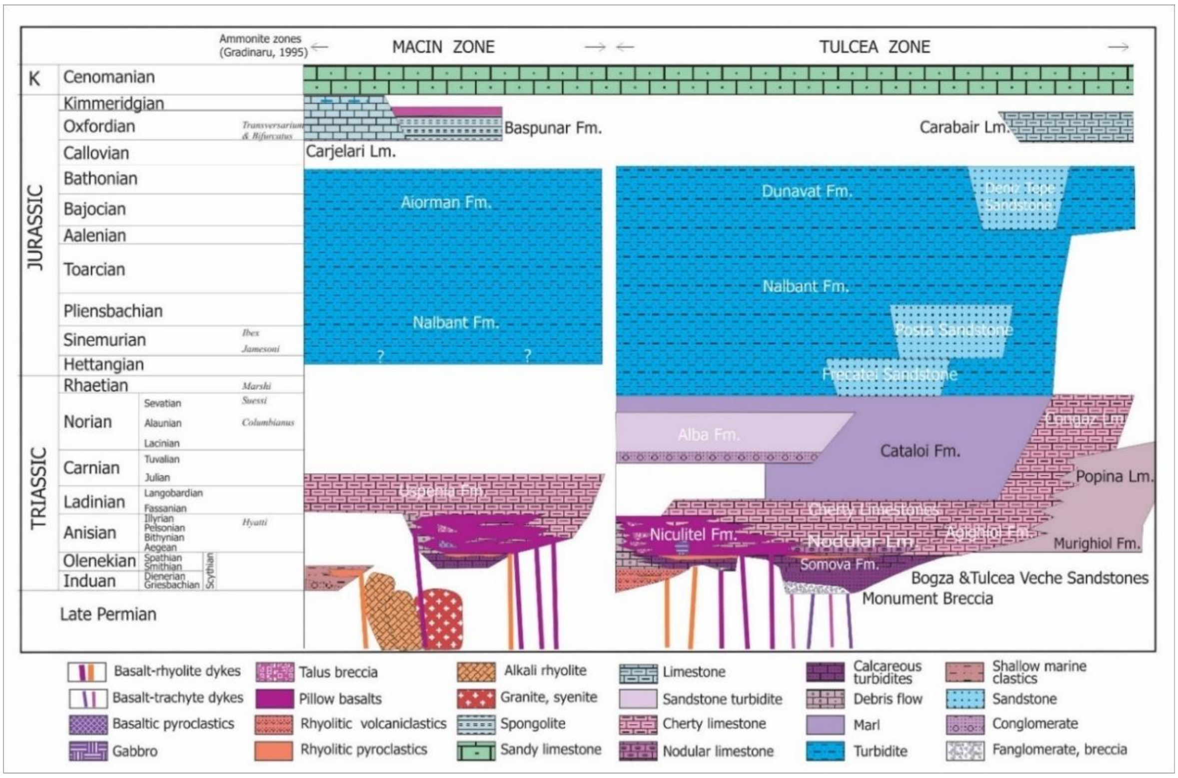


Fig. 4. Lithological chart for the Triassic-Jurassic deposits of Tulcea zone (after Mirăuţă, 1982; Grădinaru, 1984, 1995; Baltres, 1993; Mirăuţă *et al*., 1993; Antonescu, unpublished data 1978). Timescale of Gradstein & Ogg (1996).

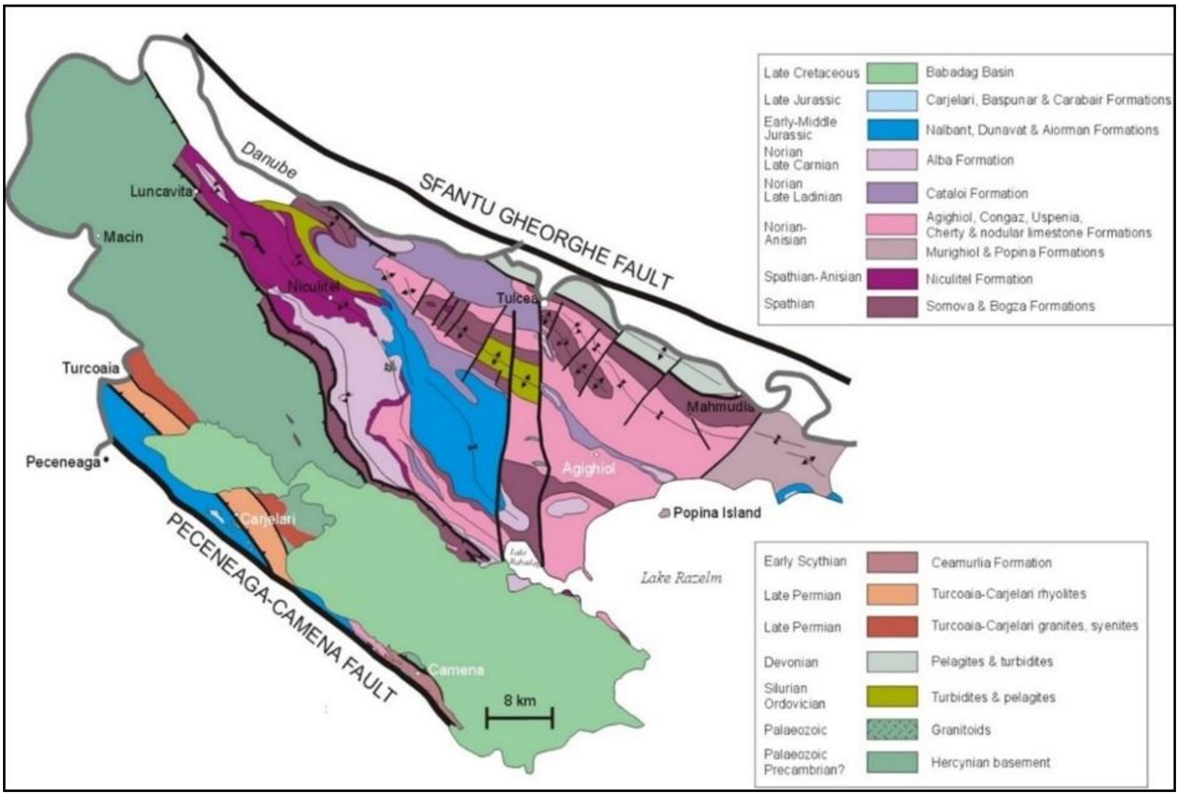


Fig. 5. Quaternary subcrop map showing the distribution of the Permo-Mesozoic formations of North Dobrogea (modified after Seghedi, 2001). Except for the Late Permian extension-related magmatic rocks, the Paleozoic formations from the Măcin zone are figured together as Hercynian basement.

The post-tectonic cover of the North Dobrogea orogen is represented by the Late Cretaceous deposits („the Babadag basin”) that unconformably overlay various Paleozoic and Triassic-Jurassic rocks (Figs. 2, 5). The break in sedimentation corresponds to the early and middle Cretaceous. Scarce remnants of a paleo-weathering crust, preserved on top of Paleozoic and Triassic formations, or below the Cenomanian calcarenites, were ascribed to the Aptian (Rădan, in Seghedi *et al*., 1988; Rădan, 1989). The late Cretaceous succession, up to 750-800 m thick, includes mainly Cenomanian to Coniacian shallow marine calcareous or calcareous-detrital sediments (Mirăuţă & Mirăuţă, 1964; Szász, 1985; Szász & Ion, 1988). Their biostratigraphy is based on ammonites, inoceramids, foraminifera and nannoplankton (Szász, 1985; Szász & Ion, 1988; Ion & Szász, 1994; Ion *et al*., 1997). In its south-eastern part, the late Cretaceous succession seals the Peceneaga-Camena Fault (Fig. 5), directly overlaying the Ediacaran basement of Central Dobrogea.

Central Dobrogea

Central Dobrogea is situated between the Peceneaga-Camena and the Capidava-Ovidiu Faults (Fig. 1) and represents the elevated part of the Western Moesian Platform. Notably, the Moesian Platform is the tectonic unit which covers almost the entire southern Romania (except its south-westernmost part), and it is divided in the West and East Moesia by the Intramoesian Fault (Fig. 1). The Moesian Platform is separated from the Romanian Carpathian chain by the Pericarpathian Fault, oriented W-E and dipping N (Săndulescu, 1984).

West Moesia is characterized by a Variscan cratonization of the basement. East Moesia (which includes the Central and Southern Dobrogea) recorded a Neoproterozoic, Ediacaran (“Baikalian-Cadomian”) and Caledonian (“Ardennian”) cratonization (Murgeanu & Patrulius, 1973; Oaie *et al*., 2005). Within Central Dobrogea, the Ediacaran folded basement (anchimetamorphic turbidites), crops out on large areas, except a narrow zone next to the Peceneaga-Camena Fault, where a NW trending antiformal fold exposes plagiogneisses and amphibolites of the Altin Tepe Group, with a Late Neoproterozoic (Cadomian) amphibolite facies metamorphism (Giuşcă *et al*., 1967), dated recently on monazites at 570±13 Ma (Săbău & Negulescu, 2017, unpublished presentation). The relationships of the Altin Tepe Group rocks and Ediacaran turbidites is tectonic, marked by greenschist facies mylonites (Mureşan, 1971) (Fig. 6).

The Ediacaran turbidites display a dominant green coloration (dark green in quarries), but locally they are bluish or purple. Because of their dominant green color, the turbiditic succession was initially described as “the green rocks”, and later “the Green Schist series”. However, the turbidites do not display the low-grade foliation with metamorphic phyllosilicates, typical for greenschists facies metamorphic rocks, but show instead a slaty cleavage, penetrative only in fine grained lithologies. As the name “Green Schists” does not reflect the real metamorphic grade of rocks but rather their color, the name of Histria Formation was proposed for the turbiditic succession in order to avoid confusion (Seghedi & Oaie, 1994, 1995).

Based on detailed sedimentological studies and mapping, the stratigraphy of Histria Formation includes three members, with lower and upper mid-fan turbidites, separated by a middle member of distal turbidites (Seghedi & Oaie, 1995; Oaie, 1998, 1999) (Fig. 6).

The lower, sandstone dominated Beidaud Member, up to 2500 m thick, consists of channelized, coarse grained middle fan turbidites, associated with depositional lobe deposits of the lower fan. The middle, Haidar Member, of maximum 500 m thickness, consists of purple and green siltstones and mudstones, forming amalgamated Tcde, Tde and Tce Bouma divisions. In this member, thin, centimetric beds of fine-grained sandstones seldom occur. The great thickness of these fine-grained successions and their facies associations suggest distal turbidites (Seghedi & Oaie, 1995; Oaie, 1999), representing lower fan lobe fringe and basin plain sediments. A probable explanation for this lithological change is a rise in sea level (Oaie *et al.*, 2005). The upper, Sibioara Member, is again made of coarse-grained, sandstone-dominated successions, interpreted as channelized middle fan turbidites; they are locally associated with conglomerates, representing upper fan channels. This upper member always starts with a coarse-grained, conglomerate facies of the upper fan, directly overlying the fine grained, thinly laminated, distal turbidites of the Haidar member.

Sandstone petrofacies of Histria Formation indicates fedspatho-lithic and litho-feldspathic sandstones, supplied by two major sources: a bimodal volcanic source (basalts and rhyolites) and a continental source, which yielded metamorphic rocks (Oaie, 1999; Oaie *et al*., 2005). Rounded and subrounded clasts prevail, indicating sediment recycling.

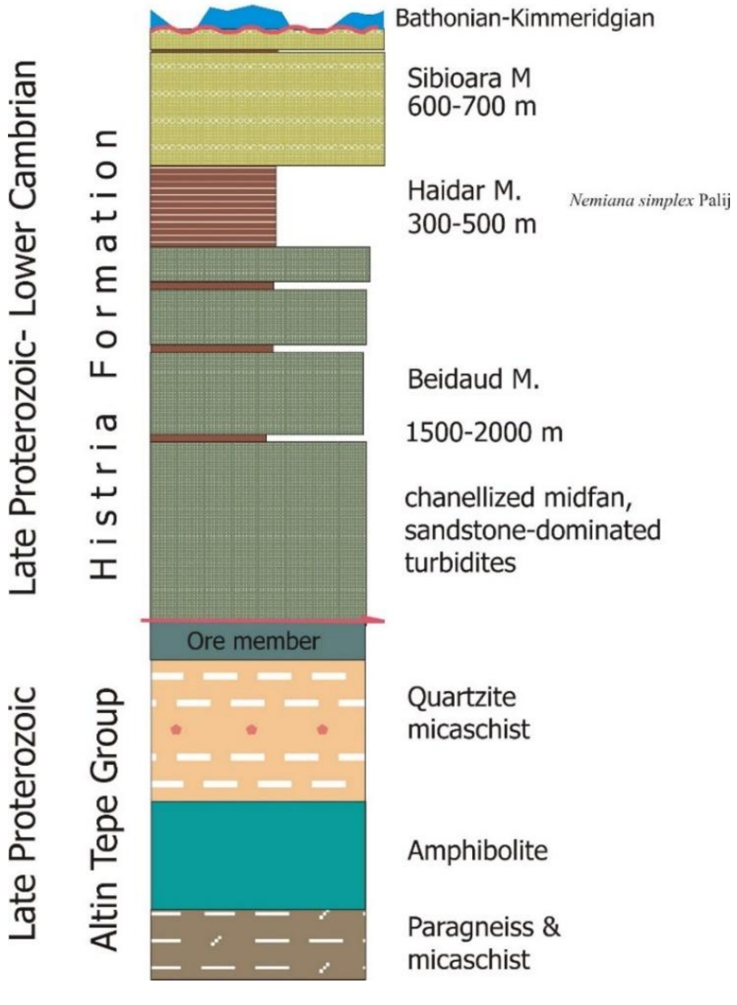


Fig. 6. Lithological chart for Central Dobrogea (after Seghedi *et al.*, 2005). The brown line separating the Altin Tepe Group from the Histria formation is a tectonic contact, a shear zone, interpreted either as a nappe contact (Mureșan, 1971; Balintoni & Balica, 2016), or as an extensional fault (Seghedi *et al.*, 1999).

Syn-diagenetic or very low-grade metamorphic minerals (pyrite, chlorite, epidote) typically form in sandstones, while detrital chlorite is still preserved in the associated siltstones and mudstones (Oaie *et al*., 2005).

Over large areas, the Histria Formation shows extremely well preserved sedimentary structures, both on bed surfaces (Mirăuţă, 1964; Jipa, 1968, 1970) or inside beds (Oaie, 1998, 1999). Scarce trace fossils include tiny burrows in the distal turbidites of the Haidar member, but also thick, irregular bioturbations filled with sandstones (Oaie, 1992). In few localities, ichnofauna is preserved on bedding surfaces of pelitic intervals (Oaie, 1992) (Fig. 7) and imprints of soft-bodies organisms have been also found (Oaie 2010; Oaie et al., 2012; Saint Martin *et al*., 2013) (Fig. 8a, b).

Fig. 7. The Casimcea nature reserve, a Natura 2000 site, hosts enigmatic biogenic traces. The first trace fossils, found on bed surfaces in fine-grained turbidites (Oaie, 1992), resemble the ichnogenous *Scalarituba*. Photo Dr. Gheorghe Oaie.

Fig. 8. Ediacaran biota from the Histria Formation.

a. *Beltanelliformis*, found by Dan Jipa and identified by Saint Martin & Saint Martin (2018). Photo Prof. Jean-Paul Saint Martin.

b. Soft bodied organism, identified as *Nemiana simplex* Palij (Oaie, 1992). Photo Amarjit Sidhu.

c. *Aspidella* trace found in Colţanii Saxanului Hill, Casimcea nature reserve. Photo Prof. Jean-Paul Saint Martin.

Other traces are 5-10 cm long, slightly elliptic imprints on the bed surface, showing concentric grooves (Fig 8c). They resemble holdfasts of leaf-type primitive organisms, used to fixate on the sea bottom (Saint Martin *et al.*, 2013), forms frequently found in the Ediacaran turbidites of Newfoundland, Canada, or Australia. These concentric traces seem to represent *Aspidella*, interpreted as an anchor or holdfast of *Charnia* (Gehling *et al*., 2000), a frond type marine organism known from the Charnwood forest in England and the Canadian Newfoundland peninsula.

Both the depositional age and the time of deformation are well constrained in the Histria Formation. Palynological assemblages indicate accumulation during the upper Precambrian-lower Cambrian (Iliescu & Mutihac, 1965). In distal turbidites, the imprint of a soft-bodied organism was recovered and identified as *Nemiana simplex*, suggesting an Ediacaran fauna (Oaie, 1992) (Fig. 7b). The youngest group of detrital zircons from Ediacaran sandstones concentrate between 622-579 Ma (Żelaźniewicz *et al*., 2009) and 633-583 Ma (Balintoni *et al.*, 2011), interpreted as depositional ages. Burial metamorphism, showing intermediate features between diagenetic textures and very low-grade metamorphism, occurred around 573 Ma (K-Ar data, Giuşcă *et al*., 1967).

The platform cover of Central Dobrogea

In outcrops, the Ediacaran basement is covered by remnants of upper Jurassic sediments of a carbonate platform (Casimcea Formation, Drăgănescu, 1976). This is mainly composed of bioconstructed reefal limestones, and calcarenites in some regions (Bărbulescu, 1974; Drăgănescu, 1975), accumulated during the upper Bathonian and Oxfordian-lower Kimmeridgian. The ”Casimcea syncline” is the largest outcrop area of the shallow marine Jurassic deposits. This platform structure, deforming only the epicratonic cover, shows highly asymmetric limbs and an eccentric axis located to the south-west (Drăgănescu *et al*., 1978, 1979).

In places, scarce remnants of a pre-Bathonian weathering crust formed on Ediacaran basement rocks are found underneath the Bathonian calcarenites and calcirudites lying locally at the base of the carbonate platform succession (Rădan, 1994). The age of this paleo-weathering crust is not precisely known, but the mineralogical record indicates that it has formed during climatic conditions favorable to intense chemical and biochemical weathering (Rădan, in Seghedi *et al*., 1999). Geological evidence indicates that this weathering crust was largely eroded before the Bathonian.

Boreholes west of the Danube have pierced the Paleozoic cover of Central Dobrogea (Iordan, 1999; Ion *et al*., 2002). This suggests that the block of Central Dobrogea had been uplifted at the end of the Variscan events and was subjected to a long period of erosion throughout the Permian-Bajocian.

The Southern Dobrogea

The *Southern Dobrogea* is located south of the Capidava-Ovidiu fault. The basement of this part of the East Moesian Platform was identified only in boreholes. It is composed of orthogneisses, showing a dominantly medium grade Orosirian metamorphism (2002 ± 22 Ma) (Săbău & Negulescu, 2016) and Neoproterozoic remobilisations. The orthogneiss is overlain by a Paleoproterozoic banded iron formation (Giușcă *et al*., 1967) and a Neoproterozoic volcano-sedimentary succession (Kräutner *et al*., 1988). The sedimentary cover consists of thick Paleozoic-Mesozoic successions (Chiriac, 1968; Iordan, 1999 and references therein).

The outcrops from the Southern Dobrogea are only composed of Cretaceous and Tertiary sediments (Chiriac, 1968). In particular the Cretaceous recorded a complete development of all the stages

(Avram *et al*., 1993; Ion *et al*., 2004; Melinte, 2006). The Berriasian and Valangian stages are represented by marine shallow water carbonates in the eastern part of the Southern Dobrogea and mainly by fluvial and continental deposits in the western part (Fig. 9). The Hauterivian-early Aptian interval is characterized by the presence of marine shallow water carbonates. A fluvial-lacustrine sedimentation took place within the middle-late Aptian (Rădan & Bratosin, 1977; Rădan & Vanghelie, 1990), and it was followed by upper Cretaceous glauconitic sandstone and chalk deposits.

The Paleogene deposits are mainly constituted of Eocene siliciclastic deposits (Ypresian in age), which contain rich large foraminiferal (especially nummulite) faunas (Bombiță, 1964). The Neogene deposits of the Southern Dobrogea, which crop out southwards of the Ovidiu-Capidava Fault, belong to the middle-upper Miocene and to the Pleistocene.

The Miocene is represented by the Paratethys stages Sarmatian, and the Pontian (*pro parte*); their deposits are mainly made by oolitic limestones, clays, sands, marls and diatomites. The middle Miocene is characterized by a marine environment (inner shore) and grades up, towards the upper Miocene, into deltaic and fluvial deposits (Tătărâm *et al*., 1977; Ion *et al*., 2004). The Pliocene (the regional Paratethyan stage Pontian *pro parte*, Dacian and Romanian) deposits mainly developed along the right side of the Danube, in low topographic areas, and are represented by the sedimentation of continental red clay sequences (Andreescu *in* Avram *et al*., 1996). The Pleistocene is characterized by the sedimentation of continental red clay sequences (early Pleistocene in age), and by a large development of the loess deposits (during the middle to late Pleistocene interval).

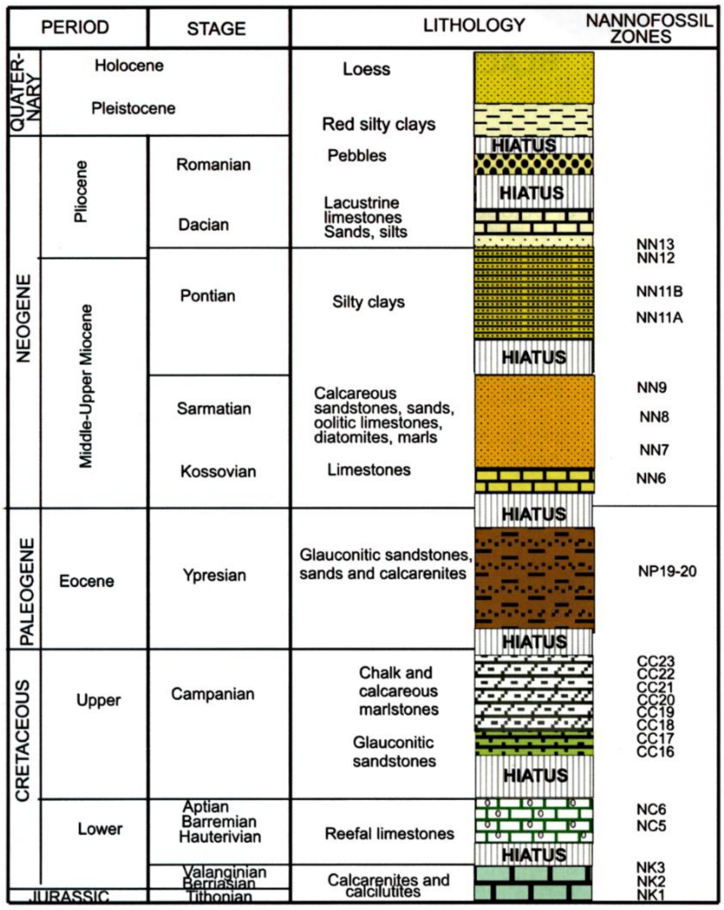


Fig. 9. The main Cretaceous to Pontian marine lithological units that crop out in the South Dobrogea and their biostratigraphy based on calcareous nannofossils (after Melinte, 2006). NK calcareous nannofossil zones after Bralower *et al*.'s Zonation (1989); NC calcareous nannofossil zones after Roth's Zonation (1983); CC calcareous nannofossil zones after Sissingh's Zonation (1977); NP and NN calcareous nannofossil zones after Martini's Zonation (1971).

2. GEOMORPHOLOGY OF THE DANUBE DELTA

The Danube River is one of the most important European waterways flowing 2,857 km across the continent from the Schwarzwald Massif down to the Black Sea. The Danube is listed after the River Volga as the second longest river in Europe. Its drainage basin extends to 817,000 km^2; more than 15 countries share the Danube catchment area and about 76 million people live within this area.

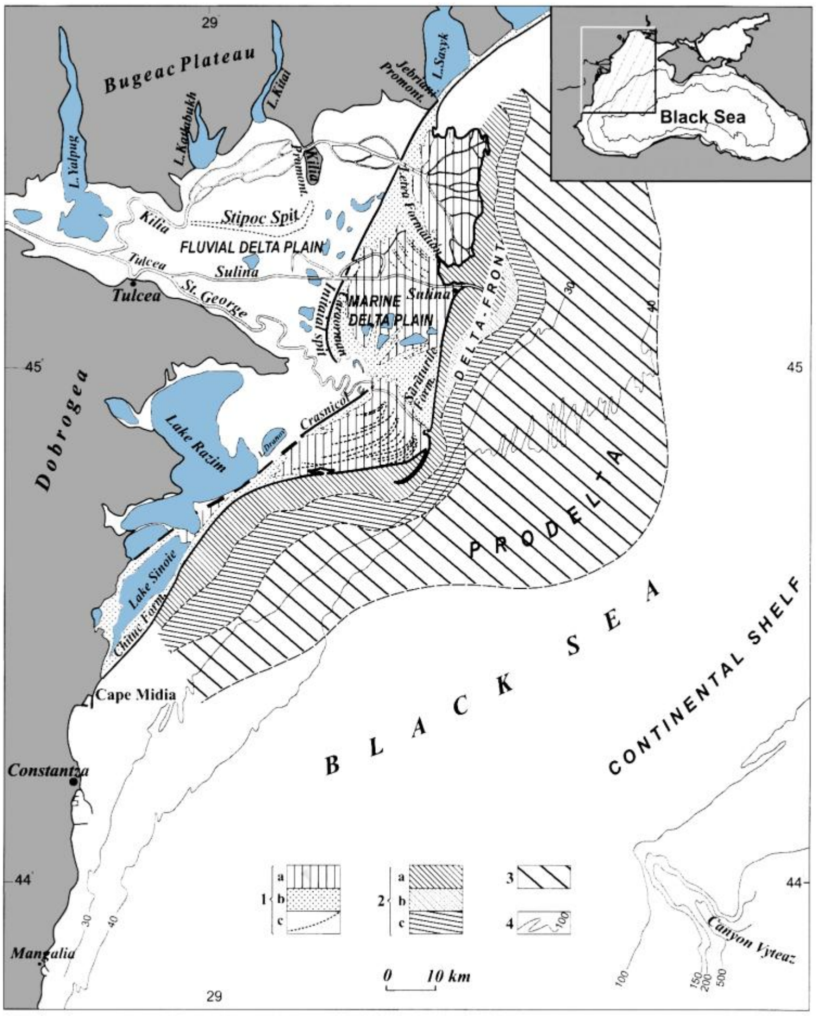


Fig. 10. Major morphological and depositional units of The Danube Delta (after Panin, 1989): 1. Delta Plain: a) Fluvial Delta Plain; b) Marine Delta Plain; c) Fossil and modern beach-ridges and littoral accumulative formations built up by juxtaposition of beach-ridges; 2. Delta Front: a) Delta Front platform; b) Relics of the Sulina Delta and its Delta Front; c) Delta Front slope; 3. Prodelta; 4. Depth contour lines. The three major depositional systems of the Danube Delta are characterized as follows: the delta plain, with a total area of about 5,800 km^2, from which the marine delta plain area is of 1,800 km^2; the delta front with an area of ca. 1,300 km^2, divided into delta front platform (800 km^2) and delta-front slope (ca. 500 km^2), extending off-shore to a water depth of 30-40 m; the prodelta lies at a depth of 50-60 m off shore, at the base of the delta-front slope, covering an area of more than 5,500 – 6,000 km^2.

The Danube Delta is situated in the north-western part of the Black Sea, between 44˚25’ and 45˚30’ northern latitude and between 28˚45’ and 29˚46’ eastern longitude, being bordered by the Bugeac Plateau to the North and by the Dobrogea orogenic area to the South. The Delta is one of the main elements of the Danube River – Danube Delta – Black Sea Geosystem. The Danube Delta can be divided into three major depositional systems: the delta plain, the delta front and the prodelta (Fig. 10).

To these is to be added the Danube deep-sea fan placed beyond the shelf break reaching from several hundred meters water depth down to the abyssal plain (just over 2,200 m).

The delta development is controlled by: the Danube sediment input (the average sediment discharge is ca. 40.106 t/y, of which 4-6.106 t/y sandy material); the prevalence of winds from the northern sector (40-50% of instances) of the delta front area; the predominance of southward oriented marine currents; the long shore sediment drift directed also towards the south; the relatively important value of wave power etc. The interaction of these factors determines the delta morphological type, the geometry of the volumes of deltaic deposits, the asymmetry of the deltas of the Danube distributaries and their development and evolution. The Danube Delta overlaps the Predobrogea Depression, which, in turn, lies mainly on the Scythian Platform (Fig. 11). The Delta edifice is built up of a sequence of detrital deposits of tens to 300-400 m in thickness formed mainly during the upper Pleistocene (in the Karangatian, Surozhian, and Neoeuxinian stages of the Eastern Paratethys) and the Holocene (Fig. 12).

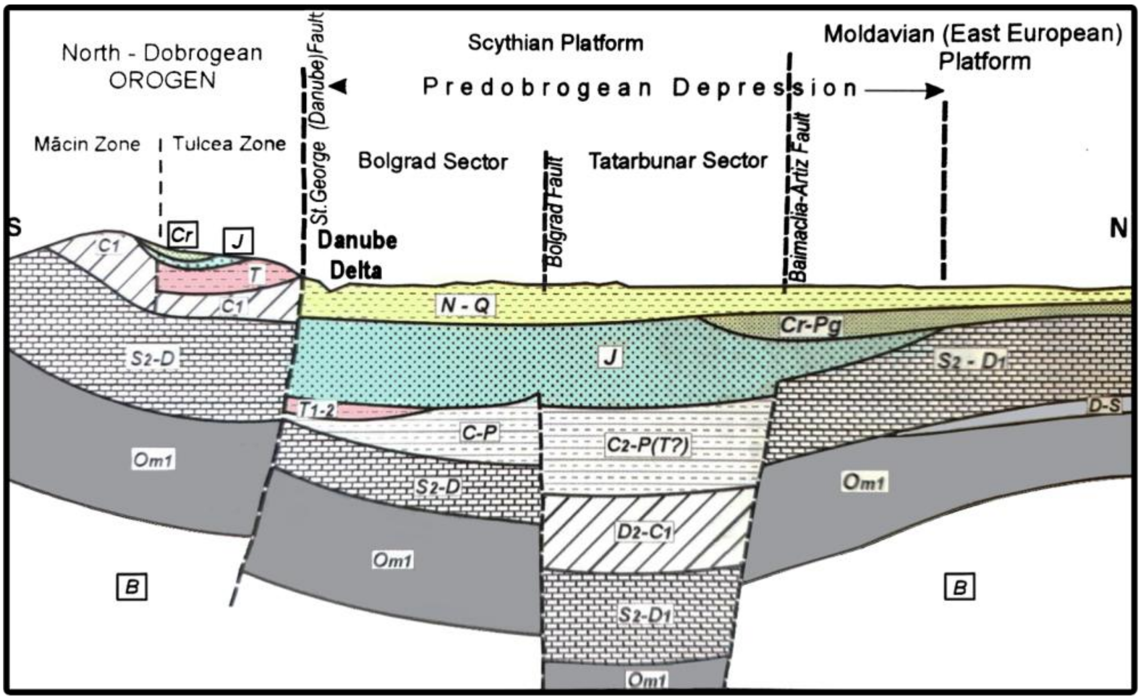


Fig. 11. Geological section showing the location of the Danube Delta within the major structural units of the region (after Liteanu & Pricăjan, 1963). Legend: B – Basement; O – Ordovician; S – Silurian; D – Devonian; C – Carboniferous; P – Permian; T – Triassic; J – Jurassic; Cr – Cretaceous; Pg – Paleogene; N – Neogene; Q – Quaternary.

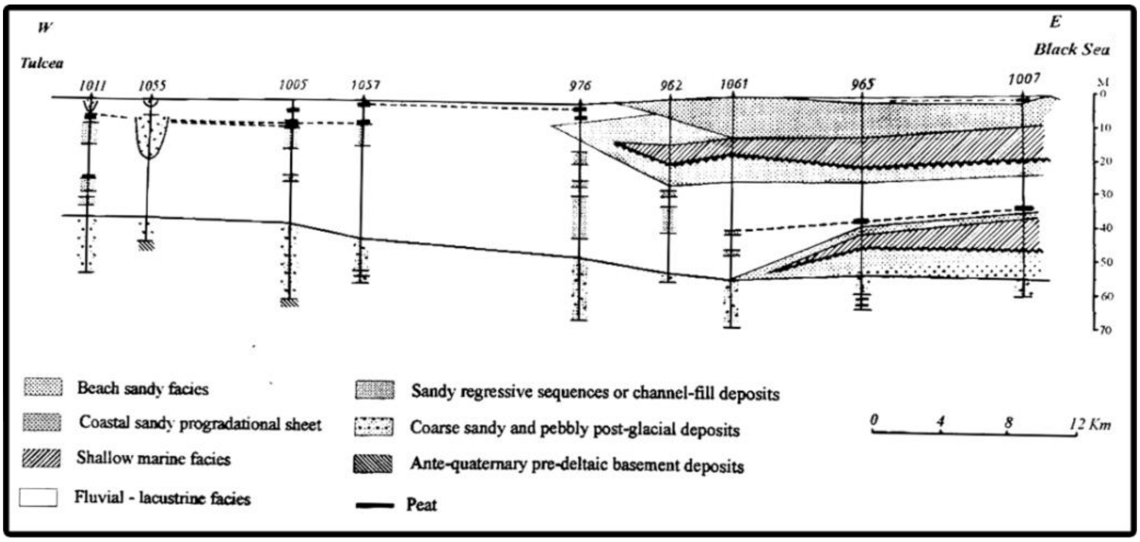


Fig. 12. Schematic cross section through the Danube Delta (after Panin *et al*., 2004).

The Holocene evolution of the Danube Delta occurred in several main phases, summarized as follows (Panin *et al*., 2004): (1) the formation of the Letea-Caraorman initial spit, 11,700-7,500 yr. BP; (2) the St. George I Delta, 9,000-7,200 years BP; (3) the Sulina Delta, 7,200-2,000 years BP; (4) the St. George II and Kilia Deltas, 2,000 years BP – present; (5) the Coşna-Sinoe Delta, 3,500-1,500 years BP (Fig. 13). Sediments in the Danube Delta display the following main facies (Panin, 1989): (I) marine littoral deposits of two types: type “a”, formed by the longshore drift from the North (from the mouths of rivers Dniester, Southern Bug and Dnieper), and type “b”, of Danubian origin; (II) lacustrian littoral deposits, forming the Stipoc and Roşca-Suez lacustrian spits; (III) fluvial deposits, genetically related to the Danube distributaries system, include several types: bed-load and mouth-bar deposits, sub-aqueous and subaerial natural levees deposits, crevasse and crevasse-splay deposits, point bar and meander belt deposits, decantation deposits into intra-deltaic depression and inter-distributary area etc.; (IV) marsh deposits; (V) loess-like deposits (Fig. 14).

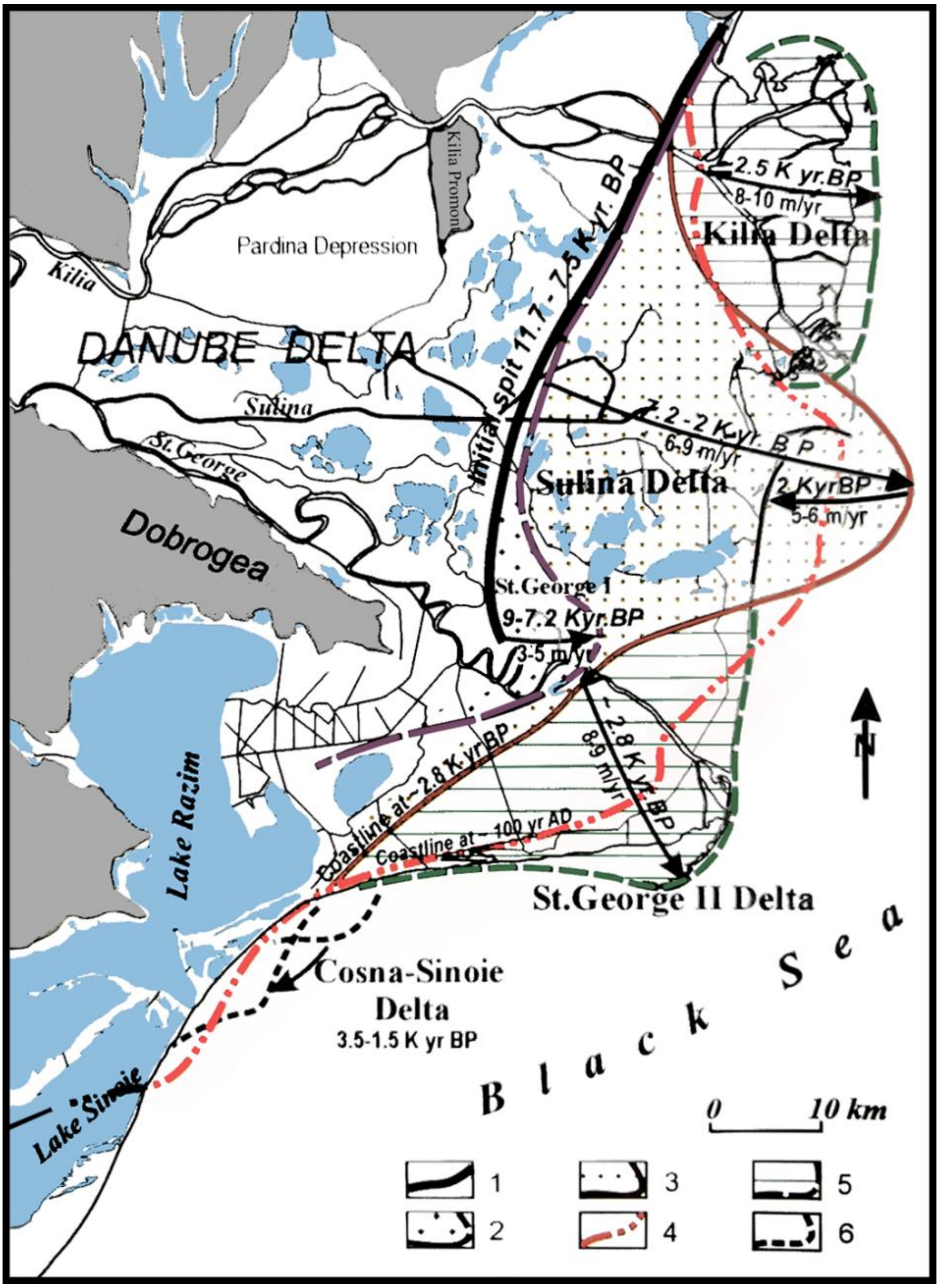


Fig. 13. The Danube Delta evolution during the Holocene and correspondent changes of coastline position (after Panin, 1997).

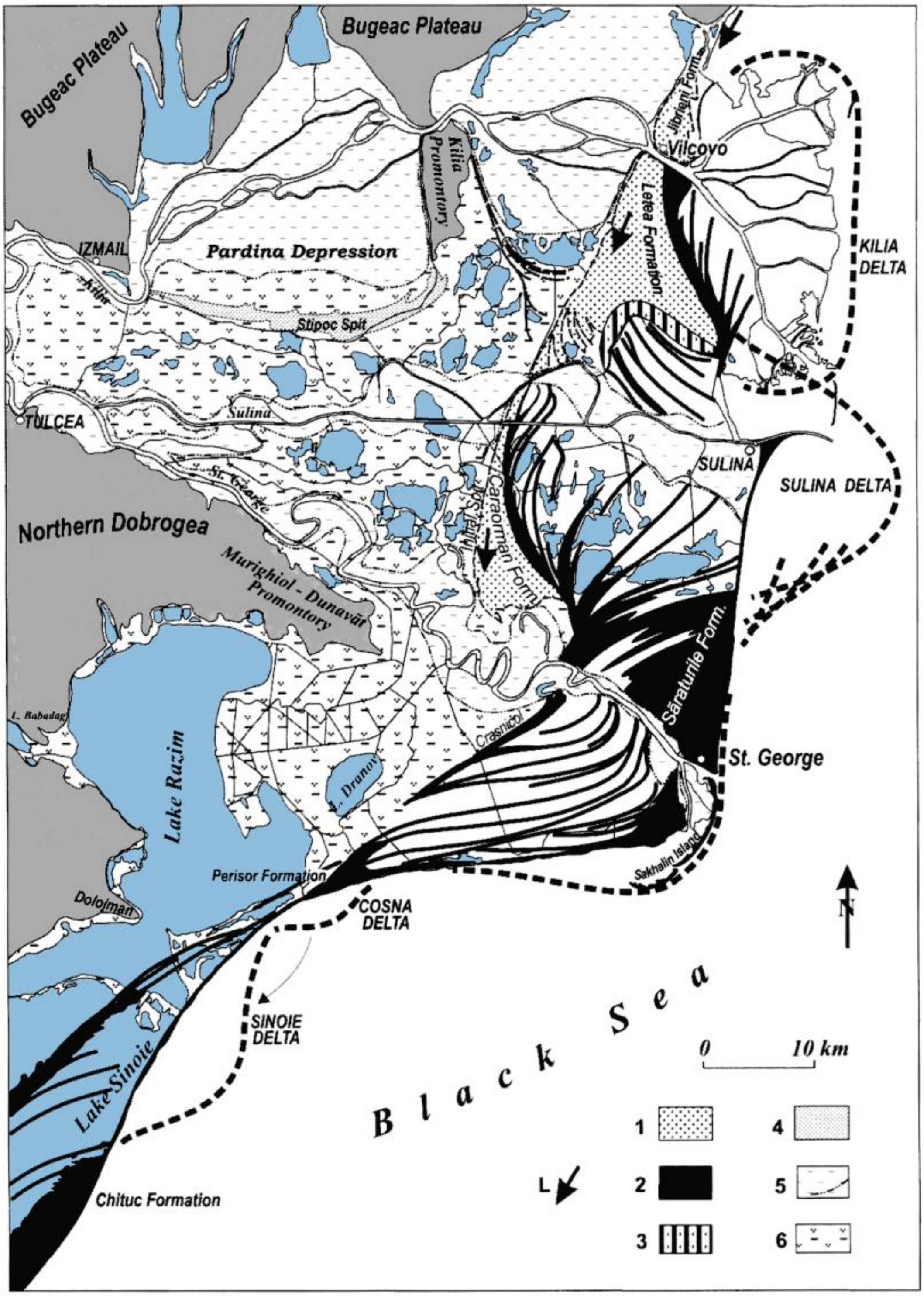


Fig. 14. Areal distribution of the main types of deposits within the Danube Delta territory (after Panin 1989). 1, marine littoral deposits of type “a”, formed by the littoral drift from the rivers Dniester and Dnieper mouths; 2, marine littoral deposits of type “b”, of Danubian origin; 3, deposits of littoral diffusion, formed by mixing of “a” and “b” types; 4, lacustrine littoral deposits; 5, fluvial meander belt deposits; 6, interdistributary depression deposits; L, direction of the longshore sediment drift.

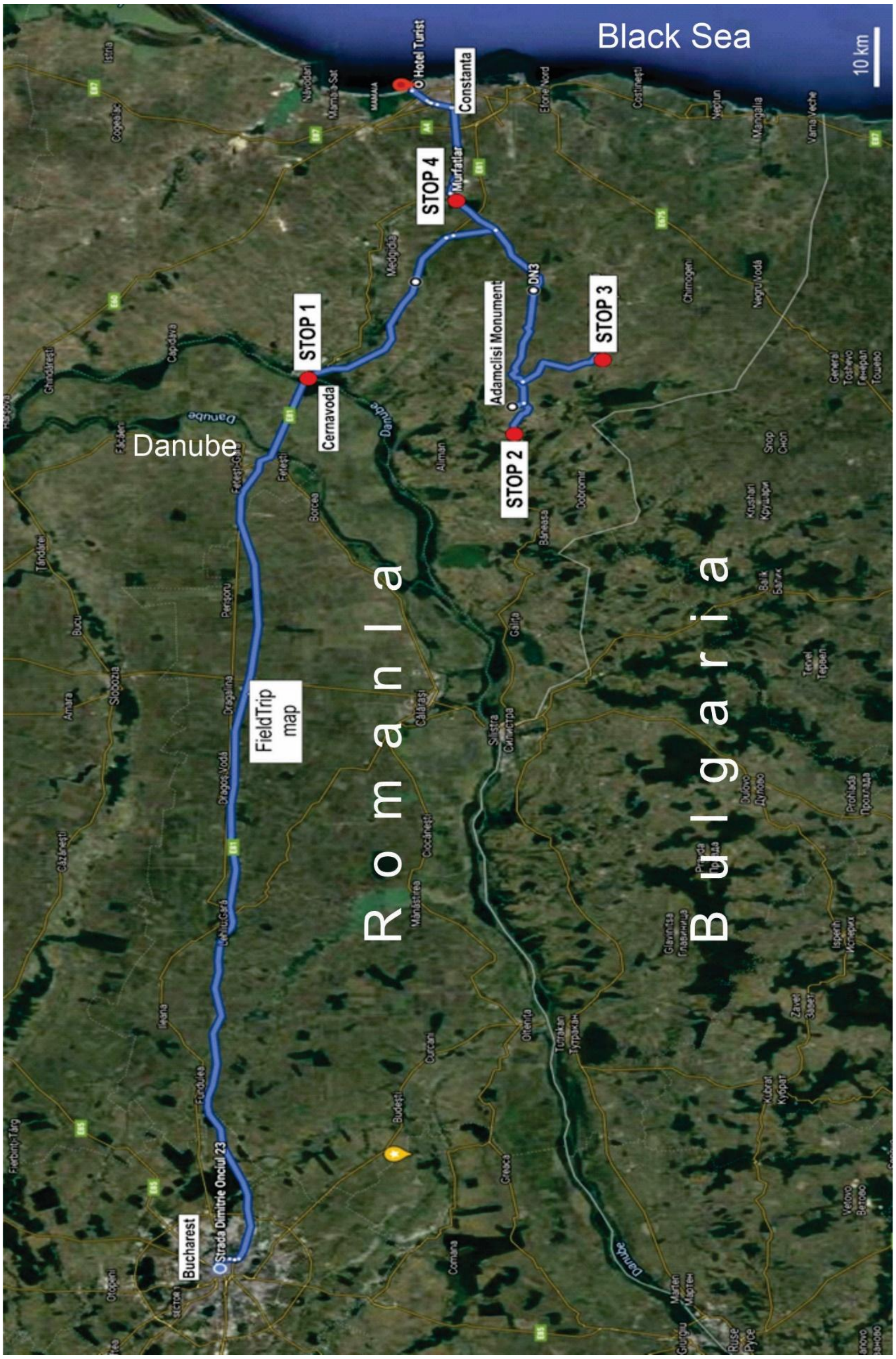


Fig. 15. Stops of the First day (itinerary from www.google maps)

3. DESCRIPTION OF STOPS

DAY 1

Southern Dobrogea

From Bucharest up to Feteşti, the motorway crosses the western part of the Moesian Platform, described as the Wallachian Platform (Fig. 15). Geomorphologically, this part is known as the Romanian Plain. The Moesian Carbonate Platform is a major tectonic unit of the Romanian territory, limited to the north by the Carpathian Foredeep and to the south by the Balkan Orogen (Săndulescu, 1984). The Moesian Platform represents a Precambrian block incorporated in the Epihercynian Platform of Central Europe (Săndulescu, 1984) and it is composed by two main domains, the "Dobrogean" and "Valachian" parts separated by the crustal scale Intramoesian Fault (Visarion *et al*., 1988).

The Moesian unit includes Jurassic and Lower Cretaceous deposits cropping out in its eastern part, in Central and South Dobrogea, and representing a wide range of depositional settings, from continental deposits to inner and middle shelf shallow marine carbonates, interlayered with Purbeckian evaporites (Dragastan, 2001; Stoica, 1997; Seghedi & Stoica, 2011). The route crosses the Danube River that is divided in this region into two: the Old Danube and the Borcea branch, forming between these arms Balta Ialomiței island. The two cities on the Danube banks are Feteşti on the left and Cernavodă on the right (Fig. 8). This bridge connects two Romanian Provinces, Muntenia to the W and Dobrogea (=Dobruja) to the E.

Over the river branches there is a bridge (historical monument) designed by the Romanian engineer Anghel Saligny between 1890 and 1895, with a total length (with viaducts) of 4,088 m. The Anghel Saligny Bridge complex has been exclusively used for almost a century, until 1987, when the new Cernavodă Bridge complex, built next to it, was inaugurated. The Cernavodă town (in Thracian *Axiopa*, in Bulgarian *černa voda* – black water) houses the single nuclear power plant in Romania, consisting of two CANDU reactors that provide about 18% of Romania's electrical energy output. Besides, the Danube-Black Sea Canal, opened in 1984, extends from Cernavodă to Agigea and Năvodari (at the N Romanian Black Sea). In the hills around Cernavodă there are well-known vineyards, producers of Chardonnay wine.

<u>Stop 1</u> *Cernavodă Fossil Site – Lower Cretaceous limestones*

Coordinates: N 44°20'17"; E 28°02'01"

This is a protected area of national interest which corresponds to IUCN III category (natural reserve of paleontological type), situated on the administrative territory of Cernavodă town, extended on around 3 ha (Fig. 16). This site was studied since the end of the 19th Century and the beginning of the 20th Century (complete references in Dragastan *et al*., 1998). In the 20th and 21st centuries detailed biostratigraphy and paleoenvironmental reconstruction of the carbonate shelf have been produced based on macrofossils, microfossils (foraminifers, ostracods and calcareous algae) along with microfacies analysis (Neagu *et al*., 1997; Stoica, 1997; Dragastan *et al*., 1998; Neagu, 2000; Seghedi & Stoica, 2011).

The exposed Cernavodă Formation includes three members (Fig. 17). These are, in the stratigraphic succession, Hinog Member (upper Berriasian), Aliman Member (Valanginian) and Vederoasa Member (lower Hauterivian), according to Avram *et al*. (1988, 1996) and Dragastan *et al*. (1998).

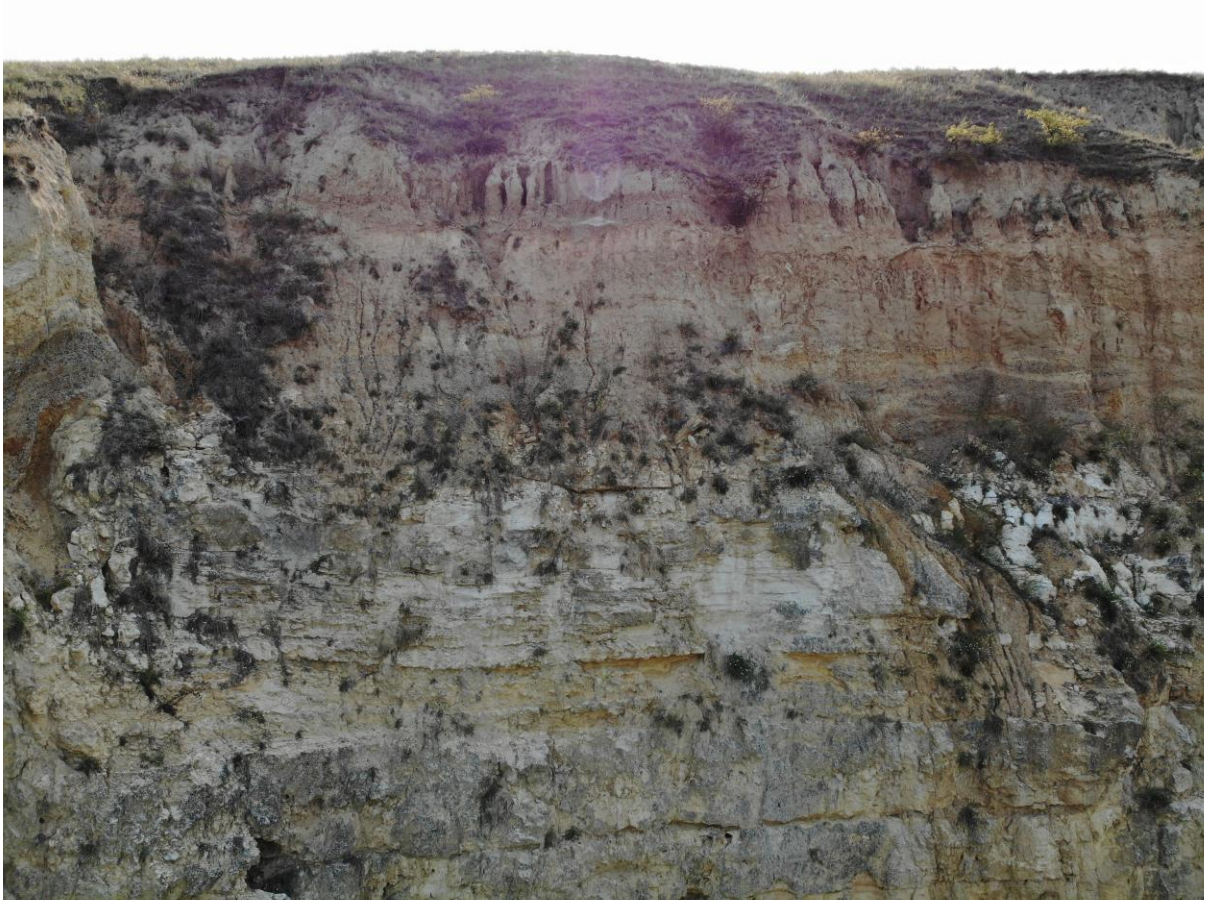


Fig. 16. Limestones exposed in the Natural protected area Cernavodă Fossil Site. The top of the outcrop is represented by yellowish loess deposits (late Pleistocene-Holocene). Drone Photo by Dr. Gabriel Ion, August 2019.

The Berriasian Hinog Member is made by conglomerates, grey marly limestones, oosparites, sandy clays and limestones. This lithology encloses gastropod taxa such as *Harpagodes pelagi* and *Saulea neocomiensis*.

The Valanginian Alimanu Member of the Cernavodă Formation unconformably overlies the Hinog Member, being mainly composed of limestones. The lower Valanginian succession starts with brecciated marly limestones, followed by grey micritic limestones, whitish-yellowish oolitic limestones, limestones with *Nerinea* sp., marly clays, lenses of pachyodont limestones (*Matheronia baksanensis* and *M. valanginiensis* identified by Masse) and 3 levels of reefal buildups of patch-reef type bioconstructed by demosponges. On bed surfaces, ferruginous crusts with subvertical, branching bottle-neck perforations, corresponding to *Gastrochaenolites* ichnogenus are present (Dragastan *et al*., 1998; Seghedi & Barbu, 2001).

The lower Valanginian succession ends with a tabular reefal buildup. Pachyodont bivalves (*Monopleura valanginiensis* and *M. baksanensis*), as well as *Ampullina* and *Nerinea* gastropods are commonly encountered, covered by *Lithocodium* and *Bacinella* algal crusts (Dragastan *et al*., 1998).

The upper Valanginian depositional interval, unconformable and transgressive onto the lower Valanginian succession, is made by detrital sequence, with angular litho- and bioclasts of limestones, and microconglomeratic gravels, overlain by thin oolithic sands, followed by pelsparitic limestones interlayered with clays and overlain by a massive reef. The latter includes crusts of demosponges, pachiodont shells (*Matheronia baksanensis*) and gastropods (*Nerinea, Purpuroidea, Leviathania, Harpagodes, Ampullina,* among other taxa).

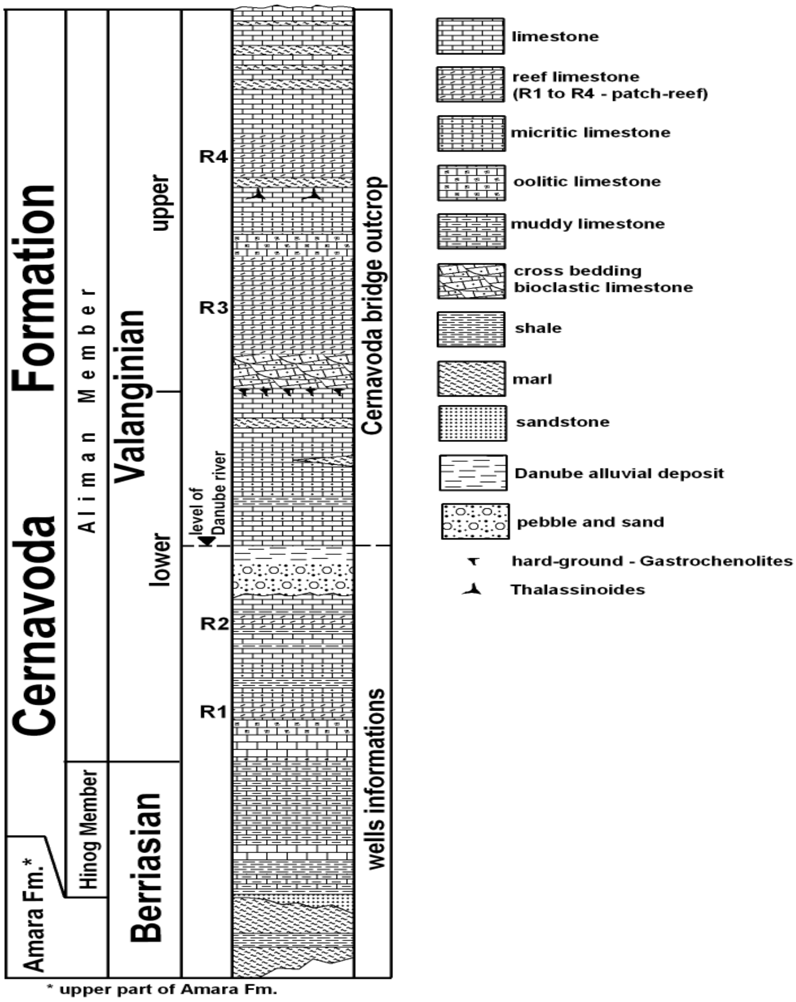


Fig. 17. Lithostratigraphy of the Cernavodă Fossil Site (after Neagu *et al*., 1997, redrawn by Barbu and published in Seghedi & Stoica, 2011).

The lower Hauterivian depositional interval, namely the Vederoasa Member of Cernavodă Formation, described in various sites from the South Dobrogea, was recently found in the Cernavodă Fossil Site (Dragastan, 2001; Dragastan *et al*., 2014). The Hauterivian is transgressive on the Valanginian, and it is mainly composed of oolitic limestones with *Ostrea* shells, and white-yellowish limestones, reddish in places. Reef structures are placed at the top of the lower Hauterivian succession.

An accurate age of the lower Cretaceous deposits was assigned based on agglutinated foraminifers (Neagu *et al*., 1997; Neagu, 2000a; 2000b). The calcareous nannofossils yielded a Berriasian-Valanginian age (Melinte & Mutterlose, 2001), indicated by the presence of NK1, NK2 and NK3 biozones, respectively *Nannoconus steinmannii steinmannii*, *Cretarhabdus angustiforatus* and *Calcicalathina oblongata*. According to several authors (Dragastan *et al*., 1998 and included references), the late Hauterivian-Barremian interval is missing. Along the Danube-Black Sea Channel, the Cernavodă Formation is unconformable covered by Aptian-Albian continental-lacustrine, brackish and marine deposits (Avram *et al*., 1991, Dragastan *et al*., 2014). It was assumed that the shallow marine carbonate deposits cropping out in the Cernavodă Fossil Site were sedimented in an intertidal to upper subtidal environments (Tătărâm *et al*., 1977; Neagu *et al*., 1997), based on the common presence of typical intertidal taxa, such as patellids and neritids, along with upper subtidal organisms, i.e., pachyodonts, ostreids, pleurotomariids and brachiopods.

<u>Stop 2</u> *Urluia Quarry – Middle Miocene (Sarmatian) deposits in shallow marine facies*

Coordinates: N 44°5'38.832"; E 27°54' 6.9798"

From the middle Miocene (i.e., regional Sarmatian stage), the current territory of Romania was situated in three distinct Paratethyan domains (Fig. 18):

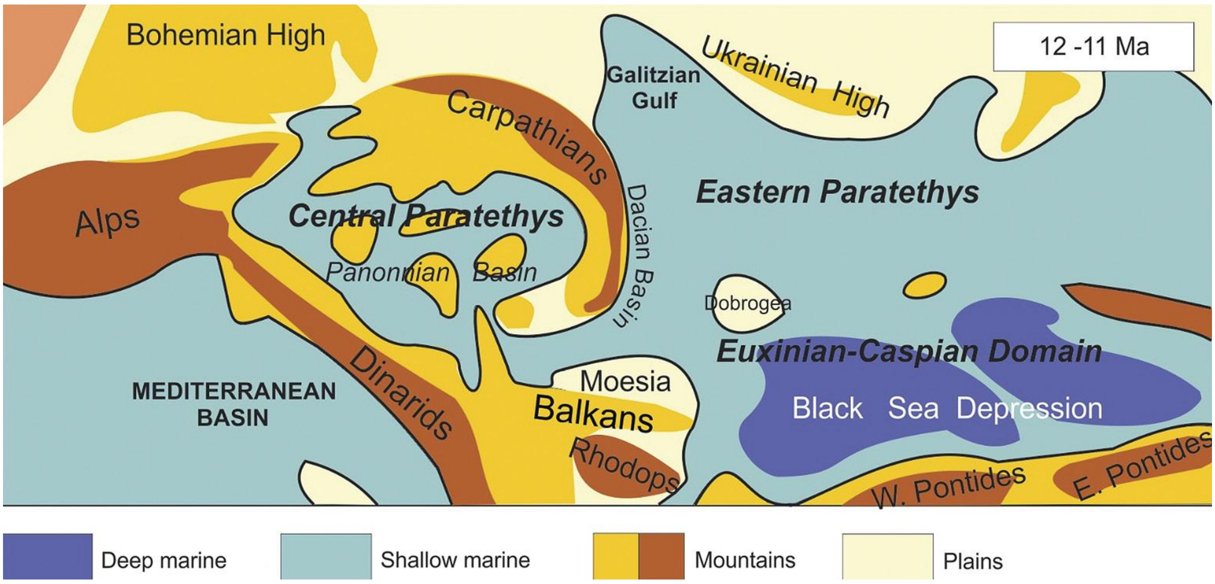


Fig. 18. Paleogeography of the Paratethyan Domain and surrounding areas during the Sarmatian (modified after Popov *et al*., 2004, Jipa & Olariu, 2013 and Briceag *et al*., 2018).

(i) The Central Paratethys, including the intra-Carpathian areas, such as the Transylvanian region;
(ii) the Eastern Paratethys, which encloses the extra-Carpathian regions, described under the name of the Dacian Basin;
(iii) the Euxinian-Caspian Basin (Andrusov, 1917; Marinescu, 1978; Popov *et al*., 2004; Jipa & Olariu, 2013). Good outcrops of the Sarmatian have been described from the southern extremity of Dobrogea (in Romania) (Atanasiu, 1940; Chiriac, 1968; Ionesi & Ionesi, 1971), located during those times in the western part of the Paratethyan Euxinian Domain.

The Sarmatian shallow carbonate platform (Fig. 19) transgressively overlies lower Cretaceous (Barremian to Albian), upper Cretaceous and Paleogene (Eocene and Miocene) sediments. Lithologically, the Sarmatian of the South Dobrogea is mainly characterized by the deposition of sands, rudites, organogenic limestones, oolitic limestones, clays and marls (Chiriac, 1968; Ionesi & Ionesi, 1971).

Across the lower Sarmatian to the middle Sarmatian boundary interval, i.e., between the Volhynian and Bessarabian substages, the paleoenvironment changed; a shallow semi-enclosed sea with brackish faunas covered the area of Dobrogea (Tătărîm *et al*., 1977; Ion *et al*., 2002), characterized by temporary connections with open marine basins (Briceag *et al*., 2018). Hence, the brackish paleosetting progressively shifted to a continental-lacustrine one in the late Sarmatian (Kersonian substage).

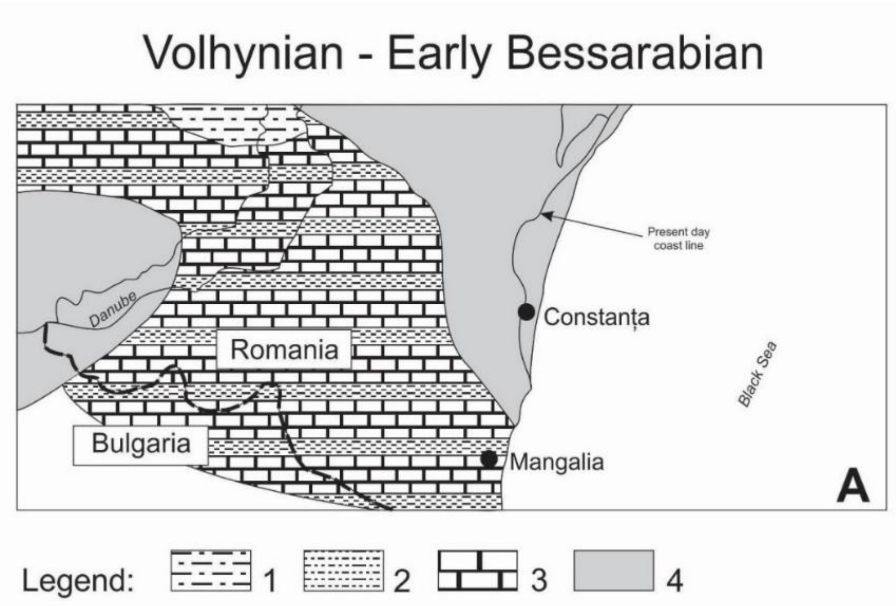


Fig. 19. Sarmatian lithofacies map of the Southern Dobrogea in SE Romania (modified after Ion *et al.*,2002); Volhynian-early Bessarabian;
Legend: 1 – siltstones; 2 - marls; 3 - limestones; 4 – emerged land.

West of Adamclisi locality there are several old quarries that yielded diatomites and limestones. Among them, the Adamclisi and Urluia are the most expanded. In the Urluia Quarry, middle Miocene (Sarmatian) deposits, belonging to the middle part of the Paratethyan Sarmatian substage (i.e., Bessarabian) are exposed. These are unconformably covered by Pleistocene deposits, mainly composed of red clays, sands and loess. The reddish clays are assumed to be early Pleistocene in age and (Ghenea *et al*., 1978, 1984) possibly, the afore-mentioned deposits have a residual-eluvial and alluvial origin and may be seen as a paleosol. Most probably, the reddish sediments have been formed subaerially in a warm and humid climate.

Fig. 20. The middle Miocene (Sarmatian) deposits of the Urluia Quarry, topped by Pleistocene red clays, sands and loess. Photo Dr. Gabriel Ion, August 2019.

Within the middle Miocene deposits of the Urluia Quarry (Fig. 20), two distinct lithostratigraphic units occur:

- The oldest units represented by the Clay and Diatomite Formation, which is earliest to early middle Sarmatian in age (the substages Volhynian - early Bessarabian). This unit, made by brown to blackish bentonite clays and diatomites, interbedded with micritic limestones, whitish to grey marlstones and calcareous sandstones. This unit corresponds to the “Lower Limestones Horizon” and *pro parte* the “Diatomite-Bentonitic Horizon” (Chiriac, 1960).
- The Cotu Văii Formation, middle Sarmatian in age, i.e. Bessarabian (Ghenea *et al*., 1978; 1984; Ionesi & Ionesi, 1971), contains mainly carbonate rocks such as marly limestones, oolitic limestones, calcilutites, calcarenites, and bioclastic calcirudites. This unit contains fossil faunas in generally composed of bivalves (most common taxa of *Mactra*, *Sarmatimactra* and *Obsoletiforma*).

Close to Stop 2, the Tropaeum Traiani historical monument, fortress, museum and the Roman castrum may be seen. Tropaeum Traiani was built in 109 in Moesia Inferior, to commemorate the victory of the Roman Emperor Traian over the Dacians, in the winter of years 101-102, in the Battle of Adamclisi. In the vicinity of the Tropaeum Traiani Monument, a museum that contains, among other artifacts, parts of the original edifice, is situated. The history of battles between the Romans and the autochthonous Dacian population is described in 48 metopes hosted by the museum. A Roman Castrum is also preserved, including the Tropaeum Traiani City wall. Interestingly, the metopes are made by limestones of different types and ages. Therefore, lower Cretaceous limestones, along with Eocene nummulitic limestones and middle Miocene limestones, i.e., mainly Sarmatian oolithic limestones along with coquinas with bivalves such as Mactra and Sarmatimactra have been used.

<u>Stop 3</u> *Limestone walls at Petroşani – upper Cretaceous and middle Miocene deposits*

Coordinates: N 44°0’32.6232”; E 28°3'14.3424"

This is a protected area of national interest which corresponds to IUCN III category (geological reserve), extended on around 4.8 ha and situated on the administrative territory of the Deleni commune, 0.5 km south of the Petroşani Village, on both sides of the Urluia valley. The protection refers to the upper Cretaceous, i.e., Cenomanian Peştera Formation and the middle Miocene, Sarmatian (middle substage Bessarabian) Cotu Văii Formation, transgresively overlaying it (Fig. 21).
The lower part of the *Limestone walls from Petroşani* are composed of lower Cenomanian glauconitic sandstones and conglomerates. These deposits start with a basal, poorly sorted conglomerate, followed by sands or quartzose sandstones with parallel or crossed stratification. Toward the upper part of the Cenomanian succession, quartz- and glauconite-rich calcareous sandstones are present. Bioturbations (mainly *Planolites* and *Thalassinoides*) frequently occur. The early Cenomanian age of the Peştera Formation was assigned based on macrofaunas, mainly ammonites and inoceramids, and microfaunas, such as foraminifers (Chiriac, 1981; Szász, 1983; Avram *et al*., 1988; Ion & Szász, 1994; Ion *et al*., 1997; Avram *et al*., 1993). Accordingly, the Peştera Formation belongs to the Mantelliceras mantelli Ammonite Zone and contains index species foraminifers, i.e., *Rotalipora appeninica* and *R. globotruncanoides*.

Fig. 21. Limestone walls at Petroşani; A: General view of the upper Cretaceous (Cenomanian) sediments transgressively overlain by the by middle Miocene (Sarmatian) deposits; B: Detail of Cenomanian glauconitic limestones. Photos Dr. Gabriel Ion, August 2019.

The middle Sarmatian (Bessarabian) deposits are composed by shelly limestones, calcarenites and calcareous sandstones, with thin clay intercalations. The aforementioned lithology is rich in mollusk shells, especially bivalves such as *Sarmatimactra vitaliana*, *S. eichvaldi*, *Tapes tricuspis* and *Obsoletiforma* spp.

<u>Stop 4</u> *The Murfatlar Quarry – upper Cretaceous chalk*

Coordinates: 44°9'59.39"N; 28°24'16.35"E

In the Murfatlar (= Basarabi) locality, the stratotype of the Murfatlar Formation is exposed in a large quarry, made up of grey-whitish argillaceous chalks, overlain by yellowish clays and whitish, massive chalky limestones towards the top (Avram *et al*., 1998; 1993; Ion *et al.*, 1997). The calcareous nannofossils indicate a Santonian – late Campanian age (Ion *et al*., 1998), similar with the one assigned based on planktonic foraminifers, from Dicarinella asymmetrica up to Globotruncana ventricosa zones (Neagu, 1987). A rich macrofauna is preserved, including echinoids (*Micraster* spp., *Offaster pillula*), bivalves (*Inocermanus muelleri*), brachiopods, bryozoans and belemnites (*Belemnitella mucronata*). This unit is transgressively overlaying older deposits.

The locality of Murfatlar is famous for its churches, crypts and tombs carved into a chalk hill, all dating from the 9th-11th century. From the late 7th until the beginning of the 11th century, this territory was part of the First Bulgarian Empire. The inscriptions found here are in Greek and Old Slavic languages. The Murfatlar area is also famous for the vineyards producing organic white, red and rosé wines.

DAY 2

Central Dobrogea

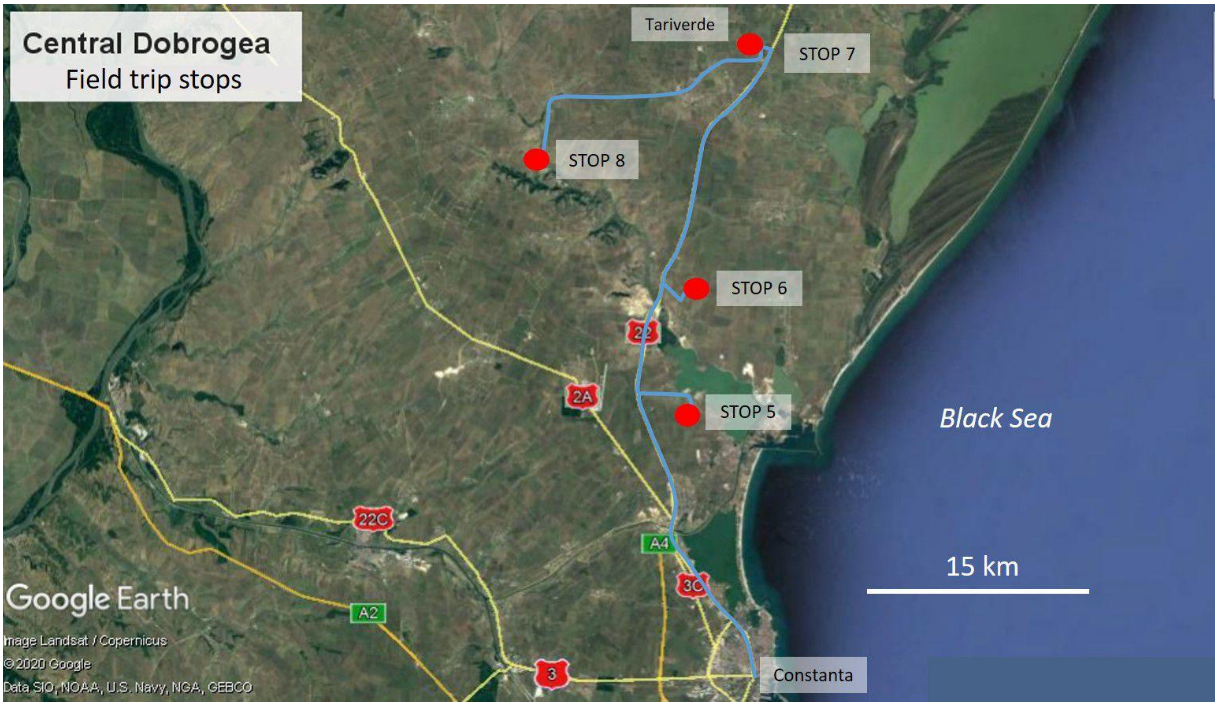


Fig. 22. Stops of the second day of the field trip, figured on a Google Earth image of Central Dobrogea.

Stop 5 *Sibioara Quarry – Ediacaran (Neoproterozoic) anchimetamorphic turbidites of Histria Formation*

Coordinates: 44°20'53.04"N; 28°34'0.31"E

Sibioara is one of the active quarries along the southern shore of Lake Taşaul in Central Dobrogea (Fig. 22). The quarry is emplaced into the upper member of the Histria Formation turbidites. The turbidites include both sandstone-dominated successions, often forming upward fining sequences up to 50 m thick, and coarse and very coarse-grained sandstone beds to microconglomerates and conglomerates. The beds are thin, around 20cm, and the sandstone facies is massive or displays a slight grading. The thin interbeds are mainly Tcd Bouma divisions, displaying ripple cross-laminations and parallel laminations. Conglomerate beds are 1-2 m thick, lens-shaped bodies, often with erosional bases and with frequent rip-up clasts of siltstones or mudstones (Fig. 23).
The turbiditic succession is deformed by E-W trending, upright open folds (Mirăuţă, 1969; Drăgănescu *et al.*, 1978). Slaty cleavages related to this deformation are steeply dipping and visible only in the fine-grained facies.

Fig. 23. Large block extracted from the Sibioara quarry, showing coarse grained channel facies conglomerate overlying medium grained sandstones and grading upwards (to the right) into microconglomerates. Photo Dr. Antoneta Seghedi.

The first palynomorphs from the Histria Formation were identified in samples collected from the Cicracic valley, close to Sibioara. They indicate a Late Neoproterozoic-Early Cambrian age for the Histria formation (Iliescu & Mutihac, 1975; Vaida, unpublished report). Detrital zircon ages, ranging between 622-579 Ma (Żelaźniewicz *et al*., 2009) and 633-583 Ma (Balintoni *et al*., 2011) confirm the Ediacaran depositional age of turbidites. Based on these ages, an Amazonian paleocontinental affinity was suggested for the turbidites (Balintoni *et al.*, 2011).

Stop 6 *Piatra geosite – Ediacaran (Neoproterozoic) turbidites with sedimentary and enigmatic (possibly biogenic) structures preserved on bed surfaces*

Coordinates: 44°24'54.87"N; 28°34'8.45"E

Piatra is a most remarkable geosite in the Ediacaran turbidites of the Histria Formation, being extremely rich in various traces on bed surfaces (Oaie, 1999; Oaie *et al*., 2012). The outcrops are located on a small ravine north-east of the Piatra village, in thin bedded distal turbidites, dipping 10-20° South.

Well exposed, large bedding surfaces, most of them showing attenuated current ripples, display various sedimentary structures, both mechanical and possibly biogenic. Current and chevron marks are the most frequently observed (Oaie, 1999), but some enigmatic traces, some very wide, others irregular or curved, also occur (Fig. 24). An *Aspidella* type trace was also observed (Seghedi *et al*., 2018).

Fig. 24. Bed surface with atenuated ripples and various current marks, along with some curved, enigmatic traces at Piatra. Photo Dr. Antoneta Seghedi.

<u>Stop 7</u> *Tariverde – Ediacaran anchimetamorphic turbidites of Histria Formation with ripple marks intersected by slaty cleavages*

Coordinates: 44°34'18.39"N; 28°35'25.92"E

Outcrops on the right side of the Cogealac valley in Tariverde village show large bedding surfaces in distal turbidites, with extremely well-preserved ripple marks (Fig. 25). Mainly longitudinal ripples occur, some showing double crests. Other sedimentary structures visible on bedding planes are drag marks and wrinkles. Known in Ediacaran deposits from India and Australia, wrinkle structures are interpreted as evidence for preservation favorized by the existence of microbial mats (Saint Martin *et al*., 2011).

The penetrative slaty cleavage visible in the entire succession does not obliterate the sedimentary structures. The cleavage planes are steeply dipping, and bedding planes are frequently overgrown by tiny, twinned calcite crystals, 1-2 mm long, which grow normal to bedding.

Measurements of ripple marks and various current marks ((Jipa, 1968, 1970; Oaie, 1999) represent the basis of the paleocurrent map of the Histria Formation. Detailed paleoflow measurements indicate a major source located to the SE, with minor input from north, probably from intrabasinal highs (Oaie, 1999). The solid discharge was redistributed by longitudinal currents flowing from E to W. The depositional basin of the turbidites is interpreted as a peripheral foreland basin, with only its internal part preserved in the exposed Histria Formation (Oaie, 1999; Oaie *et al*., 2005).

Fig. 25. Several bedding planes (S_0) showing longitudinal ripples in Histria Formation at Tariverde; the picture was taken in April 2009, when the late Prof. Dolf Seilacher visited the outcrop. Photo Dr. Gheorghe Oaie.

Stop 8 *Cheile Dobrogei, upper Jurassic sponge reefs of the Casimcea Formation, the platform cover of the Ediacaran basement*

Coordinates: 44°30'6.89"N, 28°25'28.04"E

The outcrops of Oxfordian (lower upper Jurassic) carbonate rocks in Central Dobrogea record a remarkable sponge-algal biostromal/bioherm complex and its associated fauna. The best exposures of this complex can be seen in Cheia Gorges, south of the village of Cheia, a geological reserve currently included in a Natura 2000 site, the Cheia Jurassic Reefs. These limestone deposits represent the lower (Visterna) member of the Casimcea Formation (Drăgănescu, 1976). Their Oxfordian–Kimmeridgian age is established based on ammonite faunas (Bărbulescu, 1969, 1970, 1974, 1979; Chiriac *et al*., 1977; Bărbulescu, in Dragastan *et al*., 1998). Detailed biofacies analysis was performed by Herrmann (1994, 1996), while the biozonation was largely defined by Bărbulescu (in Dragastan *et al*., 1998).

In Cheia Gorges, bioherms are dominated by microbialites and calcareous sponges. They form cylindrical structures, 30 m in diameter and 20-30 m high (Fig. 26). Their hollow internal part is filled with a limonitic calcareous breccia (Fig. 27). The sponges are representatives of the class Demospongea while microbialiths are produced by cyanobacteria (Drăgănescu, 1976; Gaillard, 1983; Herrmann, 1996). The cylindrical bioherms have formed near the distal margin of a carbonate ramp, in water depths below normal wave base. Sponge biostromes have formed seaward and coral patch reefs and lagoonal deposits have formed landward of the sponge–algal bioherms.

The benthic fauna other than sponges are represented by several invertebrate groups: sponges, serpulid worms (on the lower surfaces of sponges and on the upper part of the algal crusts), cemented craniacean brachiopods (*Lacunosella*, *Moeschia* and *Argovithyris)*, terebratulid, rhynchonellid, and thecideiid brachiopods, bryozoans, cidarioid and irregular echinoids, pelecypods, gastropods, cephalopods (ammonites, belemnites), crinoids (Bărbulescu, in Dragastan *et al*., 1998) and decapods (4 taxa of crabs, dominated by *Goniodromites*, with several species) (Feldmann *et al*., 2006).

Fig. 26. Bioherms in Cheia Gorges represent cylindrical structures separated by stratified biostromes. Photo Dr. Antoneta Seghedi.

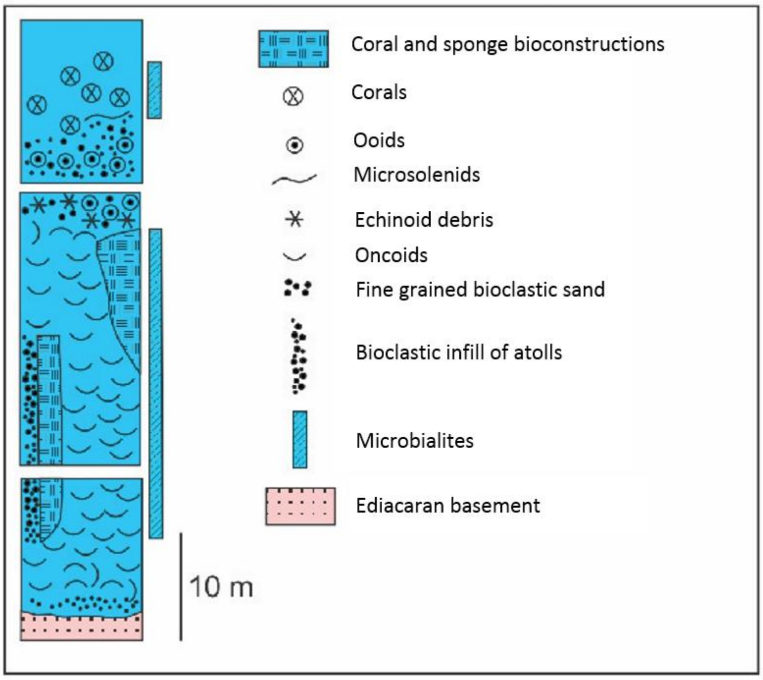


Fig. 27. Lithological column of the Oxfordian deposits in Cheia Valley (redrawn after Herrmann, 1994).

The Cheia Oxfordian reefs belong to the European limestone belt known as "the megafacies of upper Jurassic sponges" (Matyja, 1976), which develops discontinuously in the central and southern part of Europe from Portugal and Spain to Poland and Romania. Remnants of this belt help reconstruct the northern margin of the Paleotethys Ocean during the late Jurassic (Ungureanu & Barbu, 2004).

DAY 3

North Dobrogea

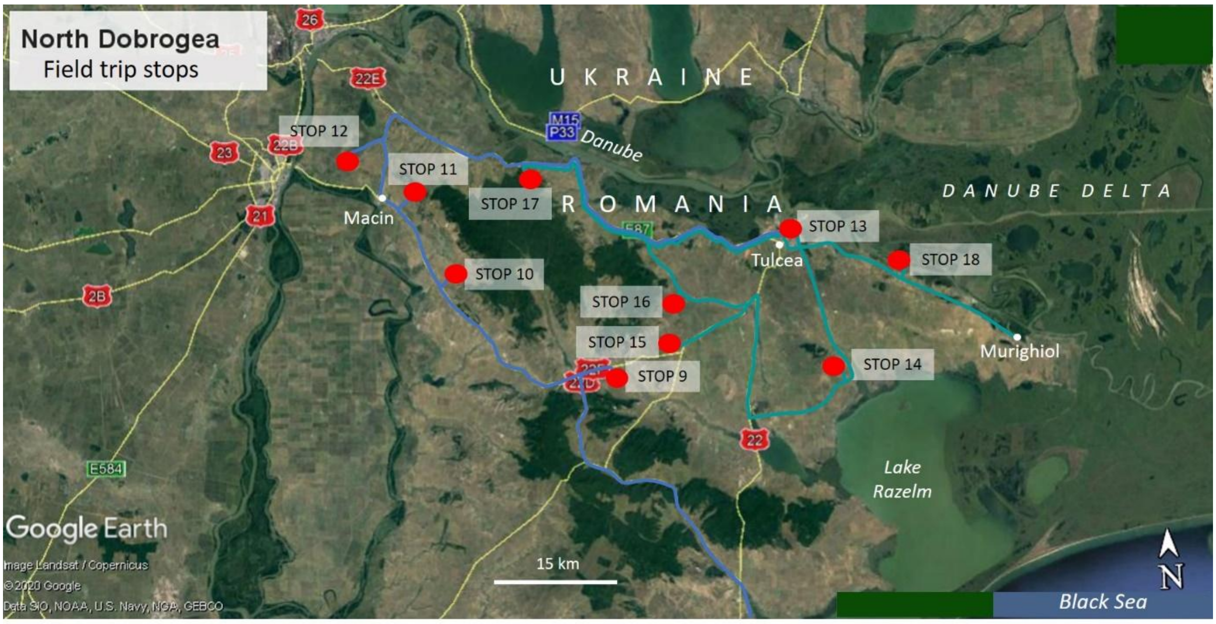


Fig. 28. Stops (red circles) and routes of the third and fourth days (in blue and green respectively) of the field trip in North Dobrogea, figured on a Google Earth image.

<u>**Stop 9**</u> *Consul Hill section – early Triassic limestone turbidites & rhyolites, overthrust onto late Triassic terrigenous turbidites*

Coordinates: 45° 1'51.92"N; 28°29'57.43"E

The Consul zone represents a narrow Cimmerian tectonic unit constrained by two high-angle faults with northeastward vergence: Luncaviţa-Consul Fault in the west and Meidanchioi-Iulia Fault in the east. The Luncavita-Consul Fault is the front of the Măcin unit, which exposes mostly pre-Triassic formations of North Dobrogea. Along this fault, low-grade metamorphic rocks of the Boclugea Group overthrust the Skythian (Spathian) deposits of the Somova Formation. The front of the Consul unit is the Meidanchioi-Iulia Fault, emplacing the Somova Formation onto Norian terrigenous turbidites (Alba Formation).

Along the right bank of the Taiţa valley, 4 km east of Horia village (Fig. 28), the northern slope of the Consul hill represents an E-W section through several thin, high-angle thrust folds in rocks belonging to the Măcin, Consul and Niculiţel Cimmerian units (Fig. 29). Mylonitic quartzites of the low-grade metamorphic Boclugea Group (late Neoproterozoic) are exposed on the western slope of the Coasta Păsunii Hill, along the Luncaviţa-Consul Fault. The quartzites overthrust limestone turbidites of the Somova Formation, but in outcrops to the north and in boreholes to the south, mylonitic quartzites overlie terrigenous clastics of the Bogza Formation (lower Triassic – Griesbachian).

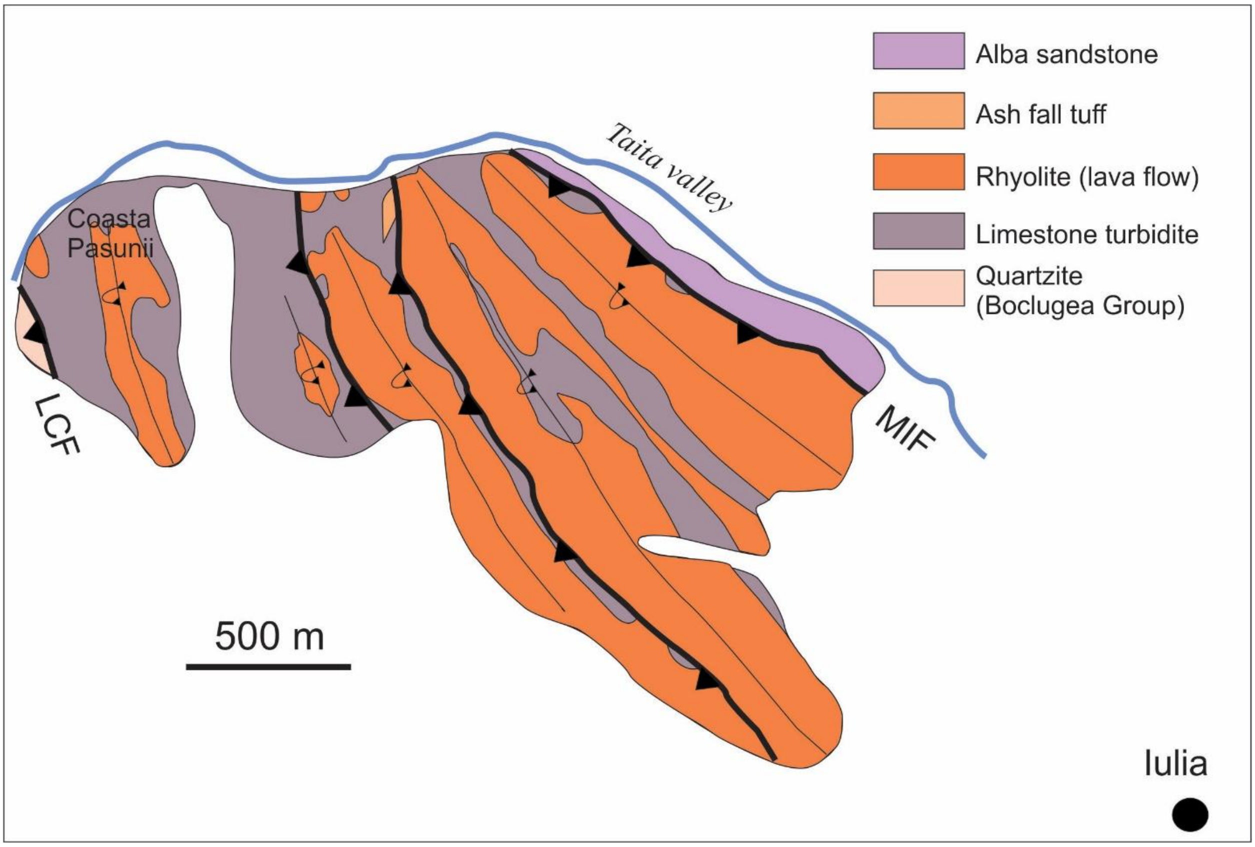


Fig. 29. Geological map of the Consul Hill, showing the structure of the Consul unit, with several thrust-folds in the early Triassic limestone turbidites and rhyolites, which ultimately overthrust the late Triassic terrigenous turbidites (redrawn after Seghedi *et al*., 1990). Abbreviations: LCF – Luncavița-Consul Fault; MIF – Meidanchioi-Iulia Fault.

The Somova Formation (Baltres, 1982) includes limestone turbidites overlain by rhyolites (Seghedi *et al*., 1990). In the Consul Hill, limestones attain 500 m in thickness, but southward, in borehole 5 Iulia, these rocks exceed 1600 m in thickness (Baltres *et al*., 1984). The limestones of the Somova Formation (Baltres, 1982) represent deep water debrites, turbidites and pelagites, emplaced by sedimentary mechanisms implying essentially gravitational transport processes (Baltres, 1982, 1993). Debrites or calcirudites are coarse grained, resedimented accumulations, representing the lower parts of the graded sequences and commonly showing progressive or abrupt transitions to calcarenites. Calcarenitic turbidites occur as beds up to 10 cm thick, displaying complete or incomplete Bouma divisions. Calcisiltitic and calcilutitic turbidites are more frequent and consist of graded and laminated divisions (E_1 and E_2 Piper divisions). The pelagites represent slowly accumulated inter-turbiditic aphanitic limestones, often intensely reworked by bioturbation (Baltres, 1993).

Magmatic rocks are restricted mainly to the top of the Somova formation and include lava flows and tuffs (Seghedi *et al*., 1990). Rhyolitic tuffites, interbedded in distal turbidites at the top of the limestone succession, consist of a mixture of igneous clasts (devitrified volcanic glass, alkali feldspar, plagioclase and quartz crystalloclasts) and carbonatic material (recrystallized lithoclasts), locally with minor metamorphic quartz or granitic clasts. The amount and grainsize of the volcanic material is variable. In the vicinity of the limestone-rhyolite boundary, the main rhyolitic body contains metric beds of rhyolitic tuffs, consisting of glass shards, crystals, crystalloclasts and pumice, within a fine grained, devitrified rhyolitic groundmass. Grainsize range is 2-3 mm. The slight grading occurring in places indicates their deposition as ash fall tuffs.

The main rhyolitic body shows various structural features, from massive (most widespread) to perlitic, brecciated and fluidal (typical for the basal parts of the rhyolites overlying the limestones). The former glassy or microgranular groundmass is devitrified and the perlitic and brecciated varieties show pervasive secondary chloritisation. The effusive facies for most of the rhyolites is suggested by petrographic evidence (Savul, 1935). Association with tuffs, large areal extent and lack of any thermal and/or metasomatic effects in the adjacent limestones constitute additional evidence for an effusive origin (Seghedi *et al*., 1990). Aphanitic or porphyritic rhyolites display flow structures that usually parallel limestone bedding. Evidence for a subaequous environment of emplacement is indicated by rhyolite structure and low crystallinity, the presence of perlites (indicating rapid cooling) and the absence of thermal or metasomatic contact effects in limestones next to rhyolites. The perlitic parts represent the outer crust of the flow, quenched in contact with water and subsequently disrupted and included in the main flow mass. All the above features suggest that most rhyolites represent a single important effusive episode, with subaequous consolidation of the acid volcanic material. Geochemical data indicate that rhyolites are calc-alkaline rocks formed in intraplate setting from magmas with crustal derivation (Seghedi *et al*., 1992).

The Somova Formation accumulated during the Late Scythian (Spathian) as indicated by foraminifera assemblages with *Meandrospira iulia* (Premoli Silva), *M. dieneri* Kristan-Tollman and *Glomospira silensis* Dager (Baltres, in Baltres & Mirăută, 1987). Conodont assemblages support the same age (Mirăută, in Seghedi *et al*., 1986).

Both limestones and rhyolitic rocks of the Somova Formation show evidence for deformation in two events (Seghedi *et al*., 1990): (1) a first deformation, producing recumbent folds and thrust-folds with north-east vergence (considered post-Liassic – Cimmerian); the slaty cleavage associated to this deformation occurs in the hinge areas of major folds and is more obvious in limestones than in rhyolites, due to the difference in competence and great thickness of the latter; (2) a second folding event locally deforms the first slaty cleavages in limestones; its interference with the major folds creates a “dome and basin” pattern in the Consul unit (Seghedi *et al*., 1990) (Fig. 29).

In the easternmost part of the Consul Hill, the Somova Formation rhyolites overthrust the terrigenous turbidites of the Alba Sandstone along the Meidanchioi-Iulia Fault. Rhyolites above the fault are highly sheared and show a penetrative mylonitic foliation over several meters from the contact with the turbidites. The Alba Sandstone (Mirăuţă, 1966a), represents typical terrigenous turbidites with carbonate cement. This is a sandstone dominated succession, with only local coarse facies (breccias, conglomerates) or limestones and pelites. The mostly coarse, grey colored rocks when fresh, show brown coloration due to weathering. Dish structures are commonly seen, as well as the trace fossil *Zoophycos*. The Norian age is based on the typical planctonic bivalve fauna from the pelitic intervals interbedded with sandstones (Baltres *et al*., 1992; Antonescu & Baltres, 1998).

Stop 10 *The Priopcea section – Hercynian overthrusts of early Paleozoic metamorphic rocks (Megina Group and Priopcea Quartzites) onto Silurian deposits*

Coordinates: 45° 7'34.58"N; 28°14'47.07"E

In the northern part of the Priopcea Hill, two Hercynian thrust-sheets of the metamorphic basement (Megina and Boclugea Groups) overthrust the Silurian deposits (Fig. 30).

The Silurian deposits (Cerna Formation, Patrulius *et al*., 1973) consist of two members: a lower member of black, pyritic limestones and an upper member dominated by grey shales with thin sandstone or limestone interbeds (Mirăuţă, 1966a) (Fig. 31). The Silurian age is poorly constrained. It was ascribed on scarce macrofaunal evidence (corals, crinoids, tentaculites, graptolites) (Simionescu &

Cădere, 1908; Mirăuţă & Mirăuţă,1962; Mirăuţă, 1966a). Recent U-Pb geochronology on detrital zircon samples from the Silurian deposits yielded a Llandovery age (435 Ma) for the younger zircons, interpreted as the depositional age (Balintoni & Balica, 2016).

Sandstones interbedded in the Silurian shales are largely quartz - muscovitic, clast supported rocks, in places with crinoidal impressions or ossicles. Tight folds in the Cerna Formation are scarcely visible, either in the lower, black limestone member (Fig. 31), or in outcrops where sandstone beds or brown limestones interbeds occur.

The Priopcea quartzites represent a low-grade succession of metapsammitic - metapelitic rocks (quartzites, muscovite-quartzite schists and phyllites), ascribed to the Cambrian and/or Ordovician based on stratigraphic criteria (Mirăuţă & Mirăuţă, 1962) and lately on detrital zircon data (Balintoni & Balica, 2016).

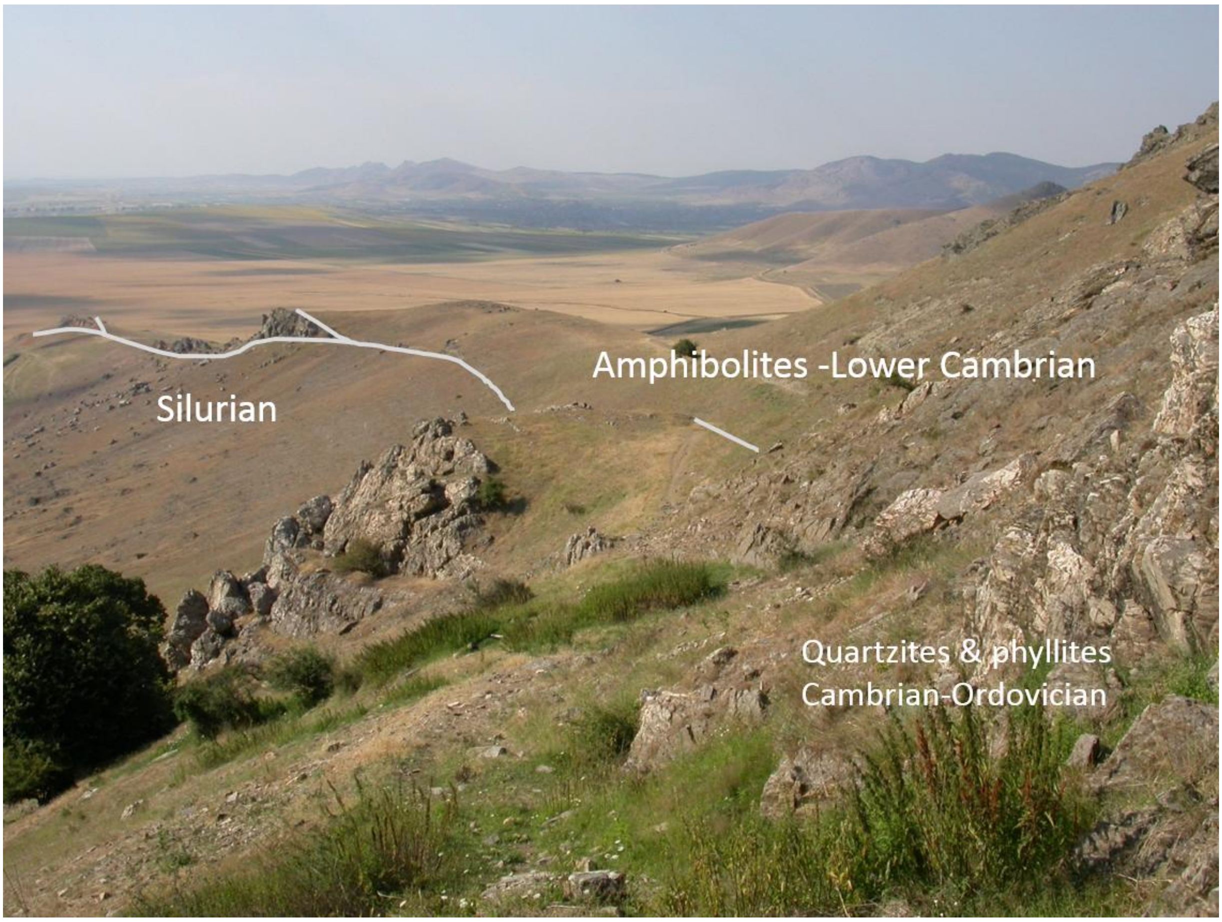


Fig.30. View of the northern part of the Priopcea Hill, showing a reverse stratigraphy: the Megina amphibolites (lower Cambrian) overthrust the Priopcea quartzites and phyllites (Cambrian-Ordovician), and both overthrust the Silurian slates. The hills in the background consist of Late Paleozoic granitoids of the Greci and Pricopan massifs emplaced into the late Carboniferous-early Permian Carapelit Formation. Photo Dr. Antoneta Seghedi.

The Megina Group includes polyphase metamorphic rocks dominated by amphibolites with MORB geochemistry (Seghedi, 1999; Crowley *et al.*, 2000). Their lithological association with quartzites, micaschists, acid metavolcanics rocks and orthogneisses indicates that the oceanic crust protoliths, formed by oceanic spreading, were probably subducted under an island arc.

Megina Group rocks show a complex sequence of metamorphic and deformational events (Seghedi, 1985, 1999). The oldest metamorphic event is a medium pressure metamorphism in amphibolite facies conditions. This early event is strongly overprinted by an upper greenschist (epidote-amhibolite) facies metamorphism, following pervasive mylonitisation at mid-crustal levels. The present structural pattern of the Megina Group rocks resulted by pervasive deformation of the S_2 foliations and associated lineations at shallow crustal levels. The distinct structural style and distinct P-T conditions of D_2 and D_3 deformations, as well as the obvious superposition of D_3 onto D_2 indicate that the two deformational events are separated in time.

Fig. 31. Outcrops in Silurian lithologies exposed on the western slope of the Priopcea Hill. Left: black, pyritic limestones (grey in weathered surface), with a penetrative, steeply dipping slaty cleavage and tight intrafolial folds; right: dark shales, displaying a steep, penetrative slaty cleavage S_1 (dipping left) which is intersecting a steep bedding S_0 (dipping right). Photo Dr. Antoneta Seghedi.

A system of normal faults with NS or NNE-SSW trend deforms the tectonic contact between Megina and Boclugea rocks in the Priopcea hill (Mirăuţă, 1966a; Seghedi, 1986).

U-Pb detrital zircon geochronology indicates that the depositional age of the Megina Group rocks is Cambrian (Balintoni & Balica, 2016). According to monazite dating of metapelites, the age of amphibolite facies metamorphism of these rocks is Hercynian, ranging around 330-280 Ma (late Carboniferous-early Permian) (Seghedi, 2003, unpublished data; Săbău & Negulescu, 2017, unpublished data).

Stop 11 *Quarry on Suluk valley – late Paleozoic biotite granite, showing mylonitic borders and intruded by Triassic dolerite dykes*

Coordinates: 45°14'43.13"N; 28°11'24.03"E

The Pricopan massif is an elongated body of large grained, pink or white biotite granites. It was emplaced into the Carapelit Formation probably in the early Permian. The Pricopan granite is slightly younger than the highly differentiated Greci massif, as proven by field relations south of Suluk, where the granite displays a thin cooling border against the Greci diorites and the Carapelit hornfelses they intrude (Seghedi, 1977). The granite preserves dioritic xenoliths typical for the Greci massif, as well as xenoliths of hornfelses formed on sandstones of the Carapelit Formation. Typically, the Pricopan granite is rich in centimetric, slightly elongated xenoliths of porphyritic, biotite rich, grey microgranites.

The Pricopan massive is bordered by two high angle faults. The western border of the granite is a large proto-mylonitic and mylonitic zone, trending NW-SE and dipping to NE (Fig. 32). There is a progressive increase in deformation toward the western margin, along which Orliga Group mylonites, Silurian slates or hornfelses on sandstones-siltstones of the Carapelit Formation are exposed (Seghedi *et al*., 1980). Close to the fault at the western border, the granite is turned into lower greenschist facies ultramylonites and banded phyllonites, showing a steeply dipping stretching lineation of the former biotite. In borehole 1Măcin and in a few outcrops at the western border of the Pricopan massif, strongly sheared Greci diorites occur, transformed into actinolite - epidote - albite - scapolite mylonites.

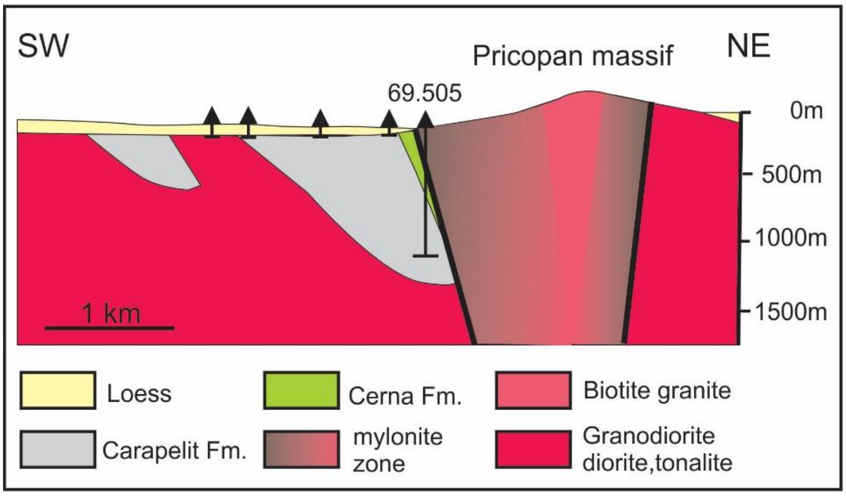


Fig. 32. Cross section in the Pricopan massif, showing its wedge shape and the two bordering faults marked by LS mylonites and phylonites: the eastern border against the granodiorites and diorites of the Greci massif and the western border against the Silurian shales (Cerna Forma-tion) and late Carboniferous-early Permian continental clastics (Carapelit Formation), intruded by Greci type granitoids. Redrawn after Seghedi *et al*. (1999).

The eastern border of the granite is a narrower and steeper ultramylonitic (phyllonitic) zone, with a penetrative mylonitic foliation showing the same NW-SE trend, but dipping steeply to the SW. Fault rocks within this zone are LS mylonites, with a steeply dipping stretching lineation of the biotite, which recrystallized as an aggregate of chlorite, epidote and stilpnomelane. In both mylonitic zones, rock microfabric suggests low temperature deformation, with strain accommodated by various mechanisms of dislocation creep. The presence of mylonitic borders with different dips suggests that the Pricopan massif is a major pop-up structure, formed by inversion of a crestal collapse graben during Cimmerian events (Seghedi *et al*., 1999).

The granitic body is intruded by the system of steeply dipping basic and acid dykes, which are usually highly mylonitic and display the same steeply dipping stretching lineation, resulted by extreme deformation of the primary, magmatic clinopyroxene (Seghedi *et al*., 1980). The dyke swarm is related to the Lower Triassic extension of the Hercynian basement of North Dobrogea (Nicolae & Seghedi, 1996; Seghedi, 2001).

<u>Stop 12</u> *The Orliga Promontory – Variscan metamorphic rocks in the core of Orliga anticline*

Coordinates: 45°17'22.85"N; 28° 6'32.29"E

The Orliga Group, with a restricted outcrop area north and north-east of the Măcin town, is dominated by muscovite-biotite quartzite schists, with thin interlayers of micaschists and plagiogneisses. Minor amphibolites, biotite gneisses and pegmatites (migmatized rocks), as well as calcsilicate marbles occur in its eastern outcrop area (Seghedi *et al*., 1980). Based on the lithological association, the initial sediments of Orliga Group represent an accretionary wedge, with limestones associated to sands and shales, and scarce slices of the basaltic oceanic crust.

Garnet and kyanite-bearing mica quartzites are exposed in the Orliga Promontory, where recumbent folds of the metamorphic foliation are visible (Fig. 33). Rocks of Orliga Group show evidence of three metamorphic and deformational events (Seghedi, 1983, 1999): 1) isoclinal recumbent folding (B_1), in medium pressure conditions of the almandine amphibolite facies (index minerals garnet, kyanite, staurolite, sillimanite in metapelites, and hornblende, andesine, garnet in metabasites) (Seghedi, 1975); 2) tight to isoclinal folding (B_2) in steeply dipping or northward inclined upright folds, with a penetrative axial planar S_2 foliation, E-W trending with northward dip, responsible to the current structure of the Orliga Group; S_1/S_2 intersection gives a constant horizontal or slightly westward plunging lineation of micas on S_2 planes; 3) subsequent shallow level deformation of the S_2 foliation planes resulted in crenulation cleavages S_3 normal to the E-W structure, accompanied by steep, mesoscopic microfolds, with strong kinking of micas and localized mylonite zones with quartz recrystallisation.

Fig. 33. Recumbent folds of the metamophic foliation of Orliga Group rocks in Orliga Promontory.
Photo Dr. Antoneta Seghedi.

U-Pb monazites dating of monazite inclusions in biotites from Orliga Promontory indicate a Variscan age of the amphibolite facies metamorphism (326-282 Ma) (Seghedi *et al.*, 2003). Detrital zircon dating yielded youngest zircons of middle Cambrian age (Balintoni & Balica, 2016), interpreted as the protolith age of the succession.

Day 4
North Dobrogea

Stop 13 *Hora Tepe, Tulcea – relations between the Lower Triassic (Werfenian) fanglomerates and the Devonian siliceous deposits*

Coordinates: 45°11'12.65"N; 28°48'52.22"E

Lower Triassic fanglomerates are exposed in the Hora Tepe (Monument Hill) in Tulcea (Fig. 34). Their clasts are largely represented by the underlying siliceous rocks and trachytes. The fanglomerates, separated as the Monument Hill Breccias of the Bogza Formation (Werfenian) (Baltres, 1993), represent continental, alluvial fan deposits and show a tectonic contact with the underlying Paleozoic basement.

This contact is a Lower Triassic extensional fault that controlled alluvial fan sedimentation. In the lowermost part of the Triassic, continental sedimentation within the Tulcea zone took place in several half-grabens (Baltres, 1993).

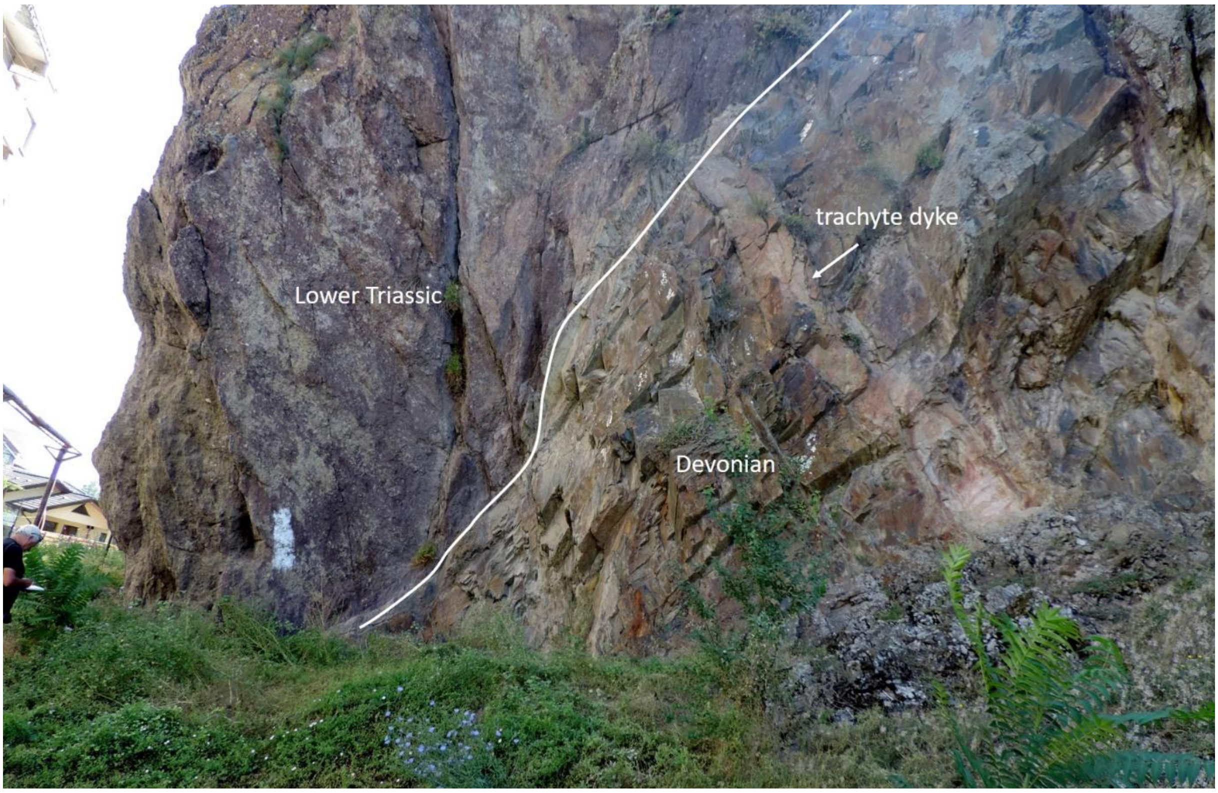


Fig. 34. Red, lower Triassic (Werfenian) fanglomerates in Hora Tepe at Tulcea, overlying Devonian siliceous deposits, the latter intruded by a steep, trachyte dyke. Photo Dr. Antoneta Seghedi.

The Variscan basement in Hora Tepe is represented by thinly bedded siliceous rocks of the Devonian Beştepe Formation (Mirăuţă & Mirăuţă, 1965; Mirăuţă, 1966b, 1967; Mirăuţă, 1971). The siliceous deposits are intruded by two steeply dipping trachyte dykes. Trachytes are pink volcanic rocks, with K-feldspar phenocrysts in a fine-grained groundmass, consisting largely of feldspar and quartz. The dykes belong to the dyke-swarm emplaced in the Tulcea type basement south of the Sfântu Gheorghe Fault (Seghedi *et al*., 1994; Seghedi & Oaie, 1995). Their age is ascribed to the Permian, based on field relations and correlation with the Permian volcanic rocks from the Permian Aluat and Sărata-Tuzla grabens of the Scythian Platform (Seghedi et al., 1994; Seghedi 2001, 2012).

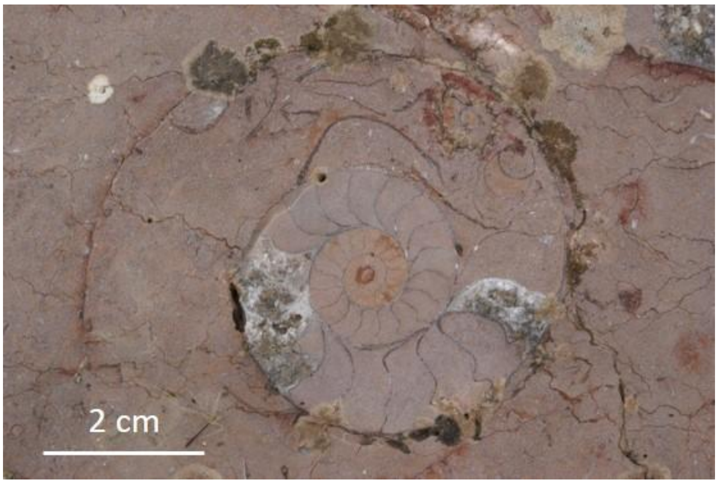
Stop 14 *Agighiol fossil site – late Anisian-middle Carnian cephalopod-rich pink limestones*

Coordinates: 45° 1'37.37"N; 28°51'52.91"E

In the eastern part of Tulcea zone north of Lake Razelm, the Agighiol area is classical for the development of the Triassic deposits in Alpine, Tethyan-type litho-and biofacies. The almost complete succession from late Olenekian (Spathian) to late Carnian is documented here by rich ammonoid faunas. Due to its richness in fossils, as well as scientific, educational and esthetic values, Dealul Pietros is a geological reserve, a paleontological fossil site lately included within the Agighiol Hills Natura 2000 site.

Bedding surfaces on the eastern slope of Dealul Pietros display abundant sections of ammonoids and orthoceratids (Fig. 35). The massive bedded limestones show large-sized late Anisian ptichytids (Grădinaru, 2000). The limestone sequence which straddles the Ladinian-Carnian boundary is the most impressive both as richness, as well as high diversity of the ammonoid faunas. This is the classical locality which delivered the rich ammonoid faunas described in the monographs of Kittl (1908) and Simionescu (1913a). Simionescu (1913b) also collected here three ichtyosaur vertebrae.

Fig. 35. Anisian cephalopods from Dealul Pietros fossil site. Photos Dr. Antoneta Seghedi.



Ceratite

Orthoceras

The Hallstatt-type, pink or red, massive bedded limestones from Dealul Pietros form a continuous succession from the late Spathian to the early Carnian, overlying a lower, light colored dolomite sequence (Grădinaru, 2000). The limestone microfacies shows a progressive transition from late Spathian-early Ladinian wackestones to late Ladinian-middle Carnian packstones. Spectacular laminations, indurated surfaces and neptunian dykes filled with sediments or secondary carbonate represent evidence of condensed sedimentation. Mottling structures are also common in the most part of the Hallstatt-type sequence.

The fossil remains are represented especially by ammonoids and seldom by pelecypods, brachiopods or crinoids (Patrulius *et al*., 1974). Microfaunas are represented by conodonts (Mirăuţă, in Patrulius *et al*., 1974) and ostracods (Crasquin-Soleau & Grădinaru, 1996; Sebe, 2013; Sebe *et al*., 2013) for the whole Hallstatt sequence, while rich foraminifera faunas occur in the late Ladinian - middle Carnian sequence. The ammonoid faunas, although very rich and diverse, are concentrated in some levels in the lower middle Anisian sequence and rather uniformly distributed in the late Anisian to middle Carnian succession. Seven fossil levels were described in the eastern part of Dealul Pietros, and the fauna correlated with ammonoid zones from the Triassic of the Mediterranean Tethyan area and/or from Western Nevada: Kocaelia Zone of the Mediterranean Triassic (Hyatti Zone in Nevada) for the Bythinian (early-middle Anisian); Paraceratites Zone of the early Illyrian – (late Anisian), Eoprotrachyceras Zone of the Late Fassanian (early Ladinnian), Protrachyceras Zone of the Longobardian (late Ladinian) and Trachyceras Zone of the Julian (early Carnian) (Grădinaru, 2000).

Considering that in the Mediterranean Triassic facies, the Aegean (early Anisian) to Bithynian (early middle Triassic) interval is poorly dated on ammonoids, the succession from the Agighiol Hills is extremely important for the improvement of the standard stratigraphic scale of the Mediterranean Triassic.

Among the ammonoid fauna, some taxa known only from the boreal realm were identified, that would contribute to correlation between the Arctic and Tethyan chronostratigraphic divisions of the Anisian. In the northern part of Dealul Pietros, the ammonoid fauna of the Anisian Hyatti Zone is associated with benthic skeletal material (echinoderm sklerites, sponge spicules and brachiopods) (Grădinaru,1996), and this association of bio- and microfacies indicates a microbial-sponge bank paleoenvironment.

Stop 15 *Nalbant Formation – Nalbant village*

Coordinates: 28°36'42.81"E; 28°36'45.86"E

The stratotype of the lower Jurassic Nalbant Formation was described in a ravine in Nalbant village (Patrulius *et al*., 1974). The Nalbant Beds (Sinemurian-lower Pliensbachian) represent a 300 m thick succession of distal turbidites (C and D turbidite facies), a mixture of complete (Ta-e) and base-missing Bouma sequences (Tb-e, Tc-e) capped by well-developed shale divisions (Grădinaru, 2000). The basal divisions include graded or laminated, quartzose sandstones and siltstones. They are interpreted as a distal midfan and outer fan apron. Lower bed surfaces show various types of mechanic or biogenous sedimentary structures, while the upper bed surfaces contain abundant vegetal remains. Macrofossils were not recovered from the Nalbant Beds and their age was established by correlation with other Lower Jurassic formations from North Dobrogea, considered to represent isochronous heteropic formations of the lower Jurassic Nalbant fan (Grădinaru, 2000). The poor microfloral assemblage, consisting mainly of vegetal tissue, with rare pollen and microspores, identified in this location, yielded *Stereisporites pefforatus*, *Dictyophyllidites* sp., *Auritulinasporites* sp., *Undulatisporites* sp., *Classopolis classoides*, *Monosulcites* sp., which are associated with species found in Hettangian-Pliensbachian interval (Antonescu & Baltres, 1997).

Stop 16 *Lower Jurassic terrigenous turbidites with Zoophycos traces at Poşta*

Coordinates: 45° 6'35.27"N; 28°36'42.81"E

According to Grădinaru (1984), the lower Jurassic deposits of the Poşta Sandstones (Patrulius *et al*., 1974) are included in the Jurassic Nalbant Formation, dominated by dark siltstones and clays, in several places with sandstones forming turbidite fans (Fig. 36). A thick succession of terrigenous turbidites dominated by sandstones is exposed south of Poşta locality.

On an exposed thickness of 90 m, the upper member of the Teliţa Formation and part of the Nalbant Fan (Grădinaru, 1984), the Poşta Sandstone (Sinemurian-Lower Pliensbachian) consists largely of amalgamated, medium-to thick-bedded, massive, argillaceous quartzose sandstones and siltstones (B_2 and C turbidite facies). Sandstone beds show a discontinuous grading, from medium grained sandstone with parallel lamination to fine grained sandstone and mudstone, with abundant traces of *Zoophycos*, seldom *Rhizocorallium* and *Ophiomorpha* (Baltres & Antonescu, 1998). Tae Bouma divisions seldom occur, consisting of thick-bedded, graded argillaceous quartzose sandstones topped by very thin shale divisions.

The Poşta sandstone was ascribed to the Sinemurian-lower Hettangian based on Rotiforme, Jamesoni and Ibex ammonite Zones, (Grădinaru, 1995). Ammonites identified in Poşta sandstone include *Trepidoceras flandrini*, *Uptonia jamesoni*, *Platypleuroceras* sp., Lithoceratids, Phylloceratids, *Mytilinoceramus* spp., *Arietites* sp., *Coroniceras* sp. (Grădinaru, 1984).

Fig. 36. Ichnogenous *Zoophycos* in lower Jurassic sandstones at Poşta. Photo Dr. Antoneta Seghedi.

On a hill south of Poşta village, on the northern bank of the Teliţa river, well-preserved and abundant large sized *Zoophycos* traces (the flat variety) occur in the Poşta sandstones (Fig. 36). *Zoophycos* is an environment-indicating trace fossil, well known in the lower Jurassic sandstones of the Tulcea zone (Grădinaru & Seilacher, 2009). It represents a feeding trace through systematic exploitation of the sediment in search of organic matter. The burrowing organism is often digging helicoidal channels.

Stop 17 *Revărsarea quarry – late Scythian-Anisian pillow basalts (Niculiţel formation)*

Coordinates: 45°15'56.58"N; 28°22'29.69"E

About 1200 m south on the quarry road located at the western margin of Revărsarea village, pillow basalts of the Niculiţel Formation (Niculiţel tectonic unit) are well exposed in the walls of an abandoned quarry (Fig. 37). Both the eastern and western quarry walls show thin, lens-shaped bodies of pink limestones interbedded in the pillow basalts. Individual pillows are decimetric in size and show a vitreous, strongly brecciated matrix. Biostratigraphic studies and geological mapping in the Assan Hill revealed that early and middle Anisian limestones are interbedded with pillow basalts (Mirăuţă,1982; Savu *et al*., 1988).

The succession of the Niculiţel Formation, ascribed to the Spathian-middle Anisian based on the interbedded conodont-bearing limestones (Mirăuţă, 1982), consists of pillow and massive basalts, associated to minor pyroclastics and gabbroic bodies (Savu *et al*., 1980; Savu, 1986). The basic rocks are interbedded with thin volcaniclastic and carbonate turbidites, basaltic and epiclastic breccias and pelagic limestones (Baltres *et al*., 1994, unpublished report). The basic rocks are 350-500 m thick at Niculiţel, thinning considerably to the north, east and south-east. According to the previously cited authors, the facies associations of the Niculiţel Formation can originate in a system of marine half-grabens, with debris flows, turbidites and talus breccias controlled by an active footwall scarp, while red pelagic carbonates with markedly condensed facies represent starved axial basin environments; eastward transition to carbonate platform sediments reflect the gentler hanging-wall slope, with carbonates directly overlying the pre-rift basement.

Fig. 37. Pillow basalts in Revărsarea quarry, Assan Hill. Photo Dr. Antoneta Seghedi.

Based on geochemical features, an intraplate setting was suggested for the Niculiţel Formation, formed as flood basalts (Savu, 1986). The basalts show the same geochemical signature as the dolerite dykes from the Măcin zone and plot in the same fields on the tectonic discrimination diagrams (Nicolae & Seghedi, 1996). However, although basalts of the Niculiţel Unit have E-MORB composition, this unit does not display a typical ophiolitic stratigraphy, due to the lack of sheeted dykes, mantle tectonites and/or mafic-ultramafic cumulates (Cioflica *et al*., 1980; Savu, 1986). Niculiţel basalts show E-MORB affinity and possibly represent melts derived from depleted NMORB-type mantle sources strongly metasomatized by OIB-type components (Saccani *et al*., 2004).

As emplacement of these E-MORB-type basalts was preceded by a bimodal basalt-rhyolite volcanism (Savu, 1986), it is likely that they were extruded either during the final rifting stage of rift that became subsequently inactive, or during the crustal separation stage of a typical passive margin.

Stop 18 *Pârlita abandoned quarry– Devonian subduction complex intruded by dykes of the basalt-trachyte bimodal association*

Coordinates: 45° 8'1.15"N; 28°58'7.90"E

Located north of Victoria village, 400 m north from the Tulcea-Murighiol national road, the Pârlita abandoned quarry exposes siliceous shales and cherts, associated with distal turbidites (Fig. 38). Both successions were and ascribed to the Devonian based on conodonts identified in thin, pelagic limestones, interbedded in places in the siliceous deposits (Mirăuţă & Mirăuţă, 1965; Mirăuţă, 1967; Mirăuţă, 1971) and included in the Beştepe Formation (Patrulius *et al.*, 1973). Siliceous shales are rich in radiolarians, variously replaced by secondary silica. Distal turbidites show tight, vertical folds, and often trace fossils.

Several steeply dipping, thin trachyte dykes intrude the Devonian deposits in the northern wall of the quarry (Fig. 38), but an alkali basalt dyke is also observed in the eastern wall. They belong to a system of E-W to ESE-WNW steeply dipping dykes of the basalt-trachyte bimodal association (Seghedi, in Baltres *et al*., 1989; Seghedi & Oaie, 1995).

The subvolcanic rocks display intense hydrothermal alteration which obscures their initial petrography. Trachytes are porphyritic rocks, with alkali feldspar phenocrysts in a feldspar dominated matrix, showing weaker hydrothermal alterations, with incipient replacement of feldspars by carbonates and illite. Basic rocks are basalts, in places olivine basalts, with dolomite and iron carbonates replacing olivine and pyroxene, associated with various amounts of chlorite and secondary silica (Seghedi *et al*., 1994). A pre-Triassic age of the dyke-swarms is indicated by field relations (Mirăuţă, 1966 b). The dykes intrude all the Paleozoic successions from Tulcea zone, including various lithologies of the Devonian Beştepe Formation (Seghedi *et al*., 1994) and are reworked in lower Triassic conglomerates. The bimodal association shows similar petrographic and geochemical features, and even the same type of extensive hydrothermal alteration, as the Permian volcanic rocks from the Scythian Platform, where the Permian age is proven both by K-Ar radiometric ages and relations with the fossil-bearing Permian red-beds (Neaga & Moroz, 1987).

Geochemical features of the dyke rocks, despite migration of silica and even of immobile elements (Zr, Ti), reveal the bimodal character of magmatism and suggest intraplate setting (Seghedi *et al*., 1994). The bimodal association and the presence of high K trachytes suggest magma generation in a complex extensional setting, probably transitional between an active continental margin and extensional (transtensional) intraplate setting.

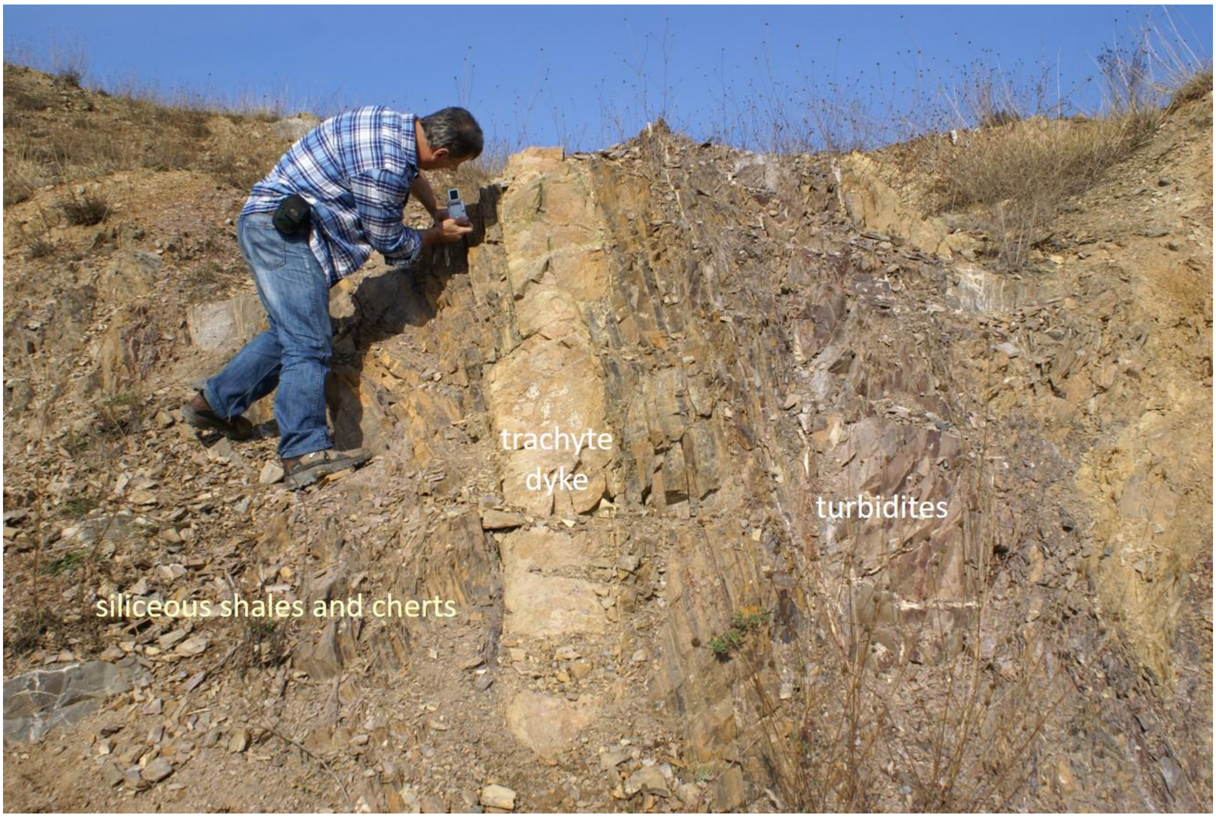


Fig. 38. Beştepe Formation in Pârlita Quarry. Distal turbidites and siliceous shales and cherts, intruded by a trachyte dyke ascribed to the late Permian. Photo Dr. Antoneta Seghedi.

Accreted terranes of the Beştepe Formation are well exposed in the Beştepe Hills, situated to the east. The lithological association of the siliceous deposits is dominated by bedded chert and siliceous shales; minor iron carbonates and iron sulphates occur as thin (subcentimetric) interbeds in the siliceous deposits, as documented by several boreholes in the Beilia Mare Hill (west of Beştepe village) (Seghedi *et al*., 1993). The thickness of pelagic deposits attains 700 meters in boreholes, as the result of tectonic thickening. Siliceous deposits are rich in radiolarians, variously replaced by diagenetic silica or carbonates.

Distal turbidites form a thin bedded sequence, dominated by thin Tcde Bouma divisions. Lithofacies Tc consists of ripple laminated calcareous limestones, usually with thicknesses of 1-3 cm (Oaie & Seghedi, 1994). Thicker orthoquartzitic sandstone beds are restricted to the contact with the pelagic sequence. In turbidites, biogenic sedimentary structures are preserved mostly at the top of the calcareous sandstone beds, associated with ripple marks and flute casts (Mirăuţă & Mirăuţă, 1965; Mirăuţă, 1967). In a few localities, the trace fossils occur in pelitic rocks. The trace fossil assemblage of the Beştepe turbidites, including *Chondrites*, *Helminthoida*, *Helminthopsis*, *Protopalaeodiction*, etc., belongs to the Nereites zone (Oaie, 1989; Oaie & Brustur, 1999), which is found at the distal parts of the turbiditic fans. They indicate accumulation at water depths exceeding 800 m. Other evidence for deep water deposition of the turbidites include: thin sandstone beds; good lateral bed continuity; low sandstone/mudstone ratios; ichnofauna confined to the upper parts of the Bouma division, between divisions Td and Te, and locally to Te (Oaie & Seghedi, 1994; Oaie, 2001). Sedimentological features suggest that the Beştepe turbidites have accumulated in a sediment starved, ponded basin (Seghedi & Oaie, 1995; Seghedi, 2012).

The complex map pattern of the Beştepe terrane is the result of complex deformation and erosion. The structural style consists of large scale, Variscan recumbent folds (Mirăuţă, 1967) and thrust-folds, refolded by normal folds (Seghedi, 2012). Recumbent folding and thrusting is accompanied by the development of a penetrative, spaced slaty cleavage with various morphologies (Seghedi & Rădan, 1989). The cleavage is penetrative in siliceous rocks, but not in turbidites, where it occurs only in places. Petrographic evidence indicates that the slaty cleavage was formed by pressure solution and cleavage formation is accompanied by recrystallisation of illite and chlorite.

The very low grade metamorphism of the siliceous deposits has occured in low temperature conditions, indicated by the local presence of prehnite+quartz veins. The slaty cleavages are overprinted by crenulation cleavages, related to Cimmerian deformation. In outcrops, mesoscopic normal folds with various morphologies are visible, usually with rounded anticline hinges and narrow, tight synclines. These second generation folds with E-W trends are related to Cimmerian refolding of the Variscan basement.

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