PALEOGENE LITHO- AND BIOSTRATIGRAPHY OF THE NE GETIC DEPRESSION (ROMANIA)

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Abstract: This paper presents the updated stratigraphy of the Paleogene from the NE of the Getic Depression, between the Cheia Valley (to the west) and Râul Doamnei Valley (to the east). The lithological units are divided and correlated based on their litho- and palaeontological, as well as geochemical features. Five sections were detailed investigated in the field and analyzed from calcareous nannoplankton point of view. The results of this study allow us to improve the lithostratigraphic nomenclature and to define a new lithostratigraphic unit - the Brăduleț Formation, which extends in the Early Oligocene-Early Miocene interval. Our work also brings new data on the sedimentology and stratigraphy of the Upper Eocene Olăneşti Formation, as well as of the Oligocene Cheia and Corbi formations. The nannofloral investigations point out the presence of the NP21, NP22, NP23, NP24, NP25, NN1 and NN2 calcareous nannoplankton zones, covering the latest Eocene (Late Priabonian) – earliest Miocene (Early Burdigalian) interval. Fluctuation in the abundance and distribution patterns of nannofloras revealed significant palaeoclimatic and palaeoenvironmental modifications in the studied area, during the Late Eocene, Oligocene and Early Miocene intervals, reflecting both regional and global changes of those times, including rapid and significant climatic modifications.

Keywords: Paleogene; litho- and biostratigraphy; sedimentology; palaeoenvironmental changes; Central Paratethys.

1. INTRODUCTION

The Paleogene is one of the most intriguing intervals in the Earth history, marked by significant changes in palaeoclimate, marine productivity and in global carbon cycle.

During the Paleocene-Eocene thermal maximum, reduced oceanic turnover, decreases in global $\delta^{13}C$ and in marine productivity were assumed (Aubry *et al*., 1996; Zachos *et al*., 2001; Bralower *et al*., 2002). Within the Early Oligocene glacial maximum, intensification of deep ocean circulation, elevated $\delta^{13}C$ and high productivity were supposed (Savin, 1977; Aubry, 1992; Zachos *et al*., 1996). Presumably, sudden changes in climate and in ocean circulation might occur as result of gradual forcing, as certain physical thresholds exceeded. The climate of the Paleogene seems to be a critical turning point from a warm, humid and high-diversity "greenhouse" world of the Paleocene - Eocene interval to glacial "icehouse" conditions of Oligocene times.

Remarkably, the palaeogeography of the European Tethys Realm significantly changes during the Oligocene, due to the isolation of the Paratethys Realm (in the Central and Eastern European regions - Báldi, 1980; Rusu, 1988; Rögl, 1998; 1999) from the Mediterranean Realm (in the Western and Southern part of Europe). The evolution pattern of Paleogene marine biotas (and especially of the most sensitive planktonic ones) mirrored the climatic deterioration and the palaeoenvironmental changes.

Paleogene deposits are widespread throughout Romania, in the Carpathian belt. One of the Romanian areas where the Paleogene deposits are well preserved, allowing, based on the nannofloral content to decipher their biostratigraphy, is the Getic Depression, located south of the Southern Carpathians (Fig.1).

Geological mapping and exploration for oil and gas in the Getic Depression drew attention for over a hundred years. Pioneer investigations of the geology of the Getic Depression belong to Ştefănescu (1897), Murgeanu (1941a; 1941b), Iorgulescu (1953) and Popescu (1954). These preliminary studies were followed by several others that elaborated on the stratigraphy of the Paleogene deposits in the Getic Depression (Fig. 2; Tătărâm, 1964; Popescu *et al.,* 1976, 1977; Jipa, 1980; 1982; Ştefănescu, 1980; Bombiță *et al.,* 1980).

Over the last decade, the investigation in the Getic Depression continued with more sedimentological and biostratigraphical works, such as Jipa (1994), Rusu *et al.* (1996) and Ryer (1998).

Notably, the most reliable biostratigraphic data for the Oligocene deposits of the Getic Depression are offered so far by the study of the calcareous nannofossils (Popescu *et al*., 1976; Rusu *et al*., 1996). The foraminifers appear to be good stratigraphical markers mostly for the Eocene deposits (Tătărâm, 1964; Bombiță *et al.,* 1980). In the Oligocene sediments they are very scarce or even absent. Yet the Oligocene macrofaunas present is quite rare, with a notable exception - the fish remains. The fish fossils are frequent and quite well preserved, but because most often taxa having a wide range are present,

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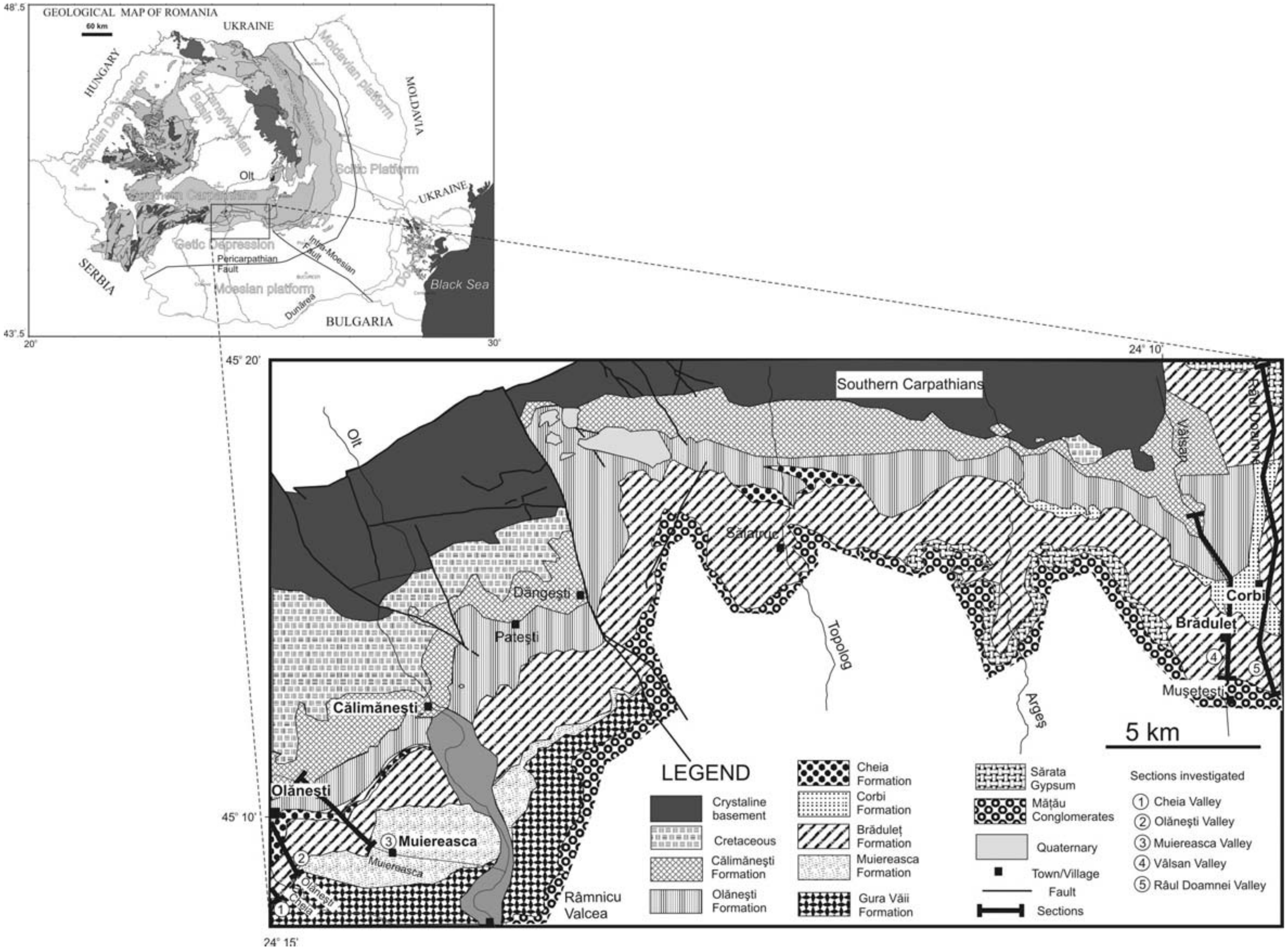

Fig. 1. Geological map of the NE Getic Depression (modified after Murgeanu *et al*., 1967; Popescu *et al*., 1976, 1977; Lupu *et al*., 1978; Bombiță *et al*., 1980; Dimitrescu *et al*., 1978; Morariu and Teodorescu, 1987, and Ryer, 1998)

they are not appropriate for high-resolution biostratigraphic studies.

The purpose of this paper is to present an updated stratigraphy of the Paleogene succession in the NE of Getic Depression, between the Vâlcea county border in the west and the Argeş county border in the east (Fig. 1).

According to the sedimentological, geochemical as well as palaeontological evidences, the lithostratigraphy of the Paleogene succession in the study area is interpreted within a new biostratigraphic framework. Based on the calcareous nannoplankton fluctuation pattern we document significant palaeoenvironmental and palaeogeographical changes that characterized the Paleogene evolution of the Getic Depression.

2. GEOLOGICAL SETTING

The Getic Depression is situated within a foreland basin formed in front of the South Carpathians, in response to the flexural loading of the Moesian Platform (Fig. 1). The Getic Depression covers a large part of the foredeep depozone (DeCelles and Giles, 1996) of the South Carpathian foreland basin, which presently is highly deformed (Săndulescu, 1984, 1988; Mutihac, 1990; Motaş *et al.*, 1995; Dicea, 1996; Maţenco, 1997; Răbăgia and Maţenco, 1999). The Pericarpathian Fault marks the boundary between the Getic Depression and the Moesian Platform, while the Intra-Moesian Fault separates the Getic Depression from the South Carpathians (Săndulescu, 1984).

In the northern part, the basement of the Getic Depression consists of a nappe system in which four major tectonic units are recognized: the Danubian, Severin, Getic and Supragetic Nappes (Săndulescu, 1984). In the south, the Moesian Platform represents the basement of the Getic Depression.

Following the Cretaceous deformation in the Carpathians (Săndulescu, 1984), the South Carpathians entered a phase of large-scale dextral deformation that caused E-W contraction and subsidence in the Getic Depression (Ratschbacher *et al.,* 1993).

The Paleogene of the Getic Depression, exposed west of Olt Valley, is characterized, according to Popescu *et al.* (1996) by the succession of the following lithostratigraphic units (Fig. 2): (1) the Călimăneşti Conglomerates, (2) the Cheia Conglomerates, (3) the Pucioasa type Marls, and (4) the Muiereasca Sandstone. In addition, massive sandstones (the Corbi Sandstone) are exposed in the eastern part of the Olt Valley.

3. DATA BASE AND METHODS

Several stratigraphic sections were measured at five localities: Cheia, Olăneşti, Muiereasca, Vâlsan and Râul Doamnei valleys (Fig. 1). A simplified composite stratigraphic section measured at each locality is shown in Fig. 3.

Petrography of sandstones and conglomerates was undertaken in thin-sections and acetate peels analized under Nikon E400-POL and Zoom Nikon SMZ 800 microscopes, respectively. The organic-matter content was estimated based on the”*loss on ignition*” method of Heinri *et al.* (2001), where 5 grams of sample are in combustion at 550°C for four hours. The carbonate content (%$CaCO_3$) was analyzed with an Eijkelkamp calcimetre.

All measured successions were sampled for calcareous nannoplankton investigations. The nannofloral studies focused on qualitative analysis and were carried out under a microscope with cross-polarized light, at 1600x magnification. The calcareous nannofossil taxonomy follows Perch-Nielsen (1985).

4. RESULTS

4.1 Olăneşti Formation

Stratigraphy and sedimentology

The Olăneşti Formation (Ryer, 1998), also described as the Lower Marls (Murgeanu, 1941b) or the Olăneşti Marls (Popescu *et al.*, 1976) continuously crops out between the Cheia Valley to the west and Râul Doamnei Valley towards the East (Fig. 1, Fig.3). The stratotype of this unit is exposed along the Olăneşti Valley, where the succession is 350 m thick. Over the entire study area this unit ranges in thickness from 200 m in the Cheia Valley to a maximum of 600 m in the Muiereasca Valley (Fig. 3).

The lower boundary of the Olăneşti Formation shows a gradual transition to the underlying Călimăneşti Formation. North of the Brădet Village, Jipa (1980, 1982) mapped a 40-50 m thick, westward thinning matrix-supported conglomeratic unit, termed the *Tilloid Conglomerates Level*, which extends over 40 km between Vâlsan and Olt (Sălătruc) valleys, were is ~ 0.55 m thick. This unit is characterized by a very poor sorting. Locally, it contains some very large clasts, up to 3 m length. Notably, in the Muiereasca Valley, several thin coarse-grained beds are scattered throughout the lower 30-50 m in the lower part of the Olăneşti Formation. These beds do not share the same characteristics that Jipa (1982, 1984) described within the Tilloid Conglomerates Level. The *Tilloid Conglomerates Level* could be used as lithological marker for the boundary between the Călimăneşti Conglomerates and the Olăneşti Marls.

At the top, the Olăneşti Formation is both disconformity overlain by the Cheia Formation (towards the west) and by the Corbi Formation (towards the east). In the central part (i.e. Topolog and Argeş valleys), the Olăneşti Formation, is paraconformity overlain by a sequence of bituminous mudstones, described herein as the Brăduleţ Formation.

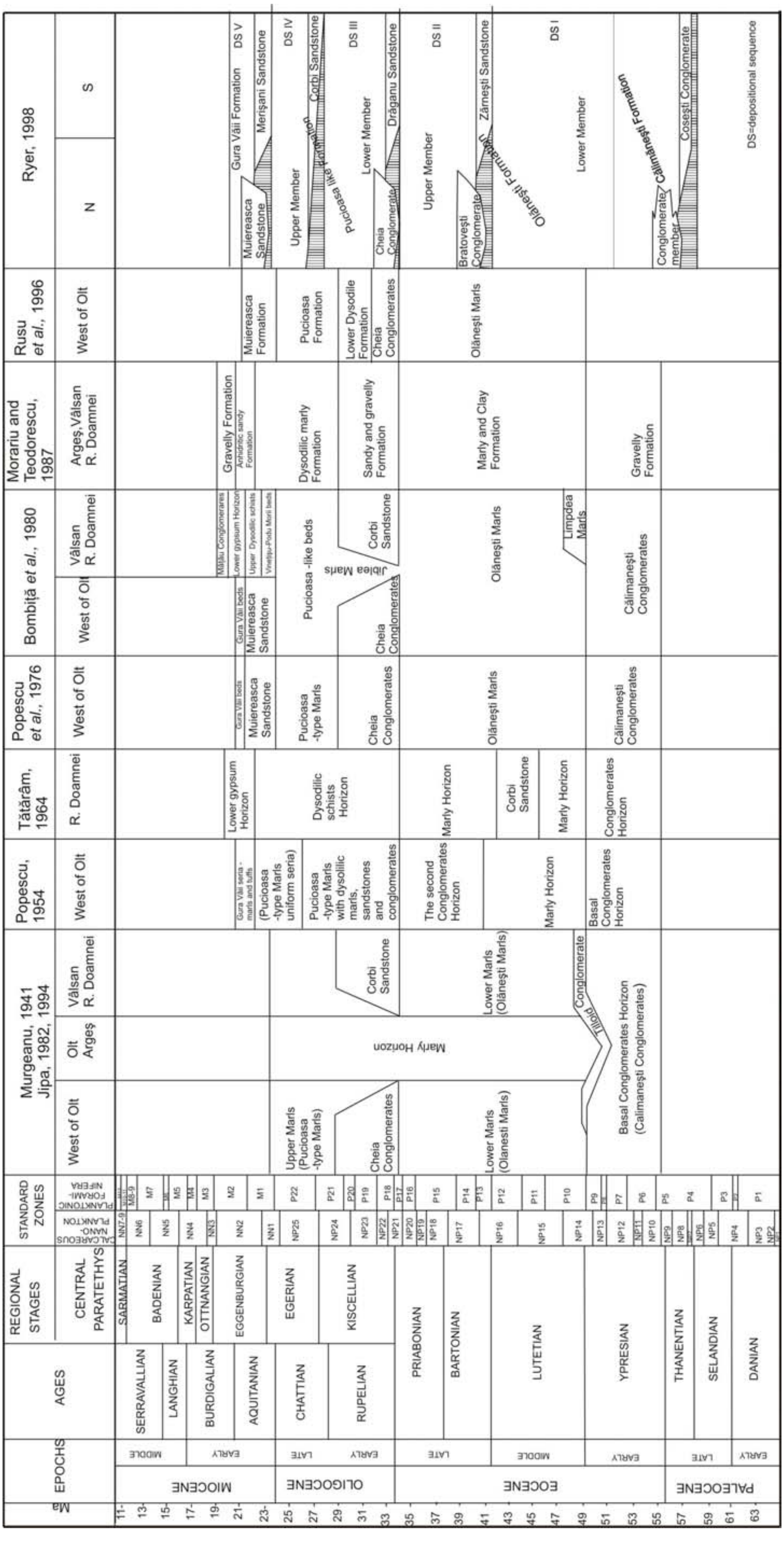

Fig. 2. Main previous stratigraphical schemes of the Paleogene of the NE Getic Depression

On the Vâlsan Valley, the Olăneşti Formation is composed of mudstones (up to 70%), sandstones (20%) and subordinate conglomerates. The lower part of the Olăneşti Formation consists of approximately 45 m of matrix-supported conglomerates and gravely mudstone. The matrix is muddy, locally reaching up to 90% of the rock volume.

The sandstones of the Olăneşti Formation range in thickness from centimeters to decimeters. In the lower 45 m of the succession, individual sandstone beds associated with clast-supported conglomerates may reach metric thickness. The sandstones are lithic to sublithic and occasionally contain abundant bioclastic material (*nummulite* shells). The most common sedimentary structures include massive, normal grading as well as parallel stratification/lamination, frequently yielding current ripples (Jipa, 1980, 1982, 1994, Ryer, 1998; Roban *et al.,* 2005 b). Trace fossil assemblages include *Ophiomorpha sp.* and *Skolithos sp.* (Brustur, pers. comm.) typical of shallow marine environments.

In the Olăneşti Formation, the conglomerates (up to 3 m in thickness) have a lenticular shape and display normal grading and intense scouring at their lower bounding surfaces. Their polymictic clast composition reflects the proximity to the source area located in the rising of the Southern Carpathians.

Thin beds of laminated mudstones represent the finest grain-size recorded in the Olăneşti Formation and have 3 - 10 % $CaCO_3$. Sometimes they form couplets with silt laminae. The organic mater contents are of 3-5%.

The Olăneşti Formation was divided, in the Vâlsan Valley section, into three distinct lithofacies (Morariu and Teodorescu, 1987; Roban *et al.*, 2005 b):

(1) gravely mudstones, (45 m), occurring only at the base, with conglomerates and sandstone levels;

(2) heterolithic thin-bedded sandstones and mudstones, having a recurrent character. Two sequences of this lithofacies, the oldest in the middle part of the Olăneşti Formation (75 m in thickness) and the youngest towards the top (175 m in thickness) were recognized;

(3) mudstones and siltstones (85 m in thickness), with parallel lamination, interbedded with rare massive or parallel laminated sandstones. This lithofacies is present in the middle part of the Olăneşti Formation.

<u>Biostratigraphy and age</u>

Based on its foraminiferal and nannofloral content, the Olăneşti Formation was previously assigned to the Lutetian and Priabonian stages (Popescu *et al.*, 1976; Rusu *et al.*, 1996). Notably, the Eocene/Oligocene boundary (the NP21 Calcareous Nannoplankton Zone of Martini, 1971) is placed in its uppermost part (Fig. 3).

The samples collected from the topmost of the Olăneşti Formation (Fig. 3) contain taxonomical diversified and very well preserved calcareous nannofloral assemblages. These assemblages include, as biostratigraphical significant nannofossils, *Discoaster saipanens* Bramlette & Riedel, *D. barbadienis* Tan, *Clausicoccus fenestratus* (Roth & Hay) Prins, *Discoaster tanii* (Bramlette & Riedel) Bukry and *Helicosphaera reticulata* Bramlette & Wilcoxon. All these taxa have their last occurrence (LO) at the top of the Eocene (lower part of the NP21 Calcareous Nannoplankton Zones – Martini, 1971; Perch-Nielsen, 1985; Krhovský *et al.*, 1993; Melinte, 1995). The above-mentioned findings indicate that the upper part of the Olăneşti Formation is latest Eocene age (Late Priabonian) and, the Eocene/Oligocene boundary is located immediately below the top of formation (Figs. 3, 4).

4.2. Cheia Formation

<u>Stratigraphy and sedimentology</u>

We apply in this paper the name Cheia Formation to designate a distinct lithologic unit, which could be map in the whole study area, stratigraphically overlying the Olăneşti Formation. Previously, the Cheia Formation was referred in the literature as the Cheia Conglomerates (Popescu *et al.*, 1976).

The type locality of this formation is the Cheia Valley (Figs. 1, 3) where it reaches a maximum thickness of 500 m. Towards the East, the Cheia Formation thins to a minimum of 20 m at Muiereasca Valley (Fig. 3). According to some authors (Popescu *et al.*, 1976; Bombiță *et al.,* 1980; Jipa, 1994), the formation completely pinches out east of Olt Valley (Fig. 1). However, Ryer (1998) described as Cheia Conglomerates a package of 75 m, located at the base of the Pucioasa like Formation in the Topolog Valley (Fig. 1).

In the Cheia and Olăneşti valleys, the boundary between the Cheia and Olăneşti formations is a disconformity one, while in the Muiereasca valley, the Cheia Formation pinch out laterally to the Brăduleț Formation (Figs. 3, 4).

The Cheia Formation is a coarse-grained unit, dominated by pebble and cobbles conglomerates that make up to 80 % of the entire succession.

Apart from pebbles and cobbles, the grain size spectrum includes also, boulders and even metre-scale blocks. In general, a relation of proportionality is observed between the grain size and the thickness of individual beds. Thus, the coarsest units have the maximum thickness in the succession. However, there are some decimetre-thick beds that, which although can be considered thin, and they contain decimetre-long clasts. These particular beds pinch out within several tens of meters.

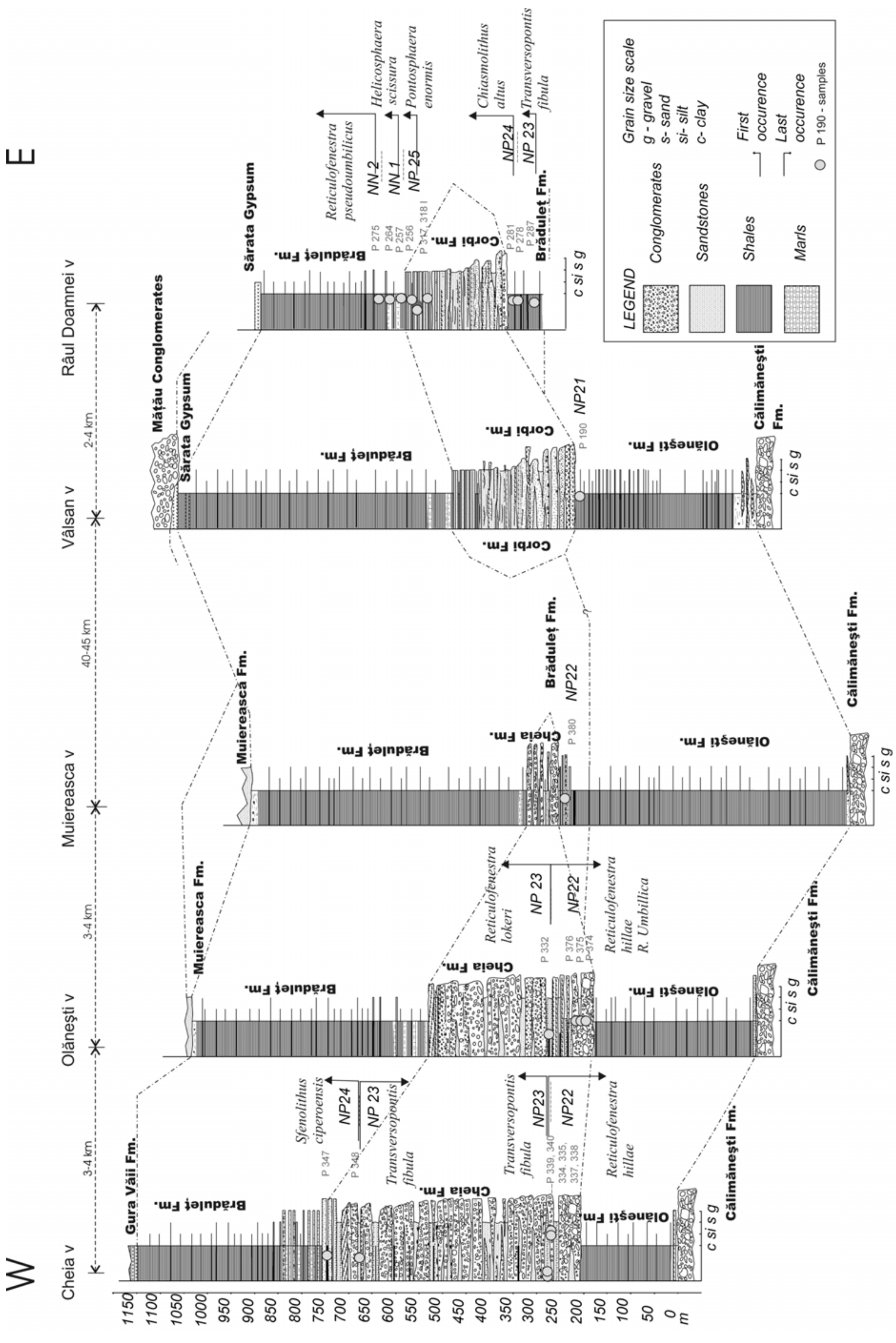

Fig. 3. Lithology and biostratigraphy of the investigated sections from the NE Getic Depression

Most of the conglomeratic beds have a wide range of thickness, from 30 cm to 5 m (due to amalgamation), while their lower surface shows scouring. Most commonly, the base of these beds is flat (Ryer, 1998).

We can suppose that the observed variation in thickness is due to the amalgamation process. We use herein the term amalgamation to indicate a mix process, erosional-constructional, leading to a depositional unit, homogenous lithologically, yielding a bigger apparent thickness, and having a multistory deposition (*sensu* Clark and Pickering, 1996).

Internal structures include massive, normal and reverse grading, diffuse parallel bedding and imbrications. The particularity of these conglomerates is rapid transition, vertically and laterally, in some meters, between types of internal structures of the same depositional unit.

Jipa (1982, 1994) also found oblique stratifications, at medium and large scale.

At Olăneşti and Cheia localities, the middle respective upper part of the Cheia Formation consists of several 30 cm up to 3 m, thick matrix-supported conglomerates that contain both extrabasinal as well as muddy intraclasts (Fig. 1). The matrix is muddy.

Petrographic analysis of the conglomerates throughout the stratigraphic succession suggested a polymictic origin with fragments of metamorphic and sedimentary rocks (limestones, marls and sandstones).

Associated with the conglomerates there are massive, and diffuse parallel bedded sandstones. Mudstones interbedded with thinly bedded siltstones and very fine-grained sandstones are present towards the top of the succession.

The general trend of the succession illustrated in the stratigraphic logs of Fig. 3 is fining and thinning upward. The megasequence consists of fining-upward cycles composed of conglomerates, sandstones and sometime mudstones. Each cycle is about 5-10 m and has a flat or sometime scoured erosional base (Ryer, 1998; Roban, 2004).

Biostratigraphy and age

Previous studies on the biostratigraphy of the conglomeratic succession, termed here the Cheia Formation, indicated that the duration of the formation extends over the entire Rupelian stage of Early Oligocene (Popescu *et al.*, 1976; Bombiță *et al.*, 1980). Rusu *et al.* (1996) indicated an earliest Oligocene (Early Rupelian) age for the Cheia Formation. According to the above-mentioned studies, the Late Rupelian (including the Rupelian/Chattian boundary interval) was characterized by the deposition of the Lower Dysodilic Shale Formation (organic-rich, bituminous shales).

The samples collected in this study at Cheia, Olăneşti and Muiereasa localities (Figs. 1, 3) contain scarce calcareous nannoplankton associations with a poor to moderate preservation.

The nannofloras from the base of the Cheia Formation are dominated by holococcoliths, such as *Isthmolithus recurvus* Deflandre *in* Deflandre & Fert, *Zygrhablithus bijugatus* (Deflandre *in* Deflandre & Fert) Deflandre, *Orthozygus aureus* (Strdaner) Brameltte & Wilcoxon and *Lanternitus minutus* Stradner, as well as by reticulofenestrids - *Reticulofenestra hillae* Bukry & Percival and *R. umbilica* (Levin) Martin & Ritzowski. Based on the identified nannofloras, the base of the Cheia Formation is placed in the earliest Oligocene (Early Rupelian) - the NP22 Calcareous Nannoplankton Zone (Fig. 3).

A significant change in nannofloral composition was noted within the lower part of the Cheia Formation. The presence of the NP23 Calcareous Nannoplankton Zone of Martini (1971) is argued by the first occurrence (FO) of the nannofossils *Reticulofenestra lockeri* Müller, *Dictyococciets ornatus* (Müller) Bistrická, *Transversopontis fibula* Gheta and *T. latus* Müller. Remarkably, these are endemic species (Báldi-Beke, 1981; Krhovsky *et al.*, 1992; Nagymarosy and Voronina, 1992; Melinte, 1993, 1995; Rusu *et al.*, 1996 b), related to the first isolation of the Paratethys Realm.

The samples collected form the upper part of the Cheia Formation (in the Cheia Valley) contain nannofloras dominated by cosmopolitan nannofossils, as *Dictyococcites bisectus* (Hay *et al.*) Bukry & Percival, *Reticulofenestra ornata* Müller, *Cyclicargolithus floridanus* (Roth & Hay) Bukry, *Pontosphaera latelliptica* (Báldi-Beke & Báldi) Perch-Nielsen and *Sphenolithus moriformis* (Brönnimann & Stradner) Bramlette & Wilcoxon. The top of the Cheia Formation contains *Sphenolithus ciperoensis* Bramlette & Wilcoxon, which is the marker species of the NP24 Calcareous Nannoplankton, latest Rupelian – earliest Chattian in age (according to Melinte, 1995, 2005).

4.3 The Corbi Formation

Stratigraphy and sedimentology

We apply the name Corbi Formation to describe a 250 m thick sandstone-dominated succession exposed in the Râul Doamnei Valley section (Corbi village - Figs. 1, 3). This unit, previously referred as the Corbi Sandstone (Ştefănescu, 1897), is exposed between the Vâlsan and Râul Doamnei valleys (Murgeanu, 1941: Figs. 1, 3; Jipa, 1980). Morariu and Teodorescu (1987) considered that the Corbi Sandstone extends also in the Argeş Valley (Fig. 2).

The lower lithological boundary of the Corbi Formation with the underlying Olăneşti Formation is a disconformity one (Fig. 3). At the upper part, the Corbi Formation fines upwards and has a gradual contact with the overlying Brăduleț Formation (Fig. 3). Overall, the Corbi Formation shows a fining upward succession.

Our study revealed that, lithologically, the Corbi Formation consists predominantly of sandstones (50%), and subordinately of conglomerates (35%) and mudstones (15 %).

The sandstone beds range in thickness from a few centimetres to several metres, this variation most probably resulting from depositional amalgamation processes. The maximum thickness measured within an amalgamated unit is 10 m. Main sedimentary structures include massive, normal grading, current and ripples, parallel lamination/stratification and scouring (rip-up clasts). Deformational soft-sediment and liquefaction structures are abundant. Several centimetres pebble stringers or lenticular pebble conglomerate beds are also present within the thick sandstones. Petrographical analysis of the sandstones revealed the predominance of lithic to sublithic types with some subordinate quartz and feldspatic sandstones present as well.

The conglomerates are more abundant in the lower part of the Corbi Formation (Fig. 3), where form several metres thick packages due to amalgamation. They are generally clast-supported conglomerates. The main structures include massive, normal grading and parallel stratification/lamination (Jipa, 1980, Ryer, 1988, Roban, 2004) and abundant scouring such as rip-up clasts. In the Vâlsan Valley section (Figs. 1, 3), within the lower part of the formation, the amalgamated thickness of the conglomeratic beds is up to 10 m. There the maximum grain size varies between boulders to cobbles, while the matrix consists of medium-sorted pebbles. The largest clasts tend to be oriented parallel with the general stratification. Above these levels as well as in the Râul Doamnei Valley, the coarse beds become thinly bedded with individual beds (*one event*) ranging between 10 cm to 1.5 m and grain size decreases to pebbles and granules. Petrographically, the conglomerates are polymictic with both metamorphic and sedimentary clasts.

In the Vâlsan Valley section, towards the lower part of the Corbi Formation (the lowest 25 m), the grain size ranges from cobbles to granules, and has a fining upward trend. In the Şiței Valley, the lower 30 m of the Corbi Formation (overlaying the Brăduleț Formation) consists of granular to pebbly conglomerates that progressively grade vertically into sandstones. Thus, the middle part of the formation is sandstone-dominated and consists of 2 to 10 m thick massive sandstones interbedded with 0.5 m thick sequences of sandstones and siltstones. The upper part of the Corbi Formation consists of vertically stacked fining upward sequences, that contain 1 to 3 m thick sandstones, capped by 1 to 3 m thick heterolithic thinly bedded sandstones and dark mudstones. The boundary between the Corbi and the Brăduleț formations is gradational and expressed by the progressive increase of the mudstone interbeds on the expense of sandstone beds.

Biostratigraphy and age

Previous studies (Murgeanu, 1941; Jipa, 1980, 1982; Bombiță *et al.*, 1980) considered the Corbi Sandstone to be a distal facies of the Cheia conglomerates, and consequently assigned it to the Early Oligocene age (Rupelian). Recently, Ryer (1998) assumed that the Corbi Sandstone is synchronous with the upper part of the Pucioasa Formation of Late Oligocene (Chattian).

The calcareous nannoplankton analysis of the Corbi Formation, collected at the stratotype locality (Corbi village on Râul Doamnei, Figs. 1, 3) revealed that the base of the Corbi Formation lies in the Upper Oligocene (Chattian) within the NP24 Calcareous Nannofossil Zone. This is indicated by the common occurrence of the nannofossils *Sphenolithus ciperoensis*, *Cyclicargolithus abisectus*, *Chiasmolithus altus*, *Dictyococcites bisectus* and *Zygrhablithus bijugatus*. The nannofloras of the Corbi Formation are diversified (up to 30 taxa were recorded), moderately preserved and contain mainly cosmopolitan taxa such as *Dictyococcites bisectus*, *Cyclicargolithus abisectus*, *C. floridanus, Sphenolithus moriformis* and *Coccolithus pelagicus* (Wallich) Schiller. It is very important to note the increase in abundance of the genera *Sphenolithus* and *Discoaster*, which are taxa confined mostly to warm surface waters, within the samples collected from the lower part of the Corbi Formation. At the upper part of this unit, the FO of the nannofossil *Pontosphaera enormis* indicates the base of the NP25a Calcareous Nannofossil Subzone of Melinte (1995).

4.4. The Brăduleț Formation

Stratigraphy and sedimentology

The name of Brăduleț Formation is firstly introduced in this study to describe a succession composed of bituminous mudstones with rare and thin sandstones beds. This succession is exposed all over the study area and was measured at all five localities (Fig. 1).

Previously, this succession was described as the Upper Marls (Murgeanu, 1941), Pucioasa-type Marls (Popescu, 1954), Dysodilic Shales (Morariu and Teodorescu, 1987), Pucioasa Formation (Rusu, *et al.,* 1996) and the Pucioasa like Formation (Ryer, 1998).

In the Muiereasca Valley, the lower part of the Brăduleț Formation was separated as a distinct unit, named the Dysodilic Shale Formation (Rusu *et al.*, 1996).

Between the Olt and the Argeş valleys, a pile of rocks, predominantly muddy, lacking sandy and/or conglomeratic intercalations, was described as a distinct lithological unit - the Jiblea Marls – by Iorgulescu (1953). In fact, our data and the field observations indicate that the lower part of the Jiblea Marls represents the top of the Olăneşti Formation, while its upper part is the base of the Brăduleț Formation, having distinct lithological features (see above).

The stratotype of the new described Brădulеț Formation is situated in the Vâlsan Valley section, near the Brădulеț village. At the type locality, the Brădulеț Formation is 500 m thick. Its thickness varies in the study area, from a maximum of 600 m measured in Muiereasca Valley to a minimum of almost 300 m, recorded at Râul Doamnei Valley (Fig. 3).

The Brădulеț Formation overlies the Cheia Formation in the western part of the study area, and the Corbi Formation in the eastern part (Fig. 3). At both extremities, the lower boundary is gradational and characterized by a progressive vertical fining from the coarse Cheia and Corbi formations (Fig. 3).

As proved by the calcareous nannoplankton content, the Cheia and the Corbi formations are contemporaneous with the lower part of the Brădulеț Formation. Due to this lateral facies variation, the Brădulеț Formation directly overlain, in the Muiereasca Valley, the Olăneşti Formation. Notably, some of the thin facies of the Cheia Formation as well as those from the upper part of the Corbi Formation are brownish mudstones and calcareous clays, being similar with those of the Brădulеț Formation (cropping out in the Şiței and Râul Doamnei valleys).

In the central part of the study area, between the Olt and the Argeş rivers (Fig. 1), we suppose that the Brădulеț Formation paraconformity overlying the Olăneşti Formation, but the contact between the two units is not exposed in any outcrop. Probably this is the reason that determined Iorgulescu (1953) to separate a single lithological unit – the Jiblea Marls - in the above-mentioned area, used also by Bombiță *et al.* (1980) - Fig. 2.

At the top, the Brădulеț Formation is disconformity bounded by the Muiereasca Formation (also called the Muiereasca Sandstone, Popescu *et al.*, 1976), between the Olăneşti and Muiereasca valleys, conformity bounded by the Gura Văii Formation in the Cheia Valley and, respectively, by the Sărata Gypsum, between the Vâlsan and Râul Doamnei valleys. In the Vâlsan Valley, the Brădulеț Formation is disconformity overlain by the Mățău Conglomerates, which locally eroded the Sărata Gypsum (Figs. 1, 3, 4).

The rock-succession of the new described by us Brădulеț Formation is mudstone-dominated (80%). Sandstone and very rare conglomerate beds add to 15%, and respectively 5% of the remaining succession. The most obvious feature observed in outcrops is the black to dark brownish colour of the mudstones.

The dark colour of the mudstones resulted from a generally high content in organic matter that averages to 4-6% with a maximum of 27%. This suggests the development of an expanded oxygen minimum layer in the basin at the time of deposition and the presence of bottom water anoxia that facilitated the preservation of organic matter. Gypsum efflorescences and sulphur-like weathering are abundant within the mudstones and probably resulted from the oxidation of the authigenic pyrite. Fish scales, teeth and bones are also common and they are similar to those commonly preserved in the Lower Oligocene Dysodilic Shale Formation of the Eastern Carpathians. Moreover, in the Muiereasca Valley section several complete fish specimens of *Eomyctophum sp.* were recovered (Constantin, 2004, pers. comm).

The carbonate content of the laminated mudstones is relatively low between 0.5 to 10% $CaCO_3$. Some marly levels of the Brădulеț Formation, from the Olăneşti, Muiereasca, Vâlsan and Râul Doamnei valleys, yielded a content of the calcium carbonate between 55 to 75%. The intervals that are richer in organic matter usually form cm-thick couplets with the more marly mudstones.

On the Vâlsan Valley there are some cm-thick bentonites derived from the chemical weathering of volcanic ashes (Morariu and Teodorescu, 1987). Unfortunately, these beds have no a laterally continuity and therefore could not be used as lithological markers.

The sandstones consist of 1 cm up to 30 cm thick beds, yielding a parallel lamination/ stratification, current ripples, lenticular and flaser bedding. In addition, abundant soft-sediment deformational structures, such as load casting and balls-and-pillows are present along the lower bounding surfaces. The sandstones are rich in lithic fragments and locally may contain abundant carbonaceous material and wood fragments.

In the Şiței Creek section, some of the beds show particular lithification and bioturbation similar to a hardground cementation (Anastasiu, 2004, pers. comm.)

The coarse-grained beds consist of 40 cm thick granular to pebbly sandstones and are characterized by scouring and lenticular shape, most probably representing event beds.

Besides its diagnostic lithological features, above-mentioned, large lithological variations were recorded laterally (from west to east) within the Brădulеț Formation (Roban *et al.,* 2005a), particularly in its lower part.

In the Cheia Valley section, the boundary between the Brădulеț Formation and the underlying Cheia Formation is placed within a 15 m thick interval consisting primarily of a matrix supported conglomerates. These conglomerates display overwhelming evidence for mass transport such as soft-sediment deformation structures (slumps), chaotic grain size distribution and abundant mudstone intraclasts. Associated with the matrix-supported conglomerates and gravely mudstones there are lenticular-shaped sandstone beds, suggesting channel morphologies. The main sedimentary structures include massive, but also normal grading, and parallel stratification in conglomerates, as well as current ripples, normal grading and parallel lamination in sandstones, which locally passed to massive ones (Bombiță *et al.* 1980; Jipa 1980, 1982; Ryer, 1998; Roban *et. al.,* 2005a). Scouring is commonly found at the base of all coarse deposits. Additionally, soft-sediment structures such as load casting and liquefaction are present. In the Cheia, Olăneşti and Muiereasca sections, towards the middle part of the rock-sequence of the Brăduleț Formation there are some lenticular matrix-supported conglomerate and sandstone beds, up to 10 m thick, (Popescu *et al.*, 1976; Bombiță *et al.,* 1980).

The coarse-grained lower part of the Brăduleț Formation, recorded in the Cheia and Muiereasca sections, pinch out to the east, where the succession is characterized by laminated dark mudstones, locally containing several cm thick (up to 0.3 m) fine-grained sandstones. Thus, the conglomeratic packages are missing in the Râul Doamnei and Vâlsan valley sections (Fig. 3).

In the Muiereasca Valley, an approximately 20 m thick matrix-supported conglomerate of diamicton aspect represents the upper part of the Brăduleț Formation. The matrix consists of mudstone and represents about 70% of the volume of the rock. The clasts are extremely large blocks, reaching a maximum of 5 m in length (Popescu *et al.*, 1976). Within this coarse package, beds of coarse-grained sandstones and pebble-conglomerates form up to 2 m thick units (Jipa, 1994).

<u>Biostratigraphy and age</u>

As evidenced by the calcareous nannofossil investigations, the whole succession of the Brăduleț Formation covers a quite large interval – the Early Rupelian – Early Burdigalian (Ithe NP22-NN2 calcareous nannofossil zones respectively).

The calcareous nannoplankton assemblages from the Muiereasca Valley and Valey Şiței Creek (below the base of the Cheia Formation, and respectively that of the Corbi Formation) were assigned to the NP22-NP24 calcareous nannofossil zones (Fig. 3), covering the Rupelian stage *pro parte,* including the Rupelian/Chattian boundary interval. As significant nannofossil events, the successive FO of *Reticulofenestra lockeri*, *Transversopontis fibula*, *Reticulofenestra ornata*, *Cyclicargolithus abisectus* and *Sphenolithus ciperoensis* were observed.

It is worth mentioning that most of the previous studies (Murgoci, 1941b; Popescu *et al.*, 1976; Bombiță *et al.*, 1980) included the black to dark brown bituminous mudstones of the lower part of the Brăduleț Formation in the Olăneşti Marls (= the Lower Marls). Rusu *et al.* (1996) separated this sequence as a distinct lithological unit and named it the Lower Dysodilic Formation. The above-mentioned authors, based on lithological similarities as well as on the age proved by nannofloras, applied the lithological nomenclature, commonly used in the Paleogene Eastern Carpathian Flysch Zone, to designate formations of the Getic Depression. Rusu *et al.* (1996) describe as 'Lower Dysodiles' a sequence which both overlies the Olăneşti Marls (in Muiereasca Valley), and the Cheia Conglomerates (in the Cheia Valley). The age assigned by Rusu *et al.* (1996) for the Lower Dysodile Formation in the Getic Depression is Late Rupelian to Early Chattian (covered by the NP23 and lower part of NP24 calcareous nannoplankton zones).

As it was underlined above, the dark-brownish deposits (cropping out in the Muiereasca Valley, between the Olăneşti Formation - at the base, and the Cheia Formation - at the top), formerly assigned by Rusu *et al*. (1996) to the Lower Dysodiles, are included herein in the Brăduleț Formation.

Based on the identification of the nannofossil *Chiasmolithus altus*, in samples collected from the top of the Cheia Formation (just bellow the base of the Brăduleț Formation, in the Cheia Valley, Fig. 3), the NP24 Calcareous Nannoplankton Zone (Early Chattian in age) was recognized. Thus, we can assume that in the Cheia section, the base of the Brăduleț Formation is placed within the Rupelian/Chattian boundary interval.

At the Corbi locality (Râul Doamnei, Figs. 1, .3), the presence of the nannofossils *Pontosphaera enormis*, *Sphenolithus ciperoensis*, *Dictyococcites bisectus*, *Cyclicargolithus floridanus* and *C. abisectus*, identified towards the lower part of the Brăduleț Formation, just above the Corbi Formation, indicates a latest Oligocene age (Late Chattian - the NP25 Calcareous Nannofossil Zone).

Based on these evidences, we can state that the base of the Brăduleț Formation is dyachronous, being situated in a large interval – Rupelian – Late Chattian.

In the stratigraphic succession of the Brăduleț Formation from the Râul Doamnei section (the Corbi village, Figs.1, 3), the successive first occurrence of the *Triquetrorhabdulus carinatus* Martini, *Helicosphaera scissura* Müller, *Coccolithus miopelagicus* Bukry and *Reticulofenestra pseudoumbilicus* Gartner indicates the presence of the NN25, NN1 and NN2 calcareous nannoplankton zones of Martini (1971), Late Chattian-earliest Burdigalian in age (Melinte, 2005).

As in the Eastern Carpathian area (Melinte, 1993), the base of the Miocene (the NN1 Calcareous Nannoplankton Zone) is characterized by significant reworking. Thus, only 25% of the total nannofloras represent *in situ* assemblages, the remaining 75% being Cretaceous, Eocene and Oligocene reworked taxa.

5. PALAEOBIOGEOGRAPHY

Both the distribution pattern and fluctuation in abundance of the Late Eocene to Early Miocene nannofloras (the NP21-NN2 Calcareous Nannoplankton Zones) recorded in the NW area of the Getic Depression reflect the environmental

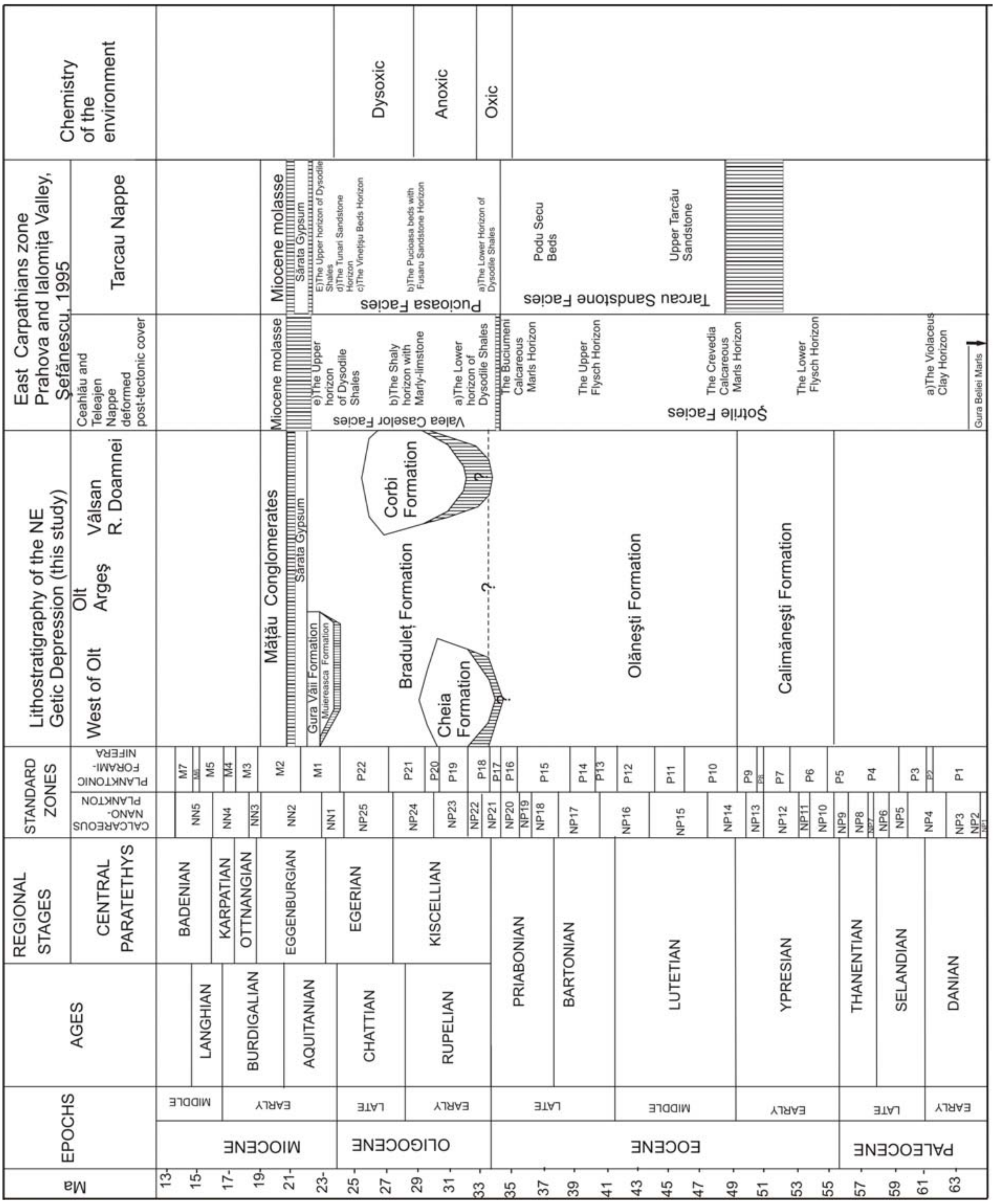

Fig. 4. Comparison between the stratigraphy of the NE part of the Getic Depression, identified in this study and the stratigraphy of Carpathian bend area (Ştefănescu, 1995)

changes, both climatic and oceanographic, that characterized the end of the Paleogene period.

The Late Eocene (Priabonian) nannofloras of the Olăneşti Formation are dominated by the *Sphenolithus* and *Discoaster* genera, which are mostly confined to the warm and well-oxygenated surface waters of the Tethys Realm. Only cosmopolitan nannofossils are present in the Upper Eocene deposits of the Getic Depression and this is a common feature for the Late Eocene nannofloras throughout the world. An open-marine environment, largely communicating with the Tethys Realm, characterized the area where the mudstones of the Olăneşti Formation accumulated.

It is widely accepted that the beginning of the Oligocene is a time of palaeoclimatic cooling (Aubry, 1992, Zachos *et al.*, 2001), that must have caused significant paleoenvironmental and biotic changes in the open seas. These changes are mirrored in the Early Oligocene nanofloras of the Getic Depression. A sharp decline in abundance of the warm water taxa *Discoaster* and *Sphenolithus* was recorded within the top of the Olăneşti Formation, followed by the disappearance of the discoasterids. The genus *Sphenolithus* is represented only by the long-ranging and diagenetical-resistant species *S. moriformis*. Endemic taxa of the Paratethys Realm, such as *Transversopontis lata*, *T. fibula*, *Dictyococcites ornatus* and *Reticulofenestra lockeri*, occur in the investigated Lower Oligocene deposits. Moreover, monospecific assemblages containing *Braarudosphaera bigelowii* (Gran & Braarud) Deflandre, a species characteristic for brackish water environments, were observed from samples of Early Oligocene age, collected from the succession exposed in the Getic Depression.

These data argue for a decrease in salinity due to the isolation of the studied area (that was part of to the Central Paratethys) from the Tethys Realm. A progressive cooling, together with a low nutrient supply could also be assumed. The occurrence of the endemic nannofossils, together with *Braarudosphaera bigelowii*, may represent the first signal of the instauration of the Early Oligocene anoxia in the area of the Getic Depresssion.

The disappearance of endemic taxa, the re-occurrence of *Sphenolithus* and *Discoaster* and, the dominance of cosmopolitan nannofossils in the Late Oligocene to Early Miocene assemblages suggest a restoration of the communication between the study area and the Tethys Realm. The increase in nannofloras diversity and abundance express environmental changes, like increased in salinity and evaporation, due to higher temperatures at the surface of the sea and, increased nutrient supply and planktonic productivity of the surface waters.

6. DISCUSSION AND CONCLUSION

A complex depositional history characterized the NE part of the Getic Depression, during the Late Paleogene. The end of the Eocene (Late Priabonian up to the Eocene/Oligocene boundary interval) is characterized by the deposition of the marine mudstones of the Olăneşti Formation over the entire study area. Beginning with the Oligocene (the Rupelian stage), organic-rich mudstones and marls of Brăduleț Formation accumulated in the NE part of the Getic Depression (Fig. 4).

Our study documented that the lower part of the Brăduleț Formation, sharing some similar lithological features with the dysodilic shales of the Eastern Carpathians, is partly synchronous with the Cheia Formation and with the lower part of the Corbi Sandstone. We can assume that the deposition of this lithostratigraphic unit started over the whole Getic Depression area in the Early Oligocene (NP22 Calcareous Nannofossil Zone), shortly after the first isolation of the Paratethys from the Tethys Realm. To the west, the lower part of Brăduleț Formation is missing due to erosion (Fig. 3). This fact is indicated by the unconformity recorded between the conglomerates from the base of Cheia Formation, and the calcareous clays of the Olăneşti Formation and, also by the absence of NP21 (upper part) and probably lower part of NP22 calcareous nannoplankton zones west of Muiereasca Valley (Fig.3).

The anoxic organic rich mudstones (dysodilic shales), deposited during the Early Oligocene, are present within the whole Carpathian belt, probably reflecting highstand conditions for that particular interval over a large area of the Southern, Eastern and Western Carpathians. However, their absence west of the Muiereasca Valley in the Getic Depression is best explained when considering the unconformity at the base of the Cheia Formation.

Despite different tectonic evolution of the Carpathian and Getic domains (Săndulescu, 1984; Ştefănescu, 1995; Mațenco, 1997), both regions are characterized by deposition of organic-rich mudstone-dominated episodes in the Early and Late Oligocene. However, there are other significant lithological differences between the two areas: in the Getic Depression the Oligocene successions are dominated by conglomerates and sandstones, the former mainly missing in the Carpathians. Taking into account these observations, we do not agree to extend the lithostratigraphic terminology from the Eastern Carpathians into the area of Getic Depression. For instance, the Upper Oligocene organic-rich mudstones, interbedded with thin sandstones, of the Getic Depression area have been previously described as Pucioasa-type Marls. Our study proves that this succession termed here the Brăduleț Formation extends in the entire interval occupied by the Pucioasa, Vinețişu and the Upper Dysodilic Shale formations (which is Chattian - Early Burdigalian in age, according to Melinte 1993) in the inner part of the Tarcău Nappe of the East Carpathians. The so-called dysodilic shales of the Getic Depresssion, named here the lower part of the Brăduleț Formation, differ also in age from the Lower Dysodilic Shales of the East Carpathians. The former spans over the Early Oligocene, while the latter was deposited in a shorter interval, within the late Early Oligocene (Melinte, 1993). The basal Oligocene (the Early Rupelian) is characterized in the East Carpathians by other two lithological units (the Bituminous Marl Formation and the Menilitic Formation) which were not recognized so far in the Getic Depression.

Based on nannofloral changes in diversity, abundance and distribution, we assume that an open-marine environment, with warm and well-oxygenated surface waters characterizes the Late Eocene, including the Eocene - Oligocene boundary interval. A restrictive oceanic circulation pattern, leading to the establishment of an anoxic regime, and cooler surface waters, could be assumed for the Early Rupelian to early Late Rupelian interval. A gradual transition from an anoxic to a hypoxic environment, together with a progressive warming, developed in the Late Oligocene (Chattian) to Early Miocene (Burdigalian) interval. The pattern of nannofloral distribution and fluctuations, recorded from the Paleogene sediments of the Getic Depression, reflects the global climatic deterioration from Eocene to Oligocene times.

We consider that this study is only the beginning of a systematic stratigraphic investigation of the Getic Depression. Further high-resolution lithological and biostratigraphical works (including also other groups of organisms - i.e. dinoflagellates) are to be carry out, in order to date more accurately the Paleogene lithostratigraphic units, cropping out in the whole region of the Getic Depression.

ACKNOWLEDGMENTS

Part of this research was supported by the Grant No.14/2004 of the Petrom National Oil Company. Relu-Dumitru Roban acknowledges International Association of Sedimentologists for a Postgraduate Grant. We are indebted to Prof. Dr. Nicolae Anastasiu, for discussions and comments on an earlier version of the manuscript. We thank to Dr. Dan Jipa, for interesting remarks and comments on earlier version of this paper, as well as for reviewing our work. Dr. Carmen Chira a kindly reviewed this manuscript. Dr. Bogdan Vârban helped us with discussions and suggestions on the stratigraphy and sedimentology, also improving the English version of this manuscript. The authors thank for field assistance to: Ovidiu Barbu, Ana Maria Muşat, Liviu Popa and Florin Stoican.

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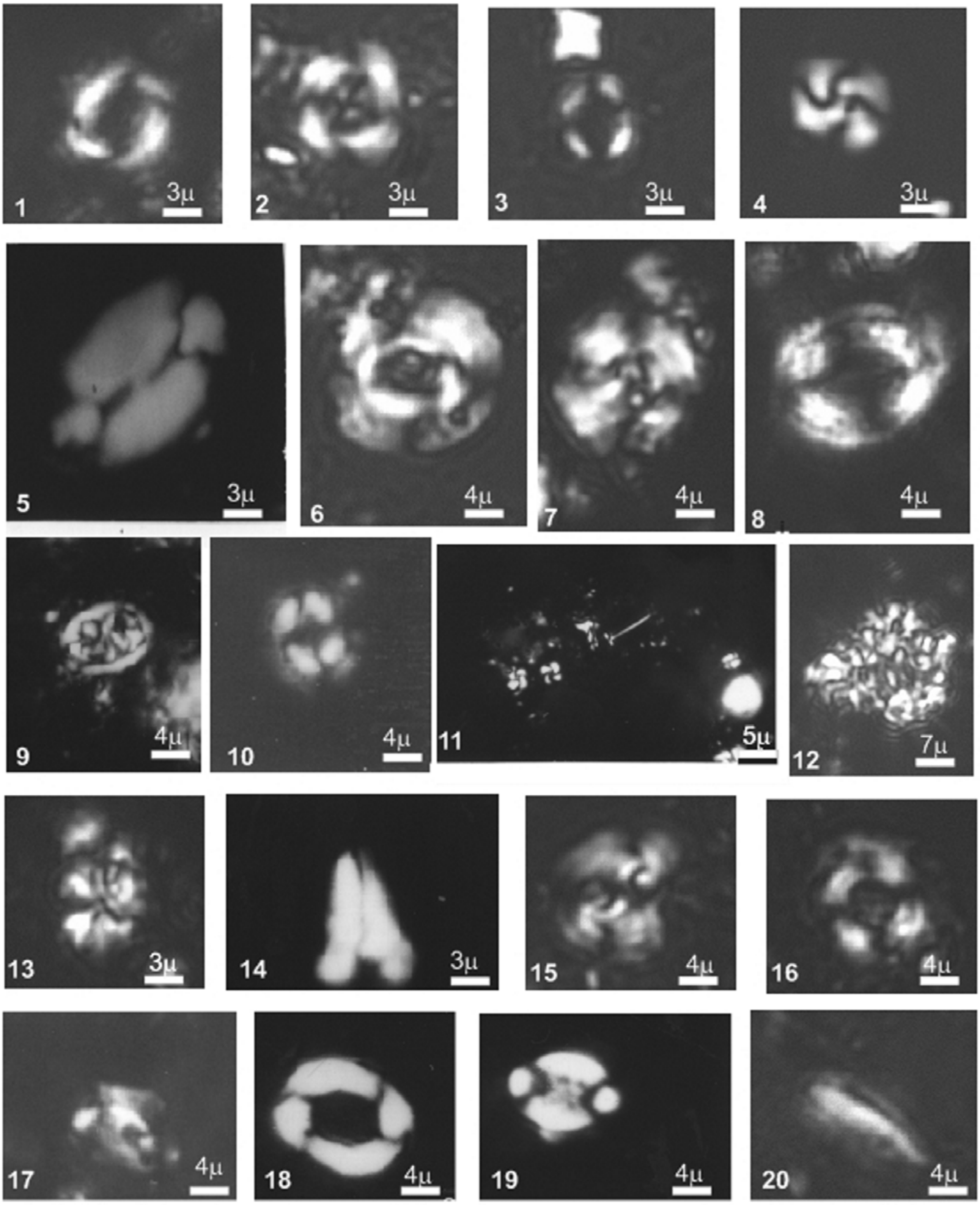
PLATE 1
LM, all microphotographs N+

1. *Reticulofenestra pseudoumbilicus* (Gartner), Early Miocene (Burdigalian), NN2 Nannofossil Zone, Râul Doamnei, Sample 275.
2. *Cribrocentrum* r*eticulatum* (Gartner & Smith) Roth & Thierstein, Eocene/Oligocene boundary interval, NP21 Nannofossil Zone, Vâlsan Valley, Sample 190.
3. *Coronocyclus nitescens* (Kamptner) Bramlette & Wilcoxon, Early Miocene (Aquitanian), NN1 Nannofossil Zone, Râul Doamnei, Sample 264
4. *Reticulofenestra scrippsae* (Bukry & Percival) Roth, Early Oligocene (Rupelian), NP22 Nannofossil Zone, Muiereasca Valley, Sample 380.
5. *Helicosphaera kamptneri* (Hay & Mohler *in* Hay *et al.*) Locker, Early Miocene (Aquitanian), NN1 Nannofossil Zone, Râul Doamnei, Sample 257.
6. *Reticulofenestra hillae* Bukry & Percival, Early Oligocene (Rupelian), NP22 Nannofossil Zone, Olăneşti Valley, sample 375.
7. *Cyclicargolithus abisectus* (Müller) Bukry, Late Oligocene (Chattian), NP24 Nannofossil Zone, Cheia Valley, sample 347.
8. *Reticulofenestra umbilica* (Levin) Martini & Ritzkowski, Eocene/Oligocene boundary interval, NP21 Nannofossil Zone, Vâlsan Valley, Sample 190.
9. *Chiasmolithus altus* Bukry & Percival, 1971, Late Oligocene (Chattian), NP24 Nannofossil Zone, Cheia Valley, sample 348.
10. *Coccolithus miopelagicus* Bukry, 1971, emend. Wise, 1973, Early Miocene (Aquitanian), NN1 Nannofossil Zone, Râul Doamnei, Sample 257.
11. Field of view with *Rhabdosphaera clavigera* Murray & Blackman - central part - and small reticulofenestrids – left down, Early Miocene (Burdigalian), NN2 Nannofossil Zone, Râul Doamnei, Sample 275.
12. *Thoracosphaera saxea* Stradner, Early Oligocene (Rupelian), NP22 Nannofossil Zone, Muiereasca Valley, Sample 380.
13. *Sphenolithus moriformis* (Brönnimann and Stradner) Bramlette and Wilcoxon, Late Oligocene (Chattian), NP24 Nannofossil Zone, Cheia Valley, sample 347.
14. *Zygrhablithus bijugatus* (Deflandre *in* Deflandre & Fert) Deflandre, Early Oligocene (Rupelian), NP22 Nannofossil Zone, Olăneşti Valley, sample 375.
15. *Dictyococcites bisectus* Hay, Mohler & Wade, Eocene/Oligocene boundary interval, NP21 Nannofossil Zone, Vâlsan Valley, Sample 190.
16. *Reticulofenestra ornata* Müller, Early Oligocene, NP23 Nannofossil Zone, Olăneşti Valley, sample 332.
17. *Helicosphaera compacta* Bramlette & Wilcoxon, Eocene/Oligocene boundary interval, NP21 Nannofossil Zone, Vâlsan Valley, Sample 190.
18. *Pontosphaera latelliptica* Báldi-Beke & Baldi) Perch-Nielsen, Eocene/Oligocene boundary interval, NP21 Nannofossil Zone, Vâlsan Valley, Sample 190.
19. *Pontosphaera enormis* (Locker) Perch Nielsen, Late Oligocene, NP25 Nannofossil Zone, Râul Doamnei, Sample 317.
20. *Triquetrorhabdulus carinatus* Martini, Late Oligocene, NP25 Nannofossil Zone, Râul Doamnei, Sample 25.

PLATE 1



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