GEOLOGICA BELGICA (2005) 8/4: 78-105.

SEDIMENTOLOGY AND SOURCE AREA COMPOSITION FOR THE NEOPROTEROZOIC-EOCAMBRIAN TURBIDITES FROM EAST MOESIA

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(8 figures, 3 tables, 4 plates)

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ABSTRACT. The basement of the Moesian Platform is extremely heterogeneous, including rocks ascribed to the Archean, Palaeoproterozoic and Neoproterozoic. Largely concealed by the Palaeozoic to Cenozoic platform cover, the basement exposed in the tectonic block of Central Dobrogea contains mainly Neoproterozoic-Eocambrian turbidites (the Histria Formation), deformed under very low grade metamorphic conditions at the end of the Neoproterozoic, in the "Cadomian" or "Baikalian" events. The Histria Formation includes channelized, midfan turbidites and distal, outer fan turbidites, forming two sandstone dominated, upward coarsening and thickening sequences, separated by a much thinner, upward fining and thinning sequence. Sedimentological, mineralogical and petrological studies are consistent with the model that the turbiditic basin was sourced by a continental margin dominated by an active volcanic arc. From various lines of evidence, a foreland basin as the tectonic setting for the turbiditic sedimentation is favoured, although a forearc origin was also proposed. The paper presents an overview of the main sedimentological, mineralogical and petrographical data on the turbidites, and discusses the palaeogeographic and geodynamic significance of the succession in a wider context of the southern margin of the East European Craton.

KEYWORDS. Central Dobrogea, foreland basin, Moesian Platform, Neoproterozoic, turbidites

1. Introduction

A belt of Neoproterozoic-Early Cambrian turbidites is lying at the western margin of the East European Craton, extending NW from the Black Sea shore in Romania to the Malopolska Massif in Poland (Murgoci, 1911; Glowacki & Karnkowski, 1962; Glowacki *et al*., 1963; Unrug *et al*., 1999) (Fig. 1A). This belt of turbidites, with a width of 40-50 km in outcrop and a total length of about 800 km, abuts to the NW against Far Eastern Avalonia, which represents the easternmost part of the Avalon terrane from Central Europe (Winchester *et al*., 2004, 2005). The belt is largely concealed, being covered by younger deposits or by the outer East Carpathian nappes. The only outcrops of these turbidites occur in Central Dobrogea, an elevated tectonic block at the northern margin of the Moesian Platform (Fig. 1A). The turbidites from Central Dobrogea were referred to as "the Green-schists series", due to their dominant green coloration (Mrazec, 1910, 1912; Murgoci, 1911; Mirăuță, 1964, 1965, 1969; Săndulescu, 1984; Visarion *et al*., 1988). As turbidites show no greenschist facies metamorphism, nor a metamorphic foliation or schistosity, the name of Histria Formation (Seghedi & Oaie, 1995) will be used further to designate this turbiditic succession. The continuity to the NW of the turbidite belt is indicated by several lines of evidence, as follows: (a) Geophysical, especially magnetic information, suggests that the belt of turbidites from Central Dobrogea prolongates NW underneath the Cretaceous-Tertiary nappes of the East Carpathians (Gavăt *et al*., 1963; Airinei, 1980); (b) Boreholes in the Outer Carpathians from Poland intercepted turbidite rocks at depths between 778 and 2760 m (Glowacki *et al*., 1963); (c) A comparative study of the turbiditic successions from borehole cores in Malopolska and exposures in Central Dobrogea revealed similarities in their sedimentology, petrography, provenance and structural features (Żelaźniewicz *et al*., 2001a, b); (d) Miocene mollasses from the East Carpathians nappes rework green clasts of these ancient turbidites, exotic to the Carpathian area (Mrazec, 1910; Murgoci, 1911; Atanasiu, 1940; Săndulescu, 1984). The source delivering exotic clasts was assumed to be a cordillera located in the foreland of the East Carpathians (Murgoci, 1911; Atanasiu, 1940). Sedimentological studies indicate that these exotic clasts were delivered from an eastern source (Mărunțeanu, 1987), confirming the foreland location of this source. We suppose that this eastern source area, now buried, possibly represented the forebulge of the East Carpathians Foredeep in Palaeogene and Miocene times, and both Central Dobrogea and Malopolska block were parts of it.

This paper presents an overview of the available sedimentological, petrographic, structural and provenance information on the Histria Formation turbidites from Central Dobrogea and discusses the geodynamic significance of the turbidite succession.

2. Geological and tectonic framework of Moesia

The Moesian Platform represents a lowland area, surrounded to the north and west by the Alpine belt of the South Carpathians and to the south by the Balkans (Fig. 1B). The Eastern platform margin is represented by the Peceneaga-Camena Fault (Mrazec, 1912), which runs NW from Lake Razelm on the Black Sea shore to the Vrancea zone of the East Carpathians bend zone and juxtaposes the exposed Neoproterozoic platform basement against the Triassic and Jurassic sediments of the Cimmerian North Dobrogea orogen (Figs 2A & 2B). Seismic data indicate that a 10 km step of the Moho occurs along this fault (Rădulescu *et al*., 1976).

The Moesian Platform consists of two segments separated by the Intramoesian Fault: East Moesia, with a Neoproterozoic ("Baikalian-Cadomian") and Caledonian ("Ardennian)" cratonization, and West Moesia with a Variscan cratonization of the basement (Răileanu *et al*., 1968; Murgeanu & Patrulius, 1973). East Moesia is bounded by two NW-SE trending crustal faults: the Peceneaga-Camena Fault to the NE and the Intramoesian Fault to SW (Săndulescu, 1984; Săndulescu & Visarion, 1988; Visarion *et al*., 1988) (Fig. 1B). To the NW, East Moesia is underthrusted beneath the East Carpathians bend zone along the Pericarpathian Fault (Airinei, 1982). Eastward, East Moesia continues in the Black Sea offshore, as indicated by geophysical and borehole data (Romanescu *et al*., 1972; Săndulescu, 1980, 1984; Cătuneanu, 1992, 1993) (Fig. 3).

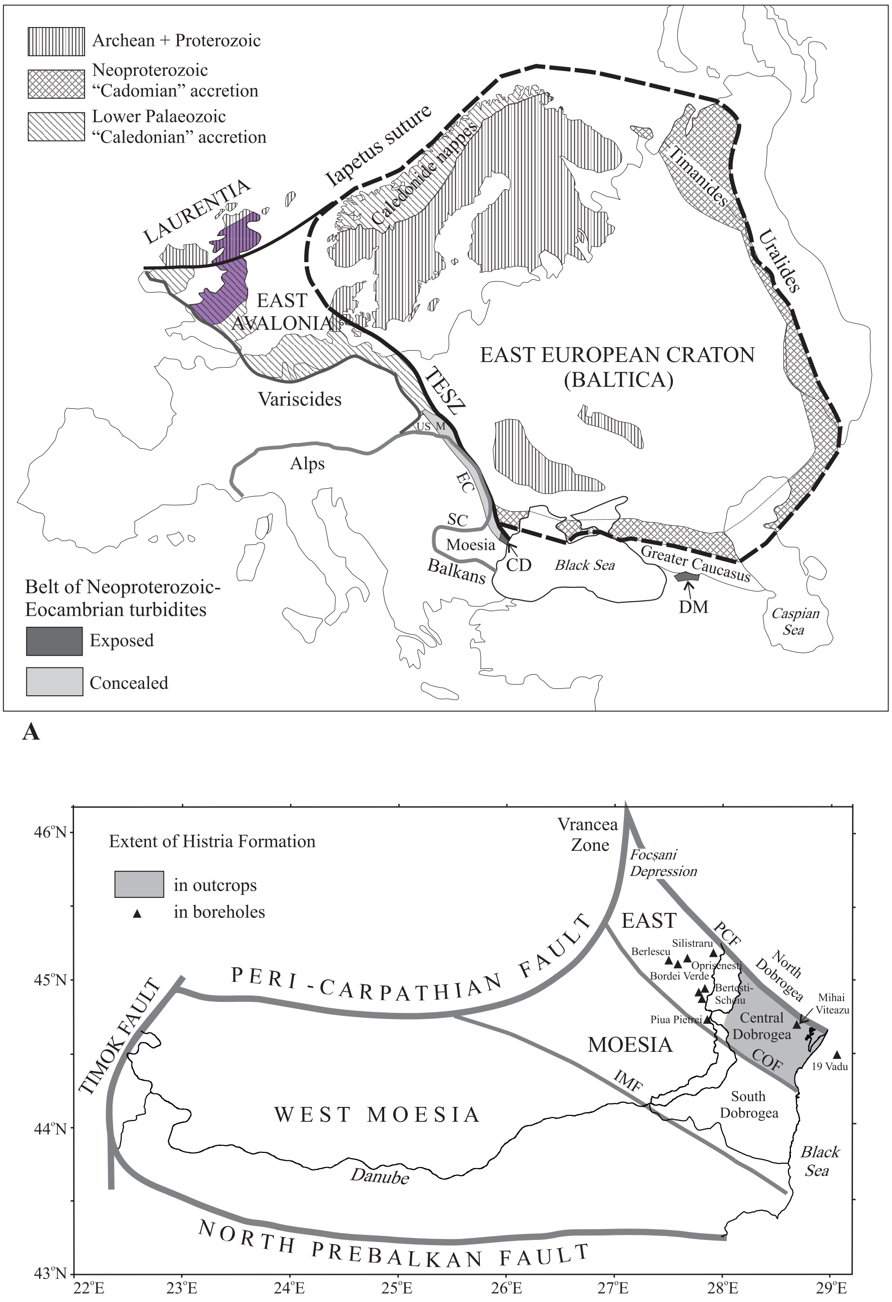


Figure 1. Location of Moesia and Central Dobrogea. A) Geological sketch map of the basement structure of Europe, showing the location of the Neoproterozoic turbidite basin at the western margin of the East European Craton. This belt is exposed in Central Dobrogea (CD) and is concealed to NW by younger sediments and by underthrusting beneath the East Carpathians nappes. TESZ: Trans-European Suture Zone; M: Malopolska; US: Upper Silesia; EC: East Carpathians; SC: South Carpathians; DM: Dzirula Massif. B) Location of Central Dobrogea within the Moesian Platform. The Neoproterozoic-Eocambrian basement is exposed only in the NE part of East Moesia, between the Danube and the Black Sea. Drillholes have reached this basement beneath Ordovician clastics (NW of the Danube) and beneath Late Jurassic limestones (in the Black Sea). PCF: Peceneaga-Camena Fault; COF: Capidava-Ovidiu Fault; IMF: Intramoesian Fault.

The only outcrops of pre-Cenozoic rocks from East Moesia occur east of the Danube in the Dobrogea highlands. Central and South Dobrogea, the main tectonic blocks separated by the Capidava-Ovidiu Fault, show a different geological constitution. Central Dobrogea represents an uplifted block, exposing the Neoproterozoic Moesian basement (Figs 1B & 2B). In South Dobrogea, the complex Precambrian basement is concealed by the Phanerozoic cover and only Cretaceous and Cenozoic formations are exposed.

West of the Danube, the rocks exposed in Central and South Dobrogea plunge beneath the platform cover. Between the Peceneaga-Camena and the Capidava-Ovidiu Faults, boreholes west of Central Dobrogea intercepted the basement overstepped by Ordovician or Silurian sediments at depths of about 2200 m, dipping steeply to NW (Murgeanu & Spasov, 1968) (Fig. 1B).

The structure of East Moesia is characterised by NW-SE trending faults, largely known from gravity and magnetic data. Apart from the main crustal faults, other faults affecting the East Moesian basement are shown in Fig. 2A: the Ostrov-Sinoe, the Istria and the Horia Faults in Central Dobrogea and the Agigea and Eforie Faults in South Dobrogea (Botezatu & Băcioiu, 1957; Gavăt *et al.*, 1965, 1967; Barbu & Vasilescu, 1967; Barbu, 1973; Airinei, 1980; Paraschiv, 1982; Botezatu *et al.*, 1984; Visarion *et al.*, 1988, 1990).

3. Metamorphic rocks of the East Moesian basement and relationship with the Histria Fm.

3.1. The basement of Central Dobrogea

In Central Dobrogea, metamorphic rocks of the Altîn Tepe Group are exposed directly south of the Peceneaga-Camena Fault, forming an antiform which closes periclinally to SE beneath the Histria Formation (Figs 2 & 6). The metamorphic succession includes metapelites and metabasites with an amphibolite facies metamorphism in staurolite grade conditions (Mureşan, 1972).

K-Ar ages in the Altîn Tepe Group range from 696-643 Ma (yielded by biotite from micaschists, Giuşcă *et al.*,

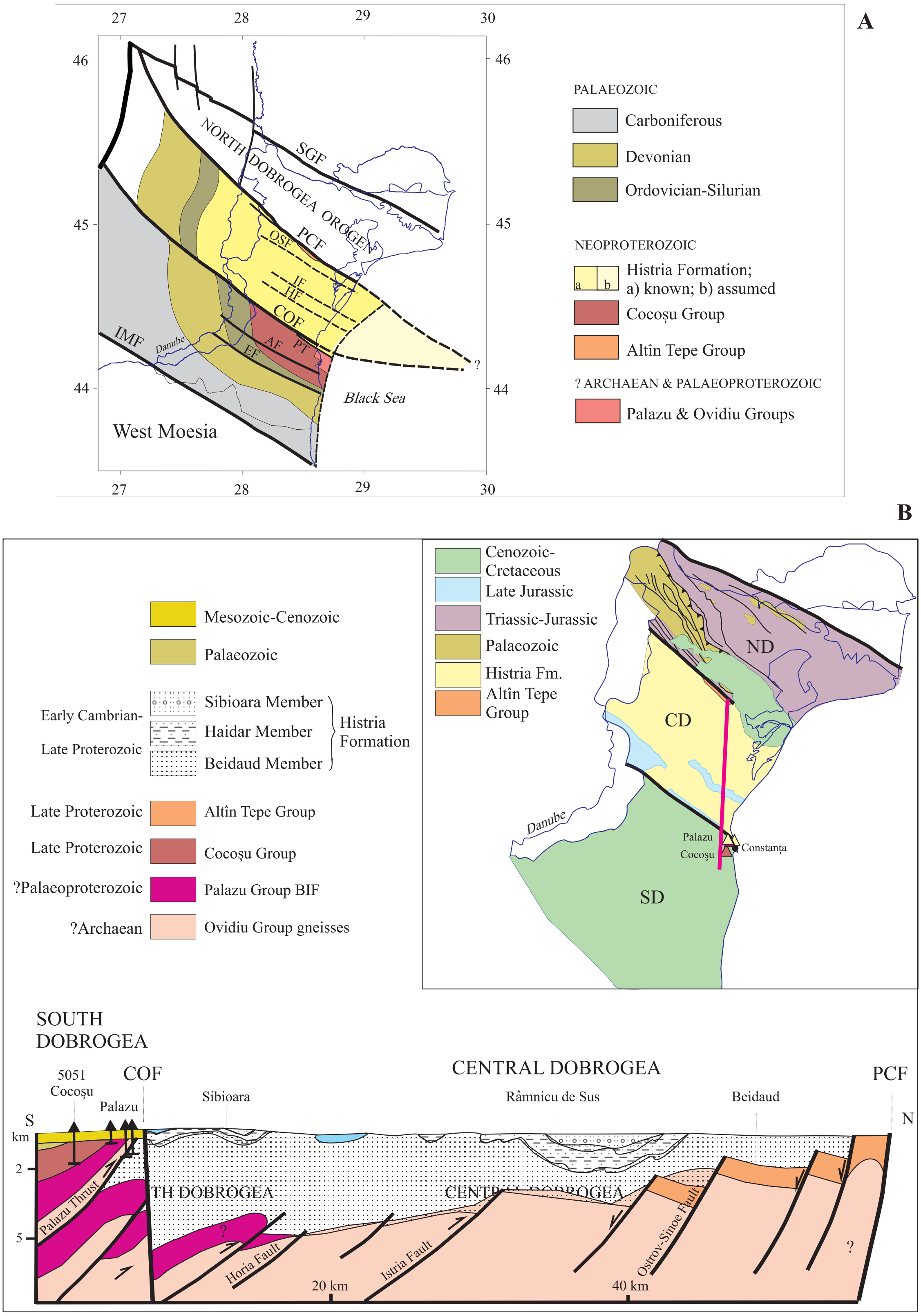


Figure 2. A) Mesozoic subcrop map of East Moesia, showing the distribution of the basement rocks and of Palaeozoic overstep successions. The basement of Histria Formation is affected by WNW-ESE running faults, known from various geophysical data (Mirăuță, 1969; Visarion *et al.*, 1979, 1988, unpublished data 1986). SGF: Sfantu Gheorghe Fault; PCF: Peceneaga-Camena Fault; COF: Capidava-Ovidiu Fault; IMF: Intramoesian Fault. B) Schematic geological cross-section across the eastern part of Central Dobrogea and the NE part of South Dobrogea (modified after Visarion *et al.*, 1988 and Kräutner *et al.*, 1988). Inset is the simplified geological map of Dobrogea. Line of cross-section is shown in red on the inset map. The progressive southward sinking of the metamorphic basement of Central Dobrogea shown in the geological section is indicated by the Δ Ta magnetic map (Visarion *et al.*, 1988; Dimitriu, 2001). The presence of the Palazu and Altîn Tepe Groups in the basement of Central Dobrogea is inferred on the basis of magnetic anomalies, produced by sources located in the basement of the Histria Formation (Airinei, 1967, 1980; Romanescu *et al.*, 1972; Visarion *et al.*, 1988, unpulished data 1986; Dimitriu, 2001). The presence of the Ovidiu Group gneisses in the same area is inferred from gravity and magnetic data (Botezatu & Băcioiu, 1957; Botezatu *et al.*, 1984), correlated with the physico-chemical properties of rocks from outcrops and borehole cores (Visarion *et al.*, 1979, 1988; Săndulescu & Visarion, 1988). ND: North Dobrogea; CD: Central Dobrogea; SD: South Dobrogea; SGF: Sfântu Gheorghe Fault; PCF: Peceneaga-Camena Fault; COF: Capidava-Ovidiu Fault; IMF: Intramoesian Fault.

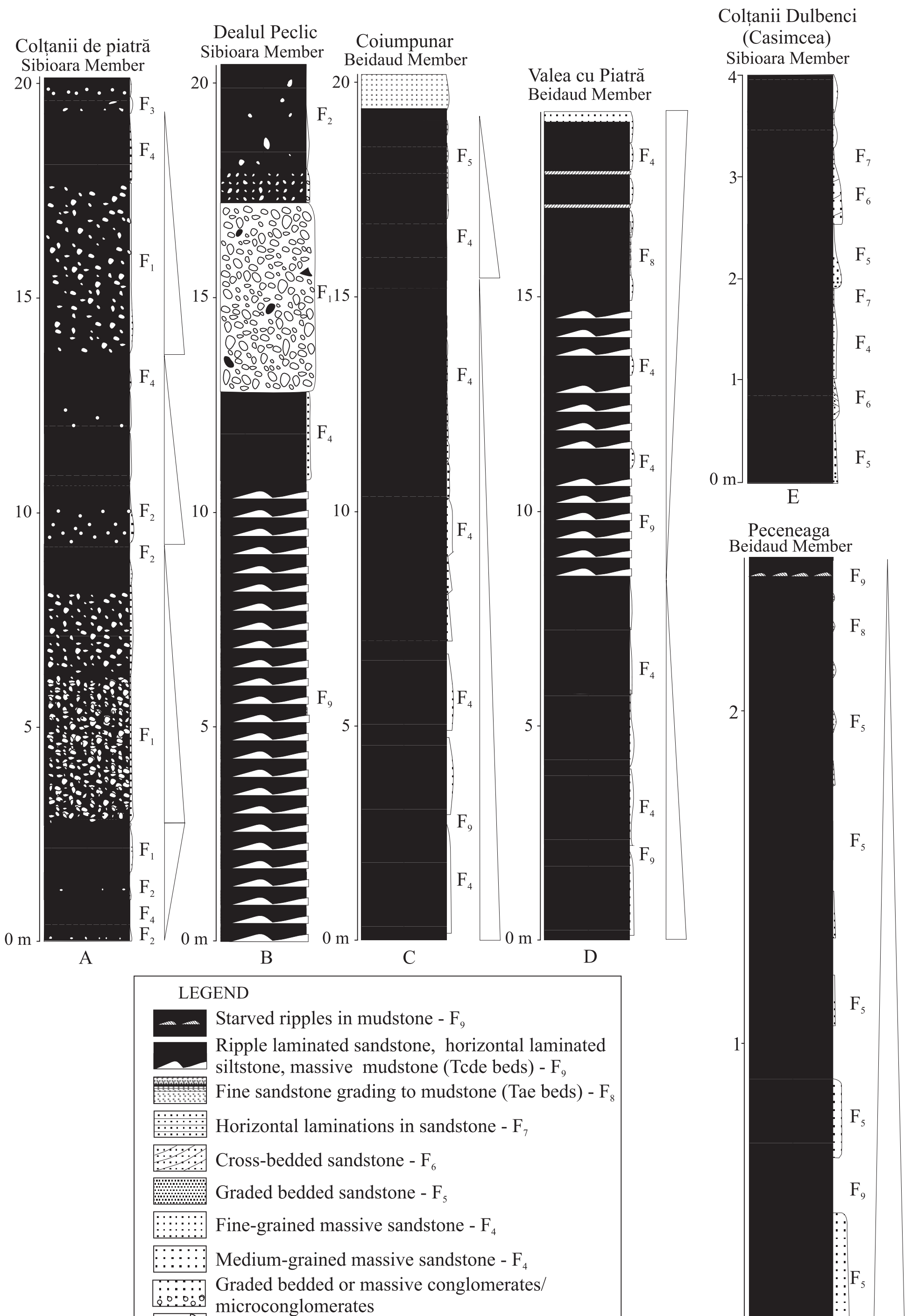


Figure 3. Lithofacies types and facies associations in coarse-grained members of the Histria Formation.
A) Three superimposed thinning and fining upward sequences of conglomerates to sandstones, indicating channel fill of the upper fan;
B) Coarse channel fill sediments on top of fine-grained, amalgamated Tcde Bouma divisions (distal depositional lobes), suggesting progradation of the upper fan;
C) Two superimposed upward thinning and fining sequences, representing channel fill sediments; vertical distribution of facies indicates progressive abandonment by lateral channel migration;
D) A lower thinning upward cycle dominated by massive sandstone beds (facies F_5), indicating channel fill sediments of the middle fan, is overlain by a thickening and coarsening upward sequence, representing a depositional lobe and suggesting progradation of the middle fan;
E) Channel deposits on the middle fan; F) Upward fining and thinning sequence indicating progressive channel abandonment.

1967) to 526 Ma (on hornblende from amphibolites, Kräutner *et al.*, 1988) (all ages recalculated by Kräutner *et al.*, 1988). Based on these K-Ar data, the metamorphism of the Altîn Tepe Group was ascribed to the Late Proterozoic (Giuşcă *et al.*, 1967; Codarcea-Dessila *et al.*, 1966; Mirăuță, 1969), or to the Early (Giuşcă *et al.*, 1969) or Middle Proterozoic (Kräutner & Savu, 1978; Kräutner *et al.*, 1988). Kräutner *et al.* (1988) interpreted the Neoproterozoic ages of the Altîn Tepe rocks as indicating partial Ar loss during their regional retrogression, connected to the very-low grade Cadomian metamorphism of the Histria Formation. A common metamorphism affecting both successions might suggest that the Histria Formation was lying on top of the Altîn Tepe Group before this metamorphic event. The presence of the Altîn Tepe rocks beneath the turbidites is confirmed by boreholes along their contact (Seghedi & Oaie, 1993). From magnetic data it is inferred that the Altîn Tepe Group extends southward in the basement of the Histria Formation at least to the Ostrov-Sinoe Fault (Fig. 2B) (Visarion *et al.*, 1988; Dimitriu, 2001 and references therein). From this geometric position it might be inffered that the Histria Formation accumulated in a basin floored by metamorphic rocks of Altîn Tepe type.

The contact between the Altîn Tepe Group and the Histria Formation is tectonic, marked by a zone of ductile, greenschist facies mylonites, which formed on the Altîn Tepe micaschists and paragneiss protoliths (Mureşan, 1971; Seghedi & Oaie, 1993). In the antiformal closure area, the mylonites display a gently SE plunging stretching lineation. This mylonitic zone was interpreted as evidence for the nappe of the "Greenschists" (Histria Formation), overthrusting the Altîn Tepe Group, which is here exposed in a tectonic window (Mureşan, 1971). However, this tectonic contact is not necessarily related to contractional deformation (Săndulescu, 1984). It might be as well related to extensional deformation at mid-crustal depths, consistent both with the geometry and the type of deformed rocks developed along the contact: older rocks with a pervasive ductile mylonitization in greenschist facies conditions, forming the core of an anticline, overlain by younger rocks with only brittle deformation. The presence of mica fishes, as well as of extended amphiboles in the Altîn Tepe mylonites along the contact, is also consistent with a low-angle extensional detachment. Possibly, the exposures of Altîn Tepe rocks in the anticline core are the result of unroofing of a metamorphic core complex (Seghedi *et al.*, 1999). The nature of the tectonic relationships between Altîn Tepe Group and Histria Formations is important for establishing the basement of the Histria Formation. In case the Histria Formation overthrusts the Altîn Tepe Group, the latter is not necessarily the basement of the former. In case of an extensional detachment, the Altîn Tepe Group represents, at least partly, the basement of the turbidites.

According to Săndulescu (1984), the age of the tectonic discontinuity between the Altîn Tepe Group and the Histria Formation can be at least coeval with the deformation of the latter in very low-grade metamorphic conditions. The tectonic discontinuity was superimposed on the surface of the former stratigraphic unconformity, separating the two rock complexes prior to the metamorphism of the former.

3.2. The basement of South Dobrogea

In the subsurface of South Dobrogea, the cratonic oldest basement was pierced by boreholes SE of the Capidava-Ovidiu Fault at depths between 430 and 800 m. The basement consists of orthogneisses (the Ovidiu Group) and an iron-bearing formation (Palazu Group) (Giuşcă *et al.*, 1976; Visarion *et al.*, 1979; Kräutner *et al.*, 1988) (Fig. 2). Based on detailed petrographic and mineralogical studies, as well as on geochronological data, these metamorphic rocks were correlated to the Archaean gneisses and the Early Proterozoic banded iron formation (Krivoy Rog) from the Ukrainian Shield (Ianovici & Giuşcă, 1961, Giuşcă *et al.*, 1967, 1969, 1976; Visarion *et al.*, 1979). The oldest K – Ar ages range between 1777 Ma (on K feldspar) and 1620 Ma (on muscovite) (Giuşcă *et al.*, 1967). A second age group includes Neoproterozoic ages (867 Ma whole-rock age on micaschist and 729-644 Ma on K feldspar from granite gneiss) (all ages recalculated by Kräutner *et al.*, 1988). A later, greenschist facies retrogression, was correlated with the late Cadomian very low-grade metamorphism of the Late Proterozoic cover (Cocoşu Group) (Kräutner *et al.*, 1988).

A volcano-sedimentary succession, situated on top of the Palazu Group and known as the Cocoşu Group (Giuşcă *et al.*, 1967; Visarion *et al.*, 1979; Kräutner *et al.*, 1988), includes syn-rift alkaline mafic volcanics and volcaniclastics with intraplate affinities (Seghedi *et al.*, 2000), overlain by a terrigenous member of fine-grained green and purple clastics. Petrographic studies indicate that the entire succession of the Cocoşu Group shows a very low-grade, subgreenschist facies metamorphism. Details on the Cocoşu Group are given in Seghedi *et al.* (this volume). A Neoproterozoic age was proposed for the Cocoşu Group by correlation with the Histria Formation, based on their very low-grade metamorphism (Mirăuţă, 1965, 1969; Giuşcă *et al.*, 1967). However, the precise age of the Cocoşu Group is unknown. Whole rock K-Ar ages of 550 Ma indicate an end Neoproterozoic age of its subgreenschist facies deformation (Giuşcă *et al.*, 1967; Kräutner *et al.*, 1988). Even if stratigraphic relations indicate an Upper Proterozoic age, it is not known if the Cocoşu Group is coeval with the Histria Formation, or if it is older, Lower Vendian or Ediacaran (Riphean). Importantly, petrographic studies indicate significant differences in lithology and tectonic setting between the Histria Formation and the Cocoşu Group.

Grain size	Facies code (Mutti, 1992)	Facies characteristics	Interpretation	Depositional mechanisms
Very coarse grained sandstones and conglomerates (Facies A of Mutti & Ricci-Lucchi, 1972	F_1	Lenticular and massive beds, erosive base, rip-up clasts, clasts at the top, mudstone matrix, weak sorting	Channellized debris-flows	Dense turbidite current
	F_2	Lenticular beds, erosive base, conglomerates to very coarse sandstones, rip-up clasts and clasts at the base of the bed, sandstone or siltstone matrix;,weak sorting, clast imbrications, normal graded bedding, unclear laminations	Channellized debris-flows	Concentrated flows
	F_3	Lenticular bed, erosive base, clast-supported conglomerates to microconglomerates, sandstone matrix, massive or graded bedding	Channel fill	High density flows
Coarse grained sandstones (Facies B, of Mutti & Ricci-Lucchi, 1972)	F_4	Amalgamated beds of coarse sandstones, shaped bases and tops, sometimes flame structures in the bases, poorly sorted, rip-up clasts at base, laminations marked by clasts, graded bedding, discontinuous parallel lamination marked by gravel size clasts	Channel fill	Supracritic flows
	F_5	Lenticular beds, amalgamated, irregular and erosive bases, coarse sandstones to microconglomerates, poorely to well sorting, with massive structure, normal graded bedding, rich in mudstone and sandstone intraclasts	Channel fill	Supracritic flows
	F_6	Tabular cross bedding with lateral continuity, flat bases, coarse sandstones, massive or as megaripples, well sorting, flat top	Channel fill	Subcritical regime flows
Fine grained sandstones Facies C, D & E of Walker & Mutti, 1973	F_7	Tabular beds, flat bases or marked by flame structures, coarse to medium sandstones; parallel laminations and normal graded bedding; disseminated pebbles	Channel fill on midfan	Subcritical regime flows
	F_8	Medium to fine-grained sandstones; good sorting, parallel laminations and normal graded bedding (T_{ae} Bouma divisions), sometimes with T_{cde}, T_{de}, T_{be} Bouma divisions	Depositional lobes	Subcritical flows in primary phase
	F_9	Very fine-grained sandstones to siltstones (T_{cde}, T_{de}, T_e Bouma divisions), well sorted, graded to massive structures	Depositional lobes on distal fan to abyssal plain	Subcritical flows (deposition from suspension)
	F_{9a}	Very fine-grained sandstones to siltstones; T_{cde}, T_{de}, T_e, e_i Bouma divisions, well sorted, starved ripples	Depositional lobe-abyssal plain sediments	Low-density current in subcritical regime (deposition from the dilute tail of the turbidity current)
	F_{9b}	Less clear sedimentary structures and coarser rocks than F_9; weaker sorting	Overbank deposits (?)	Low-density current; the bed-load transported by dragging or saltation

Table 1. Facies characteristics of the Histria Formation and their interpretation (modified after Oaie, 1999).

SE of the Capidava-Ovidiu Fault, green clastics ascribed to the Histria Formation were intercepted in two boreholes at Palazu beneath the Ovidiu gneisses (Figs 2A & 2B) (Visarion *et al*., 1979 and references therein; Kräutner *et al*., 1988). Both borehole and integrated geophysical data show that the Precambrian metamorphic basement overthrusts the Histria Formation clastics along the Palazu Thrust (Visarion *et al*., 1979 and references therein; Visarion *et al*., 1988; Krautner *et al*., 1988) (Fig. 2B). The amplitude and spatial development of the Palazu Thrust is difficult to estimate. The age of thrusting is considered to postdate the deformation of the Histria Formation (Săndulescu, 1984). In this idea, there are several possible interpretations of the age of thrusting: (a) a Cambrian age, older than the oldest (Ordovician) member of the Palaeozoic cover, consequently corresponding to the Late Cadomian or Early Caledonian deformation (Săndulescu, 1984); (b) a Lower Paleozoic, possibly Silurian age, based on borehole data in the Bordei Verde area, where one borehole intercepted the Ordovician above the Silurian deposits (Paraschiv & Paraschiv, 1978). However, the possibility that the age of thrusting is connected to the accumulation of the Histria Formation cannot be precluded. This hypothesis is consistent with the mineralogical, petrografic and provenance data existing on the Histria Formation, which indicate a source area composition and age partly similar to the basement of South Dobrogea.

4. Sedimentology of the Histria Formation

The Histria Formation is a thick succession of clastics, including coarse, sandstone dominated lower and upper members, separated by a thinner, middle member of pelitic-siltitic lithologies (Seghedi & Oaie, 1995; Oaie, 1999). Based geological and seismic data, an overall thickness of about 5000 m was estimated for the entire turbiditic succession (Mirăuță, 1969; Visarion *et al*., 1988; Pompilian *et al*., 1991).

The succession is extremely rich in sedimentary structures, well preserved over large areas, both inside beds and on bed surfaces. They include various ripple morphologies, load casts, current lineations, etc. Occasionally trace fossils are preserved, ranging from simple, silt-filled burrows and irregular burrowing, to meandering trails on bed surfaces, suggesting Nereites ichnofacies (Oaie, 1992, 1999) (Plate 1B).

4.1. Lithofacies types

Sedimentological studies show a large facies range, and detailed descriptions of facies and facies associations are presented in Oaie (1999), using the facies codes of Mutti (1992), Mutti & Ricci-Lucchi (1972), Walker (1979) and Walker & Mutti (1973). Descriptions of the main facies and their interpretation is given in Tab. 1.

Very coarse facies (equivalent to facies A of include three main types of conglomerates to microconglomerates, with lenticular bedding and erosional bases (Figs 3A & 3B). Poorly sorted, massive conglomerates, with a pelitic matrix (F_1), represent deposition from mass-flows in channels on submarine turbiditic fans. Conglomerates with internal structure (normal grading, laminations, clast imbrication, etc.) and a sandstone matrix (F_2) represent products of mass deposition from superconcentrated turbidite flows in channels located on the upper fan. Clast supported conglomerates to microconglomerates with a sandstone matrix (F_3), often grading to coarse sandstones, represent the infill of supra-cone channels, deposited from the bed-load of turbiditic currents.

The coarse facies is formed by sandstones representing the bed-load of the turbidity currents in an upper flow regime, accumulated in channels on the middle fan (equivalent to facies B of Mutti & Ricci- Lucchi, 1972; Walker & Mutti, 1973) (Figs 3C, 3D & 3E). Poorly sorted coarse sandstones, with discontinuous laminations marked by gravel sized clasts, form laterally continuous beds with sharp bases and tops (F_4); both the base and top may be flat or undulated, with flame structures due to differential loading. Coarse sandstones to microconglomerates, poorly or well sorted, massive or normal graded (F_5), form lenticular beds with irregular, erosive bases and flat tops, rich in mudstone or siltstone intraclasts. These represent typical channel fills of middle fans. Well-sorted coarse sandstones, with tabular cross bedding or mega-ripples (F_6), represent sediments of low density turbiditic flows accumulated in channels on the middle or lower fan (Fig. 4E). Coarse to medium sandstone beds, showing parallel lamination or normal grading (F_7), represents channel deposits from traction carpets (Tab. 1).

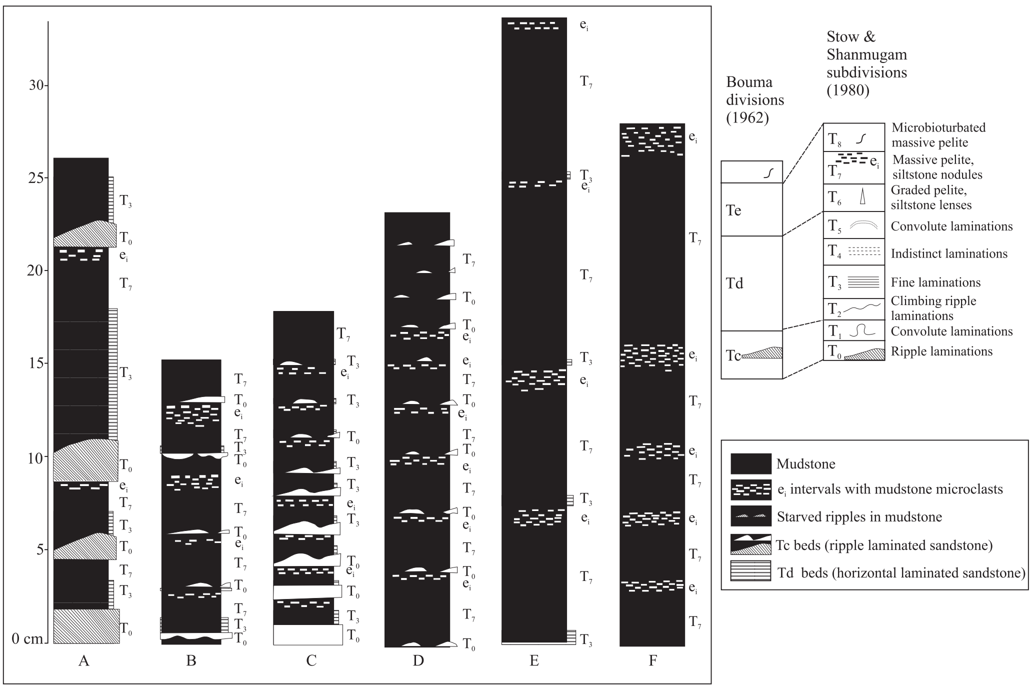


Figure 4. Vertical lithofacies associations in fine-grained turbidites of Haidar Member (Casimcea Valley at junction with Călugareni Valley). A) Random repetition of T_0-T_3-T_7-e_i beds; B) Association of T_0-T_3-T_7-e_i beds with upward thinning and fining of T_0-T_3 beds and upward thickening of T_7 beds; C) Association of T_0-T_3-T_7-e_i beds, with typical load-casted ripples, and thick e_i beds; D) Association of T_0-T_7-e_i beds, with starved ripples; E) Superimposed T_3-T_7-e_i beds with upward thinning of T_3 beds and thickening of T_7 beds; F) Association of T_7-e_i beds showing upward thickening of T_7 beds.

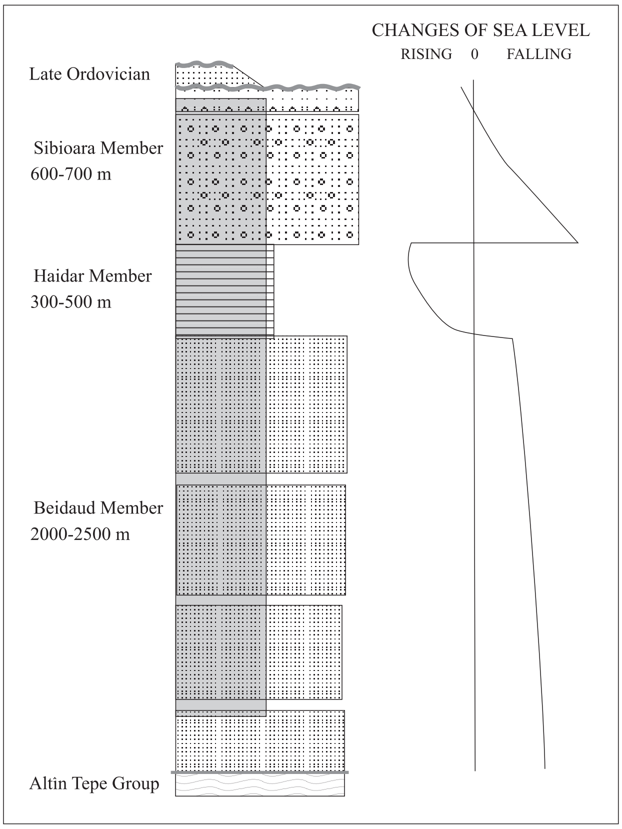


Figure 5. Schematic stratigraphic log of the Histria Formation (not to scale). The coarse lower and upper members (Beidaud and Sibioara members) are sandstone dominated, middle-fan turbidites; the middle Haidar member separating them consists exclusively of fine-grained, distal fan – abyssal plain turbidites. Turbidites show a tectonic relationship with the underlying Altîn Tepe Group metapelites and amphibolites and are unconformally overlain by Ordovician fine-grained clastics (only in boreholes NW of the Danube).

Medium to fine-grained sandstones, with good sorting, normal grading or horizontal laminations (F_8) represent T_{ae} Bouma divisions, and only seldom T_{bcde}, T_{cde} or T_{de} Bouma divisions (classical turbidites of Walker & Mutti, 1973) (Fig. 4D). The fine-grained facies includes fine-grained distal turbidites (facies association D and E according to the classification of Mutti & Ricci-Lucchi, 1972, classical distal turbidites of Walker & Mutti, 1973).

Very fine-grained sandstones to siltstones, well sorted, with parallel, cross or convolute lamination display base-missing Bouma sequences; typically T_{cde} and T_{de} are dominant, but T_{be} and T_e beds also occur (F_{9a}) (classical distal tubidites; Mutti, 1992) (Fig. 4). Depositional mechanisms imply sediment transportation as bedload, or in diluted suspension, as a turbiditic cloud.

A specific feature, almost ubiquitous in the middle member of the Histria Formation, is represented by the e_i lithofacies typically occuring in the Te Bouma divisions (Oaie, 1994, 1998). Lithofacies e_i and its associations are illustrated in Fig. 4. The e_i lithofacies consists of Te pelitic beds showing at the top fine, discontinuous parallel laminations, represented by an alignment of elongated, flat microclasts, evenly disseminated or clustered as discontinuous bands and lenses, within a matrix of similar grain-size, usually mudstone. The microclasts are up to 1-3 mm long and show a pelitic composition (mudstone), similar to that of the Te Bouma divisions (Oaie, 1998). They can be distinguished from the mudstone matrix by shape and shine. In the standard sequence of the fine-grained turbidites of Stow & Shanmugam (1980), facies e_i lies at the top of their T_7 division (ungraded mud + silt) (Fig. 4). The thickness of the e_i interval ranges from a few millimetres to several centimetres. Where e_i is interbedded in more complete sequences, usual thicknesses vary from 2-5 mm to 1-3 cm. Where amalgamated T_7 subdivisions reach thicknesses of several meters or more, the thickness of the e_i interval may attain 8-10 cm. Development of facies e_i is directly related to the evolution of the distal parts of turbiditic fans (Oaie, 1998). The fine-grained, semi-consolidated bottom sediments can be eroded by bottom currents and re-sedimented as tiny aggregates instead of clay particles. Suspension or re-suspension of the sediment occurs when the water current velocity exceeds a threshold value which depends on the density, size and degree of cohesion of the particles (Wűest, 2003). Near-bottom currents, with velocity > 6 cm/s, can suspend the sediments in form of individual particles, as flocculated material (Poulos & Drakopoulos., 2004), or as millimetric flat mud microclasts (Oaie, 1998). The local bottom currents are capable to transport in suspension particles up to 12 μm, in concentrations of 0.5-5 mg/l (Shepard *et al.*, 1977). If microclasts from the Histria Formation are derived by erosion of the upper parts of distal turbidites, the presence of lithofacies e_i represents a good indicator of a distal, lower fan turbiditic environment.

4.2. Facies associations and stratigraphy

The main vertical facies associations in the coarse members of the Histria Formation occurring at outcrop scale are shown in Fig. 3. Both thinning- and fining-upward and thickening-upward trends are defined by vertical facies associations. Thinning- and fining-upward cycles are commonly used to recognize channel deposition. Thinning-upward trends are the consequence of the progressive channel abandonment (Figs 3A, 3C & 3F). In the Histria Formation, a common thickness of a channel association may reach 1-15 or 20 m, comparable to other ancient submarine fan channel deposits. Superimposed channel associations may attain 100-300 m. The deposits display successions of amalgamated sandstone beds, from 10 to 100 m, frequently associated with graded conglomerates and separated by thin interchannel deposits.

Depositional lobes (a term usually restricted to ancient fans, Shanmugam & Moiola, 1988), consist of fine-grained, distal turbidites (Tcde-Tde Bouma divisions), associated with sandstones and displaying thickening- and coarsening-upward trends; other characteristics include: absence of basal channelling, sandstone grain-size varying from coarse to fine, sheet-like geometry and common thicknesses ranges of 3-15 m (Mutti & Ricci-Lucchi, 1972; Ricci-Lucchi, 1975; Mutti, 1977; Mutti & Normark, 1987). In the Histria Formation, the thickening-upward cycles of 10-15 m, interbedded with thin sandstone beds (1-2 m) are interpreted as depositional lobes (Fig. 5D). Thickening-upward lobe-cycles are a result of simultaneous vertical aggradation at bedform scale and basinward progradation, especially in active margin settings (Shanmugam & Moiola, 1988).

Earlier studies on the thick succession of clastics exposed in Central Dobrogea interpreted them as a Neoproterozoic ("Assyntian") flysch (Atanasiu, 1940; Mirăuță 1964, 1965, 1969; Jipa, 1970). The sedimentological record indicates that the turbidites of the Histria Formation can be defined as submarine fans, representing channel and lobe complexes formed from sediment-gravity flows in a deep-sea environment, as defined by Shanmugam & Moiola (1988). Their features suggest active margin fans. As studies of eustatic control on submarine fan development suggest that channelization and related fan growth seem to occur mainly during periods of low sea level (Shanmugam *et al.*, 1985), it is possible that the coarse members of Histria Formation have formed during such periods.

Based on sedimentological information and geometric relations, the Histria Formation can be separated into three

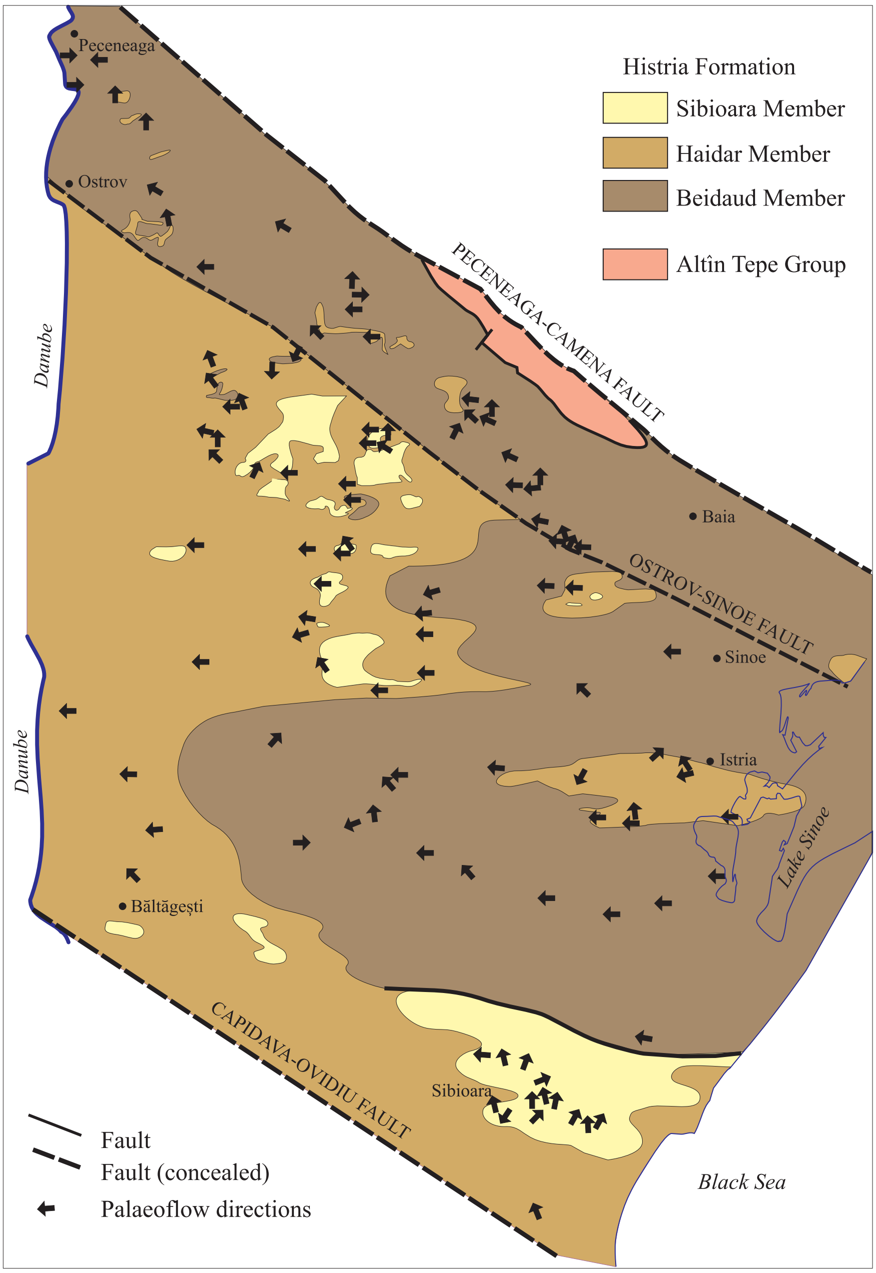


Figure 6. Palaeoflow directions measured in Tcde Bouma divisions of the Histria Formation, superimposed on a Mesozoic subcrop map showing the distribution of the three members in the Histria Formation.

members (Seghedi & Oaie, 1995; Oaie, 1999). The establishment of the overall vertical facies distribution for each member is difficult, due to the discontinuity of the outcrop areas, the possibility of lateral facies variation and folding. A schematic stratigraphic log is presented in Fig. 5, and the areal distribution of the three members is shown in Fig. 6.

The lower, sandstone dominated member, up to 2500 m thick (Beidaud Member), consists of channelized, coarse-grained middle fan turbidites, associated with depositional lobe deposits of the lower fan. In the area between the Peceneaga-Camena and Ostrov-Sinoe Faults, almost all outcrop areas consist of this lower, sandstone dominated member, occupying the hinge of a large, open anticline. On the southern limb of the anticline, scarce exposures of the middle member are locally found.

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. Thin, centimetric beds of fine-grained sandstones exceptionally occurr. 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.

The upper, Sibioara Member, is again coarse-grained and sandstone-dominated, interpreted as channelized middle fan turbidites, associated with conglomerates representing upper fan channels. This member always starts with a very coarse-grained conglomerate facies of the upper fan, directly overlying the fine grained, thinly laminated distal turbidites of the Haidar member.

The existence of a fourth member of the Histria Formation is possible, but still difficult to prove. Offshore in the Black Sea, borehole 19 Vadu (Fig. 2A) penetrated the Histria Formation on 36 m. The top of the turbidite succession, intercepted at 2113 m depth, is overlain by late Jurassic limestones, like in Central Dobrogea. Based on lithological and sedimentological descriptions (Almăşan *et al.*, unpublished data 1989), this succession appears to represent distal, lower fan turbidites, overlying a coarse - grained, conglomerate facies (Oaie, 1999). Because the gravity map shows the eastward uplift of the basement of Central

Dobrogea (Botezatu & Băcioiu, 1957; Airinei, 1980; Dimitriu, 2001 and references therein), the turbidites from the Black Sea offshore could lie in a lowermost stratigraphic position. However, in the absence of other geological and geophysical data, the stratigraphic position of these turbidites remains uncertain.

4.3. Palaeocurrent directions

Due to the very large scale of folding, the open style of folds and the very low-grade of metamorphism, the bedding surface represents the main detachment plane in the turbidites of the Histria Formation. A large variety of sedimentary structures is very well preserved on bed surfaces, as well as inside the beds. Most of the primary sedimentary structures have a mechanic, erosive origin (flute casts, rill marks, groove and chevron marks), including erosion channels marked by the presence of soft, rip-up clasts (intracalsts or intraformational remobilisations) (Figs 3A & 3B).

Constructional sedimentary structures are represented by the orientation of conglomerate elements along the main flow direction. Continuous or discontinuous vertical grain-size grading displays both fining- and coarsening- upward sorting. In coarse-grained layers, grain-size grading can be slightly contoured. Parallel or cross stratification and laminations, usually with sharp limits, are common in most of the outcrop areas. Sometimes deformational sedimentary structures, such as resulting from differential loading are found, mainly at the base of coarse-grained layers (Plate 2A, B, C).

In the coarse-grained facies, parallel bedding is in most cases absent or diffuse. Cross bedding, at mm to dm scale, appears either isolated (e.g. starved ripple) or grouped in sets and co-sets. The areas of ripple-type bedding may have straight or curved crests (Plate 3 A, B). Convolute bedding is rare, forming beds up to 10 – 15 cm thick. Rill marks, groove and chevron marks and current lineations are often found superimposed on top of pre-existing rippled surfaces (Jipa, 1970; Oaie, 1999). Several such structures are shown in Plates 2 (D) and 3 (B, C).

Palaeoflow measurements over the entire outcrop area of the turbidites, both on bedding surfaces (rill marks, chevron marks, ripple marks), as well as on sedimentary structures preserved inside the beds (imbrications, ripples), have been performed in all three members of Histria Formation (Jipa, 1968, 1970; Oaie, 1999). The palaeoflow directions measured in the fine-grained facies (Tcd Bouma divisions) of the three members are shown in Fig. 6. A dominant E-W flow pattern and W directed flows are observed in the largest part of the outcrop area. In the Sibioara member, palaeocurrent directions show a radial pattern, indicating dominantly NW, N and NE directed flows, typical for fan environments. In the Beidaud Member EW directions seem to be dominant, but NW, N and even WE directions also occur. The N, NE or NW directed flows, measured in the coarse-grained facies, represent the lateral transport directions from the source area into the basin.

Palaeocurrent directions indicate that the main transport in the basin of Histria Formation was E-W, with sediments delivered by a southerly or SE located source area (Jipa, 1970; Oaie, 1999). The N, NW and NS directions in the fine-grained Haidar Member suggest local conditions which induced a change in palaeoflow. Considering that seafloor morphology deflects the bottom currents around obstructions and funnels water through sills, it is suggested that these local conditions in the case of Histria Formation might have been structural highs. The presence of faults in the basement of Central Dobrogea (Fig. 2), indicated by geophysical data, supports the possibility of submarine highs in the sedimentary basin of the turbidites.

5. Mineralogy and petrology of the Histria Formation

Detailed mineralogical, geochemical and petrologic studies on the Histria Formation have not been done yet. Preliminary data on the scale of the entire formation provided information on the mineralogy and geochemistry of coarse- and fine-grained members, which were used to characterise the geological composition of the source areas. A vertical variation in sandstone composition within each coarse member was not observed, due to still limited petrographic information. A summary of the provenance data will give in addition information regarding the age and affinity of the source areas.

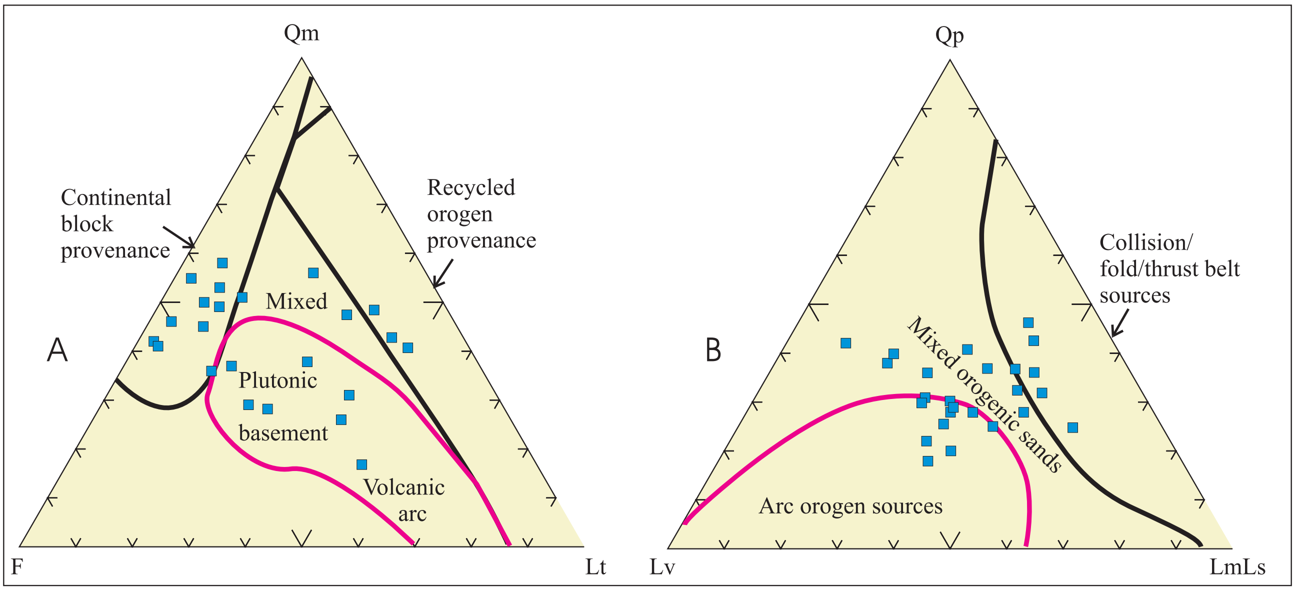


Figure 7. Petrofacies of coarse grained members of the Histria Formation. A) QFL (QmFlLt) plot for detrital modes of sandstones from the Histria Formation, indicating contributions from continental block, volcanic and plutonic sources and recycled clasts; Abbreviations: Qm = metamorphic + magmatic quartz; F (Fl) = total feldspar; Lt = total lithics, L = lithics; Qp = polycrystalline quartz; B) Tectonic discrimination diagram for sandstones, based on composition of lithics (Dickinson & Suczek, 1979; Dickinson & Valloni, 1989). The Histria Formation sandstones have a composition of mixed orogenic sands. Lv – volcanic lithics, Lm – metamorphic lithics, Ls – sedimentary lithics.

5.1. Composition of sandstones and conglomerates

5.1.1. Mineralogy

Coarse-grained member mineralogy consists of quartz (both embayed, volcanic quartz and subangular, undulose vein quartz), weathered plagioclase and K-feldspar, detrital phyllosilicates (muscovite, chlorite and biotite), epidote and various lithic fragments. Detrital epidote frequently forms large, rounded grains, consisting of prismatic or needle-like crystals either as fan-shaped, fibrous aggregates, or as granoblastic masses, suggesting a hydrothermal or metamorphic origin. Lithic clasts include fragments of magmatic, metamorphic and sedimentary rocks. Their petrography is illustrated in Plate 4. Magmatic clasts consist of granites, rhyolites, dolerites and basalts. Metamorphic rock clasts indicate a source area composed of gneisses, micaschists and quartzites, corresponding to the Precambrian basement composition of South Dobrogea. Red or black cherts (Grigoraş & Dăneţ, 1961; Anastasiu & Jipa, 1984) are reworked in the upper member and might suggest derivation from an accretionary wedge.

The petrofacies of the 24 samples from the coarse-grained members was defined using the parameters of the QFL plot (Fig. 7). Sandstone petrofacies suggest feldspatholithic and lithofeldspathic sandstones, with high P/F (ratios of plagioclase feldspar to total feldspar), low V/L (ratios of volcanic fragments to total unstable lithic clasts) and high M (mica contents). According to the QFL diagram, the major sources delivering detritus to the Histria basin were represented by a continental block and a plutonic basement – volcanic arc source (Seghedi *et al*., 2000). Only a few samples plot in the field of recycled orogen and mixed petrofacies. Such sandstone compositions occur in continental margins dominated by active volcanic arcs. Prevalence of rounded and subrounded lithic clasts indicates either a long distance of transport to the depositional basin, or, more probably, reworking in several sedimentary cycles. Coexistence of both subangular and subrounded quartz clasts indicates mixing of sediments from different sources.

In order to evaluate the tectonic context of the sandstones, the tectonic discrimination diagram based on composition of lithics was used (Dickinson & Suczek, 1979; Dickinson & Valloni, 1989). The Histria Formation sandstones plot in the central part of the QpLvLmLs diagram, in the field of mixed orogenic sands, and partly in the fields of arc orogen sources and collision/fold/thrust belt sources (Fig. 7B).

The sandstone composition indicates that the Histria Formation turbidites are highly immature lithic arkoses and volcanic arenites, rich in unstable components. Textural and mineralogical immaturity indicate quick sediment transport and burial, which prevented decomposition of feldspars and pyroxenes. The coarse-grained member composition suggests that a continental margin delivered terrigenous clasts into the basin. The presence of acid and basic volcanic clasts, together with pyroxenes, indicates a volcanic source, active at least periodically, along with a source area dominated by metamorphic rocks and granites. Source areas with such composition characterize continental margins dominated by active volcanic arcs (Seghedi *et al*., 2000, 2001).

Considering the great thickness of the Beidaud and Sibioara Members, it is obvious that a considerably larger number of samples should be analyzed in order to obtain reliable interpretation of source area composition and tectonic setting of the Histria Formation. As the basalt-rhyolite association is common both in convergent, subduction or collision-related settings, and in extensional or transtensional settings, chemical analyses of the volcanic clasts would be necessary, or at least of sandstone samples, would help to discriminate the tectonic setting of the turbidite sources.

5.1.2. Heavy mineral assemblages

In coarse-grained facies rocks, heavy minerals assemblages are disseminated throughout the bed. In Tc Bouma division of turbidites, heavy minerals are disseminated or concentrated along parallel or cross laminations and show good sorting, with both rounded and subangular-angular shapes. This suggests that erosion and transport of heavy minerals took place from source areas located at various distances to the sedimentary basin, some of them possibly intrabasinal, by reworking below sea level. The heavy mineral assemblages from Histria Formation sandstones were studied by Codarcea (in Seghedi *et al*., 2000), and the conclusions of these studies are presented below.

The heavy mineral fraction is rich in metastable minerals. Considering their frequency, heavy minerals include: pyrite, magnetite, ilmenite, hornblende, epidote, zoisite, kyanite, biotite, zircon, pyroxene, titanite, garnet, staurolite, monazite, tourmaline, rutile, chloritoid, sillimanite, brookite, apatite (Seghedi *et al*., 2000). The heavy mineral assemblages identified are kyanite + epidote + zoisite + hornblende + pyroxene in sandstones, epidote + zoisite + hornblende + zircon in siltstones and kyanite + epidote + zoisite + hornblende in mudstones. The widespread assemblage of hornblende + epidote + zoisite + kyanite indicates that the source area consists largely of metamorphic rocks. The presence of pyroxenes in the heavy mineral assemblages suggests the presence of magmatic rocks in the source area.

5.2. Composition of pelites

Mineralogical studies by XR diffraction of pelitic fraction (‹ 2 μ) were performed on 40 samples of pelites from all three members of Histria Formation. The mineralogy of the pelitic fraction is characterized by quartz, illite, albite and chlorite. Mineralogical studies point out that the clay mineral fraction is dominated by illite (65 - 75 %) and chlorite (25 – 35 %) (Rădan, unpublished data). Montmorillonite, smectite and kaolinite may occur occasionally. Values of 2-3 for the illite-chlorite ratio indicate a source area rich in acidic rocks (Kukal, 1965).

The chemical composition of 14 pelites sampled from all three members of Histria Formation is shown in Tab. 2. Major element geochemistry shows that the fine-grained sediments of the Histria Formation are rich in SiO_2 (58.80 – 69.00 %), Al_2O_3 (13.75 – 17.70 %), and Na_2O (2.46 – 3.68 %). High FeO/Fe_2O_3 ratio (0.6-6.1%) indicates either an initial reducing depositional environment, or the subsequent reduction of iron and its inclusions in the lattices of silicates during diagenesis. A significantly low K_2O/Na_2O ratio suggests high albite content in the rocks and a long distance of pelitic sediment transport (Kukal, 1965; Englund & Jørgensen,1973).

The chemical classification system of Englund & Jørgensen (1973), based on the abundance of major elements, was used to characterize the weathering conditions within the source area of the material supplied to the palaeobasins (Fig. 8). On this diagram, the pelitic rocks of the Histria Formation show the composition of unweathered sediments or of sediments with weak chemical weathering, rich in feldspars, micas and chlorite, with subordinate contents in montmorillonite and kaolinite (Englund & Jørgensen, 1973). The field of the Histria Formation overlaps with the field of black shales, mainly Scandinavian alum shales, as well as that of the Late Proterozoic-Eocambrian turbidites (Brøttum Formation) from southern Norway (Fig. 8). The Brøttum Formation is a thick, monotonous alternation of greywacke sandstones and shales, assumed to be deposited from turbidity currents, from Precambrian source rocks of granitic to gabbroic-anorthositic composition. The mineralogy of the clay fraction, similar to that of the Histria Formation pelites, includes quartz, illite, albite and subordinate amounts of chlorite (Englund & Jørgensen, 1973).

The trace element contents of 4 samples of pelites from the Haidar member (Tab. 3) are comparable to the "average shale" values of Turekian & Wedepohl (1961), but their abundances are generally lower for Cu, Pb, Zn, Sn, Ni, Co, Cr, V and La, and a little bit higher for Sc, Y, Yb and Zr. The variation and the mean values of the analysed trace elements suggest the influence of a terrigenous-volcanic source area dominated by more or less weathered acidic rocks (Kukal, 1965). The La/Sc ratio ranges from 1.36 to 2.38; this value is characteristic for shales derived from the Upper Continental Crust, in agreement with the above mentioned hypotheses Turekian & Wedepohl (1961). This is also consistent with data from the Vendian pelites in Malopolska from Jachowicz *et al.* (unpublished data 2002), who established that on discriminant diagrams, characteristic ratios of index elements (Th-La-Sc, Sc-Th-Zr/10, Ti/Zr-La/Sc, La/Y-Sc/Cr, K_2O/Na_2O, TiO_2-Ni) indicate a supply from a continental arc which was dominated by felsic rocks.

5.3. Provenance data

U-Pb SHRIMP analyses on zircons, performed on samples from all three members of the Histria Formation from Central Dobrogea (Żelaźniewicz *et al.* 2001a, b) yielded: Archean (2.8 Ga), Grenvillian (1.2-1.0 Ga) and Neoproterozoic (720-550 Ma) age groups. The oldest Neoproterozoic magmatic zircons are interpreted to mark the onset of rifting along the Teisseyre-Tornquist margin of Baltica at 720-680 Ma, while younger igneous zircons and metamorphic overgrowths may indicate collisional orogenic events in a continental arc setting at 640-580 Ma (Żelaźniewicz *et al.*, in preparation). The Grenvillian-aged zircons suggest derivation of the source area from the 1.3–0.9 Ga Grenvillian belt within the supercontinent Rodinia (Żelaźniewicz *et al.*, 2001a), which was formed by Laurentia, the Sunsas part of Amazonia and the Sveconorwegian part of Baltica (Nance & Murphy, 1994; Friedl *et al.*, 2000; Murphy *et al.*, 2000; Casquet *et al.*, 2001). The same age groups were yielded by the coeval turbidites in the Malopolska block, but there a Palaeoproterozoic age group is also present (Żelaźniewicz *et al.*, 2001a, b).

Oczlon *et al.* (2005) provide an alternative interpretation of the zircon data presented above, favouring a peri-Gondwanan provenance. The detrital zircon spectrum from the Histria Formation is suggested to indicate a contemporaneous active/accretionary margin and Archean/Grenvillian sources unrelated to Baltica, probably related to Avalonia.

Table 2. Chemical composition of pelites from the Histria Formation.

Nr.	Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	H_2O^+	CO_2	S	Total	F_2O_3 tot	LOI
1	4741	74,70	0,63	10,83	2,10	2,38	0,09	1,42	1,28	1,55	2,58	0,11	2,04	0	0,06	99,77	4,74	2,10
2	Sib3	73,23	0,48	11,56	1,77	2,26	0,07	1,52	1,00	2,45	3,24	0,12	1,33	0,59	0,08	99,70	4,28	2,00
1	4774	66,67	0,73	14,90	2,09	2,89	0,13	1,50	1,85	2,28	3,27	0,21	2,02	1,15	0,05	99,74	5,30	3,22
4	35/95	64,95	0,66	15,74	3,64	2,64	0,11	1,63	1,69	2,08	3,85	0,12	2,45	0	0,10	99,66	6,57	2,55
5	4711	63,71	0,68	16,52	2,50	4,21	0,15	1,95	0,80	2,40	3,42	0,15	3,06	0	0,07	99,62	7,18	3,13
6	B	63,05	0,62	18,09	2,79	3,06	0,12	1,64	0,64	2,96	3,08	0,10	2,97	0,37	0,06	99,55	6,19	3,40
7	4784	62,42	0,66	13,19	2,94	2,56	0,20	1,36	5,31	1,53	3,73	0,11	2,33	3,16	0,04	99,54	5,78	5,53
8	16	59,84	0,33	18,40	1,13	2,48	0,06	1,32	0,76	10,88	2,46	0,31	1,33	0	0,22	99,52	3,88	1,55
9	12	59,50	0,99	17,05	3,70	1,84	0,12	3,99	2,63	2,01	4,89	0,26	2,51	0	0,10	99,59	5,74	2,61
10	17	57,51	0,67	16,48	3,53	3,35	0,12	3,24	3,20	4,84	4,47	0,44	1,82	0	0,12	99,79	7,25	1,94
11	MV85	57,07	0,62	14,91	2,42	3,10	0,26	2,13	6,47	1,80	4,18	0,43	2,12	4,02	0,09	99,62	5,86	6,23
12	8	54,82	1,40	17,44	3,95	4,01	0,20	2,75	2,66	3,36	5,87	0,59	2,50	0	0,11	99,66	8,40	2,61
13	V1	49,57	0,82	12,77	2,89	3,65	0,14	5,44	7,27	3,50	0,61	0,40	2,80	9,73	0,09	99,68	6,94	12,62
14	13	48,21	1,33	15,86	9,82	0,63	0,11	4,16	6,31	1,05	6,39	0,40	1,56	3,68	0,09	99,60	10,52	5,33

Table 3. Trace elements in pelites from the Histria Formation (Haidar Member).

Sample	Location	Cu	Pb	Zn	Sn	Ga	Mo	Ni	Co	Cr	V	Sc	Y	Yb	Zr	Nb	La
4557	V. Siriu	8	2	60	2	18.3	4	14.5	66	39	87	17.5	37	2.9	240	10	31
4538	V. Nuntasi	20	2	58	u	16.5	traces	22	11	53	120	21	47	3.11	290	10	39
4567	V. Casimcea	7	6	50	2	18.5	traces	16.5	10	45	100	18.5	37	3.1	290	17.5	44
4554	V.Siriu	9	2.5	366	2.5	27	traces	6.5	5.5	17	110	22	40	3.2	270	u	30

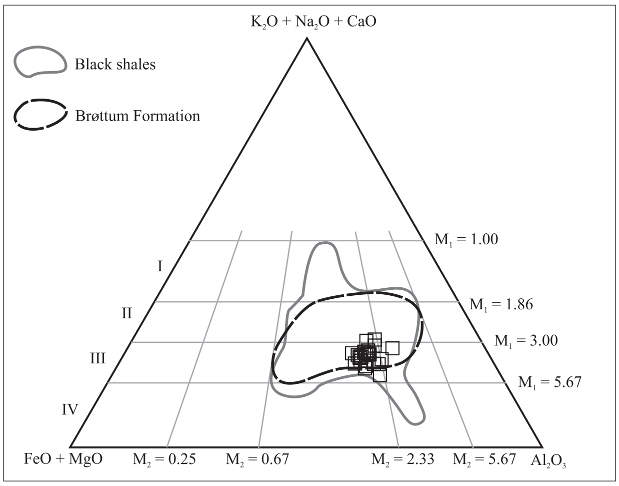


Figure 8. Englund & Jørgensen (1973) diagram used to characterize the chemical composition of pelites analysed from the Histria Formation. The variation in the degree of weathering is expressed as M_1 = $FeO+MgO+Al_2O_3/K_2O+Na_2O/CaO$ and M_2 = $Al_2O_3/FeO+MgO$. Unweathered sediments have low values of M1 and M2, while more weathered sediments have higher values.

6. Age of sedimentation

In outcrops in the southern part of Central Dobrogea, above an erosional unconformity at the top of the Histria Formation, discontinuous patches of Middle – Upper Jurassic limestones occur. They represent remnants of a largely eroded carbonate platform cover (Fig. 4). Erosion of the older, Palaeozoic cover strata from the top of the Histria Formation in Central Dobrogea took place at least before the Bathonian, as turbidites are overlain by scarce remnants of a kaolinitic weathering crust, preserved in places below the Bathonian clastics of the platform cover (Rădan, 1994).

The Late Proterozoic-Early Cambrian age of deposition is constrained by stratigraphic relations in boreholes and by palynological associations yielded by turbidites. In several boreholes west of the Danube, the folded turbidites are unconformably overlain by flat-lying Ordovician quartzitic sandstones and shales dated with graptolites (Murgeanu & Spasov, 1968; Iordan, 1981, 1992, 1999). This indicates a pre-Ordovician age of sedimentation and deformation of the Histria Formation (Mirăuță, 1965).

Palynological assemblages of Acritarchs, from the entire outcrop area, indicate a Late Proterozoic – Early Cambrian age. The palynological association identified in the upper member of the turbiditic succession includes (Iliescu & Mutihac, 1965; Vaida, unpublished data): *Protosphaeridium* sp., *P. tuberculiferum* Tim., 1966, *P. densum* Tim., 1966, *Polyedrosphaeridium bullatum* Tim., 1966, *Turuchanica* sp., *Podoliella irregulare* Tim.,1973, *Stictosphaeridium* sp., *S. sinapticuliferum* Tim., 1966, *Symplassosphaeridium* sp., *Orygmatosphaeridium* sp., *Favososphaeridium* sp., *F. favosum* Tim., 1966, *Zonosphaeridium* sp., *Micrhystridium* sp., *Synsphaeridium* sp., *S. sorediforme* Tim., 1959, *S. conglutinatum* Tim., 1959, *Kildinella hyperboreica* Tim., 1966, *K. tschapomica* Tim., 1966, *Trachysphaeridium* sp., *T. laminaritum* (Tim., 1966) Vidal, 1974, *T. levis* (Lopukh., 1971) Vidal, 1974, *Trematosphaeridium* sp., *Tetraedrixium elegans* Tim., 1973, *Dictyosphaeridium tungusum* Tim., 1966, *Pterospermopsimorpha mogilevica* Tim., 1973, *Leiomarginata* sp., ? *Tasmanites* sp., *Cymatiosphaera* sp., *Lophosphaeridium* sp., *Baltisphaeridium* sp., *Leiosphaeridia* sp.

Microspores *Leiominuscula rugosa*, *Margominuscula rugosa* and *Lophominuscula prima*, indicate a Precambrian age (Iliescu & Mutihac, 1965), while *Protoleiosphaeridium* sp. and *Trachyoleiotriletum* sp. indicate Early Cambrian, as in Malopolska (Glowacki & Karnkowsky, 1962). Other palynomorphs are confined exclusively to the Early Cambrian (Olaru *et al*., 1990): *Pulvinosphaeridium antiquum*, *Leiosphaeridia culta*, *L. dehisca*, *L. pilonifera*, *Baltisphaeridium dubium*, *B. acerosum*.

The macrofaunal evidence found in the distal turbidites of the haidar member is a medusoid imprint identified as *Nemiana simplex,* Palij, 1974 (Oaie, 1992). The trace is slightly elliptical, with maximum and minimum diameters of 42 mm, respectively 37 mm (Plate 1A). A comparison to other similar traces, found in Australia (Wade, 1969, 1972), North America, England (Jenkins, 1992) and the East European Platform (Fedonkin *et al*., 1983), suggests that it might represent a very primitive soft-bodied bentonic producer, bag-shaped and devoid of tentacles, of which findings are extremely rare. In Ediacaran quartzites, such form was named “a floating zooid” (Fedonkin *et al*., 1983). The existence of *Nemiana simplex,* Palij, 1974, a form similar to the Ediacaran Fauna, suggests an Upper Precambrian age for the Histria Formation (Oaie, 1992, 1999). The deep-water occurrences of Ediacaran Fauna are known principally in North America and England (Jenkins, 1992). The rarity of traces of organic activity found (Plate 1B) also suggests an age older than the the faunal explosion of the Cambrian, which occurred at 530 – 515 Ma (Carlovicz, 1997), or 545 Ma (Nash, 1995).

7. Deformation and structure

The main folding system deforming the turbidites consists of E-W trending, open normal folds (B_1) (Mirăuță, 1969), parallel to the main palaeoflow directions (Fig. 7). Antiformal B_1 folds often show periclinal closures toward W. The folds form large, northward verging, westward plunging anticlinoria and syclinoria, with amplitudes up to thousands of meters (Mirăuță, 1965, 1969). These large-scale structures are accompanied by secondary folds with amplitudes of hundreds or tens of meters. Outcrop scale folds are exceptions. Folding is accompanied by the development of an axial-planar slaty cleavage (S_1) (Mirăuță, 1969; Seghedi & Oaie, 1995). The cleavage is penetrative in fine-grained lithologies, and spaced or often absent in the coarser, sandstone or conglomerate members. Cleavage shows a general E-W orientation, with steep dips (70^o-80^o) to the north or south, parallel or fanning in fold hinges.

Locally, a second folding system (B_2) with NNW-SSE trend overprints the B_1 folds. B_2 folds show amplitudes of meters to tens of meters and are well developed on both sides of the Ostrov-Sinoe Fault.

In several areas, the slaty cleavage is pervasively deformed by small-scale kink bands and often by crenulations and crenulation cleavages. These areas typically occur in two strips of hundreds of meters wide south of the Peceneaga-Camena Fault and north of the Capidava-Ovidiu Fault, as well as in the area of Lake Sinoe (Seghedi & Oaie, 1993). This superimposed deformation may be the result of reactivation of the main strike-slip faults bordering Central Dobrogea (Drăgănescu, personal communication).

The age of the main deformation and very low-grade metamorphism of the Histria Formation ranges between 573-518 Ma, as suggested by whole rock K-Ar cooling ages on pelites from borehole Mihai Viteazu, located on a steep fold limb in the lower member of the Histria Formation (Giuşcă *et al*., 1967; ages recalculated by Kräutner *et al*., 1988). The above K-Ar ages indicate that deformation of the turbidites took place in the Late Proterozoic-Early Cambrian and was ascribed to the “Assynthan” or “Baikalian” events at the end of the Neoproterozoic (Giuşcă *et al*., 1967; Mirăuță, 1969).

The Neoproterozoic metamorphism took place in very low-grade, subgreenschist facies conditions. Values of the 'illite crystallinity index' ranging from 2.3 to 12 and indicating anchimetamorphic conditions are consistent with petrographic and structural data. In scarce outcrops in the lower member south of the Peceneaga-Camena Fault, chloritoid crystallization in sandstones indicates that lower greenschist facies conditions were attained locally in the lower member, in this case in the core of a large, open anticline.

8. Extent and basement of the sedimentary basin

The eastward extent of the Histria Formation is known from borehole and geophysical data. In the Black Sea offshore, one borehole bottomed in Histria Formation turbidites at depths below 2000 m (Almăşan *et al*., unpublished data 1989) (Fig. 2A). Integration of borehole and seismic data (Cătuneanu, 1992, 1993) indicates a progressive eastward deepening of the Histria Formation in the offshore area, where it is overlain by Late Jurassic sediments as in the outcrop area in Central Dobrogea. The offshore eastward stepwise deepening of the turbiditic succession is controlled by a system of N-S trending faults, probably belonging to the West Black Sea Fault system. It is not precisely known how far east the turbidites continue into the Black Sea and opinions vary considerably. Magnetic data suggest that the Moesian basement suddenly terminates at about 50 km east from the shore (Romanescu *et al*., 1972). A larger offshore extent, of about 100 km east from the shore, is suggested by correlation of seismic sections with borehole information (Cătuneanu, 1993) (Fig. 2A). Săndulescu (1980, 1984) suggests that the Histria Formation prolongates in the West Black Sea basin up to the area with basaltic crust, to reappear on the eastern side of the Black Sea in the Dzirula massif of the Transcaucasus (Fig. 1A).

Geological and seismological data suggest that the depositional basin of the Histria Formation was floored by a heterogeneous continental crust. As previously presented, although in outcrops the contact with the Altîn Tepe Formation is tectonic, magnetic and gravity anomalies suggest the continuation of the Altîn Tepe Group rocks in the basement of the Histria Formation as far south as the Ostrov-Sinoe Fault (Visarion *et al*., 1988). Gravity and magnetic data indicate that south of the Ostrov-Sinoe Fault the basement of the turbidites is made up of rocks with densities comparable to the Ovidiu gneisses from South Dobrogea (Dimitriu, 2001) (Fig. 2B).

Magnetic anomalies in the Black Sea offshore north of the COF (Airinei, 1967; Romanescu *et al*., 1972), contrasts in electric conductivity revealed by magneto-telluric soundings at depths of about 4-6 km, along with results of potential field modelling, all suggest the northward prolongation of the Palazu banded iron formation in the subsurface of Central Dobrogea up to the Istria Fault (Visarion *et al*., 1988) (Fig. 2B).

9. Discussion

9.1. Eustatic and climatic control on the deposition of the Histria Formation

The main factor in the development of a submarine fan is the global lowering of sea level (Shanmugam & Moiola, 1982, 1984). The major controls on global sea level changes are tectonism and glaciation (Shanmugam & Moiola, 1982, 1988). Correlating sea level changes (Vail *et al*., 1977) with sea-floor spreading and related climate changes, it was concluded that at the Precambrian/Cambrian boundary the sea level was 50 m lower than today (Shanmugam & Moiola, 1982; Condie, 1997; Miall, 1984, 1997). An important sea level rise started in the Late Precambrian and terminated in the Ordovician (Condie, 1997). The low sea level favoured the development of marine turbidite systems at global scale (Shanmugam & Moiola, 1982). The cold climate during this time-span, correlated with the presence of strong submarine currents, favoured the development of dominant coarse fans (Mutti, 1985). The Histria Formation turbidites, accumulated during Late Precambrian-Early Cambrian, are part of these coarse turbidite systems. The rarity of traces of organic activity in Histria Formation turbidites is consistent with a cold climate during their accumulation.

Strong erosion of the shelf area, tectonic uplift of the source area and strong submarine currents are the main controls on turbidite sedimentation (Kolla & Macurda, 1988). The low-stand favours strong erosion and continuous uplift of the source area and results in a large amount of sediments transported in the heads of submarine canyons. Depocenters, initially emplaced on the coast, prograde to the offshore and favour the development of submarine fan systems. This results in episodic sea-level changes, spanning 10.000 to 10.000.000 years (Kolla & Macurda, 1988). In the case of the Histria Formation, the succession of the three members could be similarly explained by variations in turbiditic current activity, influenced by sea level changes, climate and sea-floor spreading. The coarse-grained Beidaud and Sibioara Members might be controlled by low-stand, while the fine-grained Haidar Member suggests a rise in sea-level (Oaie, 1999). In modern ocean basins, such conditions are related to active continental margins.

The development of dominantly coarse-grained submarine fans is influenced primarily by lowstands of sea level in most tectonic settings (Mutti, 1985; Kolla & Macurda, 1988; Shanmugam & Moiola, 1988). In the classification of the mentioned authors, the Histria Formation would belong to the turbiditic systems of type I, formed at low sea level, in conditions of strong hydrodynamical and morphological gradients, which favour the formation of channels filled with coarse-grained sediments and of thick, sandy depositional lobes.

A distinctive feature of the fine-grained turbidites of the Haidar Member is the variegated colour occurring over large areas, with a purple colour of the pelitic layers and a green colour of the thin interbedded ripple-laminated siltstones or fine-grained sandstones. There are several possible explanations for the colour changes: (a) they might reflect pulsations of the turbidity current activity, eroding source areas with different geochemical composition; (b) difference in sedimentation rates, knowing that greater sedimentation rate enables preservation of organic carbon, mobilization of manganese and reduction of Fe^{3+} to Fe^{2+} in order to produce the green coloration of sediments (Kerr, 2001); (c) the degree of bottom water oxygenation (Wagreich & Krenmayr, 2005), well oxygenated waters determining red, while less oxygenated waters determining green colours; in the case of the Haidar Member, the presence of pyrite in the fine-grained, green sandstones and siltstones suggests reducing environment.

9.2. Tectonic setting of the Histria Formation

Sedimentological studies indicate that the turbidites of the Histria Formation can be defined as submarine fans, representing channel and lobe complexes formed from sediment-gravity flows in a deep-sea environment. Absence of carbonates indicates basin depths below the CCD. Although trace fossils are scarce, the presence of meandering traces on the surface of pelitic lithofacies resembling the Nereites ichnofacies suggests deep water environments.

The overall sedimentological features and the dominance of sandstones suggest that the Histria Formation turbidites accumulated on a fan developed in an active-margin setting. Active margin fans are usually small, sand-rich and show depositional lobes (Shanmugam & Moiola, 1988). Two models were suggested for the tectonic setting of the Histria Formation: a forearc basin (Seghedi & Oaie, 1993), or a foreland (foredeep) basin (Seghedi & Oaie, 1994, 1995; Oaie, 1999; Żelaźniewicz *et al.,* 2001a).

There are several tectono-stratigraphic attributes of foreland basin (foredeep) development, as revealed by Ori *et al*. (1986): foreland basins migration front of an advancing thrust pile; thrust highs, with impressive vertical displacements, can be eroded to produce a local sediment source; sediment dispersal patterns are mainly parallel to the long axis of the basin, although lateral supply is important; segmentation of the foredeep into discrete depocentres as a result of contemporaneous tectonism and diapirism, probably due to basement control.

The model of a foreland basin setting of the Histria Formation turbidites is based on the following: location north of the northward vergent Neoproterozoic thrust wedge, preserved in the subsurface of South Dobrogea; overall upward-coarsening facies association, consistent with a thrust system prograding towards its foreland; mixed sandstone petrofacies; dominant axial sediment transport, parallel to the long axis of the basin, as well as lateral supply. The presence of blind thrusts in the basin fill, represented by submarine highs related to thrust fronts which might have acted as intrabasinal sources, are inferred on the basis of structural and geophysical data (Fig. 2B). A convergent setting is indicated by pre-Ordovician deformation of the basin fill into large, open folds, parallel to the basin axis. The structural style and the very-low grade metamorphic conditions of this deformation are consistent with a foredeep setting. The dominance of sandstone facies suggests that the basin was effectively oversupplied with respect to its size, a feature that appears common to many foreland basin successions (Pickering *et al*., 1988). E-W fold directions suggest N-S directed compression. A slightly oblique convergence is suggested by the NW-SE orientation of the faults present in the basement of the turbidite basin (Fig. 2B).

A possible younger analogue of the ancient turbidites of Histria Formation is the lower and middle Eocene Hecho Group, a submarine fan from the Pyrenees formed in a foredeep basin with a width comparable to that of Histria Formation. The deep-marine turbidite facies are well documented in many papers, among which the best known are by Mutti (1977, 1985), Mutti *et al*., (1985) and Labaume *et al*. (1985). The eastern part of the Hecho Group comprises six, deep marine, channel levee complexes, while the western part consists of fan-fringe and basin-plain deposits. The main transport path for the clastics was axial and from east to west. Along the length of the foreland basin, structural highs were active during sedimentation.

Sandstone petrography and geochemistry of the pelites indicate that the turbidite basin of the Histria Formation was sourced by a continental margin. Further petrological studies are required in order to draw reliable conlusions relate to the significance of the bimodal volcanic rocks. They could have been related to an active volcanic arc (Seghedi & Oaie, 1995; Oaie, 1999), or to an intraplate volcanism connected to extension.

Zircon provenance data indicate that the source area was composed of Archaean and Grenvillian age metamorphic rocks (Żelaźniewicz *et al.,* 2001a, b), which is consistent with petrographic observations. Considering that coeval turbidites from the Malopolska Massif yielded Palaeoproterozoic zircons, it is possible that more analyses could document their presence in Central Dobrogea too. The presence of Grenvillian rocks in the source area suggests that the active continental margin sourcing the turbidite basin was probably the Avalonian-Cadomian volcanic arc at the northern coast of Gondwana.

10. Acknowledgements

We thank our reviewers Rudy Swennen and Timothy Debacker for their insightful and helpful reviews for this manuscript. Gheorghe Oaie acknowledges the support from the organizers of the Avalonia-Moesia symposium and workshop. The synthesis of data regarding the Histria Formation was accomplished in project CERES 3-195/12.11.03 of the Ministry of Education and Research and in the framework of a bilateral project BIL 01/34 between the Science and Innovation Administration of the Ministry of the Flemish Community of Belgium (Ghent University and Katholieke Universiteit Leuven) and the Department of European Integration and International Relations of the Ministry of Education and Research of Romania (Geological Institute of Romania).

References

AIRINEI, Ş., 1967. Cupluri de anomalii gravimetrice-magnetice deschise pe litoralul românesc al Mării Negre. *Studii şi cercetări de geologie, geofizică, geografie, seria Geofizică*, 5, 1: 147-154.

AIRINEI, Ş., 1980. Radiografia geofizică a subsolului României. *Editura stiintifică şi enciclopedică*, Bucureşti, 258 p.

AIRINEI, Ş., 1982. Raporturi geodinamice între Placa Moesică şi arcul carpato-balcanic pe teritoriul României. *Studii şi cercetări de geologie, geofizică, geografie, Seria geofizică*, 20: 3-10.

ANASTASIU, N. & JIPA, D., 1984. Source areas of the Assynthic flysch deposits in Central Dobrogea massif (Romania). *Anuarul Institutului de Geologie şi Geofizică*, LXIV: 445-453.

ATANASIU, I., 1940. Privire generală asupra geologiei Dobrogei. *Lucrările Societății Geografice Dimitrie Cantemir, Iaşi*, 1-89.

BARBU, C., 1973. *Structura adânca a părții de est a platformei Moesice (teritoriul R.S.R.), cu privire la zonele de acumulare posibile pentru petrol şi gaze ale Paleozoicului*. Unpublished PhD thesis, University of Bucharest.

BARBU, C. & VASILESCU, E., 1967. Tectonique du soubassement premesozoique de la plate-forme Moesique (territoire Roumain). Asociation Geologique Carpatho-Balkanique. VIII-eme Congres, Belgrade, septembre 1967. *Raport Geotectonique*, 35-40.

BOTEZATU, R. & BĂCIOIU I., 1957. Anomalia gravității in Dobrogea centrală. *Buletinul Societății Academice al RPR, Secția Geologie-Geografie*, II, 2, 2: 237-252.

BOTEZATU, R., AIRINEI, S., CHIRIAC, M. & VISARION, M., 1984. Evolutia ideilor cu privire la raporturile dintre Dobrogea Centrala si Dobrogea de Sud. *Studii şi cercetări de geologie, geofizică, geografie, seria Geologie*, 29: 18-23.

CARLOVICZ, M., 1997. Was Cambrian explosion the result of the wondering continents? *EOS*, 78, 36: 381 – 382.

CASQUET, C., BALDO, E., PANKHURST, R.J., RAPELA, C.W., GALINDO, C., FANNING, C.M. & SAAVEDRA, J., 2001. Involvement of the Argentine Precordillera terrane in the Famatinian mobile belt: U-Pb SHRIMP and metamorphic evidence from the Sierra de Pie da Palo. *Geology*, 29: 703–706.

CĂTUNEANU, O., 1992. The geology of the Black Sea Romanian shelf of Central Dobrogea type. *Revue Roumaine de Geologie*, 36: 78-81.

CĂTUNEANU, O., 1993. *Geologia şelfului românesc din prelungirea Platformei sud – dobrogene şi a Masivului central Dobrogean, cu implicații asupra perspectivelor sale petrolifere*. Unpublished PhD thesis, University of Bucharest.

CODARCEA-DESSILA, M., MIRĂUŢĂ, O., SEMENENKO, N.P., DEMIDENKO, S.G. & ZEIDIS, B., 1966. Gheologhiceskaia interpretatia dannîh polucennîh pri pomosci K-Ar metoda po absoliutnomu vozrastu kristaliceskih formatii iujnih Carpat i Dobrudji. *Tr XII sess. Kono opredelenia absoliutnovo vozrasta gheologhiceskih formaîi pri NZANDDr*, Moskva.

CONDIE, K.C., 1997. *Plate Tectonics and Crustal Evolution*. Butterworth-Heinemann, Oxford, UK, 282 p.

DICKINSON, W.R. & SUCZEK, C.A., 1979. Plate tectonics and sandstone composition. *Bulletin of American Association of Petroleum Geologists*, 63: 2164-2182.

DICKINSON, W.R. & VALLONI, R., 1989. Plate settings and provenance of sands in modern ocean basins. *Geology*, 8: 82-86.

DIMITRIU, R.G., 2001. *Geological structure of the Romanian continent-sea transition zone as inferred by geophysical data modeling*. Unpublished Ph.D. Thesis, University of Bucharest, 285 p (in Romanian).

ENGLUND, J.-O. & JØRGENSEN, P., 1973. A chemical classification system for argillaceous sediments and factors affecting their composition. *Geologiska Foreningens I Stockholm Fohandlingar*, 95, I: 87-97.

FEDONKIN,M.A.,GNILOWSKAIA,N.B.,JANKAUSKAS, T.V., KIRJANOV, V.V., LENDZIAN, K., MIERZEJEWSKA, G. & VOLKOVA, N.A., 1983. *Upper Precambrian and Cambrian paleozoology of the East European Platform*. House Wydawnictwa Geologiczne, Warszaw.

FRIEDL, G., FINGER, F., MCNAUGHTON, N.J. & FLETCHER, I.R., 2000. Deducing the ancestry of terranes: SHRIMP evidence for South America-derived Gondwana fragments in central Europe. *Geology,* 28: 1035–1038.

GAVĂT, I., AIRINEI, Ş., BOTEZATU, R., SOCOLESCU, M., STOENESCU, S. & VENCOV, I., 1963. Structura geologică profundă a teritoriului R.P.R. după datele actuale geofizice (gravimetrice şi magnetice). *Studii şi cercetări geofizice*, 1, 1: 7-34.

GAVĂT, I., AIRINEI, Ş., BOTEZATU, R., SOCOLESCU, M., IONESCU, S. & VENCOV, I., 1965. Contributions de la gravimetrie et de la magnetometrie en l'etude de la structure profonde du la territoire de la Roumanie. *Revue Roumaine de Géologie, Géophysique et Géographie, Serie Géophysique*, 9, 1: 81-109.

GAVĂT, I., CORNEA, I., GHEORGHE, A., GASPAR, R.& TOMESCU, L., 1967. La structure de la soubassement moesien entre le Danube et les rivieres du Buzau, Ialomita et Siret. *Revue Roumaine de Géologie, Géophysique et Géographie, Serie Géophysique*, 11, 1: 23-32.

GIUŞCĂ, D., IANOVICI, V., MÎNZATU, S., ŞOROIU, E., LEMNE, M., TĂNĂSESCU, A. & IONCICĂ, M., 1967. Asupra vârstei absolute a formatiunilor cristalofiliene din forlandul orogenului carpatic. *Studii şi cercetări de geologie, geofizică, geografie, seria geologie,* 12: 287-297.

GIUŞCĂ, D., SAVU, H., BERCIA, I. & KRĂUTNER, H. G., 1969. Sequence of tectono-magmatic pre-alpine cycles on the territory of Romania. *Acta Geologica Acad. Sci. Hung*., 13: 221-234.

GIUŞCĂ, D., VASILIU, C., MEDEŞAN, A. & UDRESCU, C., 1976. Formaţiunea feruginoasă kareliană din Dobrogea (La formation ferugineuse carelienne de Dobrogea). *Studii şi cercetări de geologie, geofizică, geografie, Geologie*, 21: 3-20.

GLOWACKI, E. & KARNKOWSKI, P., 1962. Comparison of the Precambrian (Riphean) of the Middle Carpathian Foreland with a Series of Green Schists of Dobrudja. *Kwarta Geologica*, Warszawa.

GLOWACKI, E., KARNKOWSKI, P. & ZAK, C., 1963. Precambrian in the basement of the Carpathian Foreland and the Holly Cross Mountains. *An. Soc. Geol. Pol*., XXXII, 3: 321-339.

GRIGORAŞ, N. & DĂNEŢ, T., 1961. Contribuţii la cunoaşterea şisturilor verzi dobrogene. *Studii şi Cercetări de Geologie, Academia RPR*, 6, 3: 541 – 553.

IANOVICI, V. & GIUŞCĂ, D., 1961. Date noi asupra fundamentului cristalin al Podişului Moldovenesc şi Dobrogei. *Academia RSR, Studii şi cercetări geologice*, VI, 1: 153-159.

ILIESCU, V. & MUTIHAC, V., 1965. Consideraţii asupra posibilităţilor de corelare a unor depozite din fundamentul zonei Tulcea cu formaţiuni cutate din Dobrogea centrală. *Dări de Seamă ale Institutului Geologic,* LI/1: 243-249.

IORDAN, M., 1981. Study of Silurian and Devonian faunas from the Eastern part of the Moesian Platform. *Memoriile Institutului de Geologie şi Geofizică,* 30: 115-221.

IORDAN, M., 1992. Biostratigraphic age indicators in the Lower Paleozoic successions of the Moesian Platform of Romania. *Geologica Carpathica,* 43, 4: 231-233.

IORDAN, M., 1999. Biostratigraphy of the Paleozoic in the Moesian Platform. Excursion Guide of the Joint Meeting of EUROPROBE TESZ, PANCARDI and Georift Projects "Dobrogea – the interface between the Carpathians and the Trans-European Suture Zone", Tulcea, 1999: Geology and structure of the Precambrian and Paleozoic basement of North and Central Dobrogea. Mesozoic history of North and Central Dobrogea. *Journal of Tectonics and Regional Geology,* 77, Supplement no.2: 18-23.

JENKINS, R. J. F., 1992. Functional and ecological aspects of Ediacaran assemblages. *In:* J. H. Lipps & P. W. Signor (eds.), *Origin and Early Evolution of the Metazoa*. Plenum Press, New York, 131-177.

JIPA, D., 1968. Azimuthal Relationships between Cross-Stratification and Current Markings of central Dobrogea, Romania. *Journal of Sedimentary Petrology*, 38, 1: 192-199.

JIPA, D., 1970. Cercetări sedimentologice în depozite Proterozoic superioare (Şisturi verzi) din Dobrogea. *Anuarul Institutului Geologic*, XXXVIII: 51-110.

KERR, R. A., 2001. Rising global temperature, rising uncertainty. *Science*, 292, 5515: 192-194.

KOLLA, V. & MACURDA, D.B. Jr., 1988. Sea level changes and timing of turbidity current events in deep sea fan systems. *In* Wilgus *et al*. (Eds.), Sea level changes: an integrated approach. *SEPM Special Publication*, 42: 381-392.

KRÄUTNER, H.G. & SAVU, H., 1978. Precambrian of Romania. Materials of the IGCP Project *22 "Precambrian in Younger Fold Belts",* Prague, 5-38.

KRÄUTNER, H.G., MUREŞAN, M. & SEGHEDI, A., 1988, Precambrian of Dobrogea. *In*: V. Zoubek, ed., *Precambrian in younger fold belts*, John Wiley, New York, 361-379.

KUKAL, Z., 1965. Geochemical development of Early Paleozoic sediments of Central Bohemia and its relation to the changes of sedimentary environment. *Geol. Sbornyk*., 14: 209-219

LABAUME, P.E., MUTTI, E., SEGURET, M. & ROSELL, J., 1985. Megaturbidites: a depositional model from the Eocene of the SW-Pyrenean foreland basin, Spain. *Geo-Marine Letters*, 7: 91-101.

MĂRUNŢEANU, M., 1987. *Studiul geologic al bazinului inferior al Tazlăului Mare*. Unpublished PhD thesis, University of Bucharest.

MIALL, A.D., 1984. *Principles of sedimentary basin analysis*. Springer Verlag, New York, 490p.

MIALL, A. D., 1997. *The geology of stratigraphic sequences*. Springer-Verlag, Berlin, 433p.

MIRĂUŢĂ, O., 1964. Şisturile verzi din regiunea Dorobanţu – Măgurele (Dobrogea Centrală). *Dări de Seamă ale Comitetului Geologic*, L, 2: 259-272.

MIRĂUŢĂ, O., 1965. Stratigrafia şi tectonica şisturilor verzi din regiunea Istria - Băltăgeşti (Dobrogea centrală). *Dări de Seamă ale Comitetului Geologic*, LI, 1: 257-276.

MIRĂUȚĂ, O., 1969. Tectonica proterozoicului superior din Dobrogea centrală. *Anuarul Institutului Geologic*, XXXVII: 31-36.

MRAZEC, L., 1910. Discussion sur l'existance des roches vertes en Carpathes. *Dări de Seamă ale Institutului Geologic*, II.

MRAZEC, L., 1912. Sur la ligne de chevauchement Peceneaga – Camena. *Dări de Seamă ale Institutului Geologic*, III: 162-165.

MUREŞAN, M., 1971. Asupra prezenței unei ferestre tectonice în zona Şisturilor verzi din Dobrogea Centrală (regiunea Altîn Tepe). *Dări de Seamă ale Institutului Geologic*, LVII, 5: 127-154.

MUREŞAN, M., 1972. Studii aspra zăcământului cu pirită de la Altîn Tepe (Dobrogea Centrală). II. Pozitia stratigrafică a mineralizației. *Dări de Seamă ale Institutului Geologic*, LVIII, 2: 25-61.

MURGEANU, G. & SPASOV, H., 1968. Les Graptolites du forage Bordei Verde (Roumanie). *Bulgarian Academy of Sciences, Comity of Geology, Bulletin of Geological Institute, Series Paleontology*, K.H. 17: 229-239.

MURGEANU, G. & PATRULIUS, D., 1973. Le paleozoique de la Plate-forme Moesienne. *Carpatho-Balkan Geological Association, Congress VI*, I, 2, Warszaw, 237-245.

MURGOCI, G., 1911. Cercetări geologice în Dobrogea nordică. *Anuarul Institutului Geologic al României*, V, VI: 77-81.

MUTTI, E., 1977. Distinctive thin-bedded turbidite facies and related depositional environments in the Eocene Hecho Group (southcentral Pyrenees, Spain). *Sedimentology*, 24: 107-131.

MUTTI, E., 1985. Turbidite systems and their relations to depositional sequences. *In* Zuffa, G.G. (Ed.), *Provenance of Arenites*. D. Reidel Publishing Company, Dordrecht, 65-93.

MUTTI, E., 1992. *Turbidite sandstones*. Instituto di Geologia. Universita di Parma (Italy), 273.

MUTTI, E. & NORMARK, W.R., 1987. Comparing examples of modern and ancient turbidite systems: Problems and concepts. *In* Legett, J.K. & Zuffa, G.G. (Eds.), *Marine Clastic Sedimentology: Concepts and Case Studies*. Graham and Trotman, London, 1-37.

MUTTI, E. & RICCI-LUCCHI, F., 1972. Le torbiditti dell'Apenino settentrionale: introduzione all analisi di facies. *Memorie di Societa Geologica Italiana*, 11: 161-199.

MUTTI, E., REMACHA, E., SGAVETTI, M., ROSELL, J., VALLONI, R. & ZAMORANO, M., 1985. Stratigraphy and facies characteristics of the Eocene Hecho Group turbidite system, south-central Pyrenees. In *Fieldtrip guide book of the 6th* European Meeting International Association of Sedimentologists, Lleida, Spain, excursion no. 12: 521-600.

MURPHY, J.B., STRACHAN, R.A., NANCE, R.D., PARKER, K.D., FOWLER, M.B., 2000. Proto-Avalonia: a 1.2–1.0 Ga tectonothermal event and constraints for the evolution of Rodinia. *Geology,* 28: 1071–1074.

NANCE, R.D. & MURPHY, J.B., 1994. Contrasting basement isotopic signatures and the palinspastic restoration of peripheral orogens: Example from the Neoproterozoic Avalonian-Cadomian belt. *Geology,* 22: 617–620.

NASH, M., 1995. When life exploded. *Time Magazine*, 146: 23.

OAIE, G., 1992. Traces of organic activity in the Greenschist Series of central Dobrogea (Romania). *Studii si Cercetări de Geologie*, 37: 77-81.

OAIE G., 1994. Necesitatea protejării urmelor de activitate organică de vârstă Precambrian superior din cadrul Seriei şisturilor verzi din Dobrogea centrală. *Ocrotirea naturii şi a mediului înconjurator*, 37/2: 133-138.

OAIE, G., 1998. Sedimentological significance of mudstone microclast intervals in Upper Proterozoic turbidites, Central Dobrogea, Romania. *Sedimentary Geology*, 115: 289-300.

OAIE, G., 1999. *Sedimentologia si tectonica Seriei sisturilor verzi din Dobrogea centrala si prelungirea lor in acvatoriul Marii Negre*. Unpublished PhD thesis, University of Bucharest, 105.

OCZLON, M.S., SEGHEDI, A. & CARRIGAN, C.W, 2005. Avalonian and Baltican origins for the Moesian Platform (Southern Europe, Romania/Bulgaria) (submitted).

OLARU, L., ERHAN, V. & ANTOCHE, M., 1990. Observații palinologice şi petrografice asupra şisturilor verzi din Dobrogea. *Analele Universității "Al. I. Cuza", Geologie*, XXXVI, Iib: 309-316, Iaşi.

ORI, G.G., ROVERI, M. & VANNONI, F., 1986. Plio-Pleistocene sedimentation in the Apenninic-Adriatic foredeep (Central Adriatic Sea, Italy). *In* Allen, P.A. & Homewood, P. (eds.), *Foreland basins*. International Association of Sedimentologists Special Publication, 8: 183-198, Oxford, Blackwell Scientific.

PARASCHIV, D., 1982. Cu privire la structura de adâncime a Platformei Moesice la est de meridianul Bucureştiului. *Studii şi cercetări de geologie, geofizică, geografie, seria Geologie*, 27: 41-56.

PARASCHIV, D. & PARASCHIV, C., 1978. Zona şisturilor verzi şi relatiile ei cu celelalte unități ale vorlandului Carpaților orientali din România. *Studii şi cercetări de geologie, geofizică, geografie, seria Geologie*, 1, 33: 44-57.

PICKERING, T.K., HISCOTT, R.N. & HEIN, F.J., 1988. *Deep Marine Environments – clastic sedimentation and tectonics*. Unwin Hyman, London, 416.

POMPILIAN, A., RĂDULESCU, F. & BITTER, M., 1991. Cercetări seismice in Dobrogea Centrală şi Sudică. *Studii şi Cercetări de Geofizică*, 29: 21-30.

POULOS, S.E. & DRAKOPOULOS, P.G., 2004. Current activity in the Thermaikos Gulf continental margin, in relation to modern sedimentation processes. *Bulletin of the Geological Society of Greece*, XXXVI: 1103-1110.

RĂDAN, S., 1994. Relicts of paleo-weathering crusts in Dobrogea. *Analele Universității Bucureşti, Seria Geologie (Abstracts volume)*, XLIII: 28.

RĂDULESCU, D., CORNEA, I., SĂNDULESCU, M., CONSTANTINESCU, P., RĂDULESCU, F. & POMPILIAN, A., 1976. Structure de la croute terestre en Roumanie. Essai d'interpretation des études seismique profondes. *Anuarul Institutului de Geologie si Geofizică*, L: 5-36.

RĂILEANU, G., PATRULIUS, D., MIRĂUȚĂ, O. & BLEAHU, M., 1968. Etat actuel des connaisances sur le Paleozoique de Roumanie. *Anuarul Institutului Geologic*, XXXVI: 63-84.

RICCI - LUCCHI, F., 1975. Depositional cycles in two turbidite formations of northern Apennines (Italy). *Journal of Sedimentary Petrology*, 45: 3-43.

ROMANESCU, D., ROŞCA, V. & SOARE, A., 1972. Recherches magnetometriques sur le platforme continental de la Mer Noire au large des cotes roumaines. *Revue Roumaine de Geologie, Geophysique et Geographie, serie Geophysique*, 16, 1: 103 – 107.

SĂNDULESCU, M., 1980. Analiza geotectonica a catenelor alpine situate in jurul Marii Negre occidentale. *Anuarul Institutului de Geologie şi Geofizică*, LVI: 5 – 54.

SĂNDULESCU, M., 1984. *Geotectonica României*. Editura Tehnică, Bucureşti, 336.

SĂNDULESCU, M. & VISARION, M., 1988. La structures des plateformes situées dans l'avant-pays et audessous des nappes du flysch des Carpathes Orientales. *Studii Tehnice şi Economice, Seria D, Prospecțiuni Geofizice*, 15: 61-68.

SEGHEDI, A. & OAIE, G., 1993. Structural style along the Peceneaga-Camena strike-slip fault zone. *Terra abstracts, abstracts supplement no 2 to Terra Nova*, Graz, 5: 32.

SEGHEDI, A. & OAIE, G., 1994. Tectonic setting of two contrasting types of pre-alpine basement: North versus Central Dobrogea. Abstracts volume, ALCAPA II Conference, supplement no 1 to *Romanian Journal of Tectonics and Regional Geology*, 75: 56.

SEGHEDI, A. & OAIE, G., 1995. Palaeozoic evolution of North Dobrogea. *In:* Săndulescu, M., Seghedi, A., Oaie, G., Gradinaru, E. & Rădan, S., *Field Guidebook, Central and North Dobrogea*, IGCP Project no. 369 "Comparative evolution of PeriTethyan Rift Basins", Mamaia 1995, 75p.

SEGHEDI, A., OAIE, G., RĂDAN, S. & CODARCEA, V., 2000. Source Areas for the Late Proterozoic Turbidites from the Moesian Block. *TESZ & PACE Conference, Zakopane, Poland, Abstracts volume*: 75.

SEGHEDI, A., OAIE, G., IORDAN M. & VAIDA, M., 2001. Correlation of the Vendian Basins along the Southern Margin of Baltica. *EUROPROBE Conference Ankara, Turkey, Abstracts*: 70-71.

SEGHEDI, A., BERZA, T., IANCU, V., MĂRUNȚIU, M. & OAIE, G. (this volume). Neoproterozoic terranes in the Moesian Basement and in the Alpine Danubian Nappes of the South Carpathians. *Geologica Belgica*.

SEGHEDI, A., OAIE, G., IORDAN, M., AVRAM, E., TATU, M., CIULAVU, D., VAIDA, M., RADAN, S., NICOLAE, I., SEGHEDI, I., SZAKACS, A. & DRAGANESCU, A., 1999. Excursion Guide of the Joint Meeting of EUROPROBE TESZ, PANCARDI and GEORIFT Projects: "Dobrogea – the interface between the Carpathians and the Trans-European Suture Zone": Geology and structure of the Precambrian and Paleozoic basement of North and Central Dobrogea. Mesozoic history of North and Central Dobrogea. *Romanian Journal of Tectonics and Regional Geology*, 77, suppl. 2, 72 p.

SHANMUGAM, G. & MOIOLA, R.J., 1982. Eustatic control of turbidites and winnowed turbidites. *Geology*, 10: 231-235.

SHANMUGAM, G. & MOIOLA, R.J., 1984. Eustatic control of the calciclastic turbidites. *Marine Geology*, 56: 273-278.

SHANMUGAM, G. & MOIOLA, R.J., 1988. Submarine fans: Characteristics, Models, Classification and Reservoir Potential. *Earth-Science Reviews*, 24: 383-248.

SHANMUGAM, G., MOIOLA, R.J. & DAMUTH, J.E., 1985. Eustatic control of submarine fan development. *In* Bouma, A.H., Normark, W.R. & Barnes, N.E. (Eds), *Submarine Fan and Related Turbidite Systems*. Springer Verlag, New York, N.Y., 23-28.

SHEPARD, F.P., McLOUGHLIN, P.A., MARSHAL, L.F. & SULLIVAN, G.G., 1977. Current meter recordings of low-speed turbidity currents. *Geology,* 5: 297-301.

STOW, D.A.V. & SHANMUGAM, G., 1980. Sequences of structures in fine-grained turbidites: comparison of the deep sea and ancient flysch sediments. *Sedimentary Geology*, 25: 23 – 46.

TUREKIAN, K.K. & WEDEPOHL, K.H., 1961. Distribution of the elements in some major units of the Earth crust. *Geological Society of America Bulletin*, 72: 175-192.

UNRUG, R., CZESŁAW, H. & CHOCYK-JAMIŃSKA, M., 1999. Easternmost Avalonian and Armorican-Cadomian Terranes of Central Europe and Caledonian-Variscan Evolution of the polydeformed Kraków Mobile Belt: geological constraints. *Tectonophysics*, 302: 133-157.

VAIL, P.R., MITCHUM, R.M., JR. & THOMPSON, S. III, 1977. Seismic stratigraphy and global changes of sealevel, part 4: Global cycles of relative changes of sea level. *In* Payton, C.E. (Ed.), *Seismic Stratigraphy – Applications to Hydrocarbon Exploration*. American Association of Petroleum Geologists Memoir, 26: 83-97.

VISARION, M., MAIER, O., NEDELCU-ION, C. & ALEXANDRESCU, R., 1979. Modelul structural al metamorfitelor de la Palazu Mare, rezultat din studiul integrat al datelor geologice, geofizice şi petrografice. *Studii şi Cercetări de Geologie, Geofizică, Geografie, seria Geofizică*, 17: 95-113.

VISARION, M., SĂNDULESCU, M., STĂNICĂ, D. & VELICIU, Ş., 1988. Contributions a la connaissance de la structure profonde de la plateforme Moesienne en Roumanie. *Studii Tehnice şi Economice*, *Seria Geofizică*, D, 15: 211-222.

VISARION, M., SĂNDULESCU, M., ROŞCA, V., STĂNICĂ, D. & ATANASIU, L., 1990. La Dobrogea dans le cadre de l'avant-pays carpatique. *Revue Roumaine de Géophysique*, 34: 196-225.

WADE, M., 1969. Medusae from Uppermost Precambrian and Cambrian sandstones, Central Australia. *Paleontology*, 12, 3: 351 – 365.

WADE, M., 1972. Hydrozoa and Scyphozoa and other Medusoids from the Precambrian Ediacaran Fauna, South Australia. *Paleontology*, 15, 2: 196 – 225.

WAGREICH, M., KRENMAYR, H. G., 2005. Upper Cretaceous oceanic red beds (CORB) in the Northern Calcareous Alps (Nierental Formation, Austria): slope topography and clastic input as primary controlling factors. *Cretaceous Research*, 26: 57-64.

WALKER, R.G., 1979. Facies and Facies Models, general introduction. In Walker, R.G. (Ed.), *Facies Models*. Geoscience Canada, Reprint series, 1: 1-7.

WALKER, R.G. & MUTTI, E., 1973. Turbidite facies and facies associations. *In* Middleton, G.V. & Bouma, A.H., *Turbidites and Deep-Water Sedimentation*. Pacific Section Society of Economists, Palaontologists and Mineralogist, Los Angeles, California, 119-157.

WINCHESTER, J.A. & PACE TMR NETWORK, 2002. Palaeozoic Amalgamation of Central Europe: new results from recent geological and geophysical investigations. *Tectonophysics,* 360: 5-22.

WINCHESTER, J.A., YIGITBAS, E., BOZKURT, E. & PHARAOH, T.C., 2004. Microcontinent-Promontory interactions: is there a link between Avalonia, Moesia and the Istanbul Block, NW Turkey ? *Avalonia-Moesia Symposium and Workshop, Ghent/Ronse 2004, Abstractbook*: 47.

WINCHESTER, J.A., PHARAOH, T.C., VERNIERS, J., IOANE, D. & SEGHEDI, A., 2005. Accretion of Avalonia and the Armorican terrane assemblage to the East European craton. European Lithosphere Dynamics *(in press)*.

WÜEST, A., 2003. Bottom boundary mixing. *Aquatische Physic II. Transport and mixing in natural waters*. EAWAG, Kastanienbaum, 5.77-5.85.

ŻELAŹNIEWICZ, A., SEGHEDI, A., JACHOWICZ, M., BOBIŃSKI, W., BUŁA, Z. & CWOJDZINSKI, S., 2001a. U-Pb SHRIMP Data Confirm the Presence of a Vendian Foreland Flysch Basin Next to the East European Craton: *EUROPROBE Conference Ankara, Turkey, Abstracts*, 98-101.

ŻELAŹNIEWICZ, A., SEGHEDI, A., JACHOWICZ, M. & BOBIŃSKI, W., 2001b. A Neoproterozoic orogen at the present-day SW margin of the East European Craton. *EUG XI Conference - Strasbourg, France, Abstracts* Vol. 6, 1, 525-526.

ŻELAŹNIEWICZ, A., SEGHEDI, A., BUŁA, Z. & BOBIŃSKI, W., in preparation. Fragments of a Neoproterozoic orogen derived from a Grenvillian-type crust at the Teisseyre-Tornquist margin of Baltica: U-Pb SHRIMP data from southern Poland and Romania .

Manuscript received on 31.03.2005 and accepted for publication on 15.11.2005.

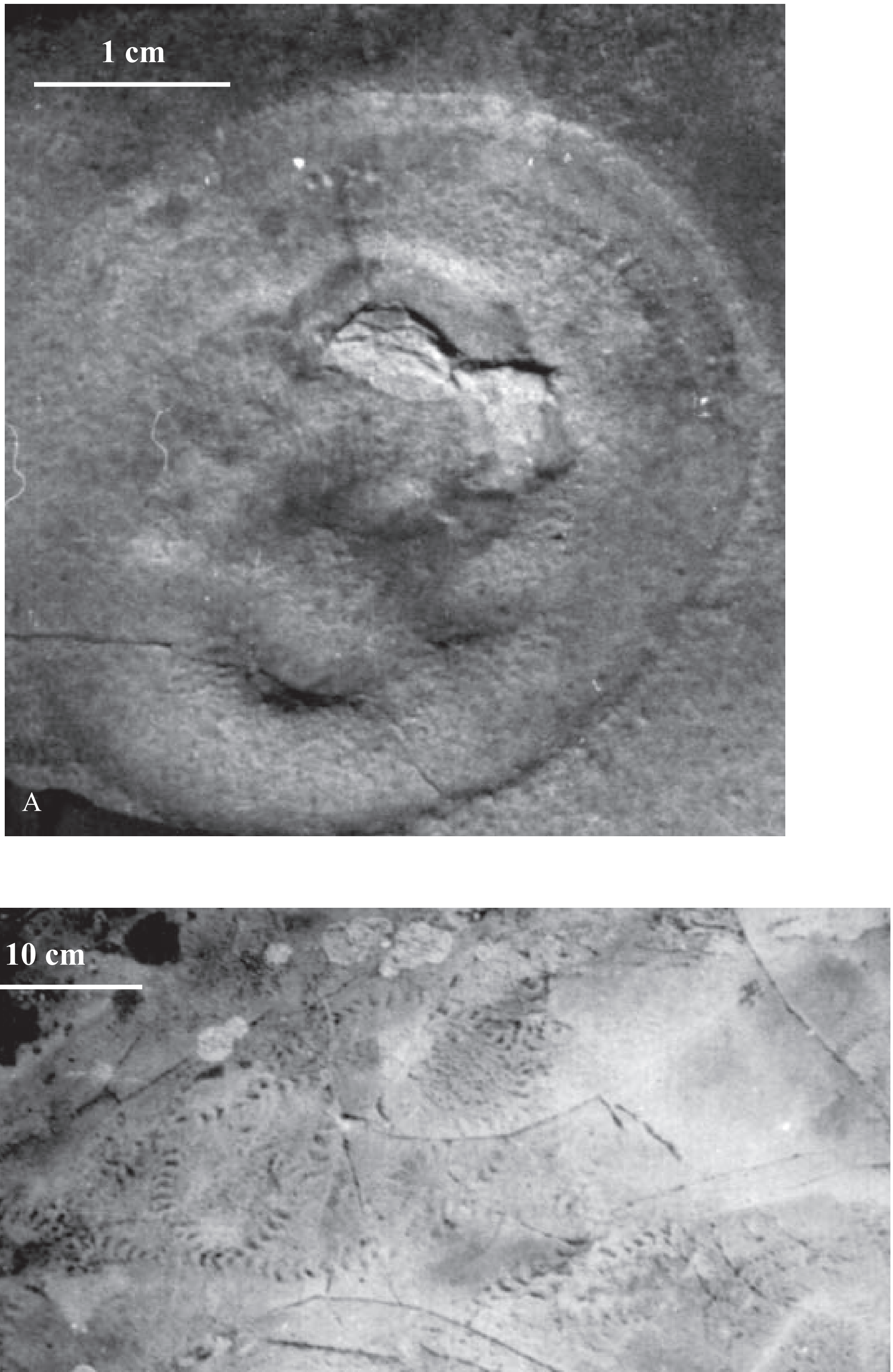


Plate 1. Traces of organic activity in Te beds (Haidar Member) of the Histria Formation. A) impression of *Nemiana simplex* Palij 1974 in fine-grained turbidites, Casimcea valley north of village Razboieni; B) unidentified meandering trails on the bed surface in pelites, left tributary of Casimcea valley NE of Casimcea locality (after Oaie, 1999).

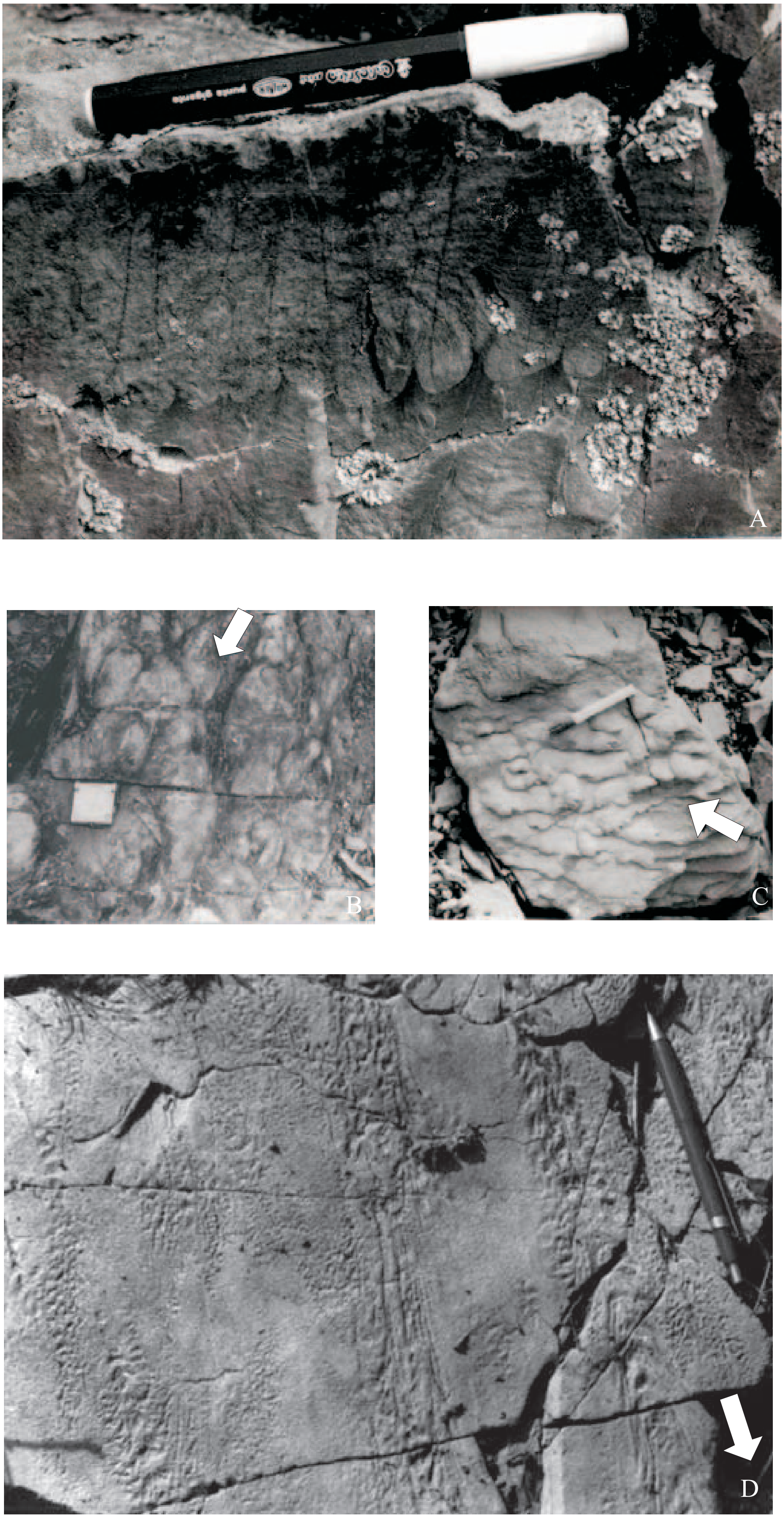
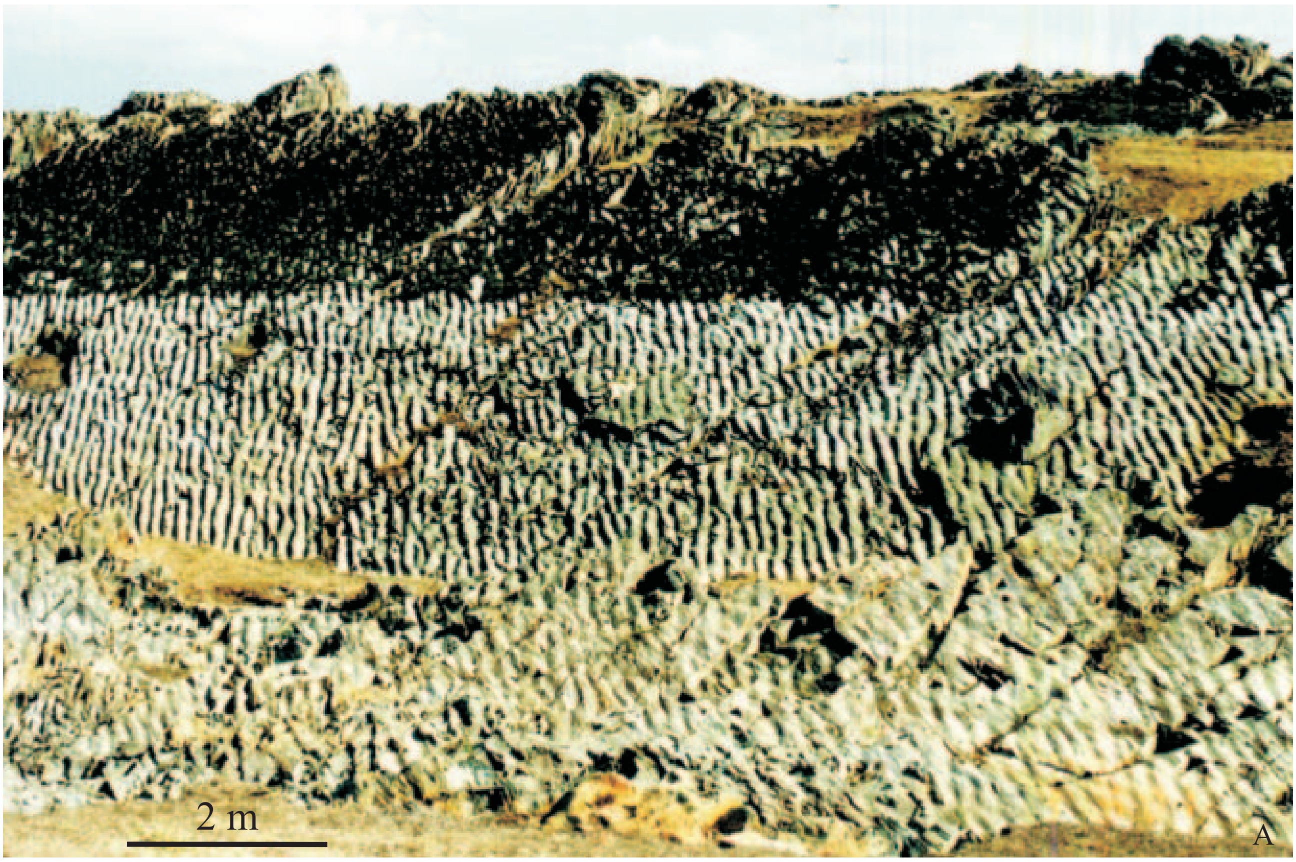
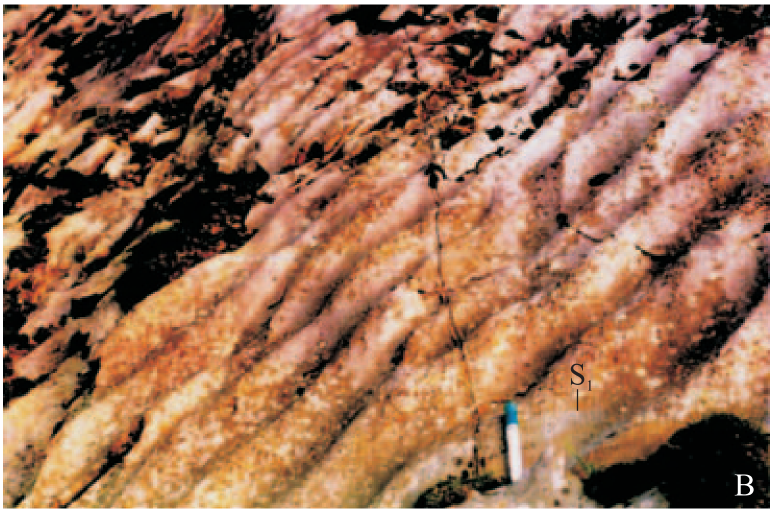


Plate 2. Sedimentary structures in turbidites (Haidar Member) of the Histria Formation. A) Flame structures developed due to differential loading at the base of a sandstone bed overlying a pelite (Te) bed; B & C) Load casts; D) Rill marks and chevron marks (right corner) on bed surface. Arrows indicate the flow direction.





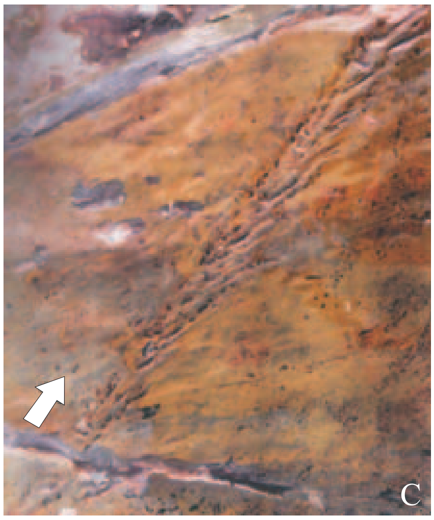


Plate 3. Sedimentary structures on the bed surface in fine-grained turbidites (Haidar Member) of the Histria Formation. A) Large bedding surface with ripple marks parallel to the current direction; B) Drag mark overprinting ripple marks on a bedding plane; note the fine intersection lineation between the rippled surface (S_o) and the slaty cleavage (S_1); C) Chevron marks on a bed surface north of Piatra village; the arrow indicates the palaeoflow direction.

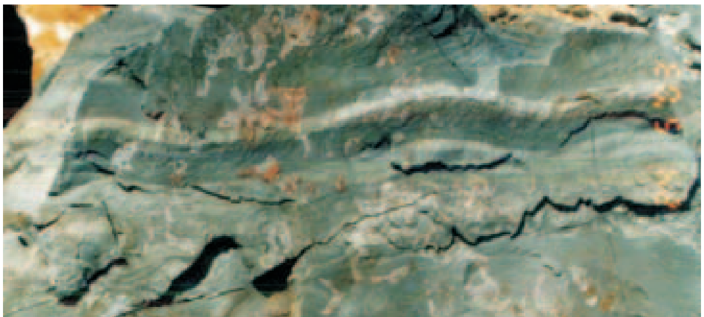


Plate 4. Coarse - grained facies of the Histria Formation, Sibioara Quarry (Sibioara member). A) Clast supported orthoconglomerate with coarse, weakly sorted matrix, overlain by a reverse graded medium to coarse-grained sandstone bed; B) Poorly sorted conglomerate grading upwards into massive sandstone. Note the abundance of subangular lithic clasts (granites, gneisses, rhyolites).