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The **Romanian Journal of Stratigraphy** (Rom. J. Stratigraphy) is now at its first volume in the new form. However, the publication goes back to 1910, as the first volume of the "Dări de seamă ale Şedinţelor" (D.S.) has appeared as proceedings of geologists working with the Geological Institute of Romania. The journal (D.S.) appeared initially as a single volume (till volume 54, 1969), then with five series, the present issue being a direct continuation of the D.S./series 4 (Stratigraphy).

The editor has changed the name as follows: Institutul Geologic al României (vol. I–XXXVI, 1910–1952), Comitetul Geologic (vol. XXXVII–LII/1, 1953–1966), Comitetul de Stat al Geologiei (vol. LII/2–LV/1, 1967–1969), Institutul Geologic (vol. LV/2–LX, 1970–1974), Institutul de Geologie şi Geofizică (vol. LXI–74, 1975–1990).

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Rom. J. Stratigraphy, 1993, 75, p. 1–4

DEVONIAN SPORES IN THE RUSAIA TECTONIC WINDOW (EAST CARPATHIANS): THEIR GEOLOGICAL SIGNIFICANCE

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Key words: Devonian. Windows. Stratigraphic units. Metamorphic rocks. Spores. East Carpathians – Crystalline–Mesozoic Zone – Bistriţa Mountains.

Abstract: Middle Devonian spores have been identified in the top of the low-grade metamorphic Paleozoic of the Rusaia tectonic window. Thus, it has been proved that this part of the sequence – the Căminet Formation – is a stratigraphic equivalent of the lower part of the Devonian-Lower Carboniferous Cimpoiasa Group and therefore it does not belong to the Silurian Rusaia Group. Differences in the facial development of the Middle Devonian sedimentation in the Rusaia tectonic window and in the Rodna Mts suggest that the Paleozoic sequences of the mentioned units belong to different Infrabucovinian nappes or to tectonic complications of the same main nappe.

In the Rusaia tectonic window, west of Cîrlibaba, an Infrabucovinian tectonic unit is exposed, represented by medium-grade Precambrian metamorphics of the Bretila Group, covered by the Silurian sequence of the Rusaia Group with Variscan low-grade metamorphism (Bercia et al., 1976; Iliescu, Kräutner, 1978).

In the Paleozoic cover three main lithostratigraphic units have been identified (Fig. 1):

– *Pîrîul Omului Formation* (Ru^{PO}) (50–100 m), in a lower position, constituted of metaconglomerates, quartzites, limestones, calcschists, sericite-chlorite schists, graphite schists with detrital muscovite, chloritoid schists.

– *Rotunda Formation* (Ru^{Ro}) (700–800 m), represented by a monotonous sequence of sericite-chlorite schists in which petrographic changes are due mainly to variable amounts of quartz, albite and the local presence of chloritoid as well as of a discontinuous metaconglomerate level.

– *Căminet Formation* ($Ci^{Că}$) constituted of a dolomitic horizon (Căminet Dolomite) in the lower part and an alternation of carbonatic sericite-quartz schists with sericite-chlorite schists and graphite quartzitic schists in the upper part.

This Paleozoic low-grade metamorphic sequence transgressively covered the Precambrian basement (Kräutner, 1972). It was cut in the upper part by the Alpine Subbucovinian overthrust. The overlying Subbucovinian Nappe is represented by the medium-grade metamorphics of the Precambrian Rebra Group. In the Variscan event the Paleozoic cover was locally separated from its Precambrian basement and slipped (or moved) over the retrogressive gneisses of the Bretila Group, as it is suggested by bands of lamination and phyllonitization (Kräutner, 1972).

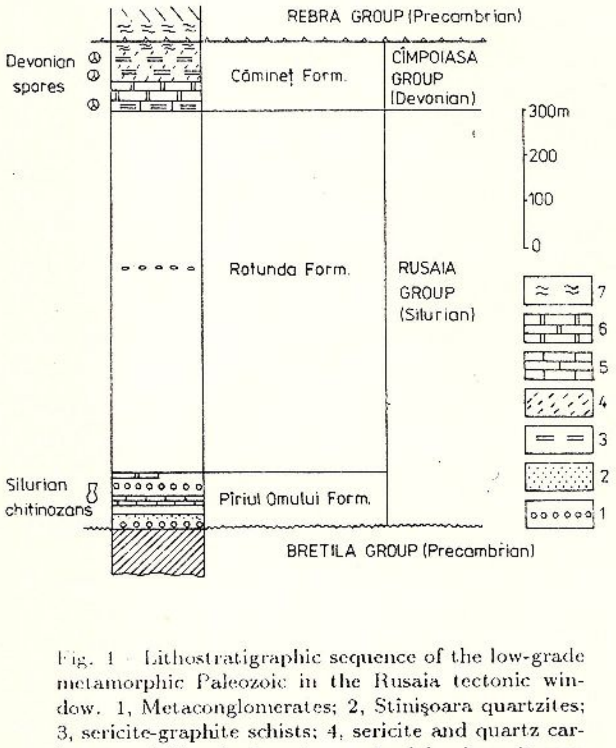

Fig. 1 – Lithostratigraphic sequence of the low-grade metamorphic Paleozoic in the Rusaia tectonic window. 1, Metaconglomerates; 2, Stinişoara quartzites; 3, sericite-graphite schists; 4, sericite and quartz carbonate schists; 5, limestones; 6, dolomites; 7, mylonites.

From the lower part of the Paleozoic sequence Silurian Chitinozoans have been reported (Iliescu, Kräutner, 1978). That is why usually the whole sequence was assigned to the Silurian, although for the upper part (Căminet Formation) an equivalence with the Devonian Cimpoiasa Group was envisaged (Iliescu, Kräutner, 1978). Thus, a new palynological research started for the Căminet Formation, whose results are reported in the present contribution.

The sampling was performed from the good and nearly continuous exposures on the road in the Bistrita Valley, upstream the Căminet Valley. The detailed sequence and the lithostratigraphic position of the samples are shown in Figure 2.

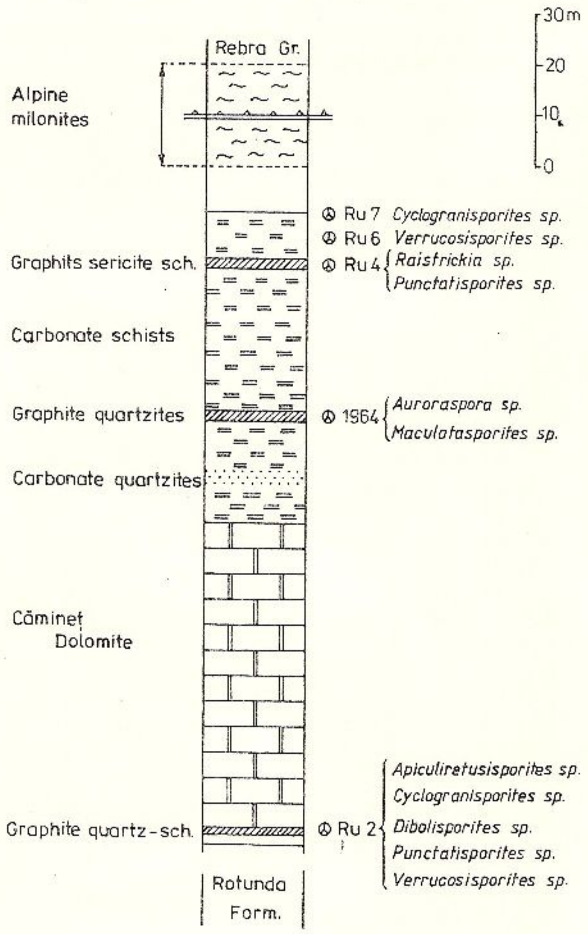

Fig. 2 – Position of the spore assemblages in the lithostratigraphic sequence of the Căminet Formation.

Devonian spores have been identified in the graphitic intercalations in the lower and middle part of the sequence as well as in the carbonatic schist in the top (Fig. 2), in the following samples: Ru_2 banded grey limestone, 1964 quartzitic graphite schist, Ru_4 sericite graphite schist (with relict detrital muscovite), Ru_6 quartzitic carbonate-sericite schist, Ru_7 carbonate sericite schist. These samples represent only about 25 per cent of the analysed material that was mostly waste. The identified palynological assemblage is scarce and badly preserved which allows only genera determination (Tiwari, Schaarschmidt, 1975; Becker et al., 1974; Jersey, 1966): *Apiculiretusispora* sp. (Ru_2), *Cyclogranisporites* sp. (Ru_2, Ru_7), *Auroraspora* sp. (1964), *Raistrickia* sp. (Ru_4), *Maculatasporites* sp. (1964), *Dibolisporites* sp. (Ru_2), *Punctatisporites* sp. (Ru_2, Ru_4), *Verrucosisporites* sp. (Ru_2, Ru_6).

According to the stratigraphic distribution of the specimens (Fig. 3) the mentioned assemblage comprises spores with evolution (1) from the Lower Devonian to the Upper Carboniferous (5 specimens), (2) from the Middle Devonian to the Upper Carboniferous (2 specimens) and (3) only during the Middle Devonian (*Maculatasporites* sp.).

Spores whose evolution starts in the Middle Devonian have been identified both in the lower part (*Punctatisporites* sp.) and in the upper part (*Punctatisporites* sp., *Verrucosisporites* sp.) of the sequence, while the Middle Devonian specimen (*Maculatasporites* sp.) is placed in the middle part of the sequence (Fig. 2). Therefore the whole sequence of the Căminet Formation may be assigned to the Middle Devonian.

Geological significance of the Devonian spores

The new palynological data confirm the stratigraphic equivalence of the Căminet Formation of the Rusaia Window with the lower part of the Cimpoiasa Group (Gura Fîntînii Formation) of the Rodna Mts, in which (Lower-Middle) Devonian spores have been identified (Iliescu, Kräutner, 1975). It allows us to recognize some differences of the facial development in the mentioned tectonic units.

Figure 4 shows the lithologic sequences of the Căminet Formation and its equivalents in the Infrabucovinian units of the Rodna Mts in their palynspastic position. A change may be pointed out from the mainly carbonatic sedimentation of the Rusaia Window to a more detrital and finally pelitic sedimentation in the west.

Facial differences between the Paleozoic sequences of most of the Infrabucovinian tectonic units are known also for the Silurian and Lower Carboniferous. Thus, it was presumed that the present picture is due to the nappe transport from different sedimentation areas. In the Valea Vinului Nappe of the Rodna Mts and in the Rusaia tectonic window such differences were not very obvious because of the rather homogeneous development of the Rusaia-type Silurian. Some peculiar development is now evident for both the upper part (Căminet Formation and Gura Fîntînii Formation) and the lower part (Stînişoara quartzite). This

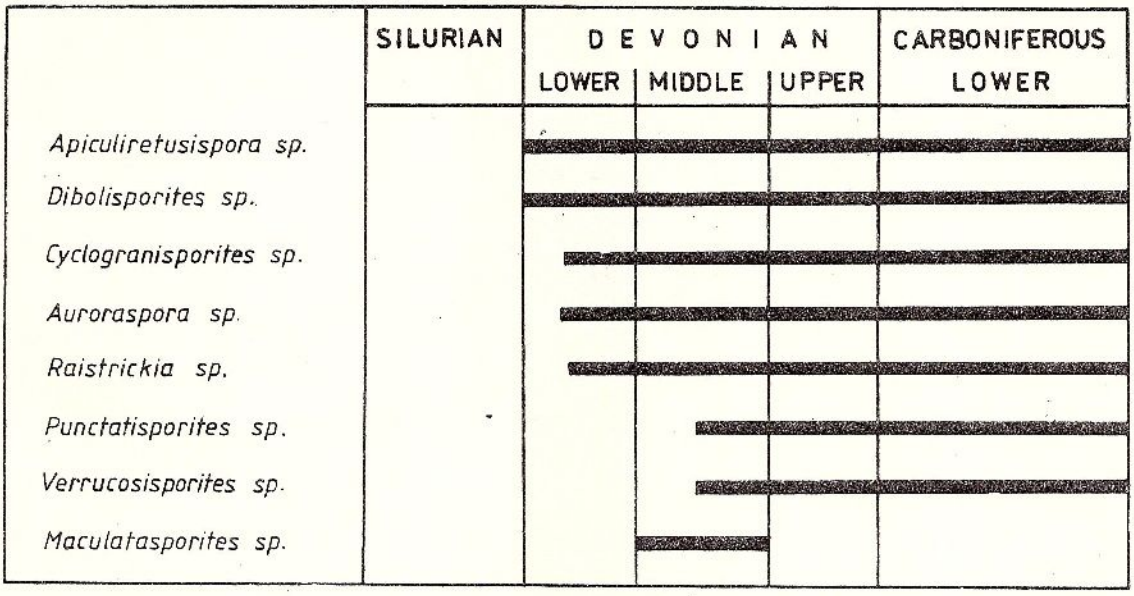

Fig. 3 – Stratigraphic distribution of the spores identified in the Cămineţ Formation (Bistriţa Valley).

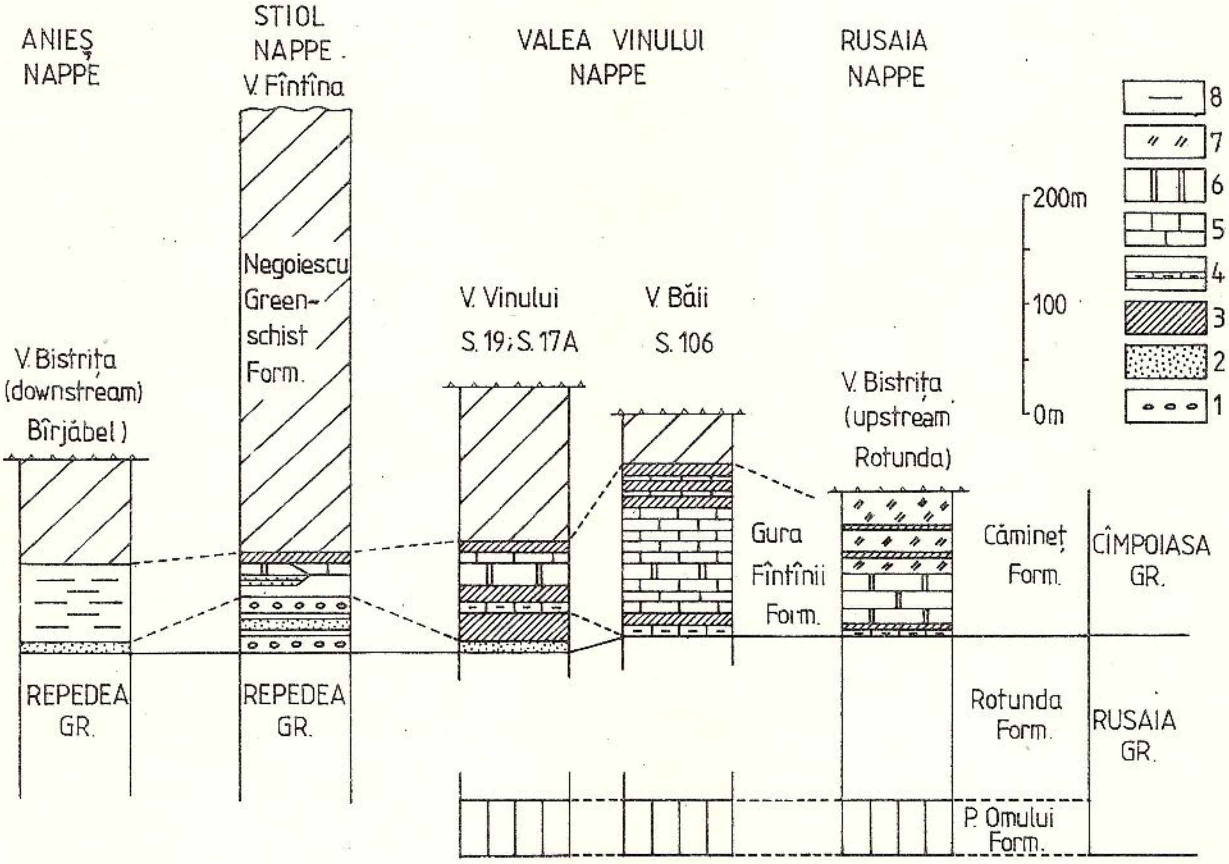

Fig. 4 – Different facial developments in the lower part of the Cimpoiasa Group in the Rusaia tectonic window (Cămineţ Formation) and in the Rodna Mts (Gura Fîntînii Formation).
1, metaconglomerates; 2, quartzites; 3, sericite schists ± graphite; 4, graphite limestones; 5, grey-white banded limestones; 6, dolomites; 7, sericite and quartz carbonate schists; 8, phyllites.

suggests that overthrust relationships may be considered also between the Valea Vinului and Rusaia units, as it was inferred by Kräutner (1972).

Figure 5 gives a model for the relationships between the Infrabucovinian units of the Rodna Mts and Bistriţa Mts. All these units are successively overridden by the Subbucovinian Nappe. The significance and importance of each of the represented overthrusts are not clear so far. It might be possible that all of them should be nappe structures, but it is obvious that not all facial differences between the Paleozoic sequences are of the same importance. For example,

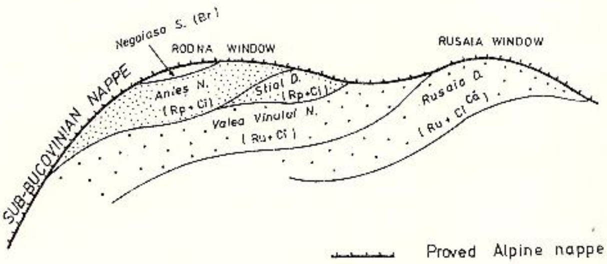

Fig. 5 – Schematic representation of the relationships between the Infrabucovinian Units in the Bistriţa Mts and Rodna Mts.
(Abreviations: Rp – Repedea; Ci – Cimpoiasa; $Ci^{Că}$ – Căminet Formation; Ru – Rusaia; Ar–St – Argestru–Steghioara; Br – Bretila; N – nappe; D – digitation; S – tectonic slide).

closer facial developments are in the Anieş and Stiol units or in the Valea Vinului and Rusaia units than between these two couples or other units in a lower position. Therefore it may be supposed that a part of the units represents complications of nappes. This means that instead of the five nappes represented in Figure 4 there could be only two main nappes and a set of tectonic slides (Negoiasa Unit) below the Subbucovinian Nappe: 1. Anieş Nappe with the Stiol Digitation; 2. Valea Vinului Nappe with the Rusaia Digitation. The Iacobeni Nappe and the Şarul Dornei Nappe, exposed more to the south, have a different Paleozoic cover represented by the Argestru–Steghioara sequence (Silur–Devon ?) and the Neagra Şarului Formation (very low-grade metamorphic Upper Paleozoic, Kräutner, 1978).

From all the mentioned Infrabucovinian units, it is only for the Iacobeni Nappe that there are proofs on its Alpine age: the molassoid Upper Paleozoic of the Neagra Şarului Formation in the subjacent Şarul Dornei Nappe.

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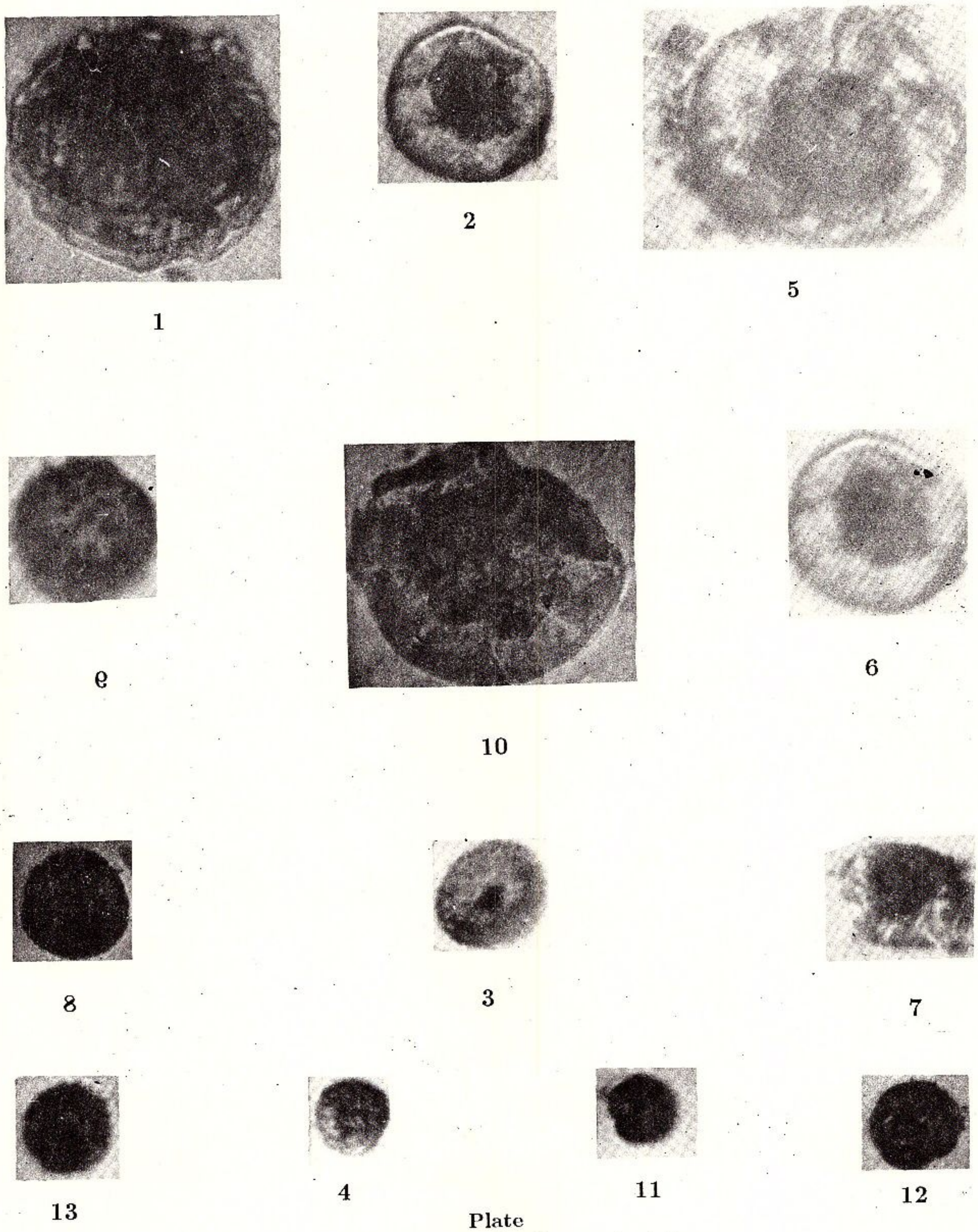

Plate

Devonian spores in the Rusaia tectonic window

Fig. 1 – *Maculatasporites* sp.
Figs. 2, 5, 6, 7, 10 – *Auroraspora* sp.
Fig. 3 – *Cyclogranisporites* sp.
Fig. 4 – *Dibolisporites* sp.
Fig. 8 – *Apiculiretusispora* sp.
Fig. 9 – *Raistrickia* sp.
Fig. 13 – *Verrucosisporites* sp.

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BIOSTRATIGRAPHIC DATA ON THE UPPER CARBONIFEROUS DEPOSITS IN THE NORTH–WESTERN PART OF THE REŞIŢA ZONE

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Key words: Upper Carboniferous. Westphalian. Stephanian. Paleoflora. Taphonomy. Biostratigraphy. South Carpathians. Sedimentary deposits. Getic and Supragetic Domains. Reşiţa Zone.

Abstract: The paper presents a few paleofloristic assemblages collected from Upper Carboniferous deposits that, according to a previous lithological classification, belong to the Lupacu Bătrîn Beds and the Lupac ones. The Lupacu Bătrîn Beds contain in their base a *Neuropteridae* – bearing assemblage and in their middle part an association including the first *Pecopteridae*. The Westphalian D – Stephanian A + B age of the Lupacu Bătrîn Beds has been assigned based on the two paleofloristic associations. The Lupac Beds contain a rich paleoflora and an *Anthracosidae* assemblage, both of them characteristic of the Stephanian C.

Introduction

The investigation of the coals in the Upper Carboniferous deposits of the Reşiţa Zone (western part of the South Carpathians) asks for detail biological and lithostratigraphical studies of these deposits. In a region with very intricate tectonics, as the one we refer to, it is particularly important to elaborate as precise as possible and easily identifiable litho-biostratigraphic markers.

In this respect, the present paper proposes a biostratigraphic model of the Upper Carboniferous deposits, based on the paleoflora collected from well defined lithological horizons. That is why paleoflora elements have been used, collected from outcrops or mining works offering very good conditions of observation, the possibility existing for the whole paleoflora known so far, including that from boreholes, to be accounted for by this model.

History of Research

The area under investigation has been of special interest since the end of last century, when the first papers were published concerning the Paleozoic in the Reşiţa Zone.

Halaváts (1893), based on paleoflora elements found in the Paleozoic deposits of the Lupac region, assigned these deposits to the Upper Carboniferous.

Mateescu (1962) studied for the first time the petrography of the upper coals in the Lupac region and drew up a list of the paleoflora associated to the coal beds.

Biţoianu (1973) elaborated a synthesis of the paleoflora of the Upper Carboniferous deposits in Romania, on which occasion she also presented a rich paleoflora association collected from the Reşiţa Zone.

Năstăseanu (1978) presented a new stratigraphic and tectonic image of the Lupac region, establishing the existence of three Upper Carboniferous complexes: Doman Beds, Lupacu Bătrîn Beds and Lupac Beds. Subsequently it has become possible to collect paleoflora from well established lithological horizons of this area.

Biţoianu (1979, unpubl. report) collected a rich paleoflora from the boreholes drilled in the region. Based on it, the author has assigned the Lower Stephanian age (A+B) to the Lupacu Bătrîn Beds and the Lower Stephanian age (C) to the Lupac Beds.

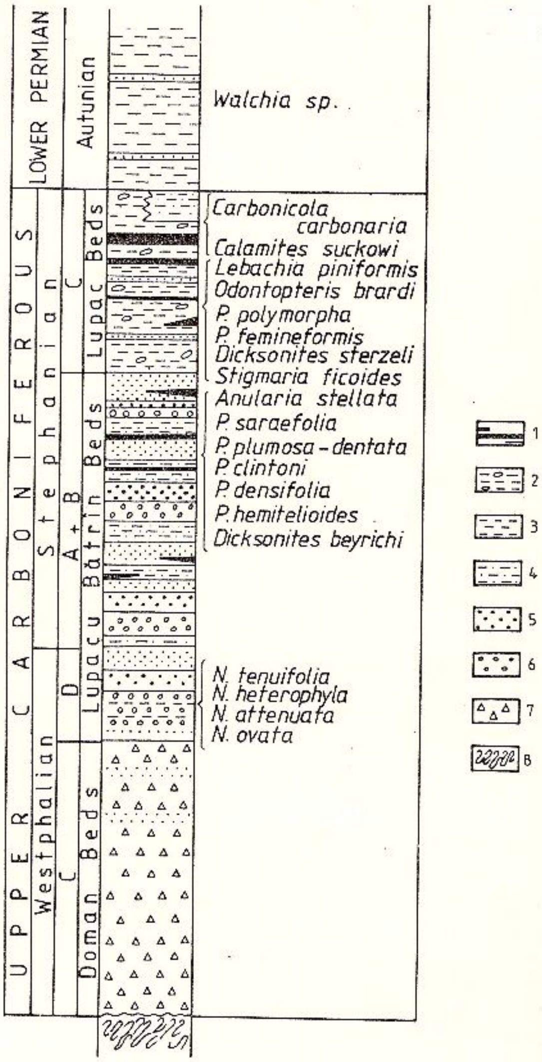
Năstăseanu (1984), making a synthesis of all the data referring to the coal deposits in Romania, showed that the Lupacu Bătrîn Beds contain fossil flora, indicating the presence of the Westphalian D and the Stephanian A+B, and the paleoflora of the Lupac Beds, confirming their Stephanian C+D age.

Negrea (1987) completed with new data the study of the paleoflora of the Lupac Beds, confirming their

Stephanian C+D age.

Stratigraphic Data

In the north-western part of the Reşiţa Zone (sensu Codarcea, 1940), the deposits of the Variscan molasse have been completely developed. From bottom to top, the following sequence can be identified: Doman Beds, Lupacu Bătrîn Beds, Lupac Beds, horizon of Walachia-bearing clays and red gritty-conglomeratic horizon (Năstăseanu, 1978). The first three lithological entities have been assigned to the Upper Carboniferous (Westphalian-Stephanian), and the remaining two to the Lower Permian (Autunian). The whole sedimentary cover is 2000–2500 m thick and unconformably overlies a metamorphic basement.



Upper Carboniferous deposits in the north-western part of the Reşiţa Zone. 1, coals; 2, black clays with concretions; 3, clays and argillites; 4, gritty clays; 5, microconglomerates and quartz-feldspathic sandstones; 6, conglomerates; 7, massive breccias; 8, crystalline schists.

The sequence of Variscan molasse deposits starts with a breccia-conglomerate deposit, 300 m thick, made up of coarse breccia, progressively passing to coarse conglomerates, with elements of crystalline schists within a detrital matrix. Thus, the Doman Beds look like a uniform lithological complex, massive in aspect, in which no paleontological elements are known. Based on the spatial relations with the upper complexes, rich in paleoflora remains, the Doman Beds have been assigned to the Westphalian C.

The sequence is continued by a progressively gritty conglomeratic complex, the Lupacu Bătrîn Beds. They are of a great lithological variety and are characterized by the presence of sedimentation cycles, beginning with coarse quartzose conglomerates, followed by microconglomerates, quartz-feldspath sandstones and black argillites with coal interbeds. This complex is 200-400 m thick and its age, based on the paleoflora it contains, was considered to be Westphalian D – Stephanian A+B (Năstăseanu et al., 1977; unpubl.report).

The final part of the succession is made up of a prevailingly clayey-silty complex, 150-200 m thick, the Lupac Beds. The passage to this complex is made progressively, by less and less coarse detrital components, so that the Lupac Beds are made up of argillites and finely bedded micaceous black siltstones, with interbeds of graywackes, coaly clays and bituminous coal beds.

A characteristic feature of the Lupac Beds is the presence of slightly ferruginous concretions, varying in size, but no more than 20–25 cm in diameter, with non-uniform distribution.

The paleofloristic assemblage of these deposits has made it possible to establish its Stephanian C+D age (Năstăseanu, 1978; Negrea, 1987).

The Upper Carboniferous deposits conformably support Autunian deposits (Antonescu, Năstăseanu, 1976).

Taphonomy

The Lupacu Bătrîn Beds paleoflora consists of imprints either in black clayey shales or in medium-grained quartzo-feldspathic sandstones that have not accurately preserved all the details of the existing flora. In both cases, the paleoflora has been collected from boreholes, out of which only samples of small sizes could be obtained.

The Lupac Beds paleoflora consists of imprints or, rarelier, of medullary moulds. The mining works in the region have supplied a rich paleobotanic material, often of big sizes and very well preserved. A peculiarity of the paleoflora in these deposits, noticed in fact equally by other researchers in various coal basins (Delvolvé, Laveine, 1985), is its intimate association with the ferruginous concretions abounding in this complex.

Paleoflora

The paleoflora has been collected from the north-western area of the Reşiţa Zone, from the Lupac and Bîrzava Valleys Basins respectively. As mentioned above, the samples we refer to have been found in the Lupacu Bătrîn Beds and in the Lupac Beds deposits.

Lupacu Bătrîn Beds Paleoflora

We point out that the macroflora assemblage from Valea Ariei belong to the Lupacu Bătrîn Beds. From the clayey-silty sequences of these beds, Biţoianu (1973) mentions the following species: *Neuropteris gigantea* STBG., *N. heterophylla* BRGT., *N. ovata* HOFFM. *flexuosa* STBG. form, *N. ovata* HOFFM., *tipica* CROOKAL., *N. tenuifolia* SCHLOTH., *Linopteris neuropteroides* (GUTB), *Validopteris* cf. *integra* GOTHAN. It is that assemblage that proves the existence of the basal part of the Westphalian D. We consider this paleofloristic list (Pl. I), to which the species *Neuropteris attenuata* LIND & HUT is added, to be the most suggestive and complete as far as the age of the deposits in Valea Ariei is concerned. Taking into account the presence of the species *Neuropteris ovata* HOFF. (Laveine, 1974) and *N. attenuata* LIND & HUT, we consider the Westphalian D age assigned to these deposits fully justified.

Detail field studies have permitted the identification of a new fossil site, on a tributary of the Bîrzava Valley, nearby the last house in the proximity of the TV relay, in quartz-feldspathic sandstones, corresponding to the middle part of the Lupacu Bătrîn Beds. This outcrop has yielded the following species: *Annularia stellata* SCHLOTH., *Pecopteris clintoni* LESQ., *P. densifolia* GOEPP., *P. hemitelioides* BRONG., *P.* cf. *miltoni* ART., *P. plumosa-dentata* BRONG., *P. saraefolia* BERT., *Dicksonites beyrichi* WEISS. (Pl. II).

The relative monotony of this assemblage and the extremely great frequence of the species *Pecopteris hemitelioides* BRONG. and *P. densifolia* GOEPP. could represent a characteristic feature of the Lupacu Bătrîn Beds paleoflora. The boreholes having pierced this complex have yielded other species too, of which the following are mentioned (Năstăseanu et al., 1977): *Neuropteris praedentata* GOTH., *N. ovata* HOFFM., *Reticulopteris germani* GIEBEL sp., *Linopteris neuropteroides* (GUTB.), H. POT., *Pecopteris feminaeformis* SCHLOTH. We have mentioned these species only for information, since, as we have already mentioned, the extremely intricate tectonics makes us be cautious as far as their assignment to the Lupacu Bătrîn Beds is concerned.

Based on the mentioned flora, we could consider the middle and the upper parts of the Lupacu Bătrîn Beds to be of Stephanian A+B age.

Lupac Beds Paleoflora

The Lupac Bătrîn Beds complex contains a very rich and diverse paleoflora and an assemblage of fresh water Lamellibranchs (Pl. III, IV).

Without claiming that it contains all the paleofloristic species in the Lupac Beds, we draw up a list of the species identified by us, that surely belong to this complex: *Stigmaria ficoides* BRONG., *Annularia stellata* SCHLOTH., *A. sphenophylloides* ZENK., *Calamites cisti* BRONG., *C. cruciatus* BRONG., *C. gigas* BRONG., *C. suckowi* BRONG., *Asterophyllites equiseliformis* SCHLOTH., *Sphenophyllum oblongifolium* GERM. & KAULF., *Pecopteris acuta* BRONG., *P. bredovi* BRONG., *P. cyathea* SCHLOTH., *P. densifolia* GOEPP., *P. hemitelioides* BRONG., *P. feminaeformis* SCHLOTH., *P. longiphylla* CORS., *P. pectinata* GERM., *P. platoni* GR. EURY, *P. plumosa-dentata* BRONG., *P. polymorpha* BRONG., *Odontopteris brardi* BRONG., *Sphenopteris matheti* ZEILL., *Dicksonites sterzeli* ZEILL., *Lebachia piniformis* (SCHLOTH.) FLOR., *Calipteris conferta* STERNB., *Cordaites* sp.

This assemblage with paleofloristic elements specific to the Upper Stephanian (C) is hosted especially in the foot-wall and the cover of the coal beds. Towards the upper part of the Lupac Beds, where there are no coal beds, the paleoflora becomes extremely poor. Examining recent mining works, we have found out, in the upper part of the Lupac Beds, lumachelle levels with *Carbonicola carbonaria* and *Anthracomya thuringensis*. The presence of these Lamellibranchs accompanied only by big fragments of *Calamites suckowi* Brong. and *C. cisti* Brong. point out a rise in the water level, which has hindered flora to grow and therefore coals to accumulate. In an outcrop on the Bîrzăviţa Valley, Stănoiu (1964) points to the presence in Permian gritty deposits of the species *Carbonicola carbonaria* GOLDFUSS and *Anthracomya* cf. *A. thuringensis* GEINITZ, associated with *Linopteris neuropteroides* BRONG.

The Anthracosidae assemblage represents a limited facies of the top of the Carboniferous. In the Reşiţa Zone the progressive passage from Upper Carboniferous deposits to Lower Permian ones is well known. It is marked by the diminution in the Stephanian flora and the massive appearance of typical Permian flora.

Conclusions

In the north-western part of the Reşiţa Zone, the most complete sequence of Upper Carboniferous deposits in Romania is developed. The presence at its upper part of coal beds has made these deposits well known in point of lithology, leading to the elaboration of a detail lithostratigraphic column (Năstăseanu, 1978). On it we have mentioned all the paleofloristic species collected from the lithostratigraphic horizons concerned, thus pointing to the stratigraphic distribution of the mentioned paleoflora (Fig.).

Without intending to be a synthesis, the present paper only presents a few assemblages whose position in the stratigraphic column has been clearly established, the Carboniferous flora from those deposits being much richer, as results, in fact, equally from other previous papers.

The paleoflora from Valea Ariei, that belongs to the basal part of the Lupacu Bătrin Beds, is very homogeneous, consisting only in representatives of the genus *Neuropteris*. In the middle part of the Lupacu Bătrin Beds a paleoflora is noticed in which the first Pecopteridae appear, the most frequent being *Pecopteris hemitelioides* BRONG. and *P. densifolia* GOEPP.

As a whole, the two assemblages of the Lupacu Bătrin Beds point out the Westphalian D – Stephanian A+B age.

The evolution of the Stephanian flora reaches its culminating point in the Lupac Beds. This paleoflora is typical of the Upper Stephanian (C), characterized by a great diversity of the genera *Calamites* and *Pecopteris*.

It is high time to mention that, as has already been noticed, we have used the name of Stephanian C for the top of the Carboniferous, for at least two reasons.

First of all, in the Romanian geological literature of the last period of time, the name of Stephanian D has also been used for a stage in which there are elements of both Carboniferous and Permian flora. But our researches have shown that the main species considered to be typically Permian, i.e. *Callipteris conferta* STERNB. and *Lebachia piniformis* (SCHLOTH.) FLOR., can be sporadically found in the whole sequence of the Lupac Beds, not only at their upper part. Secondly, this conclusion is equally supported by other researchers who have pointed to the presence of the species *Callipteris conferta* STERNB. in the Stephanian C. of Saint-Etienne (Doubinger et al., 1976, fide Doubinger et al., 1986), being thus devoid of chronostratigraphic value.

As a whole, it can be asserted that the paleoflora from the Upper Carboniferous deposits in the north-western part of the Reşiţa Zone has specific contents, but all the species known from there can be found in the synthetical lists of paleoflora in the most typical coal basins in Europe (Doubinger, 1965; Vetter, 1968).

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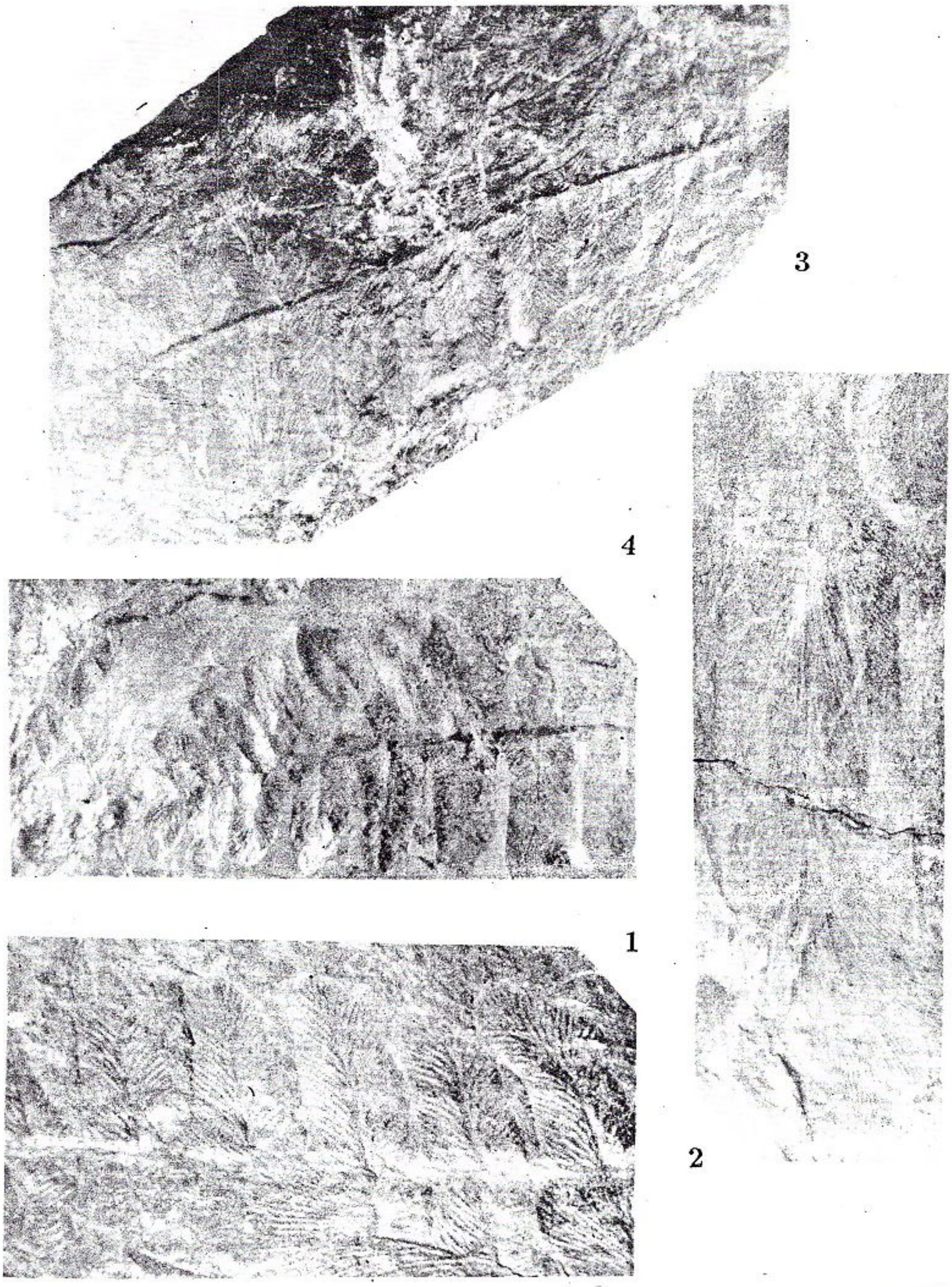

Plate I

Lupacu Bătrîn Beds Paleoflora

Fig. 1 – *Neuropteris attenuata* LIND & HUT., x 3. Valea Ariei. Private collection.
Fig. 2 – *Neuropteris gigantea* STBG., x 2. Valea Ariei. Private collection.
Fig. 3 – *Neuropteris heterophylla* BRGT., x 2. Valea Ariei. Private collection.
Fig. 4 – *Neuropteris ovata* HOFFM., x 2. Valea Ariei. Private collection.

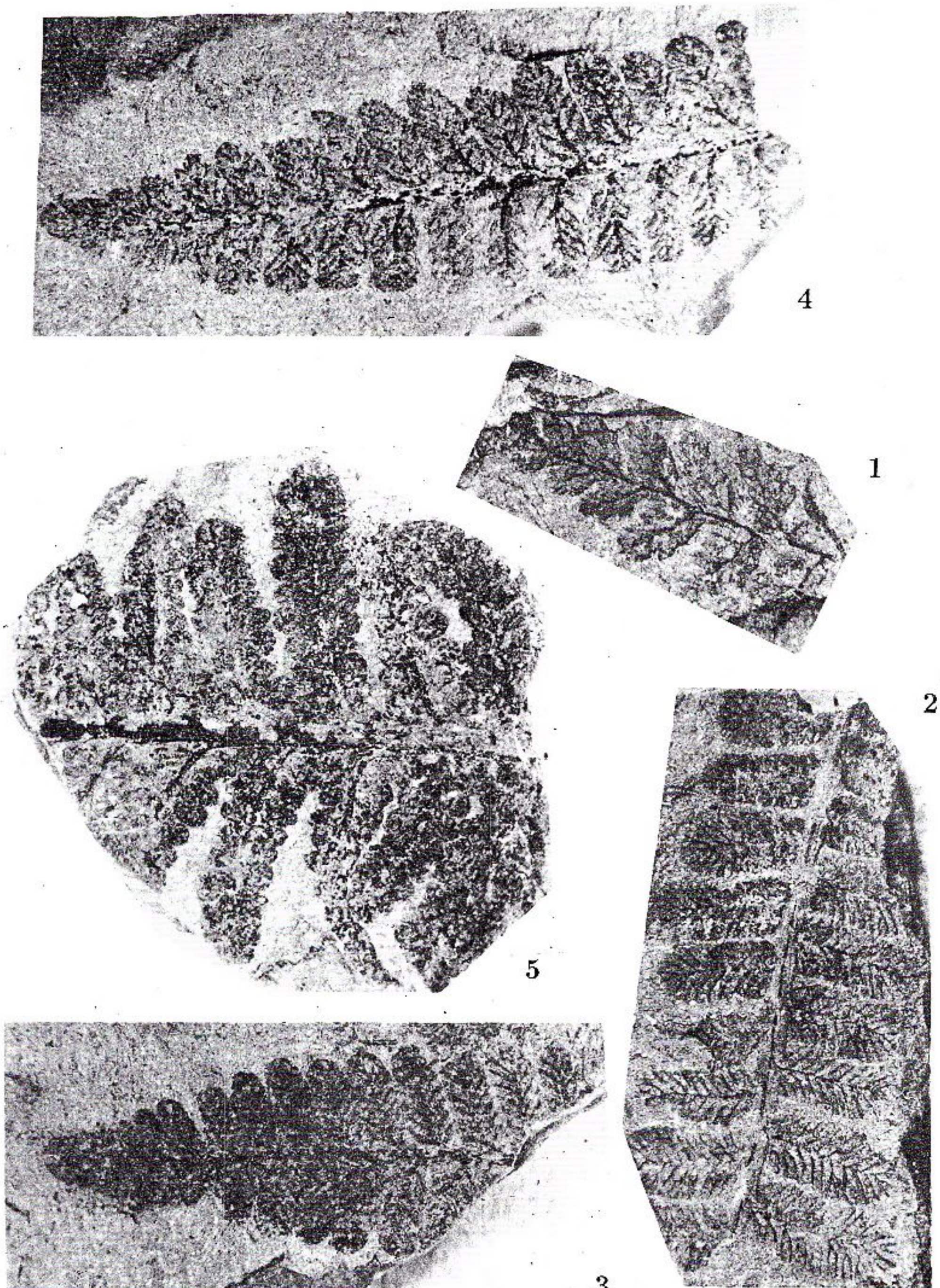

Plate II

Lupacu Bătrîn Beds Paleoflora

Fig. 1 – *Dicksonites beyrichi* WEISS, x 2.5. Tributary of the Bîrzava Valley. Private collection.
Fig. 2 – *Pecopteris densifolia* GOEPP., x 2.5. Tributary of the Bîrzavą Valley. Private collection.
Fig. 3 – *Pecopteris hemitelioides* BRONG., x 3. Tributary of the Bîrzava Valley. Private collection.
Fig. 4 – *Pecopteris hemitelioides* BRONG., x 3. Tributary of the Bîrzava Valley. Private collection.
Fig. 5 – *Pecopteris plumosa-dentata* BRONG., x 2.5. Tributary of the Bîrzava Valley. Private collection.

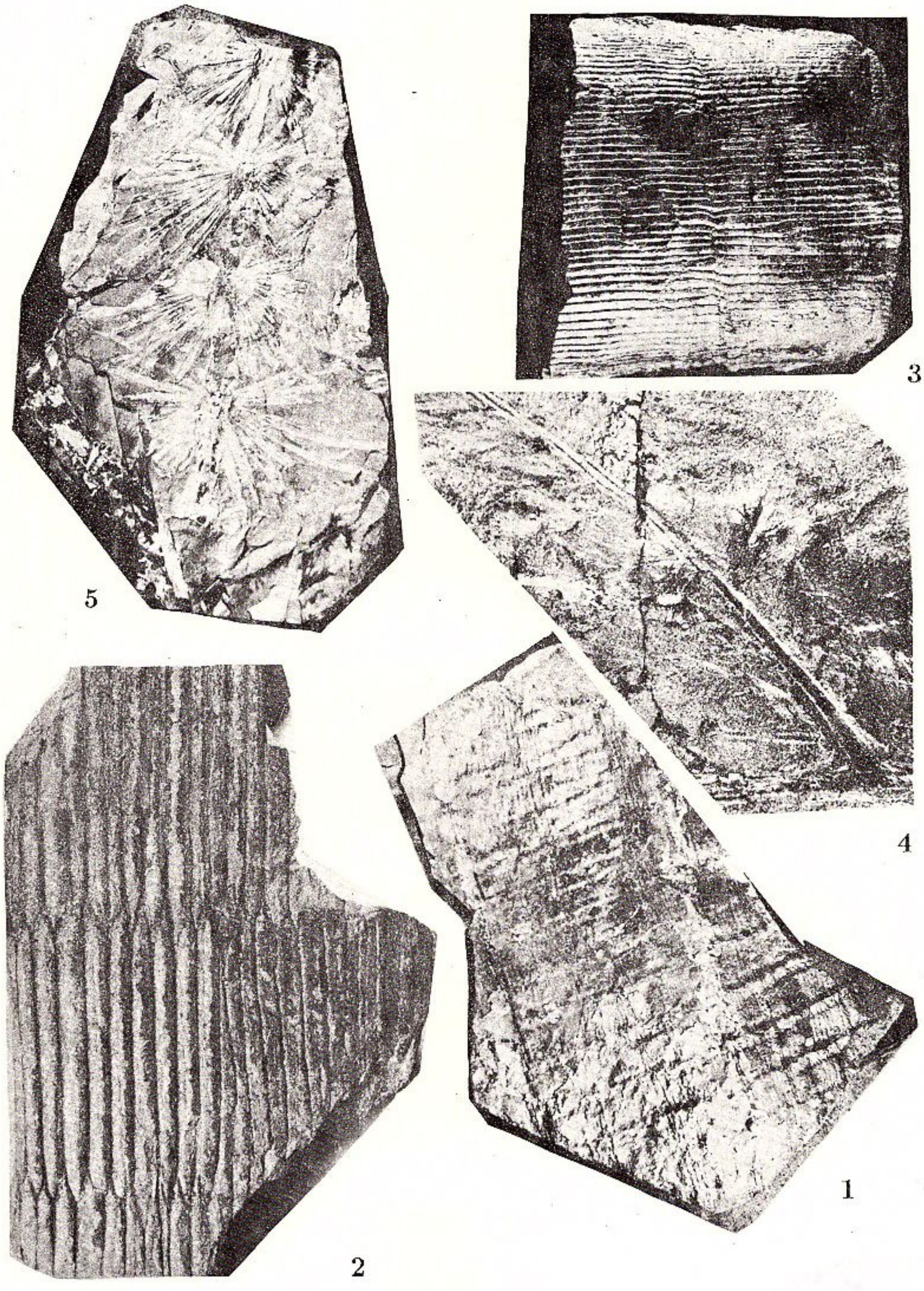

Plate III

Lupac Beds Paleoflora

Fig. 1 – *Calamites cruciatus* BRONG., x 0.5. Lupac, waste dump of gallery G 240. Private collection.
Fig. 2 – *Calamites gigas* BRONG., x 0.5. Tributary of the Bîrzăviţa Valley. Private collection.
Fig. 3 – *Calamites suckowi* BRONG., x 0.5. Tributary of the Bîrzăviţa Valley. Private collection.
Fig. 4 – *Dicksonites sterzeli* ZEILL., x 2. Lupac, waste dump of gallery G 240. Private collection.
Fig. 5 – *Annularia stellata* SCHLOTH., x 0.5. Lupac, waste dump of gallery G 241. Private collection.

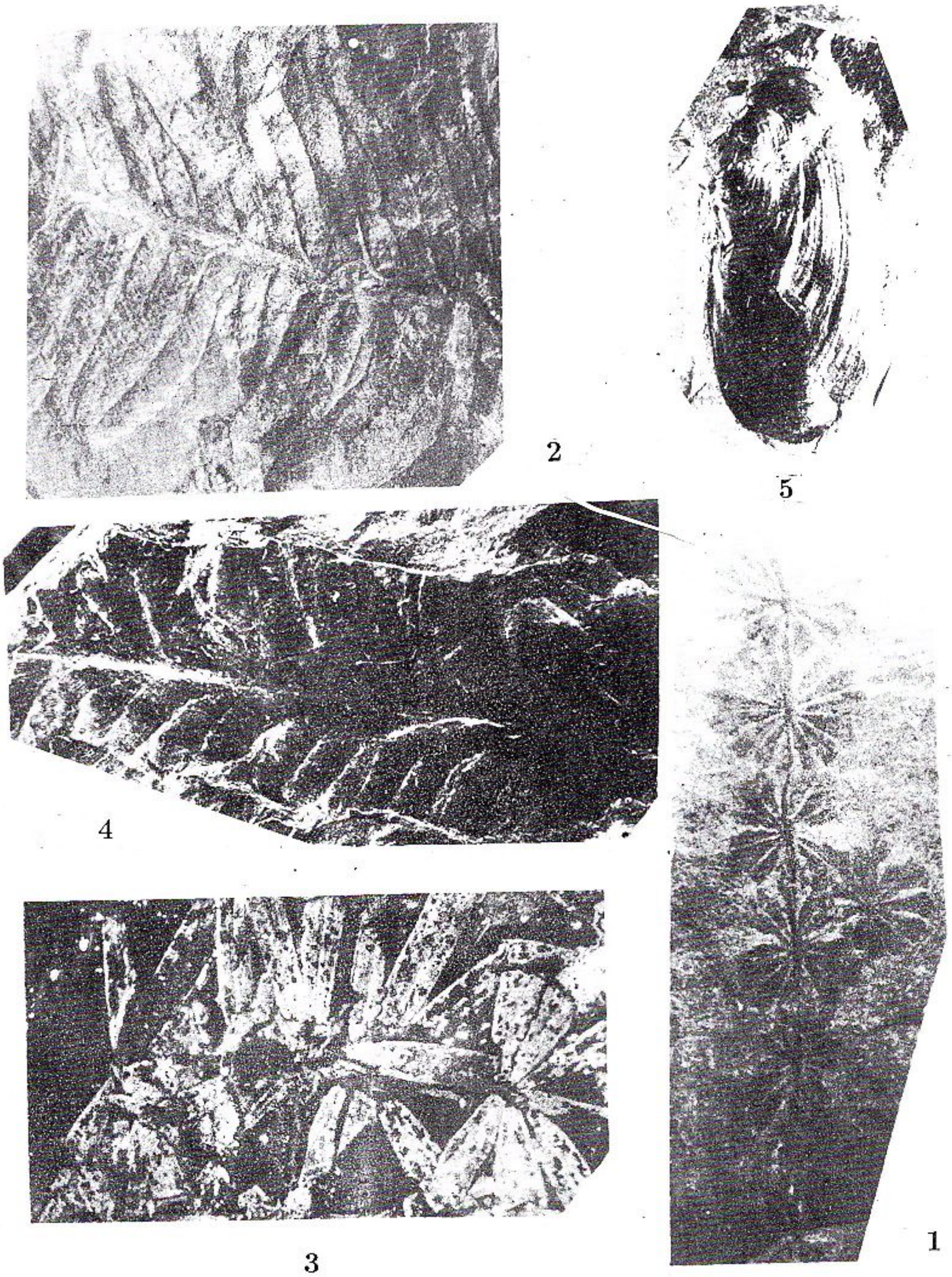

Plate IV

Lupac Beds Paleoflora

Fig. 1 – *Annularia sphenophylloides* ZENK., x 2.5. Lupac, wastw dump of gallery G 241. Private collection.
Fig. 2 – *Pecopteris longiphylla* CORS., x 2.5. Lupac, waste dump of gallery G 240. Private collection.
Fig. 3 – *Sphenophyllum oblongifolium* GERM. & KAULF., x 2.5. Lupac, waste dump of gallery G 241. Private collection.
Fig. 4 – *Calipteris conferta* STERNB., x 2.5. Lupac, waste dump of gallery G 240. Private collection.
Fig. 5 – *Anthracomya* cf. *thuringensis* GEIN., x 1.5. Lupac, waste dump of gallery G 240. Private collection.

Rom. J. Stratigraphy, 1993, 75, p. 9–15

CONSIDERAŢII STRATIGRAFICE ASUPRA FORMAŢIUNILOR GEOLOGICE DIN REGIUNEA CARANSEBEŞ – BISTRA (BANAT)

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Key words: Carboniferous. Jurassic. Cretaceous. Badenian. Sarmatian. Pannonian. Pontian. Stratigraphic units. Mollusks. Foraminifers. Ostracods. Biostratigraphy. South Carpathians – Sedimentary Getic and Supragetic Domains – Rusca Montană Zone.

Abstract: *Stratigraphic Data Concerning the Geological Formations in the Caransebeş – Bistra (Banat) Area.* The authors present their biostratigraphic and tectonic contributions. Unknown geological formations have been identified, of Carboniferous, Lower Badenian (Langhian), Upper Badenian (Kossovian), Lower Sarmatian, Pannonian s. str. and Pontian ages. The prevailingly ruptural character of the area is worth mentioning. Strike and transcurrent faults have been evidenced, that prove tectonic contracts between the Carboniferous/the metamorphic rocks, the Badenian/the metamorphic rocks, the Cretaceous/the Pontian etc.

Regiunea care face obiectul notei de faţă se situează în zona de legătură dintre depresiunea Caransebeş şi grabenul Bistra (pînă la oraşul Bistra). Datele obţinute prin prospecţiunea de mare detaliu executată de noi, cele din forajele efectuate de I.P.E.G. "Banatul", precum şi studiile paleontologice asupra probelor din aflorimente şi foraje ne-au permis obţinerea unor imagini stratigrafice noi privind acest sector.

Cercetarea geologică s-a axat de-a lungul unui secol şi jumătate în special pe studiul formaţiunilor post-tectogenetice ale depresiunii Caransebeş şi mai puţin ale grabenului Bistra. Cităm unele lucrări care au caracter de sinteză: Lubenescu, Pavnotescu, 1967, date nepubl., 1970; Marinescu et al., 1977; Marinescu, Popescu, 1987; Petrescu et al., 1987; Breban et al., 1988; Lubenescu et al. (sub tipar, 1989).

Formaţiunile geologice prezente în zonă aparţin ramei şi bazinului post-tectogenetic. Formaţiunile de ramă sînt reprezentate prin şisturi cristalofiliene (Formaţiunea de Băuţari – Savu et al,. 1977), roci magmatice şi depozite sedimentare de vîrstă paleozoică (Carbonifer) şi mezozoică (Jurasic inferior şi mediu şi Jurasic superior– Cretacic inferior) (fig. 1).

Carbonifer. Pe afluentul dreapta al pîrîului Găina, la sud de localitatea Var, au fost întîlnite cele mai vechi depozite sedimentare ale pînzei getice, reprezentînd molasa hercinică, de vîrstă carboniferă, nesemnalate pînă în prezent.

Depozitele repauzează peste paragnaisele ce formează fondul petrografic al regiunii şi suportă suita sedimentară mezozoică aparţinînd şanţului geosinclinal Rusca Montană. Menţionăm prezenţa unei zone milonitizate la contactul sedimentar/cristalin. Litologic, depozitele carbonifere sînt alcătuite din gresii bogat micafere, cenuşii, cenuşiu-negricioase, dure, cu spărtură neregulată, stratificate în bancuri decimetrice-submetrice, în care se remarcă, spre partea inferioară, un banc submetric de conglomerate şi spre partea superioară intercalaţii decimetrice de pelite cu aspect şistos (fig. 2). Acestor depozite li s-a atribuit vîrsta carboniferă atît pe baza asemănării litologice cu depozitele carbonifere din zonele Reşiţa-Moldova Nouă, Măgura Brebului etc. (Boldur, Boldur, 1962; Năstăseanu, Boldor, 1964), cît şi pe baza conţinutului palinologic. Probele palinologice prelevate din intercalaţiile pelitice, analizate de Mărgărit Maria (căreia îi mulţumim şi pe această cale), sînt sărace, identificîndu-se doar cîteva elemente fitoplanctonice aparţinînd speciilor *Hymenodiscina sincera* TSCH. şi *Hymenodiscina conspicua* TSCH. ce participă frecvent la alcătuirea asociaţiilor microfloristice caracteristice Devonianului terminal şi Carboniferului.

Jurasic inferior-mediu. Discordant peste depozitele carbonifere, la sud de localitatea Var, repauzează o stivă de depozite groase de 50–70 m, alcătuite din gresii cuarţoase siltice, cu ciment carbonatic, argilite în pachete decimetrice şi siltite cenuşiu-verzui, cu grosimi

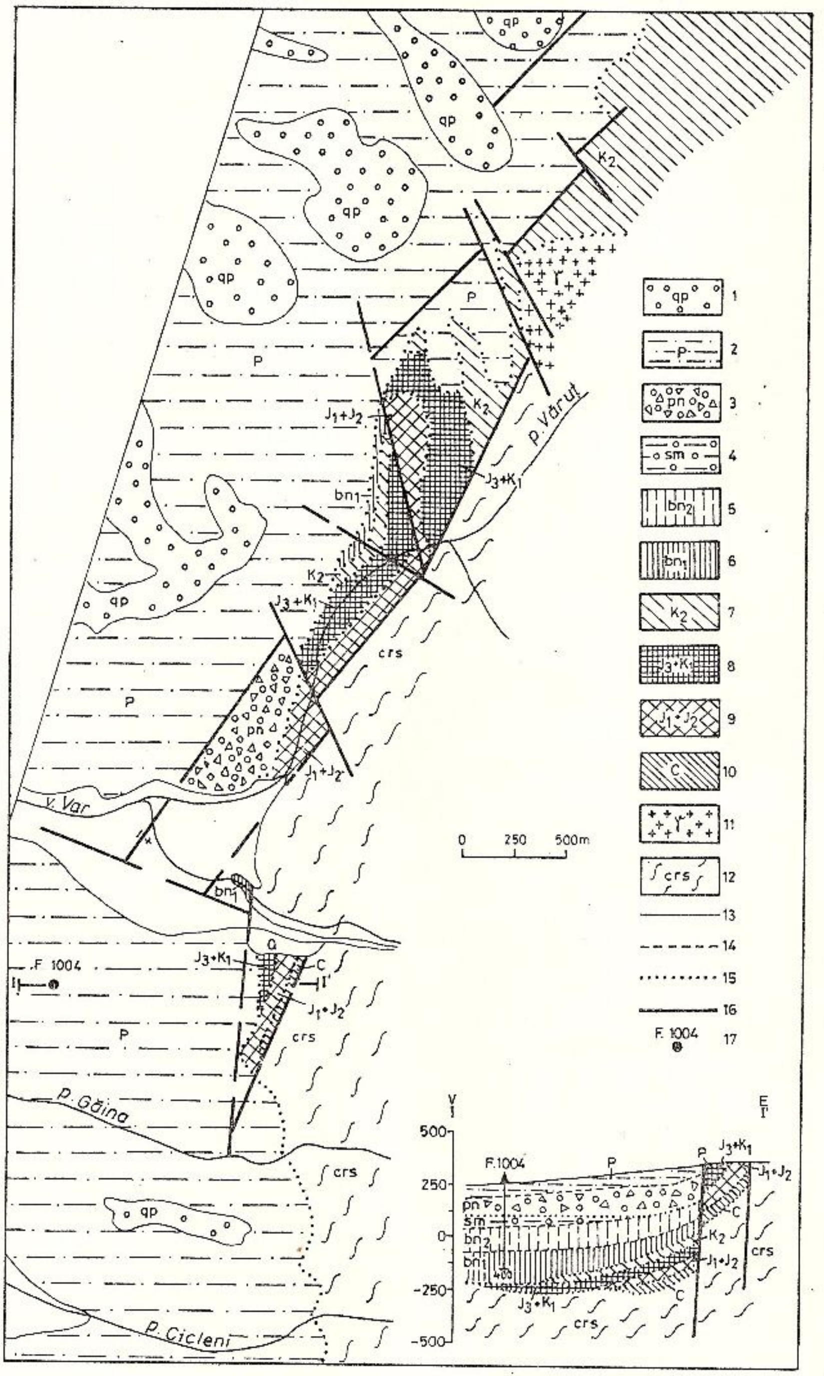

Fig. 1 – Secţiune geologică prin forajul F 1004 şi harta ramei bazinului.
1, Cuaternar (?); 2, Ponţian inferior (p); 3, Pannonian s. str. (pn); 2 + 3 = formaţiunea de Valea Bistrei; 4, Sarmaţian (sm) = formaţiunea salmastră; 5, Badenian superior (Kossovian)(bn_2); 6, Badenian inferior (Langhian)(bn_1); 5 + 6 = formaţiunea de Delineşti; 7, Cretacic superior (K_2); 8, Jurasic superior-Cretacic inferior ($J_3 + K_1$); 9, Jurasic inferior-mediu ($J_1 + J_2$); 10, Carbonifer (c); 11, formaţiuni eruptive; 12, formaţiuni cristalofiliene; 13, limită de formaţiune; 14, limită geologică normală; 15, limită de discordanţă; 16, falie; 17, foraj.

submetrice şi metrice, avînd intercalaţii subţiri de calcarenite şi calcisiltite. Alterarea conferă gresiilor o patină gălbui-roşcată.

La nord de localitatea Var, aceleaşi formaţiuni apar în contact tectonic cu şisturi cristalofiliene. Vîrsta jurasic-inferioară, conferită de Dincă (1977) s-a făcut prin similitudini cu formaţiuni asemănătoare din bazinul Haţeg. Conform hărţii geologice, foaia 104 d Muntele Mic, scara 1:50.000 (Savu et al., 1981), s-a atribuit acestor formaţiuni vîrsta jurasic inferioară şi medie.

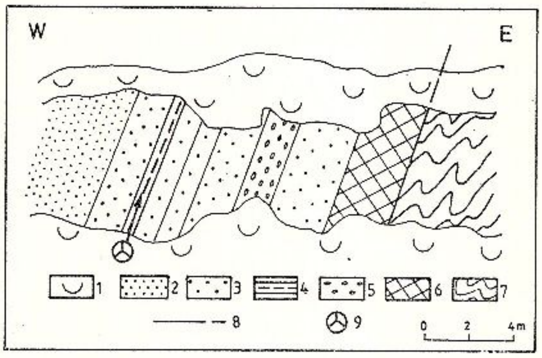

Fig. 2 – Schiţă de aflorimente (versantul drept al afluentului dreapta al pîrîului Găina.
1, depozite de pantă; 2, Jurasic inferior-mediu (gresii cuarţoase, siltite); 3, gresii micacee; 4, argile şistoase; 5, conglomerate; 3 + 4 + 5 = Carbonifer; 6, zonă milonitizată; 7, cristalin; 8, falie; 9, probă palinologică.

Jurasic superior – Cretacic inferior. Discordant, peste depozitele Jurasicului inferior-mediu, repauzează o stivă calcaroasă care are în bază un banc de calcare cenuşiu-negricioase spatice, masive, compacte, cu spărtură neregulată şi cu vinişoare de calcit, formînd reţele anastomozate, iar spre partea superioară calcare cenuşii şi cenuşiu-albicioase, dure, masive. Microfacial, acestea sînt reprezentate prin biopelsparite şi biopelmicrite cu textură brecioasă, foarte fisurate, cu resturi de alge dasicladacee indeterminabile, microoncolithe, foraminifere triturate şi rare calcisfere. Apar, de asemenea, calcare recristalizate cu faldspaţi de neoformaţie, cuarţ detritic şi minerale opace cu contur izometric (? pirită). Grosimea formaţiunii este de 50–150 m. La nord de grabenul Bistra, unor formaţiuni similare, pe baza unor elemente microfaciale, Dincă (1977) le atribuie vîrsta Dogger-Malm. Prezenţa Cretacicului inferior este semnalată pe harta geologică, scara 1:50.000 (Savu et al., 1981).

Cretacic superior. Transgresiv şi discordant peste şisturile cristalofiliene, magmatite sau depozite sedimentare mai vechi, se dispune o formaţiune alcătuită din conglomerate, microconglomerate, gresii cuarţoase cenuşiu-gălbui, gresii calcaroase, marnocalcare şi marne argiloase cu aspect şistos, avînd o grosime de 200–300 m. Întreaga suită este sterilă din punct de vedere faunistic. Vîrsta de mai sus a fost atribuită conform hărţii I.G.G., scara 1:50.000.

Cuvertura post-tectogenetică neogenă. Depozitele care alcătuiesc umplutura depresiunii Caransebeş (în acest sector) şi, implicit, a grabenului Bistra au fost întîlnite atît în aflorimente, cît şi în foraje, şi aparţin Badenianului, Sarmaţianului, Pannonianului s.rstr. = Malvensian, Ponţianului şi Cuaternarului.

Formaţiunea de Delineşti (Langhian + Kossovian), separată de Marinescu şi Popescu (1987) şi regăsită în regiunea cercetată de către noi, se dispune transgresiv şi discordant peste termenii mai vechi. Astfel, ea apare pe pîrîul Var, pe pîrîul Văruţ, precum şi în forajul 1004.

Pe pîrîul Var, la cca 700 m amonte de confluenţa cu pîrîul Văruţ, în talveg şi în ambii versanţi, de sub depozitele aluvionare, pe cca 60 m lungime, apar depozite badeniene, în contact tectonic cu şisturi cristaline (fig. 3). Petrografic, cristalinul este reprezentat prin gnaise cu biotit în care se evidenţiază filoane de cuarţ. În partea mediană a zonei de falie, cristalin/sedimentar care are o grosime de 30 m şi pe care aflorează argile albe, cu pete cenuşii şi roşii (mascate parţial de aluviuni), apare, în poziţie verticală un bloc de cca 1–1,5 m grosime de tuf bentonitic şi marne verzui-negricioase, fără resturi fosile. În aval de zona de falie, cu căderi spre ramă (fig. 3) cu valori cuprinse între 20-60^0, se observă o suită de depozite alcătuită din pachete metrice de argile cenuşiu-albăstrui, micafere, fosilifere, marne cenuşii şi argile cenuşii, nisipoase, cu intercalaţii de gresii calcaroase cenuşii, fine şi medii, micafere, fosilifere (cu resturi de plante incarbonizate şi moluşte) formînd plăci cu grosimi cuprinse între 0,15–0,40 m.

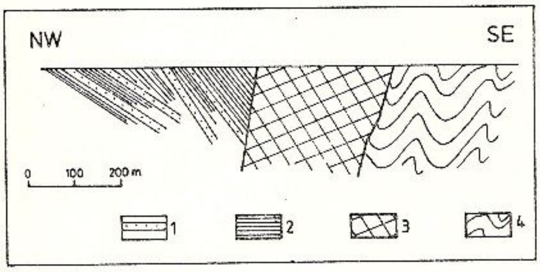

Fig. 3 – Schiţă de aflorimente pe valea Var (contact Badenian inferior/Cristalin).
1, gresii micafere fosilifere; 2, argile, marne; 1 + 2 = Badenian inferior; 3, zonă de falie; 4, cristalin.

Fauna de moluşte întîlnită anul acesta, precum şi cea recoltată de V. Lubenescu în 1967 (Lubenescu, Pavnotescu, 1970), din depozitele grezo-calcaroase, cît şi din cele pelitice este alcătuită, predominant, din gasteropode, la care se adaugă lamelibranchiate; între formele întîlnite menţionăm: *Corbula (Varicorbula) gibba* OLIVI, *Nucula (Nucula) nucleus* (LINNÉ), *N.(N.)*

sulcata BRONN., *Glycymeris pilosus deshayesi* (MAY), *Turritella (Archimediella) turris* (BAST.), *T.(Zaria) subangulata polonica* FRIEDB., *Ocinebrina sublavata striata* (EICHW.), *Murex (Haustellum) partschi* HOERN., *Conus (Chelyconus) vindobonensis* PARTSCH, *Columbella* ex. gr. *Columbella curta* DUJ.

Analiza micropaleontologică a pus în evidenţă, pe valea Var, numeroase foraminifere, ostracode şi fragmente de echinoderme. Se remarcă asociaţia de foraminifere (planctonice şi bentonice) şi ostracode caracteristice zonei cu *Candorbulina universa / Turborotalia bykovae* (= partea echivalentă zonei cu lagenide). Cităm: *Heterostegina costata* D'ORB., *Amphistegina lessonii* D'ORB., *Martinottiella kuscheriana* (CUSH.), *Textularia danae* POPESCU, *Spiroplectinella carinata* (D'ORB.), *Cycloforina reticulata* (KARRER), *Adelosina longirostra* (D'ORB.), *Stilostomella elegans* (D'ORB.), *Nodosaria bacillum* DEFR., *Dentalina paronai* (DERV.), *Lenticulina cultrata* (MONTF.), *Heterolepa dutemplei* (D'ORB.), *Bolivina vienensis* MARKS., *Brizalina dilatata* (REUSS.), *Praeorbulina transitoria* (BLOW), *Globigerinoides sicanus* DI STEFANI, *Globigerinoides triloba* (REUSS.), *Globigerinoides imaturus* LE ROY, *Globoquadrina langhiana* CITAGELATI, *Globoquadrina globosa* BOLLI, *Turborotalia bycovae* AIS., *Hermanites haidingeri* (REUSS.), *Mutilus (Aurila) cicatricosa* (REUSS.), *Loxoconcha punctatella* (REUSS.), *Falunia* ex. gr. *F. plicatula* (REUSS.).

Asociaţia paleontologică caracterizează Langhianul şi prezintă afinităţi cu cea de la Balta Sărată, Delineşti şi Lăpugiu.

În forajul 1004 a fost interceptată partea terminală a Langhianului (în intervalul 280–300 m), stabilit prin apariţia, cu o frecvenţă ridicată, a speciei *Globigerina drury* AKERS, alături de foraminiferele: *Marginulina hirsuta* D'ORB., *Lenticulina cultrata* (MONTF.), *Lenticulina arcuatostriata* HANTKEN, *Lenticulina formosa* (CUSH.), *Stilostomella elegans* (D'ORB.), *Orbulina bilobata* (D'ORB.), *Candorbulina suturalis* (BRONN.), *Globigerinoides triloba* (REUSS.), *Globigerinoides quadrilobatus* D'ORB., *Globigerina obesa* BOLLI.

Depozitele superioare ale Formaţiunii de Delineşti au fost întîlnite numai în forajul 1004 (intervalul 190–280 m). Microfauna de foraminifere şi ostracode determinată caracterizează Kossovianul bazal, reprezentat prin speciile: *Spiroplectinella carinata* D'ORB.), *Siphotextularia concava* (KARRER), *Pavonitina styriaca* SCHUBERT, *Pyrgo lunula* (D'ORB.), *Borelis rotella* D'ORB., *Baggatella gutsulica* (LIV.), *Baggatella elongata* (D'ORB.), *Baggatella subulata* (CUSH. et PARKER), *Brizalina dilatata* (REUSS.), *Reusella banatica* POPESCU, *Uvigerina brunensis* KARRER, *Uvigerina asperula* CZJZEK, *Valvulineria complanata* (D'ORB.), *Elphidium crispum* (LINNÉ), *Cribrononion flexuosum* (D'ORB.), *Cytheridea hungarica* (ZAL.), *Senesia vadaszi* (ZAL.). Asociaţia microfaunistică este caracteristică zonei cu *Pavonitina styriaca* SCHUBERT, avînd afinităţi cu microfauna de la Buituri (Popescu, 1977).

Formaţiunea salmastră inferioară – Sarmaţian (Marinescu, Popescu, 1987) a fost interceptată numai în forajul 1004, unde Badenianul terminal, reprezentat litologic prin marne şi nisipuri cenuşii, suportă nisipuri cenuşii cu intercalaţii subţiri de pietrişuri, marne şi marne nisipoase (în intervalul 150–190 m), în care s-a întîlnit, în exclusivitate, specia *Anomalinoides dividens* LUCZK., foraminifer ce caracterizează baza Volhinianului.

Formaţiunea de Valea Bistrei (Pannonian s.rstr. = Malvensian terminal–Ponţian). Depozitele aparţinînd acestei formaţiuni repauzează transgresiv şi discordant peste termeni jurasici sau sarmaţieni (în foraje). Litologic, în bază apar pietrişuri şi bolovănişuri poligene, cu un grad variat de rulare, constituite din fragmente de şisturi cristaline, roci eruptive şi sedimentare (mezozoice) prinse într-o matrice nisipoasă, gălbui-roşcată, uneori slab consolidată, nefosiliferă (forajele 1004, 1018, 1019, 1017).

Complexul pietrişurilor şi bolovănişurilor a fost recunoscut şi în aflorimente, pe pîrîul Văruţ în amonte de confluenţa acestuia cu pîrîul Var, repauzînd transgresiv şi discordant peste depozitele Jurasicului inferior-mediu. În continuitate de sedimentare se dispune un complex argilos-marnos cu fragmente de congerii şi limnocardiide evidenţiat în afloriment pe pîrîul Găina (afluent dreapta al pîrîului Cicleni), precum şi în forajele 1020, 1015, 1017, 1018 şi 1004, cu grosimi cuprinse între 20 şi 300 m.

Asociaţia macropaleontologică a Pannonianului s.rstr. a fost recunoscută numai în forajul 1015 (intervalul 366-408 m) şi forajul 1016 (adîncimea 370 m). Cităm: *Melanopsis bouei* FÉR., *Melanopsis bouei rarispina* LÖR., *Melanopsis vindobonensis* FUCHS, *Pontalmyra (Pontalmyra) promultistriata* (JEK.), *Congeria ramphophora* BRUS., *Congeria banatica* R. HÖRN.

Pannonianul s.rstr. a fost identificat în forajele 1004, 1015, 1016 şi 1017 şi pe baze micropaleontologice, întîlnindu-se, în exclusivitate, ostracode în asociaţie cu *Orygoceras*: *Hungarocypris abscisa* (REUSS., *Hemicytheria amphalodes lörenthay* (MEHES), *Candona multipora* POK., *C.(Caspiolla) elongata* SOKAĆ, *C.(C.) prochaskai* POK., *C.(Typhocypris) ornata* OLTEANU, *C.(Pontoniella) unguiculus* (REUSS.), *Loxoconcha granifera* (REUSS., *Loxoconcha hodonica* POK., *Leptocythere lacunosa* (REUSS.), *Pontoleberis altilata* STANCEVA, *Cypria sibovici* KRSTIĆ, *Cyprideis triangulata* KRSTIĆ.

Trecerea de la Pannonianul s.rstr. la Ponţian a fost surprinsă numai în forajele studiate de noi din acest sector (1004 - intervalul 50-60 m; 1016 - intervalul 350-360 m; 1000 - intervalul 350-366 m şi 1017 - adîncimea 280 m). Într-o suită argiloasă-marnoasă, pe lîngă rare fragmente de *Congeria* ex gr. *C.digitifera* şi *Paradacna abichi*, se disting *Candona (Typhlocypris) ornata* OLTEANU (taxon ce precede saltul spre noul gen ponţian *Bakunella* - Olteanu, 1986) în asociaţie cu genul *Pontoniella* (cu valvele nestriate) şi *Cyprideis triangulata* care, după Jiricek (1974, 1975) reprezintă "Infraponţianul". Situaţii similare am întîlnit şi în forajele săpate în Banatul de vest (Cornea et al., 1987) şi în "golful" Făget (Mihăilescu et al., 1987).

Partea superioară a Formaţiunii de Valea Bistrei are cea mai mare extindere în sectorul cercetat de noi şi aparţine Ponţianului. Fauna aflată în aflorimente este neconcludentă (apar fragmente de congerii şi limnocardiide). În schimb, în aproape toate forajele studiate s-au întîlnit indicii faunistice pentru separarea acestui etaj.

Ponţianului îi revin pachetele pelitice ale părţii superioare a complexului marnos-argilos şi un complex de pietrişuri suprajacent. Alcătuirea litologică este următoarea: în bază, se disting marne cenuşii, compacte, marne albicioase, argile şi marne cenuşiu-verzui, urmate de o alternanţă de marne şi nisipuri cenuşii în care apar, la diferite nivele, resturi de plante incarbonizate, faună de moluşte şi ostracode. În unele foraje succesiunea se încheie cu pietrişuri şi nisipuri cu structură încrucişată, ceea ce sugerează acumulări fluviatile (15-20 m).

Fauna identificată în forajele 1006, 1015, 1016, 1018 şi 1020 este destul de săracă, fiind alcătuită din moluşte şi ostracode la care se adaugă resturi de plante. Între moluşte cităm: *Congeria digitifera* ANDRUS., *Congeria czjzeki* (M. HÖRNES), *Caladacna steindachneri* (BRUS.), *Paradacna abichi* HOERN., *Prosodacna* sp., *Radix paucispira* (FUCHS), *Valvata* sp., *Lithoglyphus* sp., *Planorbis* sp.

În forajul 1020 (între 16,50-82,75 m) se distinge o suită de argile cenuşii şi cenuşiu-negricioase, cărbunoase, cu intercalaţii de nisipuri şi gresii cenuşiu-gălbui, micacee. Pe lîngă moluştele prezente în asociaţie, între care predomină rare gasteropode (*Valvata* sp., *Lithoglyphus* sp., *Planorbis planorbis* (LINNÉ)) şi rare lamelibranchiate (*Congeria digitifera* - exemplare de talie mică şi paradacne), apar şi resturi de plante (determinate de colegul N. Ţicleanu de la I.G.G., căruia îi mulţumim şi pe această cale) între care menţionăm: *Glyptostrobus europaeus* (BRONN.) HEER, *Acer tricuspidatus* BRONN. şi plante ierbacee.

Studiul microfaunei de ostracode accentuează prezenţa Ponţianului inferior, în asociaţie determinîndu-se următorii taxoni: *Hungarocypris pannonica* SOKAČ, *Hungarocypris hieroglyphica* (MEHES), *Candona (Caspiolla) balcanica* (ZAL.), *Candona (Pontoniella) acuminata* (ZAL.), *Candona (Pontoniella) unguiculus* (REUSS.), *Candona (Caspiocypris) pontica* SOKAČ, *Candona (Lineocypris) reticulata* (ZAL.), *Candona (Typhlocypris) ornata* (OLTEANU), *Cyprideis triangulata* KRSTIĆ, *Cyprideis macrostigma* KOLLM., *Loxoconcha subrugosa* ZAL., *Loxoconcha djaffarovi* SCHN., *Bakunella dorsoarcuata* (ZAL.), *Hemicytheria dubokensis* MEHES., *Cypria tocorjescui* HANGANU, *Pontolebris pontica* STANCEVA.

Fauna studiată aparţine Ponţianului inferior, inclusiv partea sa terminală şi începutul Portaferianului. Ne situăm, probabil, sub stratele purtătoare de cărbuni de la Visag, Darova şi Sinersig şi, în parte, sub cele de la Căvăran sau chiar sincrone cu acestea (Lubenesu, Ştefănuţ, 1986).

Cuaternarul. Depozitele cuaternare au o extindere relativ mare şi aparţin preponderent teraselor Timişului şi Bistrei şi, subordonat, formaţiunilor aluviale, proluviale şi deluviale.

Consideraţii tectonice. Regiunea studiată de noi se suprapune, între localităţile Iaz-Bistra, peste extremitatea sudică a grabenului Bistrei, iar între localităţile Cicleni şi Var, peste sectorul nord-estic al depresiunii Caransebeş-Mehadia, formată în timpul şi după tectogeneza miocenă.

Depozitele sedimentare mezozoice ale părţii sudice a bazinului Rusca Montană participă, alături de metamorfite şi plutonite, la alcătuirea ramei depresiunii Caransebeş (în zona cercetată).

Depozitele carbonifere identificate de noi la sud-est de localitatea Var, ar indica prezenţa extremităţii nord-estice a şanţului geosinclinal al Reşiţei, ceea ce ar sugera că, bazinul Rusca Montană nu s-a conturat numai ca o prelungire a şanţului, ci s-a şi suprapus parţial peste acesta.

Imaginea tectonică realizată prin coroborarea datelor obţinute prin prospecţiunea geologică de suprafaţă cu datele din foraje şi cu cele geofizice, evidenţiază caracterul predominant ruptural al depresiunii. Astfel, a fost evidenţiată prezenţa a două sisteme de falii: un sistem de falii direcţionale, cvasiparalele cu faliile principale de scufundare (nord-sud, în cazul depresiunii Caransebeş şi vest-est, în cazul grabenului Bistra) mai vechi şi un sistem de falii transversale, ceea ce a condus la o compartimentare accentuată, în blocuri, situaţie confirmată deja, în zona Buchini-Balta Sărată-Zerveşti, de lucrările de explorare-exploatare (Petrescu et al., 1987).

În partea sudică a grabenului Bistra au fost recunoscute trei falii, aparţinînd sistemului direcţional, care au putut fi urmărite între pîrîul Scoarţa, la est şi pîrîul Iezuţ, la vest.

Pe pîrîul Var a fost identificată, la contactul Badenian inferior/cristalin, o zonă de falie umplută cu argilă, avînd o lăţime de cca. 30 m, care scoate la zi, în partea mediană, argile langhiene pe cca. 1,50 m grosime.

Faliile transversale evidenţiază contactul tectonic Carbonifer/cristalin, Jurasic/Cretacic, formaţiuni mezozoice/formaţiuni neogene sau depozite mezozoice şi neogene din compartimentul nordic, faţă de depozite jurasice din compartimentul sudic. Menţionăm că, la sud-vest de dealul Făget, pe afluentul drept al pîrîului Văruţ, a fost surprinsă o falie care delimitează un compartiment în care au fost identificate depozite mezozoice (Jurasic/Cretacic) care a suferit şi o rotaţie către est.

Concluzii. Principalele contribuţii reieşite din nota de faţă sînt de ordin biostratigrafic şi tectonic.

Astfel, au fost puse în evidenţă formaţiuni geologice noi aparţinînd Carboniferului, Badenianului, Sarmaţianului, Pannonianului s.str. şi Ponţianului.

Depozitele carbonifere prezintă similitudini litologice cu cele din zonele Reşiţa–Moldova Nouă şi Măgura Brebului, dar şi o asociaţie palinologică de vîrstă paleozoică.

Badenianul inferior (Langhian, zona Candorbulina universa/Turborotalia bykovae) a fost recunoscut în aflorimente şi foraje. Numai în forajul 1004 s-a separat şi Badenianul superior (Kossovian).

Sarmaţianul bazal (zona cu Anomalinoides dividens) a fost surprins numai în forajul 1004.

Pannonianul s. rstr. (= Malvensian) are o arie de dezvoltare restrînsă în aflorimente şi foraje. Documentarea sa paleontologică s-a făcut pe bază de macrofaună (gasteropode şi lamelibranchiate caracteristice asociaţiei cu *Congeria banatica*).

Ponţianul ocupă, alături de Cuaternar, marea majoritate a ariei cercetate de noi. El este bine individualizat şi în foraje. Limita Pannonian s. rstr./Ponţian a fost trasată pe baza apariţiei unor taxoni ponţieni: *Congeria digitifera*, *Caladacna steindachneri*, *Paradacna abichi* şi o bogată ostracofaună cu *Candona* div.sp., *Bakunella dorsoarcuata* şi *Cyprideia* div.sp. Menţionăm şi cu această ocazie că, după părerea noastră, limita Pannonian s. rstr. (Malvensian)/Ponţian trebuie căutată în cuprinsul stratelor de Cîmpie (Langenfeld) (Cornea et al., 1987) şi a marnelor albe cu ostracode (Beiuş, Zarand – Istocescu, 1971) sau, poate, în baza acestor subdiviziuni considerate încă de vîrstă Pannonian s. rstr. = Malvensian.

Din punct de vedere tectonic, se distinge caracterul predominant ruptural al zonei. S-au evidenţiat sisteme de falii direcţionale şi transversale.

Faliile transversale demonstrează o serie de contacte tectonice între Carbonifer/cristalin, cristalin/Badenian, Pannonian s. rstr./cristalin, Cretacic/Ponţian (în parte) etc. Unul dintre compartimentele tectonice separate la sud-vest de dealul Făget a suferit o scufundare şi o rotaţie către est.

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PRESENCE OF THE "VERMICULATE SANDSTONE FORMATION" (PERMIAN) IN THE CENTRAL ZONE OF THE HIGHIŞ MOUNTAINS

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Key works: Permian. Sedimentary rocks. Ichnofossils. Planolites. Apuseni Mountains – Southern Apuseni Mountains – Zarand (Highiş) Mountains.

Abstract: In the paper the age of this sedimentary formation occurring in the central part of the Highiş crystalline massif is specified. The formation, known as the "Black Series" or the Cladova Formation was reported by previous researchers either to the Carboniferous or to the Carboniferous–Permian, the Permian or merely to the Paleozoic. The lithological constitution and the presence of traces belonging to worms as Planolites type lead to the conclusion that these deposits represent the "Vermiculate Sandstone Formation", that crops out also to the north-east, in the Bihor Mts.

The present paper deals with the assignment of a formation settled in the central part of the crystalline Highiş Massif, a formation referred by our predecessors to the Carboniferous, the Carboniferous–Permian, the Permian or the Paleozoic.

Our researches have been carried out in view of drawing up the geological map, scale 1 : 50 000, Şiria Sheet. They have covered the central-western area of the Highiş Mountains, generally trending west-east, between the Cuvin locality and the sources of the Şoimuş Valley.

As results from the studies carried out in the Bihor Mountains and in the Permian areas (Biharia, Pietroasa, Zece Hotare and Poiana Horea Sheets, all of them on scale 1 : 50 000) we have been interested, from the very beginning, in the Highiş Mts., in investigating a formation which Giuşcă (1962) and Dimitrescu (1967) compared with the "Black Series" in the Bihor Mountains.

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The previous researches referred the pile of sedimentary rocks cropping out in the central part of the Highiş Mountains to the Carboniferous (Lóczy, 1884, 1885; Rozlosznik, 1936). In 1962 Giuşcă pointed out the similarity of these rocks with the so called "Black Series" in the Bihor Mountains, but without being able to establish the age more precisely. In the same year, Dimitrescu assigned these rocks to the Carboniferous–Permian and showed they unconformably lie over Păiuşeni Series. Subsequently, Dimitrescu (1967), Cibotaru et al. (1975, unpubl. data) and Giuşcă (1979) referred these deposits to the Permian, Cibotaru also mentioning the presence of basic tuffs within them. In 1986, Balintoni accepted the opinion of his predecessors and placed these deposits, which he grouped under the name of the "Cladova Formation" within the Upper Carboniferous interval. Recently, Pană, Ricman (1988) considered the Cladova Formation to be made up only of a few types of thermally altered basalts and cataclazed granitoid bodies; the authors call this formation the "Cladova Complex" and assign a Paleozoic age to it.

Stratigraphic and Structural Considerations

The pile of rocks described by Balintoni as the "Cladova Formation" ("Black Series" according to Giuşcă or Dimitrescu) crops out as a stripe generally trending west-east and is about 1 km wide. It appears in the source area of the streams running north-west, towards Cuvin and Covăsinţ, and trends ENE–WSW, up to Cioaca Zăcătorii; from there on the stripe develops eastwards, over the upper courses of the Chersca and Cladova Valleys; at the Cladoviţa sources, because of intricate tectonics, the Permian sedimentary deposits crop out on a width of about 3 km.

These deposits are constituted of the following lithologic types: compact, hard black argillites, sandy clays, micaceous on the faces, grey quartzites with interbeds of basic rocks, dark in colour. The thickness of these deposits is of about 300 m. Epidote makes up nests within the rock mass or patches on the bedding faces; Giuşcă mentions the presence of olive biotite, microcline, tourmaline and magnetite as products of thermal contact metamorphism. At the sources of Căsoaia Valley, on Valea Băii, Cibotaru et al., (1975, unpubl. data) point to the existence, in the base of the Permian series, of limy pelitic rocks.

In the Capu Mlatinu area as well as in that of the sources of the Cladoviţa Valley, besides the rocks mentioned above there occur: clays and violaceous sandy clays within which coarser episodes (2–3 mm thick) alternate with fine ones (1–2 mm); violaceous or dark bluish siltstones on whose faces of splitting there are numerous worm traces. The study of the cross-sections from these rocks shows that they were deposited in a lacustrine environment with non-agitated waters. Just like in the Bihor Mts (right slope of Valea Calului, east of Băiţa Plai) and in the Highiş Mts, transitions can be noticed from violaceous rocks (argillites, siltstones, sandstones) with worm traces in the mass or on the bedding faces, to dark bluish or black rocks with the same characteristic features as the vermiculate sandstones. Within the mentioned deposits there appear basic rock veins or basic tuff interbeds, cornified sometimes; they are not thicker than 10 m. The interbedded basic rocks crop out extensively in the area of the Cladoviţa, Cladova, Chersca and Cuvin Valleys.

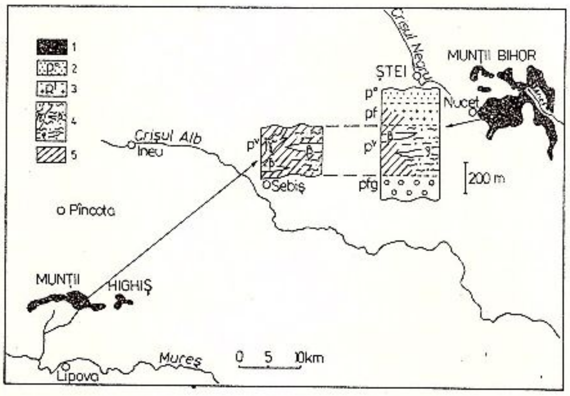
Bleahu (1962), correlating the Permian deposits in the North Apuseni Mountains, established the following order in their sequence: laminated conglomerates in the base, overlain by vermiculate sandstones and then by feldspathic sandstones, to end with oligomictic sandstones. The sequence has been adopted by all the researchers. Subsequently, Bordea, Bordea (1985) show that at the level of the Vermiculate Sandstone Formation, there also occur carbonate rocks interbeds, that are also present in the cornified equivalent of the formation ("Black Series"). The limy pelitic rocks mentioned by Cibotaru et al. in the Highiş Mts. could be placed at the same level with those mentioned by us in the Bihor Mts.

Vermiculate sandstones, well developed at the Arieşul Mare sources, have made the object of the studies carried out by Brustur (1986) that identified the presence of the ichnogenus Planolites. The formation is characterized by micaceous sandstones with numerous bioglyphs that occur all over their width and on the surface of the beds. The galleries, cylindrical or ellipsoidal, are straight or slightly curved, always with smooth surfaces. The same types of galleries also appear in the Permian rocks in the Highiş Mts., but they are constantly ellipsoidal, with compressed section. In the Permian, near the Balaton Lake, worm traces have been mentioned like the ones in the vermiculate sandstones of the Bihor Mts.

The rocks containing *Planolites* are identical with the ones encountered by us in the Highiş Mts. (on the edge south of Capu Mlatinu; at the sources of the Căsoaia Valley – on the first valley south of the Cladoveanu Summit –; on the edge north of Crucea Ţiganului – at about 500 m north of the summit –). In the Cladova Formation the initial features of the vermiculate sandstones have been maintained only where Permian deposits are wider (between Capu Mlatinu and Crucea Ţiganului).

Consequently, the rock pile investigated by us corresponds, in Bleahu's diagram, to the lower half of the Permian sequences, i.e. at the level of vermiculate sandstones. In Bihor and Codru Moma, as well as in Highiş, vermiculate sandstones have interbeds of basic rocks. We should mention that in Highiş, the ignimbritic rhyolites identified in Bihor, are absent. They are substituted by microgranites that thermally altered most of the Vermiculate Sandstone Formation.

As regards the structure of the area under investigation, we are of Balintoni's opinion, according to which the line south of the Cladova Formation ("Vermiculate Sandstone Formation") is a Pre-Alpine overthrust plane; the line concerned, considered to be a fault of Saalic age, was mentioned for the first time by Dimitrescu (1962). We assign the "Vermiculate Sandstone Formation" to the lower part of the Permian and the granites crossing it would be of Upper Permian age. It results that the southern unit was thrust over between the two intervals, i.e. the Saalic phase.



Distribution of the Permian formations in the central area of the Highiş Mts and in the southern area of the Bihor Mts.
Permian: 1, distribution of the formations; 2, Oligomictic Sandstone Formation; 3, Feldspatic Sandstone Formation; 4, Vermicular Sandstone Formation; ρ^1, ignimbritic rhyolites; β, basic rocks; γ, microgranites; 5, black series (cornified rocks).

The "Vermiculate Sandstone Formation" represents the sedimentary rocks of the Păiuşeni Series (Giuşcă, 1962; Dimitrescu, 1962) and is a digitation of the

Highiş Nappe. The exaggerated width of the Permian deposits in the Capu Mlatinu-Crucea Ţiganului area is due to reversed strike faults.

Conclusions

The pile of sedimentary and basic rocks developed between Cuvin and the sources of the Şoimuş Valley is equivalent, in age and lithological constitution, with the "Vermiculate Sandstone Formation" of the Bihor and Codru Moma Mts. The thermally altered areas can be referred to the "Black Series" situated at the level of the Vermiculate Sandstone Formation; on the geological map scale 1 : 50.000 - Biharia Sheet (Bordea et al., 1988) the areas thermally altered are considered similar either to the Feldspath Sandstone Formation or the Vermiculate Sandstone one. We appreciate the other opinions (Balintoni, 1986; Pană, Ricman, 1988) concerning the age of the formation investigated by us in the central-western area of the Highiş Mts as groundless. As for the age of acid magmatites that cross the Vermiculate Sandstone Formation, being insinuated equally on its contact with the metabasics complex in the south, it can be considered to be Post Lower Permian.

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DONNÉES BIOSTRATIGRAPHIQUES SUR LA FORMATION DE CATALOI (DOBROGEA DE NORD, ROUMANIE)

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Key words: Anisian. Norian. Stratigraphic units. Foraminifers. Conodonts. Ostracods. Biostratigraphy. Dobrogea – North Dobrogea – Tulcea and Consul–Niculiţel Zone.

Abstract: *Biostratigraphic Data on the Cataloi Formation.* The paper presents the microfauna (conodonts, foraminifers, ostracods) originating from the deposits attributed to the Cataloi Formation from the type locality (North Dobrogea) as well as from the underlying and overlying formations. Also, the chronostratigraphic value of the conodont, foraminifer and ostracod associations was discussed, while regional and interregional correlations were made.

Historique des recherches

Les dépôts qui affleurent sur la vallée de Teliţa, sur la colline de Biserica (localité de Cataloi) ont attiré l'attention de plusieurs chercheurs scientifiques par leur particularités lithologiques et fauniques.

La faune de bivalves identifiée dans la séquence schisto-argilo-calcaire de Cataloi, comportant *Halobia lommeli, H. moussoni* (in Peters, 1866), *H. fluxa* (in Redlich, 1897), *H. lucana* et *H. insignis* (in Anastasiu, 1898), où on ajoute *Posidonomya cf. alta* (in Kittl, 1908) et *Daonella tyrolensis, D. badiotica, D. pichleri, D. sturi, D. obliqua, Posidonia wengensis* (in Simionescu, 1925) a permis aux auteurs d'attribuer à ces dépôts, considérés comme équivalents des couches de Wengen, l'âge ladinien supérieur.

La faune d'ammonites prélevée de Cataloi comporte les espèces ladiniennes de *Protrachyceras* cf. *archelaus, P.* cf. *pseudoarchelaus* (in Kittl, 1908), *Sageceras haidingeri, Cladiscites diuturnus*, sur la base desquelles Mutihac (1964) détermine l'âge carnien supérieur et norien. À la partie supérieure de la séquence il y a *Rhacophyllites neojurensis* (Mirăuţă in Patrulius et al., 1974, rapport nonpubl.; Grădinaru, 1984), *Steinmannites sp., Helicites geniculatus, Distichites megacanthus* (in Grădinaru, 1984), aussi d'âge norien.

Atanasiu (1940) sépare deux faciès au niveau de la zone à *Daonella lommeli*; le premier, représenté par des calcaires rouges du type Hagighiol et le deuxième, constitué de calcaires à intercalations épaisses de schistes calcaires du type Cataloi. Plus tard, Orest et Elena Mirăuţă (1958, rapport nonpubl.) séparent un autre faciès récifogène et récifal développé dans l'île de Popina et dans la carrière de Zebil.

Baltreş (in Patrulius et al., 1976) décrit les dépôts de Cataloi du point de vue pétrographique et sédimentologique, en considérant la séquence de roches de milieu de bassin, y séparant quatre membres dont il attribue l'âge ladinien-carnien inférieur.

Sous le nom de "Formation des couches de Cataloi", Mirăuţă (in Patrulius et al., 1974, rapport nonpubl.) sépare une entité lithologique ladinien-norienne inférieure, nommée ultérieurement par Grădinaru (1976, 1984) "Formation de Cataloi" dont la partie supérieure, englobant aussi le Norien supérieur, passe graduellement au grès de Frecăţei (Rhétien-Hettangien), individualisé à la base de la Formation de Teliţa.

Sous le même nom, Baltreş (in Baltreş et al., 1984,1987, rapports nonpubl.) décrit de la zone de Somova, des dépôts similaires, d'âge compréhensif (Ladinien supérieur-Norien supérieur), situés au-dessus de la formation des calcaires à cherts en continuité de sédimentation.

Aspects lithologiques et micropaléontologiques

En ce qui suit nous présentons les aspects lithologiques et surtout ceux micropaléontologiques tant des dépots de la Formation de Cataloi, aussi bien que des dépôts du lit et du toit. Nous avons considéré nécessaire de réaliser une série d'analyses micropaléontologiques détaillées en vue d'enrichir le contenu paléontologique connu de la formation, ainsi que pour établir avec plus d'exactitude l'âge des sous-formations y inclues. Ainsi, nous avons obtenu des associations complexes de conodontes, foraminifères, ostracodes, sclérites de holothurides, crinoïdes, microproblématica, dents et écailles de poissons.

Les dépôts attribués à la Formation de Cataloi affleurent dans la colline où est emplacée l'église de la localité de Cataloi, sa base étant ouverte par le ruisseau de Teliţa. Ces dépôts peuvent être suivis aussi en amont de Cataloi, le long des affluents de droit de la vallée de Teliţa, tout comme dans la localité de Frecăţei. Un autre affleurement, présentant seulement des dépôts de la formation des calcaires à cherts est ouvert en aval de Cataloi, aussi le long de la partie droite de la vallée.

La base de la Formation de Cataloi repose en concordance au-dessus des dépôts attribués à la formation des calcaires à cherts que nous allons décrire avec son contenu microfaunique.

Formation des calcaires à cherts (Baltreş, 1987)

La succession des dépôts est formée de calcaires micritiques grisâtre-bleuâtre à concrétions siliceuses jaunâtres et de rares cherts noires qui impriment à la roche un accentué caractère noduleux. Les couches, épaisses de 5 à 50 cm, présentent une surface courbicorticale moulée de pélites verdâtres jusqu'à noirâtres qui forment des intercalations de 1 à 3 cm, très déformées. Plus en haut elles deviennent prédominantes et les calcaires apparaissent tant comme des boudins, irrégulièrement distribués dans la masse pélitique, aussi bien que sous forme de niveaux décimétriques de calcaires noirâtres à silicifications opalescentes noires. Du point de vue microfacial, la roche est formée de micrites à radiolaires calcitisés (fig. 1, pl. VI).

De la séquence susmentionnée, à une épaisseur d'approximativement 80 m, on a prélevé les échantillons 5707 à 5711, 6544 et 6545 qui nous ont fourni des associations à conodontes: *Gondolella bulgarica* (BUDUROV et STEFANOV), *Nicoraella kockeli* (TATGE), *Gladigondolella tethydis* (HUCKRIEDE), *Gl. malayensis budurovi* KOVÁCS et KOZUR, *Neohindeodella acquiramosa* KOZUR et MOSTLER, *N. dropla* (SPASOV et GANEV), *Hindeodella multihamata* HUCKRIEDE, *Hibbardella magnidentata* (TATGE), *Diplododella triassica* (MÜLLER), *Cypriodella mediocris* (TATGE), *Cratognathodus postcrognathus* MOSHER, *Veghella delicatula* (BUDUROV), *Chirodella dinodoides* (TATGE), *Enantiognathus ziegleri* (HUCKRIEDE), *Prioniodina muelleri* (TATGE), *P. latidentata* TATGE, *Ozarkodina tortilis* TATGE; on y ajoute *Kamuellerella (K.) sejmeni* GEDIK de la partie inférieure et *Gondolella hanbulogi* (SUDAR et BUDUROV) de celle supérieure (échantillon 5711).

L'association mentionnée indique l'âge anisien moyen (pelsonien) et pelsonien supérieur pour le dernier échantillon.

Les échantillons 5707 à 5710 comportent, près de conodontes, des associations de foraminifères et ostracodes. Parmi les foraminifères on mentionne: *Tolypammina discoidea* TRIFONOVA, *Earlandinita libera* (TRIFONOVA), fragments de *Nodosaria sp.* et *Dentalina sp.*, "*Vidalina*" *sp.*, des espèces sans importance stratigraphique.

La partie supérieure de la formation des calcaires à cherts (échantillon 5711) comporte une association de conodontes, où près de *Gondolella bulgarica* (sporadique au Pelsonien supérieur, absente ultérieurement) apparaît *G. hanbulogi* (qui débute au Pelsonien supérieur), les deux espèces indiquant avec certitude l'âge pelsonien supérieur.

L'association de foraminifères de cet échantillon est bien représentée surtout par: Nodosariidae (*Dentalina sp.* cf. *D. subsiliqua* FRANKE, *D. leidapoensis* KRISTAN-TOLLMANN, *Pseudonodosaria major* (BORNEMANN), *P. simpsonensis* (TAPPAN), *Lingulina borealis* (TAPPAN), auxquelles on ajoute *Ammobaculites sp. cf. A. parallellus* IRELAND et *Gheorghianina vujisici* (UROSEVIĆ et GAZDZICKI), des espèces qui suggèrent l'âge illyrien.

L'ostracofaune identifiée dans la formation des calcaires à cherts (échantillons 5707 à 5711) se ressemble à celle de type épibathiale, décrite par Kozur (1970) et considérée comme étant caractéristique pour l'Illyrien de Felsöörs (Hongrie). Ce type de faune a été rencontré ultérieurement (Kozur, 1972) à d'autres niveaux aussi (Spathien en Nepal, Trias moyen dans la Tchécoslovaquie, Norien dans les Alpes).

Ces données, coroborées avec celles d'âge offertes par les associations de conodontes ont établi que dans la Dobrogea de N cette ostracofaune de type Felsöörs apparaît dans un intervalle stratigraphique plus large (Pelsonien-Fassanien) sans être strictement caractéristique pour un certain sous-étage.

L'association déterminée dans les échantillons mentionnés comporte: *Acanthoscapha veghae* KOZUR, *A. bogschi* KOZUR, *Acratina triassica* KOZUR, *A. goemoeryi* KOZUR, *H (Healdia) anisica* KOZUR, *H.(H.) felsöörsensis* KOZUR, *Polycopsis levis* KOZUR, *Discoidella hungarica* KOZUR, *Bairdiacypris anisica* KOZUR, *Aglaiocypris aequalis* KOZUR, *Spinocypris vulgaris* KOZUR, *Bairdia anisica* KOZUR, *B. austriaca* KOZUR, *Triebelina kristanae praecursor* KOZUR, *T. nodosa* KOZUR, *Monoceratina sp.*, *Urobairdia sp.*, *Bairdiolites sp.*.

Généralement l'ostracofaune a évolué dans un milieu épibathial et euhalin. La même ostracofaune a été rencontrée aussi dans les échantillons 6544 et 6545, mais plus abondante (acme-zone).

L'association microfaunique renferme aussi des sclérites de holothurides (*Priscopedatus* sp.), crinoïdes, dents de poissons, *"Venerella" stilata* KOZUR et MOSTLER.

Formation de Cataloi (Grădinaru, 1984)

La séquence de calcaires noirs à cherts est suivie d'une lacune d'observation de presque 30 m. Ici, isolement, les dépôts quaternaires laissent lieu à un niveau de calcirudites épais d'approximativement 0,70 m, constituées de clastes arrondies de calcaires gris foncé, claire ou bruns, quelques-unes à filaments, renfermées dans le ciment de calcite sparitique où sont présentes des ammonites (pl. IV, fig. 2). Ce niveau représente un épisode de l'activité d'érosion sous-marine (Baltreş in Patrulius et al., 1976) et nous l'avons attribué à l'Anisien supérieur par corrélation avec les niveaux semblables identifiés dans la zone de Parcheş où, associée aux calcarénites fait son apparition une microfaune illyrienne. La séquence située au-dessus de la lacune est schisteuse en grande partie (sous-formation schisteuse) et comporte des marnes et des calcaires schisteux grisâtres à intercalations subordonnées de calcaires fins, noirs à aspect noduleux parfois, dépourvus de cherts. Les intercalations de calcaires noirs qui recouvrent le niveau de calcirudites dans la partie inférieure de la séquence schisteuse ont fourni (échantillons 5711 bis, 5712A, 5713) des associations de conodontes comportant: *Gondolella hanbulogi* (SUDAR et BUDUROV), *G. constricta* MOSHER et CLARK, *Gladigondolella malayensis budurovi* KOVÁCS et KOZUR et respectivement *G. constricta*, *G.cf. pseudolonga* KOVÁCS, KOZUR et MIETTO, *Enantiognathus ziegleri*, *Neohindeodella triassica riegeli* (MOSHER). La présence de l'espèce *Gondolella hanbulogi* à côté des autres conodontes de type plateforme, ainsi que l'absence de l'espèce *G. bulgarica*, indiquent l'âge illyrien inférieur.

L'association de foraminifères ne diffère pas de celle présente dans la formation sousjacente que par la présence sporadique de *Triadosphaera radiata* KRISTAN-TOLLMANN, espèce considérée comme typique pour le calcaire de Hallstatt, carnien-norien (Kristan-Tollmann, Tollmann, 1983, p. 207). En Roumanie l'espèce a une apparition sporadique, étant rencontrée dans la base du calcaire de Roşia, à Roşia (monts Apuseni) (Ladinien inférieur) (T. Gheorghian in Patrulius et al., 1976; Gheorghian, 1976), au niveau de l'Anisien–Ladinien et du Norien de Rarău, ainsi qu'au niveau du Carnien de Perşani (Mirăuţă, Gheorghian, 1978).

L'ostracofaune de la base de la formation renferme les mêmes éléments que la faune de type Felsöörs: *Acratina goemoeryi* KOZUR, *Nagyella longispinosa* KOZUR, *Healdia (Healdia) anisica* KOZUR, *H. (H.) felsööersensis* KOZUR, *Aglaiocypris aequalis* KOZUR, *Polycopsis levis* KOZUR, *Bairdiolites* sp.

Le calcaire qui a fourni l'échantillon 5713 représente le dernier niveau caractérisé par la présence du "faciès à ostracodes et nodosariidés" (T. Gheorghian in Salaj et al., 1988), rencontré dans la Dobrogea de Nord, du Pelsonien jusqu'au Fassanien inclussivement.

La suite continue par une lacune d'observation, épaisse de presque 50 m, où se développent probablement les dépôts ladinien inférieurs, dans le mêmes faciès schisteux qui est observable aussi plus en haut.

Les affleurements réapparaissent dans la base de la colline sur laquelle est emplacée l'église et sont constitués de marnes grisâtres, verdâtres en altération, où les calcaires micritiques, noirs le plus souvent, forment des intercalations subordonnées. Tant les schists, aussi bien que les calcaires comportent des ammonites (*Protrachyceras archelaus*) et des bivalves (*Daonella lommeli* et lumachelles de bivalves juvéniles) qui ont indiqué l'âge langobardien pour cette séquence schisteuse. Du point de vue microfacial la roche représente un biomicrite à radiolaires calcitisés auxquels on ajoute des filaments, considérés comme caractéristiques pour les calcaires de Hallstatt (Flügel, 1963).

L'échantillon 5714 a été prélevé des calcaires micritiques noirs à rubans bruns intercalés dans la séquence schisteuse, au-dessous du niveau à *Daonella lommeli*. Les échantillons 5715 et 5716 ont été prélevés des calcaires micritiques noirs (pl. IV, fig. 3) et des calcaires marneux grisâtres intercalés dans la séquence schisteuse à ammonites et bivalves. On y a rencontré de rares conodontes: *Gondolella inclinata* KOVÁCS, *G. foliata* (BUDUROV), *Cratognathodus kochi* (HUCKRIEDE), *Gladigondolella* sp., qui indiquent le Langobardien (en l'absence de l'espèce *Gondolella polygnathiformis*).

A l'association de foraminifères connue des niveaux inférieurs on ajoute de nouveaux éléments: *Gheorghianina anae* (GHEORGHIAN), *Ammobaculites tzankovi* (TRIFONOVA), *Turriglomina mesotriasica* (KOEHN-ZANINETTI), spécifiques au Trias supérieur, mais ayant des apparitions meme au Ladinien. L'association micropaléontologique est enrichie par des sclérites de holothurides et de nombreux exemplaires de bivalves juvéniles de type *Posidonia wengensis*.

Les échantillons 5714 à 5716 présentent une association d'ostracodes ladiniens, tels: *Aglaiocypris triassica* KOZUR, *Spinocypris vulgaris* KOZUR, *Triassocypris pusilla* KOZUR, *Ceratobairdia crassa* KRISTAN-TOLLMANN, *Praemacrocypris* sp., *Ogmoconcha tailleuri* SOHN, *Polycopsis* sp., *Triebelina* sp. La présence de l'espèce *Aglaiocypris triassica* indique l'évolution dans un milieu où a existé, à un moment donnée, un apport d'eaux plus douces (pliohalin – Kozur, 1972).

Les lentilles de calcaires micritiques noirs intercalées entre les schistes qui se trouvent au-dessus du niveau à bivalves (5717) ont mis en évidence les premiers exemplaires, en état fragmentaire, de *Gondolella polygnathiformis* BUDUROV et STEFANOV dont le début marque la base du Carnien (Krystyn, 1983). Ils sont accompagnés d'une association de foraminifères à de fréquents exemplaires de *"Vidalina"* sp., *Gheorghianina anae*, *G. vujisici*, *Turriglomina mesotriasica*; y font leur apparition aussi les premiers rovéacrinides.

La sous-formation "pseudoconglomératique" (approximativement 150 m) représente la partie terminale de l'affleurement de la colline de Biserica. Elle comporte des calcaires micritiques grisâtres, à aspect noduleux, englobés dans une matrice schisteuse qui a été inclue diapiriquement parmi les nodules de calcaire à la suite de certains processus de déformation diagénétique (Baltreș in Patrulius et al., 1976).

Les échantillons 5719, 5712, 6547 et 5718, prélevés de cette séquence nous ont fourni de rares fragments de *Gondolella polygnathiformis* BUDUROV et STEFANOV, ainsi que des associations de foraminifères à *"Vidalina"* sp., *Gheorghianina anae* et *G. vujisici*, *Ammobaculites tzankovi*, accompagnés de rovéacrinides, de rares sclérites de holothurides et des exemplaires de bivalves juvénils. L'échantillon 5718, provenant de la partie supérieure nous a offert de plus des exemplaires de *Turriglomina mesotriasica* (KOEHN-ZANINETTI), *Ophthalmidium lucidum* (TRIFONOVA) et *"Glomospira"* sp.

Turriglomina mesotriasica (pl. I, fig. 14; pl. II, fig. 5) caractériserait le Ladinien, bien que la zone déterminée pour le domaine carpatho-balcanique et hellénique soit comprise entre le Bithynien et le Cordévolien inclusif (Salaj et al., 1988); dans la Roumanie cette espèce a été rencontrée seulement au niveau du Ladinien et du Carnien inférieur;

Ophthalmidium lucidum (pl. I, fig. 7–9) est présent au niveau du Norien et du Rhétien, dans les Balcans, les Carpathes Occidentales et les Alpes autrichiens. Dans la Dobrogea elle a des apparitions sporadiques dès le Carnien;

"Glomospira" sp. (pl. II, fig. 8, 9) a été rencontrée dans la Dobrogea, dès le Ladinien, étant plus fréquente au Carnien. Il y a des exemplaires de taille variable qui pourraient être attribués au groupe *G. gemerica-kuthani* (les deux espèces étant déterminées dans des lames minces) en établissant des zones: la première au Ladinien, la deuxième au Carnien (Salaj et al., 1988). Zaninetti (in Courtin et al., 1982) signale la présence de ce groupe au Trias supérieur (Carnien–Norien) de Grèce et souligne l'impossibilité de déterminer les deux espèces entre lesquelles le passage est graduel. Quelques sections minces exécutées par nous sur des exemplaires de Dobrogea, nous ont déterminé de les attribuer au groupe susmentionné.

Les échantillons 5717 à 5721 comportent des associations carniennes d'ostracodes à *Polycope pumicosa schleiferae* KOZUR, en explosion, à côté de: *Spinocypris vulgaris* KOZUR, *Semionella brotzenorum alpina* BUNZA et KOZUR, *Mostlerella nodosa* KOZUR, *M. blumenthali* KOZUR, *Falloticythere* sp., *Ogmoconcha* sp. Y font leur apparition aussi les rovéacrinidés, dents de poissons, *"Venerella" stilata* (Microproblematica).

En amont de la localité de Cataloi, le long de quelques affluents de droit de la vallée de Telița, tout comme dans la localité de Frecăței, après une lacune, affleure la séquence supérieure de la Formation de Cataloi, comportant à la base des calcaires micritiques noirs, fortement diaclasés, à filaments et de nombreux halobies, en alternance avec des pélites jaunâtres (échantillon 5736). Cette séquence pourrait être nommé la sous-formation des calcaires à halobies.

L'association de foraminifères n'est pas significative. Elle comporte *Tolypammina* div. sp., *"Vidalina"* sp. et *Ophthalmidium* sp., près de rares sclérites et ostracodes.

Plus en haut il y a des calcaires marneux grisâtre-jaunâtres à schistosité accentuée, à de nombreuses ammonites (*Cladiscites diuturnus*) et halobies en alternance avec des calcaires micritiques gris- noirâtres, dures (échantillons 5750A, 4374, 1567) (pl. IV, fig. 4, 5). Ces échantillons comportent des associations de conodontes à *Gondolella navicula* HUCKRIEDE, *G. steinbergensis* (MOSHER) et *Metapolygnathus abneptis* HUCKRIEDE, qui indiquent l'âge norien moyen/supérieur. Y apparaissent aussi de riches associations de foraminifères à *Ophthalmidium lucidum* (TRIFONOVA), *Paraophthalmidium* sp., *Textularia jurassica* (GÜMBEL), *Variostoma* sp., *Jaculella stabilis* (BLUMENSTENGEL), *Endothyranella* sp.

Les échantillons 4373 et 5748 proviennent d'un niveau supérieur à calcaires marneux, grisâtres, fossilifères, qui forment des couches de 10 à 50 cm, à caractère noduleux englobées dans des schistes calcaires jaunâtres (pl. IV, fig. 6, 7). Les calcaires comportent une association de foraminifères à *Ophthalmidium lucidum*, *Gheorghianina vujisici*, *G. anae*, *Endothyranella* sp. et *Variostoma crassum* KRISTAN-TOLLMANN. La dernière espèce est considérée comme caractéristique pour le Sévatien (Kristan- Tollmann, 1983).

Des calcaires semblables, fossilifères, à concrétions, englobés dans les schistes marneux grisâtres, à aspect diapir, qui affleurent dans la localité de Frecăței (échantillons 4371, 4371A et B), ainsi que des calcaires noirs spathiques (pl. IV, fig. 8–10) à de petits halobies (où on a identifié *Halobia norica* MOJS.), situés au-dessus des calcaires à concrétions (échantillons 4371C et 5748A) on a obtenu des associations de conodontes, où prédominent *Gondolella steinbergensis* et

Metapolygnathus bidentatus (MOSHER) (pl. I, fig. 4), cette dernière espèce indiquant le Norien supérieur. On y ajoute les espèces de foraminifères: *Ophthalmidium lucidum*, O.sp.5, *Gheorghianina vujisici*, *Textularia jurassica*, *Variostoma crassum*, *Paraophthalmidium* sp., *"Vidalina"* sp. On doit souligner le caractère presque monospécifique (*Ophthalmidium lucidum*) de l'association présente dans l'échantillon 5748A.

Dans l'échantillon 4371A apparaît une ostracofaune norienne, très proche de celle décrite par Kristan-Tollmann (1980) des marnes intercalées entre les calcaires récifaux sévatien-rhétiens d'Isfahan, comportant: *Hiatobairdia subsymmetrica* KRISTAN-TOLLMANN, *Nodobairdia mammilata* KOLLMANN, *Cytherella acuta* URLICHS, *Paracypris cf. redcarensis* (BLAKE), *Triadohealdia ventroplana* KRISTAN-TOLLMANN, *Carinobairdia* sp.

Formation de Teliţa (Grădinaru, 1984)

En continuité de sédimentation on a des calcaires marneux gris foncé, à surface blanchie (échantillon 5735; pl. IV, fig. 11) à de nombreux gastéropodes, ainsi que des calcaires gréseux grisâtre-bruns, noduleux, à pyrite. Il y a aussi des silts calcaires à *Otapiria marshalli alpina* ZAPFE attribués au grès de Frecăţei de la base de la formation de Teliţa, d'âge rhétien (Grădinaru, 1984).

De ces dépôts proviennent des conodontes (*Gondolella steinbergensis*) et foraminifères (*Vidalina"* sp., *Ophthalmidium fusiforme* (TRIFONOVA), *Ophthalmidium* sp.5, auprès de fréquentes *Tolypamminidae*), qui indiquent un âge norien supérieur. De cette association sont absentes les espèces de conodontes caractéristiques au Rhétien.

Au-dessus des dépôts à *Otapiria* reposent des grès quartzeux et silts calcaires gris-claires, à intercalations d'argiles gris foncé à *Schlotheimia angulata* d'âge jurassique inférieur (Hettangien) attribuées aussi au grès de Frecăţei (Grădinaru, 1984). Ceux-ci sont recouverts par des grès quartzeux, argilitiques grisâtre-verdâtres, caractéristiques à la Formation de Poşta.

Conclusions stratigraphiques

Il résulte des données présentées que les dépôts attribués à la Formation de Cataloi, qui affleure le long de la rive droite de la vallée de Teliţa, entre Cataloi (au S) et Frecăţei (au N), forment une unité lithologique susceptible d'etre séparée en trois sousformations: schisteuse, "pseudoconglomératique" et celle des calcaires àhalobies. Elle est délimitée à la base par des dépôts de la formation des calcaires à cherts (ici d'âge anisien moyen) et à la partie supérieure par les dépôts de la Formation de Teliţa (Rhétien–Jurassique). La limite inférieure de la Formation de Cataloi est recouverte par des dépôts quaternaires; celle supérieure est visible, permettant l'observation de la transition lithologique graduée, de dépôts calcaires surtout à la partie terminale de la formation vers ceux détritiques de la formation du toit.

L'âge des dépôts de la Formation de Cataloi, argumenté micropaléontologiquement est compréhensif et inclut l'intervalle de temps dès l'Anisien supérieur jusqu'à la base du Norien supérieur inclusivement. De la succession décrite sont absents les éléments micropaléontologiques caractéristiques au Ladinien inférieur dont les dépôts sont recouverts par loess et sol. La présence de quelques espèces de conodontes et foraminifères norien supérieurs dans les dépôts attribués au Rhétien sur bases macrofauniques, indique une continuité de sédimentation entre les deux formations et permet aussi l'inclusion dans la base du Rhétien, d'une partie du Sévatien (Urlichs, 1973; Tollmann, 1976, 1978).

Le changement du milieu de sédimentation, de hémipélagique-épibathial, partielemment subeuxinique, propice aux faunes benthiques, au faciès de bassin, où prédominent des eaux tranquilles mais relativement impropres aux conditions de vie, est marqué par un changement possitif, qualitatif et quantitatif dans la microfaune de foraminifères et négatif dans celles de conodontes et ostracodes qui deviennent plus rares. Elles acquieront un rôle plus important à peine au Norien.

La microfaune de foraminifères présente les suivantes caractéristiques le long de la colonne stratigraphique de la Formation de Cataloi:

– le remplacement gradué du faciès à nodosariidés et ostracodes (qui évolue du Pelsonien jusqu'au Ladinien inférieur inclusivement) avec le faciès à miliolides dont les représentants apparaissent dès le Pelsonien et prédominent vers la partie supérieure de la coupe (Norien–Rhétien); les deux microfaciès coexistent dans la partie inférieure de la Formation de Cataloi;

– L'espèce *Turriglomina mesotriasica* a été rencontrée dans la coupe de Cataloi, dans les dépôts inférieurs et supérieurs au niveau à Daonella lommeli, attribués au Ladinien supérieur et respectivement au Carnien inférieur, mais aussi dans ceux considérés comme Carnien supérieurs, à *Ophthalmidium lucidum*;

– l'espèce *Variostoma crassum*, considérée caractéristique pour le Norien (Zaninetti, 1976; Kristan-Tollmann, 1983) apparaît dans la coupe de la vallée de Teliţa, en association avec les miliolides. Elle a été retrouvée, dans une même association, dans les calcaires noirs à halobies et ceux grisatre-brunâtres à *Monotis haueri* intercalés dans une séquence prédominante schisteuse qui se développe dans les collines de Coasta lui Nicu et Caracuş (au nord de la localité d'Izvoarele) tandis que les calcaires à halobies et respectivement *Monotis haueri*, des séquences calcaires qui affleurent dans la colline de Muchea Verde (à l'ouest de la localité de Poşta), ainsi qu'au sud de la localité de Rîndunica, ont fourni, à ce niveau, seulement l'espèce *Variostoma crassum*, citée aussi du faciès de Hallstatt (Salzkammergut et Hohe Wand, Autriche).

Auprès des caractéristiques de la microfaune, on doit souligner aussi l'installation précoce du faciès schisteux dans la zone étudiée, dès la partie inférieure de l'Anisien supérieur. La même particularité a été décélée dans la séquence schisteuse traversée par le forage 69802 Isaccea, où a été déterminée la présence du Carnien, du Ladinien inférieur et de l'Anisien supérieur (Mirăuţă, Gheorghian in Baltreş et al., 1989, rapport nonpubl.), succession qui parfait micropaléontologiquement la coupe de la localité type. La ressemblance lithologique entre les deux entités a facilité leur corrélation le long d'une zone dépréssionnaire située en face des affleurements des basaltes de Sarica-Niculiţel et développée entre Isaccea et Cataloi (Baltreş in Baltreş et al., 1989, rapport nonpubl.).

Les variations latérales de faciès de la Formation de Cataloi indiquent la grande diversité, dans un espace restreint, des milieux de sédimentation. Ainsi, dans l'aréal de Somova, au-dessus des dépôts attribués à la formation des calcaires à cherts dont l'âge monte probablement jusqu'au Ladinien inférieur, repose une séquence schisteuse, ladinienne supérieure. Elle réapparaît à peine à la partie supérieure du Carnien, après une période prédominée par la sédimentation calcaire du Carnien inférieur et moyen.

Dans la partie SE de la zone de Tulcea, au niveau de la partie inférieure et moyenne de la formation de Cataloi, se développent les calcaires rouges, fossilifères du faciès de Hallstatt, qui ont fourni de riches associations de conodontes et foraminifères agglutinants, sclérites de holothurides et rovéacrinides.

Plus vers l'est, vers la terre ferme qui occupait avant le Crétacé supérieur le domaine où se trouve à présent la Mer Noire, on a reconnu, au niveau du Ladinien supérieur–Carnien inférieur, le faciès de Wetterstein, interprété comme des accumulations en "offshore banks", au-dessous de la base des ondes, qui comporte surtout des miliolidés et des agglutinants (Baltreş, Gheorghian in Baltreş et al., 1980).

L'uniformité des faciès au niveau du Carnien supérieur est suggérée par la microfaune de foraminifères représentées exclussivement par des miliolidés.

Au niveau du Norien, on sépare de nouveau deux faciès, le premier plus profond, caractérisé par la présence des miliolidés et des agglutinants, le deuxième renfermant seulement les agglutinants cités du faciès de Hallstatt.

Les données micropaléontologiques et microfaciaux susmentionnées permettent d'obtenir l'image d'un bassin de dépôt situé sur un paléorelief paléozoïque qui a facilité la formation de petits bassins séparés par des seuils. L'évolution du processus de sédimentation reflète l'influence de certains mouvements tectoniques épirogéniques qui débutent dès l'Anisien inférieur, mais n'ont pas une action uniforme et non plus la même intensité dans tout le bassin. La zone investiguée se situe dans un domaine plus bas du point de vue structural, par raport à celle située à l'est et de laquelle est séparée par l'anticlinorium d'Uzum Bair (O. Mirăuţă, 1966).

La continuité de sédimentation entre le Trias et le Jurassique suggère l'emplacement de la zone étudiée dans la proximité du centre du bassin, trouvé autrefois plus vers l'ouest.

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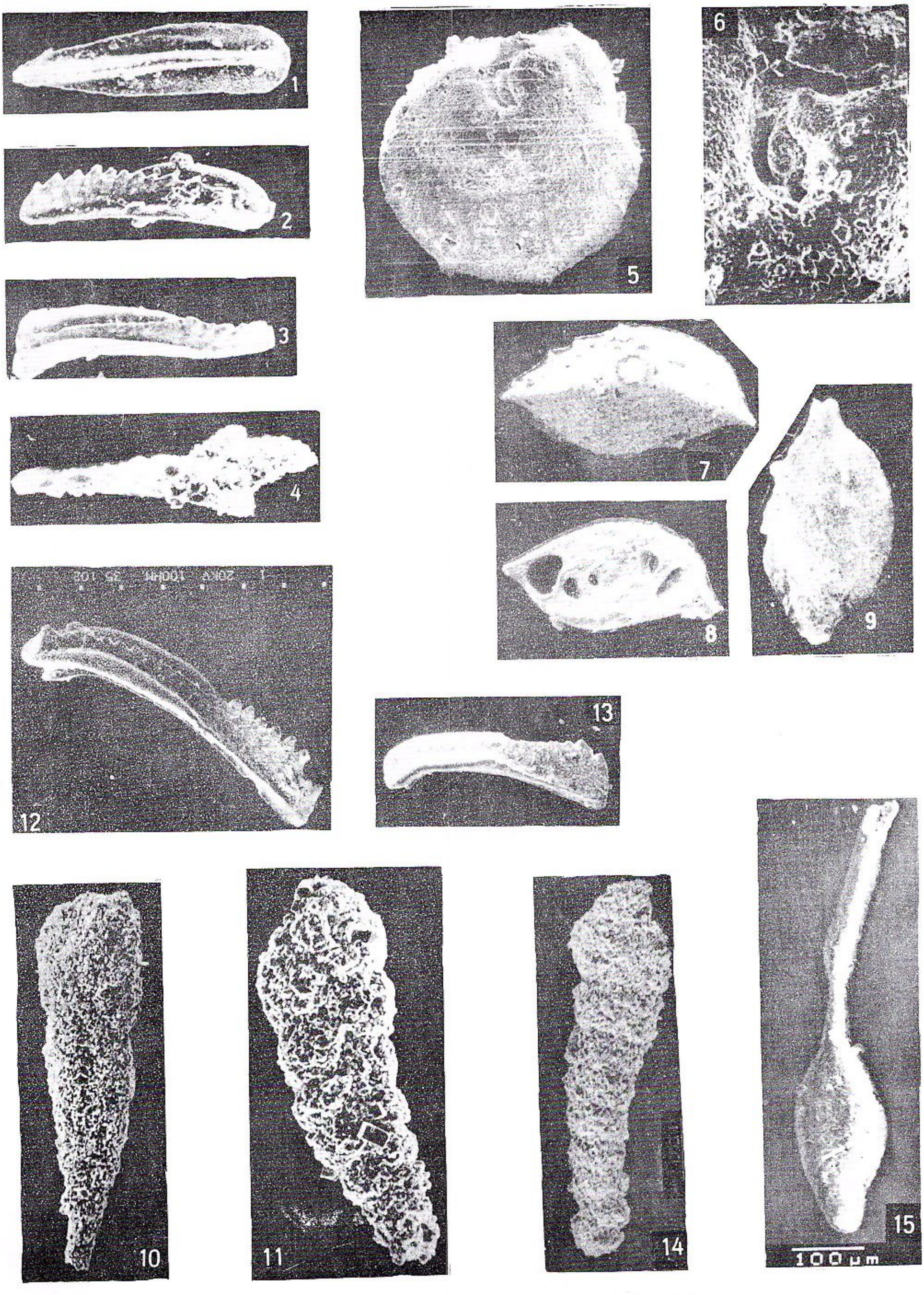
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Planche I

Figs. 1, 12 – *Gondolella foliata* (BUDUROV) – Cataloi.
Fig. 2 – *Gondolella navicula* HUCKRIEDE – éch. 1567, vallée de Teliţa.
Fig. 3 – *Gondolella steinbergensis* (MOSHER) – éch. 4371, Frecăţei.
Fig. 4 – *Metapolygnathus bidentatus* (MOSHER) – éch. 4371, Frecăţei.
Figs. 5, 6 – *Variostoma crassum* KRISTAN-TOLLMANN – diam. 0,52 mm; éch. 4371.
Figs. 7–9 – *Ophthalmidium lucidum* (TRIFONOVA) – fig. 7, vue de l'aperture, largeur 0,49 mm; fig. 8, section transversale, largeur 0,52 mm, grosseur 0,35 mm; fig. 9, longueur 0,70 mm, largeur 0,42 mm; éch. 5748.
Fig. 10 – *Jaculella stabilis* (BLUMENSTENGEL) – longueur 1,43 mm, largeur 0,38 mm; éch. 5750 A.
Fig. 11 – *Textularia jurassica* (GÜMBEL) – longueur 0,82 mm, largeur 0,26 mm; éch. 4374.
Fig. 13 – *Gondolella polygnathiformis* (BUDUROV, STEFANOV) – Cataloi.
Fig. 14 – *Turriglomina mesotriasica* KOEHN-ZANINETTI – éch. 5717, Cataloi.
Fig. 15 – ?*Ophthalmidium fusiformis* (TRIFONOVA) – éch. 5750 A.

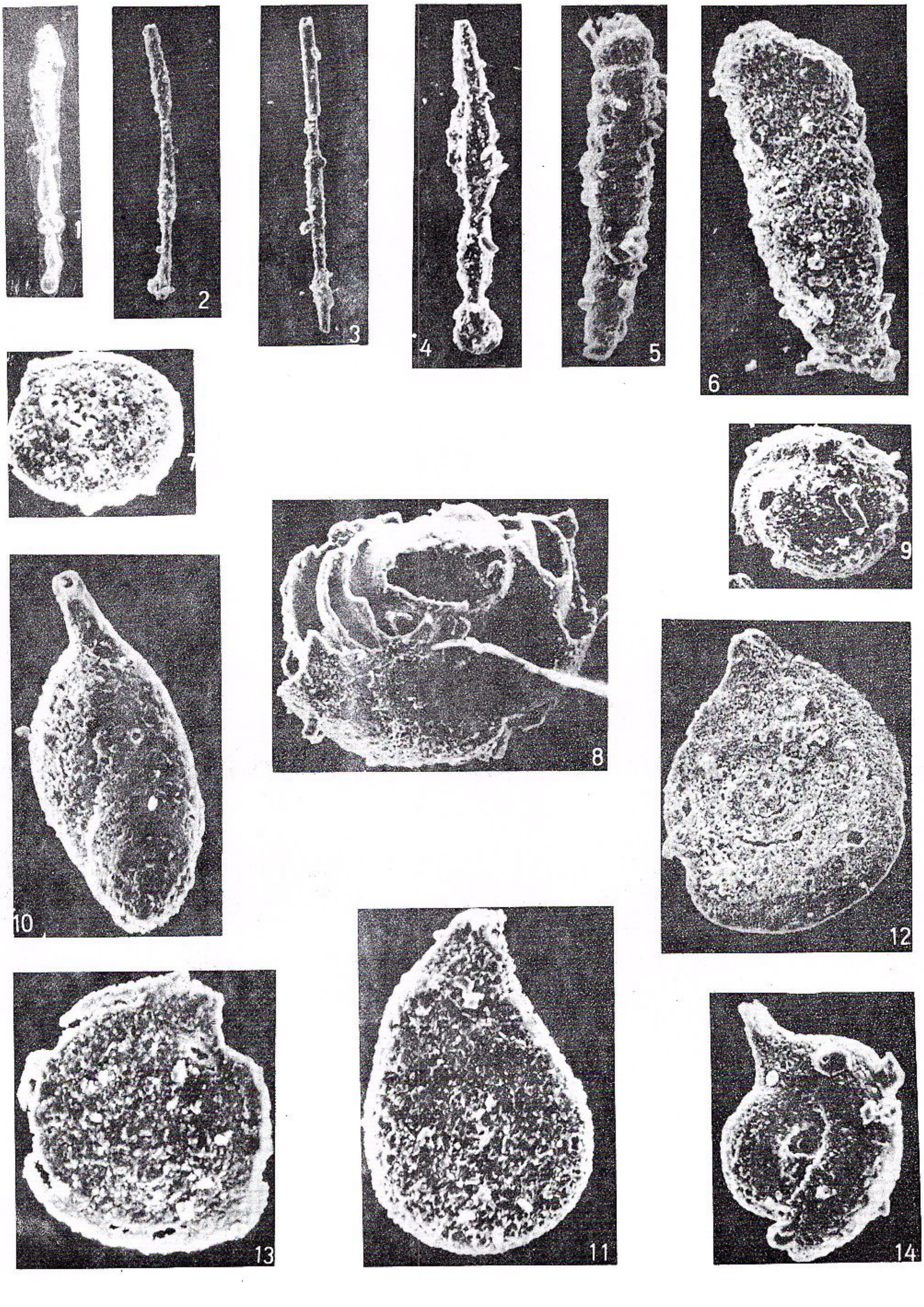
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Planche II

Fig. 1 – *Gheorghianina vujisici* (UROSEVIČ et GAZDZICKI) – longueur 0,75 mm, grosseur 0,08 mm; éch. 5715.
Figs. 2, 3 – *Gheorghianina* sp. (filiforme) – longueur 0,70 et 1,45 mm, grosseur 0,08 mm; éch. 5717.
Fig. 4 – *Gheorghianina anae* (GHEORGHIAN) – longueur 0,73 mm, grosseur 0,12 mm; éch. 5717.
Fig. 5 – *Turriglomina mesotriasica* KOEHN-ZANINETTI – longueur 0,41 mm, grosseur 0,12 mm; éch. 5717.
Fig. 6 – *Gaudryina triassica* TRIFONOVA – longueur 0,61 mm, largeur 0,21 mm; éch. 5717.
Fig. 7 – "*Vidalina*" sp. – diam. 0,19; éch. 5717.
Figs. 8, 9 – "*Glomospira*" sp. (ex. gr. gemerica-kuthani ?) – fig. 8, diam. 0,35 mm; fig. 9, diam. 0,52 mm; éch. 5717.
Fig. 10 – *Ophthalmidium fusiforme* (TRIFONOVA) – longueur 0,40 mm, largeur 0,17 mm; éch. 5748.
Fig. 11 – *Ophthalmidium* sp. 5 – diam. 0,38/0,28 mm; éch. 5748.
Fig. 12 – *Ophthalmidium* sp. 1 – diam. 0,82/0,40 mm; éch. 5748.
Fig. 13 – *Ophthalmidium* sp. – diam. 0,26 mm; éch. 5748.
Fig. 14 – *Ophthalmidium* sp. – diam. 0,29/0,21 mm; éch. 5748.

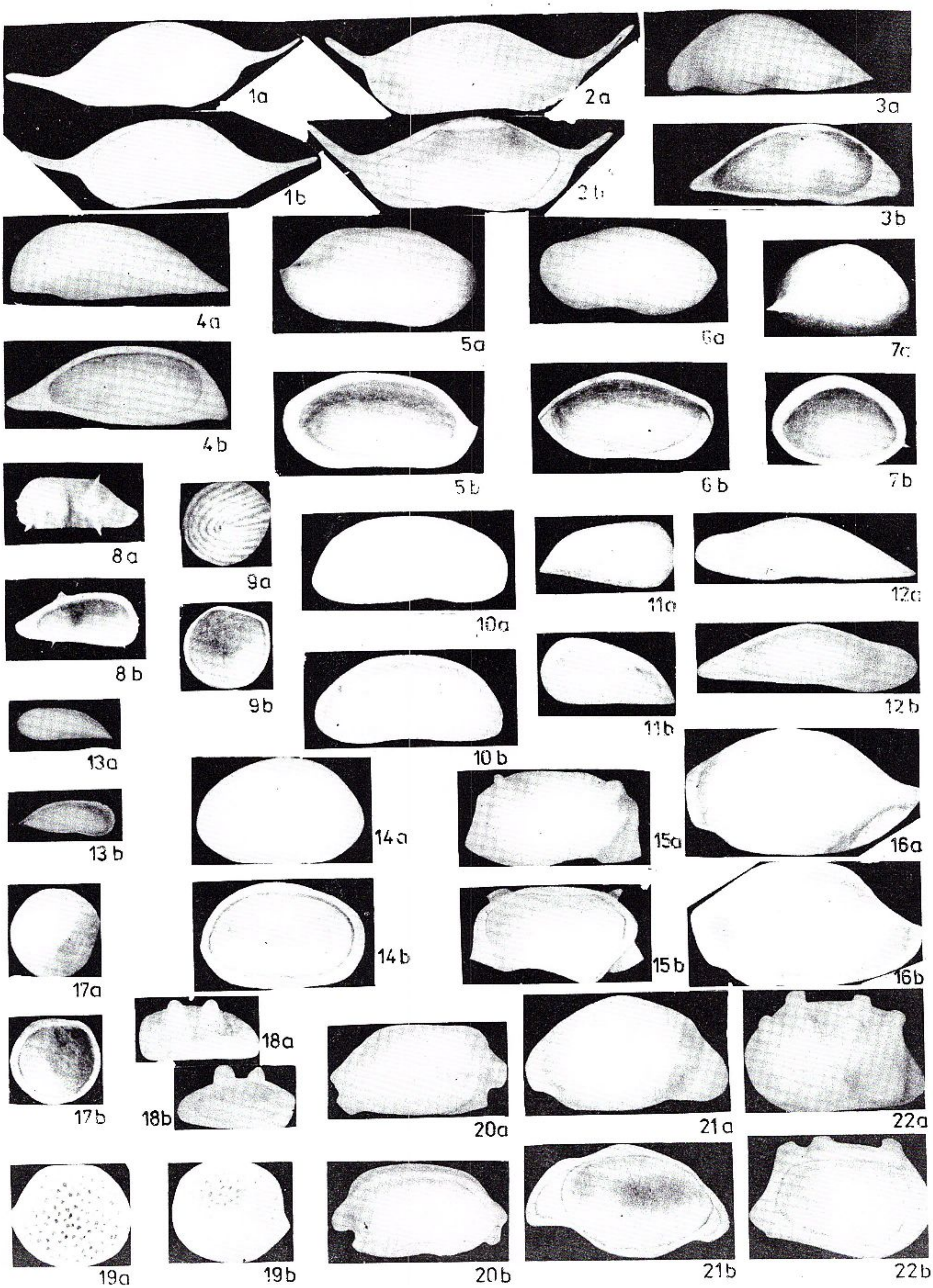
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Planche III

Fig. 1 a, b – *Acantoscapha veghae* KOZUR (x 40) –éch. 5707, Cataloi.
Fig. 2 a, b – *Acantoscapha bogschi* KOZUR (x 40) – éch. 5707, Cataloi.
Fig. 3 a, b – *Acratina triassica* KOZUR (x 40) – éch. 5707, Cataloi.
Fig. 4 a, b – *Acratina goemoeryi* KOZUR (x 40) – éch. 5707, Cataloi.
Fig. 5 a, b – *Bairdia anisica* KOZUR (x 40) – éch. 5708, Cataloi.
Fig. 6 a, b – *Bairdiacypris anisica* KOZUR (x 40) – éch. 5707, Cataloi.
Fig. 7 a, b – *Healdia (Healdia) felsöörsensis* KOZUR (x 40) – éch. 5708, Cataloi.
Fig. 8 a, b – *Nagyella longispinosa* KOZUR (x 40) – éch. 6545, Cataloi.
Fig. 9 a, b – *Discoïdella hungarica* KOZUR (x 40) – éch. 5710, Cataloi.
Fig. 10 a, b – *Aglaiocypris triassica* KOZUR (x 40) – éch. 5714, Cataloi.
Fig. 11 a, b – *Spinocypris vulgaris* KOZUR (x 40) – éch. 5714, Cataloi.
Fig. 12 a, b – *Praemacrocypris* sp. (x 40) – éch. 5715, Cataloi.
Fig. 13 a, b – *Triassocypris pusilla* KOZUR (x 40) –éch. 5714, Cataloi.
Fig. 14 a, b – *Ogmoconcha tailleuri* SOHN (x 40) – éch. 5715, Cataloi.
Fig. 15 a, b – *Ceratobairdia crassa* KRISTAN-TOLLMANN (x 40) – éch. 5715, Cataloi.
Fig. 16 a, b – *Tricbelina* sp. (x 40) – éch. 5716, Cataloi.
Fig. 17 a, b – *Polycopsis* sp. (x 40) – éch. 5716, Cataloi.
Fig. 18 a, b – *Mostlerella nodosa* KOZUR (x 40) – éch. 5718, Cataloi.
Fig. 19 a, b – *Polycope pumicosa schleiferae* BUNZA et KOZUR (x 40) – éch. 5718, Cataloi.
Fig. 20 a, b – *Carinobairdia* sp. (x 40) – éch. 4371, Frecăţei.
Fig. 21 a, b – *Hiatobairdia subsymmetrica* KRISTAN-TOLLMANN (x 40) – éch. 4371, Frecăţei.
Fig. 22 a, b – *Nodobairdia mammilata* KOLLMANN (x 40) – éch. 4371, Frecăţei.

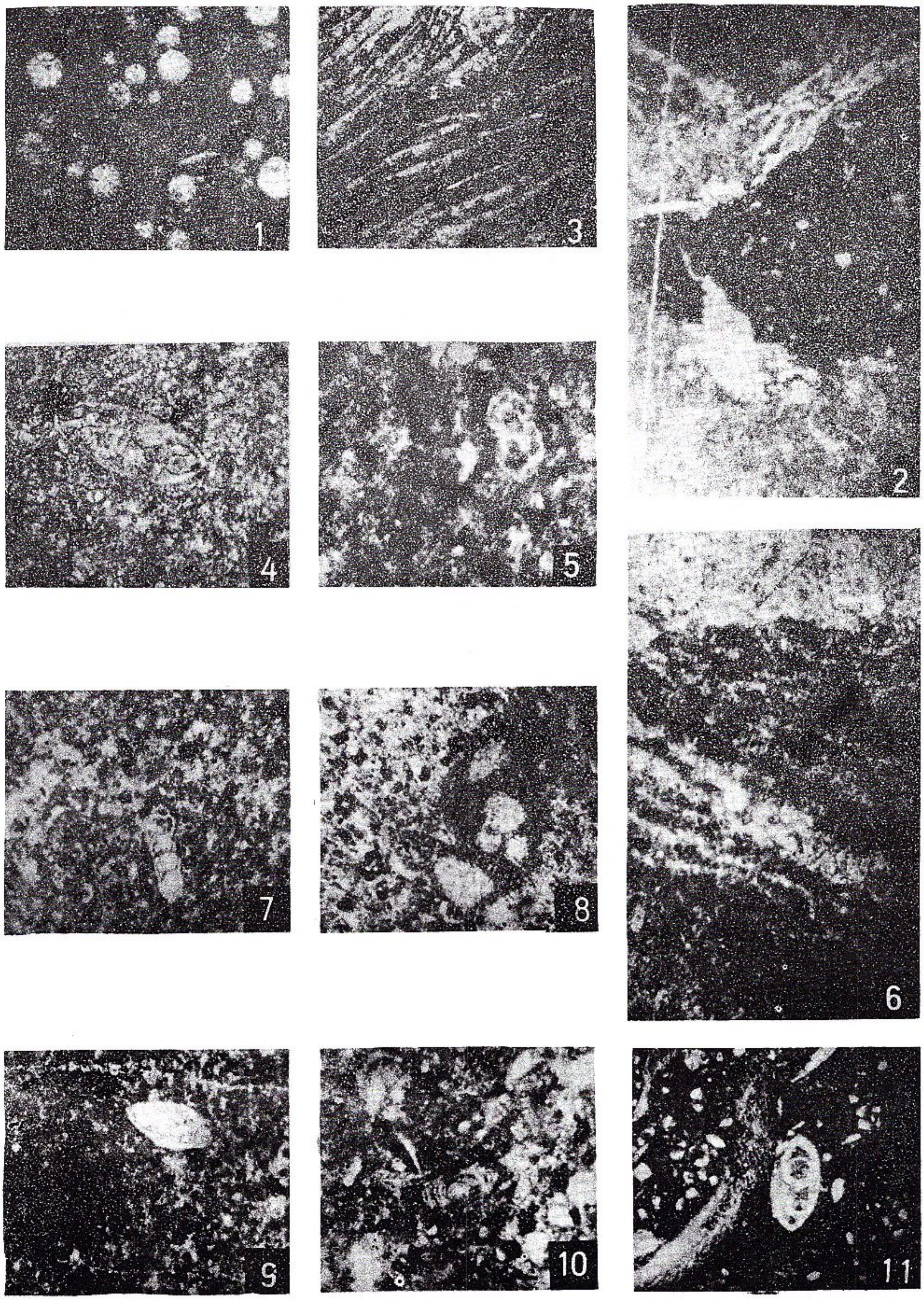
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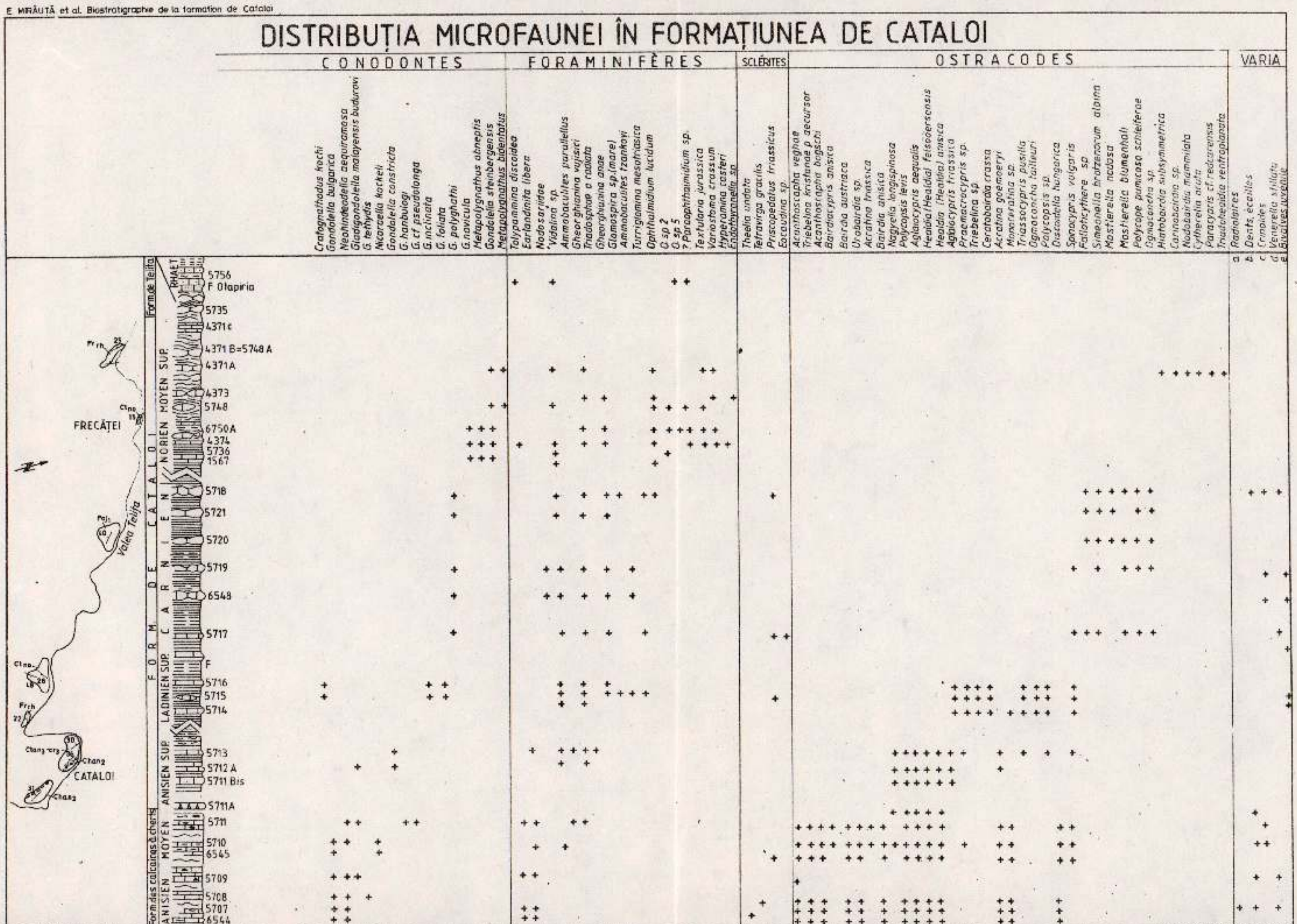
Planche IV

Fig. 1 – Micrite à radiolaires calcitisés. Cataloi, éch. 5710.
Fig. 2 – Calcirudites à clastes de micrite à radiolaires calcitisés dans un ciment microsparitique à filaments (coquine). Cataloi, éch. 5712.
Fig. 3 – Microsparite à filaments (coquilles de bivalves pélagiques). La vallée de Teliţa, éch. 1567.
Fig. 4 – Microsparite à ? *Ophthalmidium lucidum*. La vallée de Teliţa, éch. 4373.
Fig. 5 – Calcisiltite à foraminifères non-déterminés. La vallée de Teliţa, éch. 5750.
Fig. 6 – Micrite pélétale à lamelles de filaments. Cataloi, éch. 5716.
Fig. 7 – Calcisiltite à *Nodosaria* sp. La vallée de Teliţa, éch. 5748.
Fig. 8 – Calcilutite à ? *Trochammina* sp. Frecăţei, éch. 4371 A.
Fig. 9 – Biomicrite à *Lenticulina* sp. et des fissures fines parallèles remplies avec calcite. La vallée de Teliţa, éch. 4373.
Fig. 10 – Calcisiltite à miliolides. La vallée de Teliţa, éch. 5735 A.
Fig. 11 – Grès argileux fins à *Lenticulina* sp. La vallée de Teliţa, éch. 1566.



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NOUVELLES DONNÉES BIOSTRATIGRAPHIQUES CONCERNANT LES DÉPÔTS TRIASIQUES DE SASCA MONTANĂ (ZONE DE REŞIŢA–MOLDOVA NOUĂ, BANAT)

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Key words: Triassic. Anisian. Ladinian. Limestone. Foraminifers. Conodonts. Sclerites. Biostratigraphy. South Carpathians – Sedimentary Getic and Supragetic Domains – Reşiţa–Moldova Nouă Zone.

Abstract: *New Biostratigraphic Data on the Triassic Deposits at Sasca (Reşiţa–Moldova Nouă Zone, Banat).* The dolomitized limestones and the bituminous ones, cropping out in the Sasca Montană zone (Banat), that have been assigned to the Anisian, have yielded a varied microfauna of conodonts, foraminifera, holothurian sclerites etc., indicating the Middle Anisian age for the lower horizon and the Upper Anisian – Lower Ladinian one for the upper horizon.

Dans une note publiée en 1975 nous avons signalé (D. Gheorghian) une microfaune rare et peu variée, identifiée dans les échantillons prélevés de deux horizons calcaires attribués au Trias, qui affleurent dans la selle de la colline Redut (le versant méridional de la vallée de la Nera, entre les localités de Sasca Română et Sasca Montană).

L'horizon inférieur, formé de calcaires dolomitiques blanc-jaunâtre ou gris, à brachiopodes et crinoïdes, situés au-dessus des conglomérats werféniens, a été attribué (Halavàts, Schreter, 1916; Codarcea, 1940), au Werfénien supérieur ? - Anisien.

Dans les lames minces exécutées dans ces calcaires on a identifié *Glomospira densa* (PANTIC) et *Meandrospira dinarica* KOCHANSKY-DEVIDÉ et PANTIC, que nous considérions comme "appartenant à l'Anisien moyen (peut-être aussi inférieur)" (Gheorghian, 1975, p. 52).

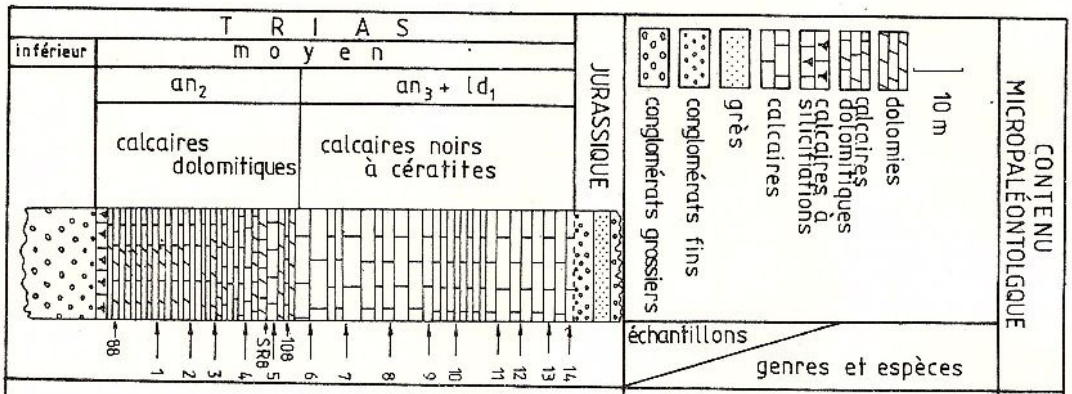
L'horizon supérieur, représenté par des calcaires marneux bitumineux, noirs, à diaclases de calcite, nommé "l'horizon des calcaires noirs à cératites", ou "l'horizon à Trinodosus" (Halavàts, Schreter, 1916), a été attribué aussi à l'Anisien. Le même âge est attesté aussi par la macrofaune identifiée par Năstăseanu (1964) et Boldur et al. (1964).

La présence d'un sclérite dans une lame mince de ces calcaires (*Theelia* aff. *T. planorbicula*) nous a déterminé de reprendre l'échantillonage de ce profil afin d'aborder l'étude de la microfaune détachée (à l'acide acétique) que nous pensions de beaucoup plus riche et variée. Pour cela on a prélevé 14 échantillons, les premiers 5 (SR 1-SR 5) des calcaires dolomitiques gris-jaunâtres (en commençant de 25 m vers le NO des bancs conglomérats werféniens du chemin qui lie Sasca Montană de Sasca Română), les autres étant des calcaires gris-noirâtres du deuxième horizon.

Un seul échantillon de ceux prélevés de l'horizon inférieur a fourni une microfaune détachée, représentée par de très fréquents moulages internes de microgastéropodes et de rares sclérites de holothurides (*Calclamnella regularis* STEFANOV et *Tetravirga* sp.) et dans une section mince (SR 5) on a retrouvé *Meandrospira dinarica*.

Les calcaires noirs à cératites de l'horizon supérieur se sont prouvés de beaucoup plus riches du point de vue microfaunique. Les échantillons SR 6–SR 14, collectés de la crête de la colline Redut, contiennent une microfaune variée, formée de nombreux sclérites de holothurides, des foraminifères et conodontes sporadiques, ensemble avec *Uvanogelia incurvata* KOZUR et MOSTLER et des "vertèbres" d'ophiurides, des pédicellaires d'équinidés, des écailles de poissons placoïdes. Il faut remarquer que toute la microfaune est très diagénisée.

Quant aux foraminifères, on remarque la présence, dans tous les échantillons, de l'espèce *Ammobaculites* sp. cf. *A. parallellus* IRELAND, que nous avons rencontrée jusqu'à présent dans notre pays seulement dans l'Anisien supérieur de la Dobrogea, des Monts Apuseni (l'Autochtone de Bihor) et des Carpathes (Cristian). Il faut remarquer que, tout comme dans les Monts Apuseni, y prédominnent les exemplaires



Genres et espèces	Groupe	88	1	2	3	4	SR8	5	108	6	7	8	9	10	11	12	13	14
Gondolella cornuta	CONODONTES									+		+						
Gondolella constricta										+	+	+			+	+		
Gondolella bifurcata										+								
Enantiognathus ziegleri										+	+	+			+			+
Prioniodina muelleri											+							
Hindeodella suevica												+						
Hibardella magnidentata												+						
Gondolella cf. pseudolonga															+			
Ozarcodina tortilis															+			
Gondolella balcanica																+	+	
Gondolella transita																		+
Gondolella bakalovi																		+
Gondolella excelsa																		+
Meandrospira dinarica	FORAMINIFÈRES	+	+				+	+	+									
Glomospira densa		+	+				+		+									
Ammobaculites cf. A. parallellus										+	+	+	+	+	+	+	+	
Glomospirella cf. G. spirillinoides										+								
Diplotreminina sp.										+								
Ammodiscus sp.										+	+							
Nodosariidae										+		+			+			
Turriglomina mesotriasica												+				+		
Earlandinita sp.													+	+	+	+	+	
Calclamnella regularis	HOLOTHURIDAE					+												
Priscopedatus triassicus										+		+		+	+	+	+	+
Priscopedatus acanticus										+		+			+			
Priscopedatus tyrolensis										+		+			+			
Tetravirga perforata										+		+			+			
Tetravirga sp.						+												
Kuehnites acanthotheeloides										+								
Staurocumites bartensteini										+								
Theelia immisorbicula										+		+			+			
Fissobractites sp.										+		+			+			
Priscopedus horridus										+		+			+			
Theelia planorbicula												+		+	+	+		
Theelia sp. 5												+						
Priscopedatus multiperforatus															+			
Achistrum monochordata															+	+		
Theelia monicae															+	+		
Acanthotheelia spinosa															+	+	+	
Theelia patinaformis															+			
Microgastéropodes-moulages internes	VARIA					+				+								
Ecailles de poissons										+	+	+	+				+	
Uvanogelia incurvata Ophiuridae										+					+	+	+	
„Vertèbres"														+	+		+	
Pedicellaires - équinidés															+	+	+	

Distribution de la microfaune dans les calcaires triasiques de Sasca Montană jeunes représentés seulement par la partie enroulée, qui prouvent en transparence une structure interne identique à celle des spécimens adultes (Gheorghian in Salaj et al., 1988).

À Sasca, cette espèce apparaît en association avec *Glomospirella* sp. cf. *G. spirillinoides* (GROZDILOVA et GLEBOVSKAJA), *Earlandinita* sp., *Ammodiscus* sp., des moulages internes de *Lenticulina*, *Frondicularia* et *Diplotremmina* sp. Il faut mentionner que dans deux lames minces (SR 8 et SR 12) on a identifié aussi des fragments de *Turriglomina mesotriasica* (KOEHN-ZANINETTI), une espèce qui marque une zone dans l'aire carpatho-balkanique pour l'intervalle Illyrien-Cordévolien (Salaj et al., 1988a).

Plus ou moins abondants, les sclérites de holothurides sont présents dans tous les échantillons des calcaires noirs. On a identifié les espèces suivantes: *Tetravirga* sp., *Praeeuphronides* sp., *Priscopedatus triassicus* MOSTLER, *P. multiperforatus* MOSTLER, *P. staurocumitoides* MOSTLER, *P. tyrolensis* MOSTLER, *P. horridus* (MOSTLER), *P.? acanthicus* MOSTLER, *Staurocumites bartensteini* DEFLANDRE-RIGAUD, *Achistrum* sp., *Theelia monicae* MOSTLER et RAHIMI-YAZD, *Th. patinaformis* MOSTLER, *Th. immisorbicula* MOSTLER, *Th. planorbicula* MOSTLER, *Th. multiradiata* KOZUR, *Kuehnites acanthotheeloides* (MOSTLER), *Acanthotheelia spinosa* FRIZZEL et EXLINE. De celles-ci, les espèces *Priscopedatus multiperforatus*, *P. horridus* et *P. tyrolensis* ont été mentionnées dans la littérature seulement pour l'Anisien supérieur, toutes les autres espèces ayant une distribution stratigraphique plus large.

Les associations de conodontes fournies par les calcaires bitumineux de l'horizon supérieur sont assez réduites quantitativement, mais assez significatives en ce qui concerne la valeur chronostratigraphique des espèces. Ainsi, dans l'échantillon prélevé de la base de l'horizon mentionné (SR 6), on a identifié les espèces *Gondolella bifurcata* (BUDUROV et STEFANOV), *G. cornuta* (BUDUROV et STEFANOV), *G. constricta* MOSHER et CLARK, *Enantiognathus ziegleri* (DIEBEL). La présence dans l'association de la première espèce qui monte jusqu'à la base de l'Anisien supérieur, ensemble avec des espèces avec une distribution plus large, permet la précision de l'âge illyrien inférieur.

Les échantillons suivants contiennent des conodontes sporadiques appartenant aux espèces *Gondolella cornuta* et *G. constricta* (qui ne dépasse pas le Fassanien inférieur), avec *Hindeodella suevica* (TATGE), *Enantiognathus ziegleri* (DIEBEL), *Hibbardella magnidentata* (TATGE).

Dans l'échantillon SR 11, avec *Gondolella constricta* il y a aussi l'espèce *G.* cf. *pseudolonga* KOVÀCS, KOZUR, MIETTO, restreinte à la base du Ladinien.

Les échantillons au-dessus de ce niveau (SR 12, SR 13) contiennent l'espèce *Gondolella balkanica* (BUDUROV et STEFANOV), évoluant dans l'intervalle Illyrien supérieur-Fassanien.

Les nouveaux éléments d'âge réapparaissent dans l'échantillon SR 14, qui contient une association de conodontes avec les espèces *Gondolella bakalovi* BUDUROV et STEFANOV, *G. transita* KOZUR et MOSTLER et *G. excelsa* (MOSHER). Les premières deux espèces sont caractéristiques pour le Fassanien supérieur.

Les éléments micropaléontologiques mentionnés ci-dessus complètent les données fournies par les associations macrofauniques et permettent des précisions chronostratigraphiques de détail. Ainsi les espèces de foraminifères *Glomospira densa* et *Meandrospira dinarica* confirment la présence, dans l'horizon inférieur, de l'Anisien moyen (Pelsonien), attesté d'ailleurs par l'espèce de brachiopodes *Decurtella decurtata* (GIRARD), un fossile de zone pour le Pelsonien (Halavàth et Schreter, 1916; Codarcea, 1940).

L'horizon supérieur a un âge compréhensif, en incluant l'Anisien supérieur (démontré aussi par la macrofaune) et le Ladinien inférieur.

L'association de conodontes fournie par les calcaires de la base de l'horizon supérieur a précisé la présence de la partie inférieure de l'Anisien supérieur. Pour l'Illyrien plaide aussi l'espèce de foraminifères *Ammobaculites sp.* cf. *A. parallellus*, identifiée au même niveau dans plusieurs régions du pays, tout comme certaines espèces de sclérites, signalées dans la littérature seulement à ce niveau.

Un élément nouveau est, dans cet horizon, la présence du Ladinien inférieur, prouvée par les associations de conodontes. Des espèces identifiées, une importance à part présente l'espèce *Gondolella balkanica*, mentionnée dans la "province balkanique de conodontes" (les Monts Balkans d'ouest – Budurov et Stefanov, 1975) et dans les Alpes du sud (Préalpes Vincentines) et la zone Recoaro, dans l'Italie du NE (Mietto, Petroni, 1979, 1980); – l'espèce *Gondolella bakalovi*, citée des Monts Balkans de l'ouest (Budurov et Stefanov, 1972) et les Monts Apuseni (nappe d'Arieşeni, Kozur et Mirăuţă, 1980); – l'espèce *Gondolella transita* identifiée dans l'Hongrie (Felsöörs, 1977) et les Carpathes de l'ouest, en Roumanie (les Carpathes Orientales – Perşani et la vallée de la Sărăţii, Mirăuţă, Gheorghian, 1978); les Monts Apuseni, dans la nappe d'Arieşeni (Kozur, Mirăuţă, 1980), en Bulgarie, les monts Golo Bârdo et la région de Suhodole (Budurov, Stefanov, 1972, 1975), dans les Alpes (Italie du NE, Mietto, Petroni, 1979, 1980) et l'Asie.

La présence des espèces *Gondolella cornuta*, *G. constricta*, *G. balkanica*, tout comme l'absence du genre *Gladigondolella* caractérisent la "province balkanique de conodontes" (sensu Budurov, 1976), mais considérée par Kozur (in Kozur, Mirăuţă, 1980) comme équivalente à la province faunique de conodontes dinarique et autriche- alpine p.p., qui présente des particularités similaires.

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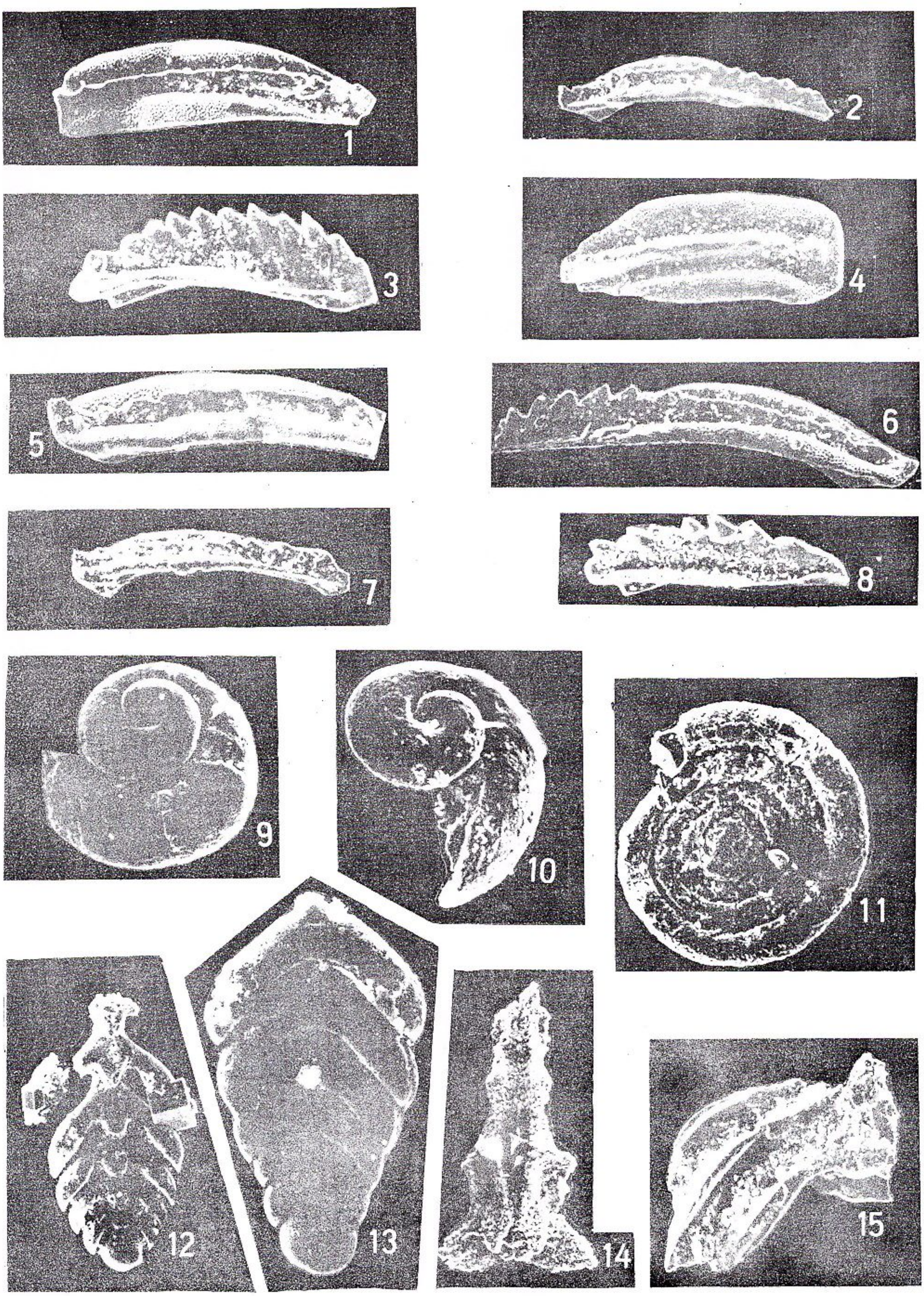
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Planche I

Fig. 1 – *Gondolella bifurcata* (BUDUROV et STEFANOV); éch. SR 6, longueur 0,80 mm.
Fig. 2 – *Gondolella constricta* MOSHER et CLARK; éch. SR 7; 0,63 mm.
Fig. 3 – *Gondolella excelsa* (MOSHER); éch. SR 7; 0,64 mm.
Fig. 4 – *Gondolella bifurcata* (BUDUROV et STEFANOV); éch. SR 7; 0,82 mm.
Fig. 5 – *Gondolella cornuta* (BUDUROV et STEFANOV); éch. SR 8; 0,87 mm.
Fig. 6 – *Gondolella constricta* MOSHER et CLARK; éch. SR 8; 0,85 mm.
Fig. 7 – *Gondolella* cf. *pseudolonga* KOVÁCS, KOZUR et MIETTO; éch. SR11; 0,68mm.
Fig. 8 – *Gondolella excelsa* (MOSHER); éch. SR 14; 0,40 mm.
Figs. 9, 10 – *Microgastéropodes*; éch. SR 4; 0,26 mm, 0,31 mm.
Fig. 11 – *Ammodiscus* sp. (moulage interne pyritisé); éch. SR 6; 0,40 mm.
Figs. 12, 13 – *Nodosariidae*; éch. SR 6; (fig. 12 = *Frondicularia* sp. 4, longueur 0,31 mm, largeur 0,14 mm; fig. 13 = *Frondicularia* sp. cf. *F. lordosa*, longueur 0,24 mm, largeur 0,15 mm).
Fig. 14 – "Vertèbre" d'ophiuride; éch. SR 11; longueur 0,66 mm.
Fig. 15 – Écaille de poisson placoïde; éch. SR 6, 0,87 mm.

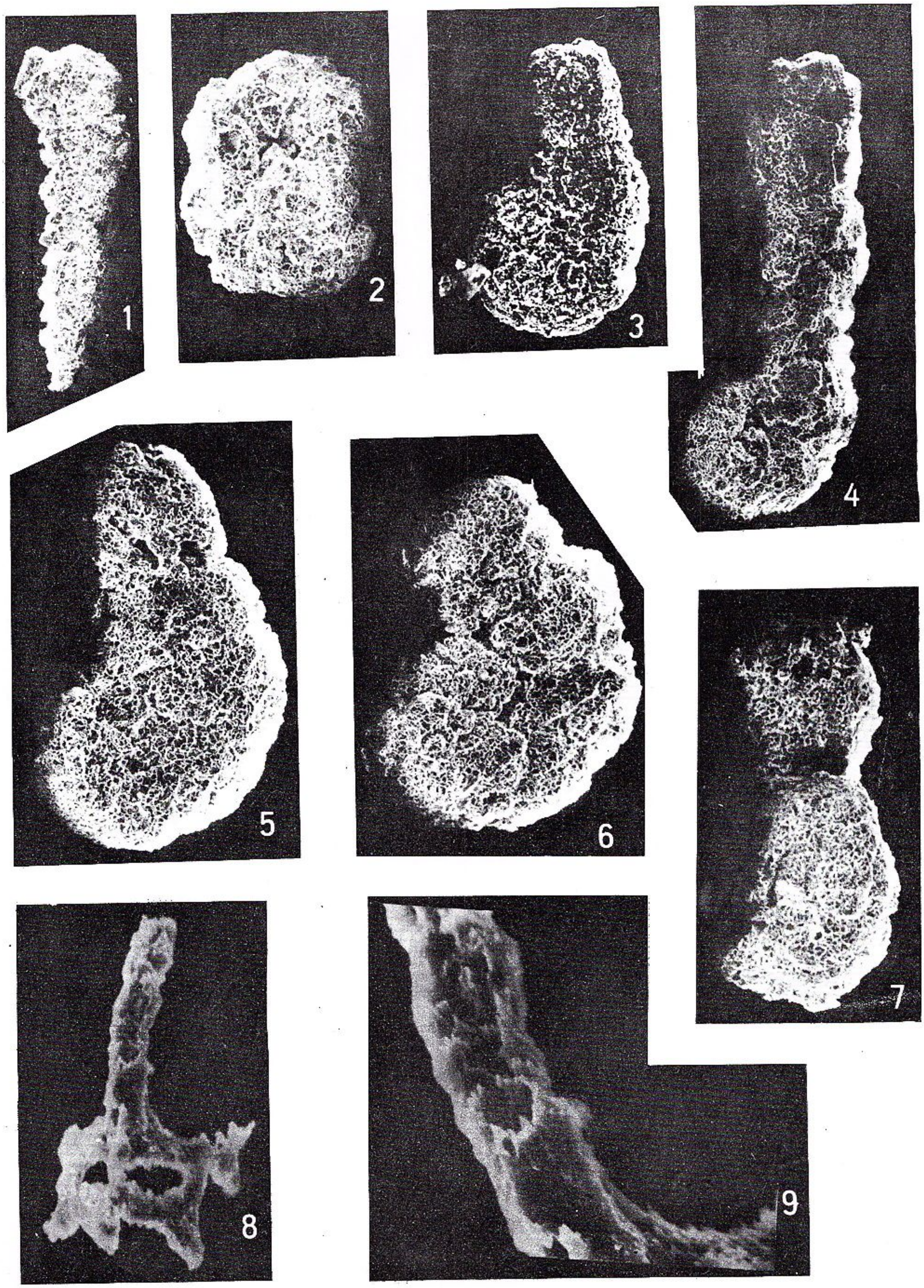

Planche II

Fig. 1 – ? *Earlandinita* sp.; éch. SR 10; longueur 0,50 mm.
Fig. 2 – ? *Trochammina* sp.; éch. SR 10; diam. 0,38/0,29 mm.
Figs. 3–7 – *Ammobaculites parallelus* (IRELAND); fig. 3, éch. SR 10; fig. 4–7, éch. SR 7; longueur 0,52–0,78 mm.
Figs. 8, 9 – *Uvanogelia incurvata* KOZUR et MOSTLER; éch. SR 11; 0,17 mm.

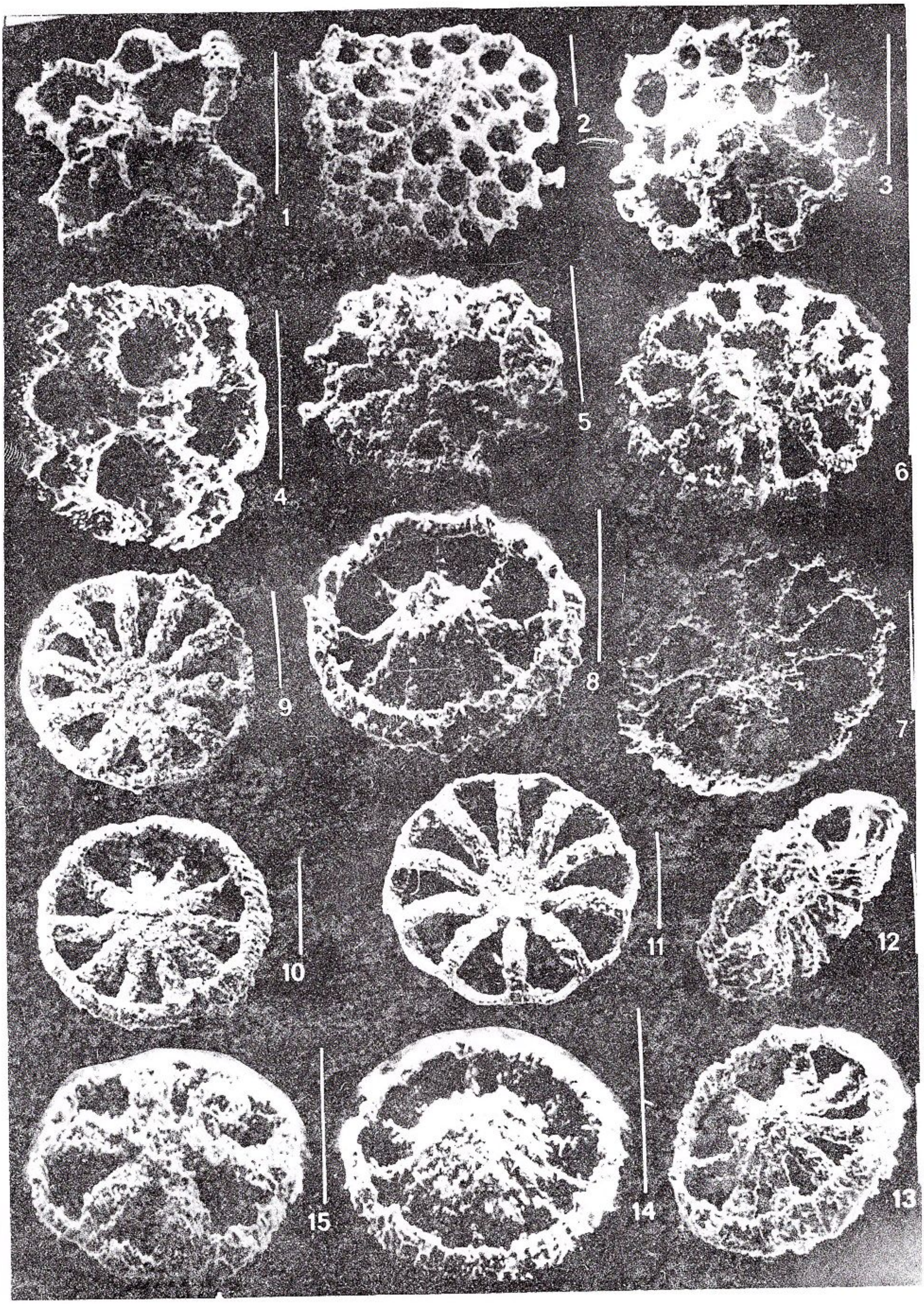
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Planche III

Fig. 1 – *Priscopedatus horridus* (MOSTLER)
Figs. 2, 3 – *Priscopedatus tyrolensis* MOSTLER
Fig. 4 – *Priscopedatus ? acanthicus* MOSTLER
Fig. 5 – ? *Fissobractites* sp.
Fig. 6 – *Theelia monicae* MOSTLER-RAHIMI YAZD
Fig. 7 – ? *Theelia* sp.
Figs. 8–10 – *Theelia immisorbicula* MOSTLER
Figs. 11–13 – *Theelia planorbicula* MOSTLER
Fig. 14 – *Theelia patinaformis* MOSTLER
Fig. 15 – *Theelia* sp. 5

Barre = 30λ.

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MICROFACIES AND MICROFOSSILS IN THE UPPER JURASSIC–LOWER CRETACEOUS LIMESTONE IN THE SOUTHERN PART OF THE PĂDUREA CRAIULUI MOUNTAINS

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Key words: Upper Jurassic. Lower Cretaceous. Limestone. Microfacies. Microfossils. Foraminifers. Calcareous algae. Diagenesis. Paleoenvironment. Carbonate platforms. Apuseni Mountains – Northern Apuseni – Pădurea Craiului Mountains.

Abstract: The study of the Upper Jurassic-Lower Cretaceous carbonate deposits at the southern part of the Pădurea Craiului Mountains has pointed out the existence of interesting micropaleontological assemblages as far as biostratigraphy is concerned, especially in the (? Upper Kimmeridgian-Portlandian) Albioara Limestone and in the (Barremian) Lower Pachyodont Limestone. Its microfacial and diagenetic features make it possible to establish the paleoenvironment in which these limestones were formed. A protected (probably lagoonal) environment corresponding the inner infralittoral realm, relatively frequently isolated from the rest of the basin, characterizes the Albioara Limestone. The Lower Pachyodont Limestone was formed mainly in the mediolittoral realm, as shown by the abundance of fenestral structures, of schzophytoids and of the miliolid Pseudotriloculina.

1. Introduction

The study carried out on Upper Jurassic-Lower Cretaceous carbonate deposits at the southern part of the Pădurea Craiului Mountains, aiming at establishing micropaleontological and microfacial markers of the formations lying immediately under and over the bauxite ores, has led to new data, whose synthesis will make the object of the present paper. The study is based on the detail analysis of more than 700 thin sections cut from samples collected from outcrops, as well as from 25 boreholes drilled in the Sclavu Pleş-Albioara-Secătura-Măgura area, north of the Roşia locality (Fig. 1).

2. Stratigraphic Setting

In the Bihor Autochthon of the Pădurea Craiului Mountains, in the Upper Jurassic-Lower Cretaceous interval, the following lithostratigraphic units (Patrulius, in Ianovici et al., 1976) have been separated: 1, Vad and Farcu Limestones (Oxfordian-Kimmeridgian); 2, Cornet, Aştileu and Albioara Limestones (Tithonian); 3, bauxites (? Valanginian); 4, Cheracea and Gastropod limestones (? Hauterivian); 5, Lower Pachyodont Limestones (Barremian); 6, Ecleja Formation (Aptian) and 7, the complex of glauconite sandstones and Upper Pachyodont Limestones (Albian).

As can be inferred from the diagram of facies relations published by Patrulius (in Ianovici et al., 1976), the Vad and Farcu limestones, on the one hand, and the Aştileu-Cornet-Albioara ones, on the other hand, represent different facies of synchronous deposits.

This stratigraphic diagram has been drawn up after a long period of study of the Upper Jurassic-Lower Cretaceous formations, from the Pădurea Craiului Mountains a study to which many researchers brought their contribution: Kräutner (1941); Patrulius (1956); Patrulius, Istocescu (1967); Bordea, Istocescu (1970); Papiu (1970); Ianovici et al. (1976); Patrulius (1976); Popa (1981); Patrulius et al. (1975, 1982, 1983, unpubl. reports); Bordea et al. (1983, unpubl. report); Bordea, Bordea (1987).

Excepting the Ecleja Formation and the Glauconite Sandstone Complex, the samples analysed by us have been collected from all the other lithostratigraphic entities.

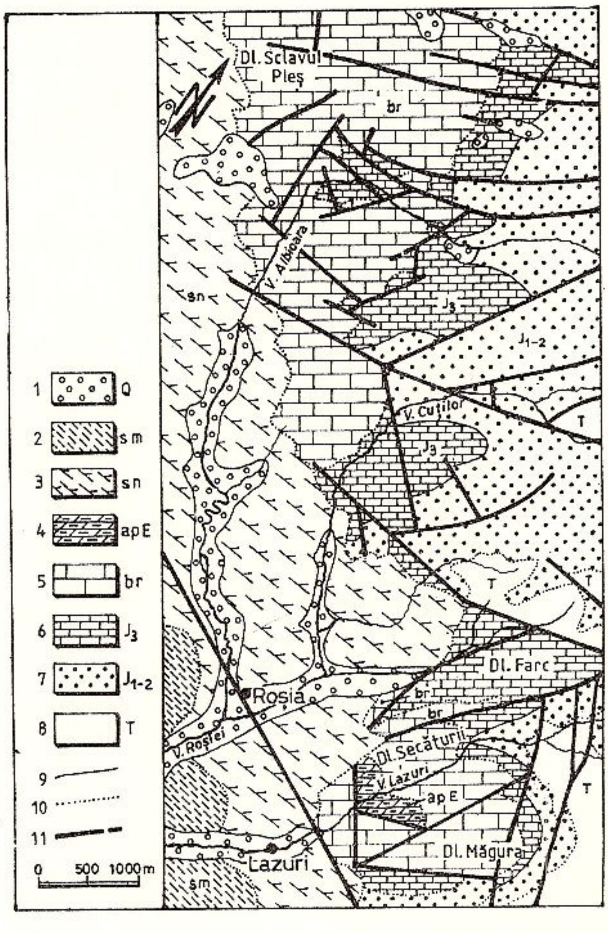

Fig. 1 Geological formations in the Sclavu Pleş Albioara-Secătura-Măgura region, according to the Geological Map of Romania, scale 1 : 50 000, Sheet 40 b (Zece Hotare) (Bordea et al., 1986), with additional data for the south-eastern area, according to Patrulius et al. (1983, unpubl. report).

1, (Quaternary) terrace deposits; 2, Sarmatian deposits; 3, Senonian marls; 4, Aptian Ecleja Marls; 5, Barremian, Lower Pachyodont Limestone; 6, Upper Jurassic limestones; 7, Lower–Middle Jurassic deposits; 8, Triassic deposits; 9, normal boundary; 10, transgression boundary; 11, fault.

3. Biostratigraphic and Microfacial Data

The stratigraphic and microfacial sequence of the Upper Jurassic–Lower Cretaceous deposits in our region of study is synthetically rendered in Figure 2.

3.1. Vad Limestone

The Vad Limestone has been intercepted by only one borehole on about 20 m thickness, being represented by tiny pelsparites (Pl. VII, Fig. 2, 3) and peletoidal microsparites, microsparites with "filaments" and (dolomitic ?) pseudosparites with terrigenous material (Pl. VII, Fig. 1). The micro-organisms in the Vad Limestone are characteristic of hemipelagic deposits (calcispheres, *Globochaetae, Lenticulina*), to which various tiny bioclasts, especially echinoderm plates are added.

There are no micropaleontological arguments for dating these limestones. Taking into account that the Vad Limestone grows into thinner and thinner from the northern towards the southern part of the Pădurea Craiului Mts, constantly preserving its age at the basal part, we are of the opinion that, in the area of study, this limestone probably corresponds to the Oxfordian.

3.2. Farcu Limestone

The Farcu Limestone, intercepted in several boreholes and in the outcrops in the region, is characterized in point of microfacies by the presence of pelmicrites/pelsparites containing echinoderm plates, of pelsparites with rare *Dasycladaceae* fragments, of coral biolithites and of peletoidal-echinodermic microsparites. Two main features of these limestones should be noticed: the abundance of echinoderm plates and their "recrystallized" aspect resulting mainly from neomorphic sparitization phenomena. The Farcu Limestone also contains lamellibranch, hydrozoan, coral, bryozoan and foraminifer fragments. It is obviously coarser in comparison with both the Vad Limestone, which it follows, and the Albioara Limestone, which it is followed by, in the stratigraphic sequence, this feature being bestowed on it first of all by its bioclast nature and size.

At the level of the passage from the Farcu Limestone to the Albioara one, granular limestones occur of the biointrapelsparite or oopelsparite type (Pl. VII, Fig. 4).

The foraminifer microfauna of the Farcu Limestone consists of: *Cornuloculina* sp. (Pl. II, Fig. 5), *Acruliammina* cf. *neocomiana* BARTENSTEIN (Pl. II, Fig. 11), *Protopeneroplis striata* WEYNSCHENK (Pl. II, Fig. 8, 10) and *Conicospirillina basiliensis* MOHLER (Pl. II, Fig. 7). Although certain forms are frequent in the Upper Jurassic (*Protopeneroplis, Conicospirillina*), the microfauna as a whole is not characteristic of a more limited stratigraphic interval. But estimating the position of the Farcu Limestone of the studied area within all the Upper Jurassic Pădurea Craiului deposits, we could admit that this stratigraphic entity is, in broad lines, Kimmeridgian in age.

3.3. Albioara Limestone

The Albioara Limestone has been intercepted by most boreholes drilled in the zones. It has also been

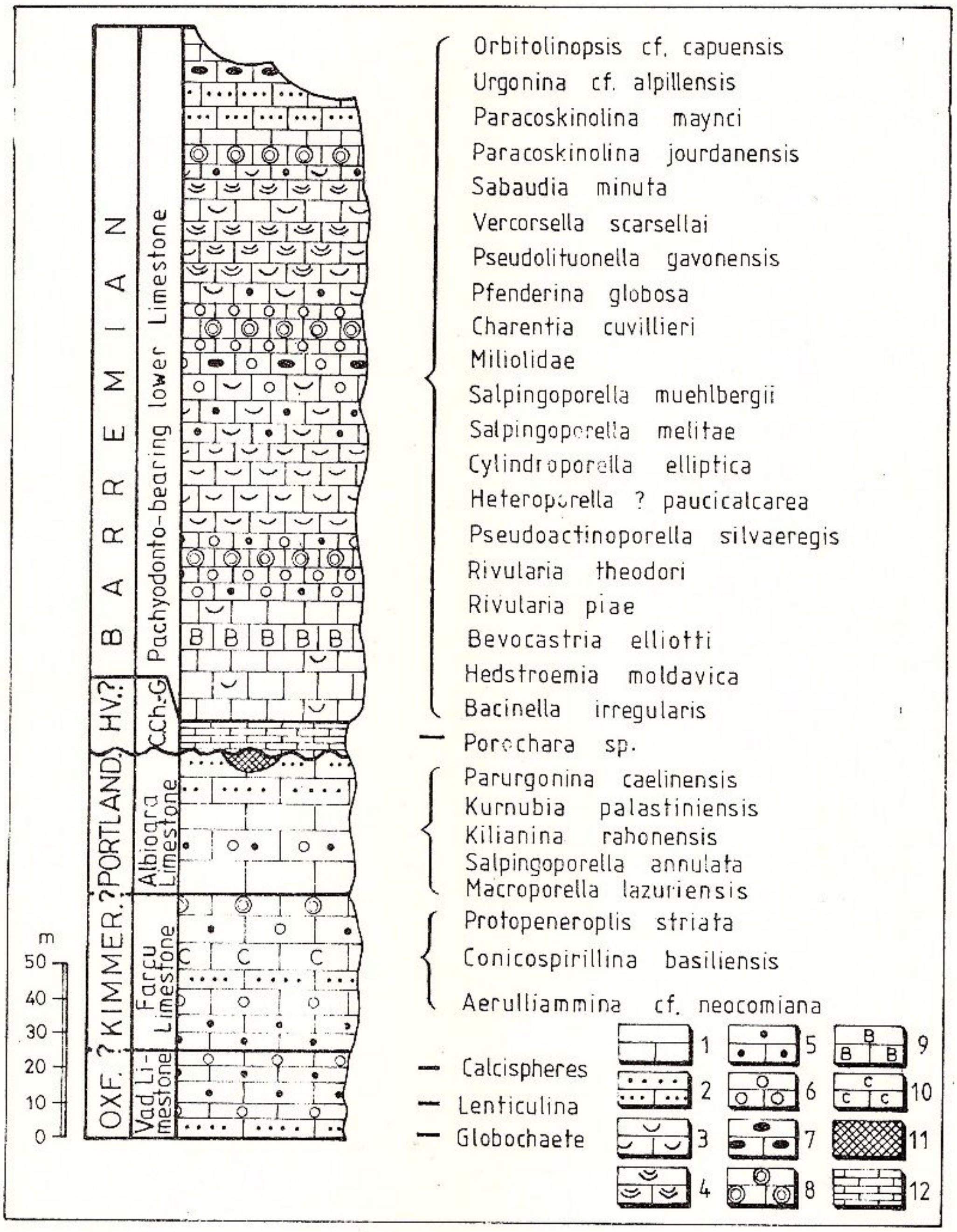

Fig. 2 Upper Jurassic–Lower Cretaceous formations in the south of the Pădurea Craiului Mountains.
1, micrites; 2, microsparites; 3, fenestral micrites; 4, laminitic fenestral micrites; 5, pelmicrites; 6, pelsparites; 7, intrasparites; 8, oosparites-oomicrites; 9, *Bacinella*-bearing biolithites; 10, coralligenous biolithites; 11, bauxites; 12, Characeae-gasteropods-bearing limestones.

identified in outcrops in the Albioara Gorge, in the Roşia Valley and in the Lazuri Valley.

In point of microfacies, the Albioara Limestones vary only little. Micritic limestones prevail, formed in a protected realm of inner infralittoral type, with the subordinate participation of pelmicrites and very rarely of pelsparites. It is micrites/oncoid-bearing micrites that prevail. It is well known that rocks microsparitic in aspect constantly occur at the uppermost part of the Albioara Limestone. Carefully examining these rocks it appears obvious that they are made up of a homogeneous mass of tiny, frequently rhombohedral calcite crystals (Pl. VIII, Fig. 1–6) (chemical analyses carried out on two samples a MgO content of less than 1 %). The crystallographic aspect of the calcite suggests the initially dolomitic nature of the carbonate sediments that have subsequently evolved, through processes of diagenetic dedolomitization, to a finely microsparite limestone.

As far as the micropaleontologic content of the Albioara Limestone is concerned, our study brings a few new data, interesting both in point of paleontology and, especially, in point of biostratigraphy. The foraminifer and calcareous alga assemblage consists of: *Parurgonina callinensis* CUVILLIER, FOURY et PIGNATTI-MORANO (Pl. IV, Fig. 1–8; Pl. VII, Fig. 5), *Kurnubia palastiniensis* HENSON (Pl. III, Fig. 1–3), *K.* cf. *variabilis* REDMOND (Pl. II, Fig. 12), *Kilianina rahonensis* FOURY et VINCENT (Pl. III, Fig. 4, 5, 8, 9), *Valvulina lugeoni* SEPTFONTAINE (Pl. III, Fig. 6, 7, 10, 11), *Trocholina* gr. *alpina* (LEUPOLD et BIGLER), *Conicospirillina basiliensis* MOHLER, *Rectocyclammina* sp. (Pl. II, Fig. 13), *Salpingoporella annulata* CAROZZI (Pl. I, Fig. 2), *Macroporella lazuriensis* BUCUR (Pl. I, Fig. 1) and oncoids with *Bacinella*-type structure (*Bacinella oncolitica* BUCUR, nomen nudum). This assemblage is also, relatively frequently, accompanied by the hydrozoan *Cladocoropsis mirabilis* FELIX.

From the assemblage mentioned above, *Parurgonina callinensis*, *Kurnubia palastiniensis* and *Kilianina rahonensis* are important in point of biostratigraphy. The three foraminifers have been frequently found in the Upper Kimmeridgian–Portlandian interval (Cuvillier et al., 1968; Schroeder et al., 1975). *Parurgonia caelinensis* has already been, very recently, identified in the Albioara Limestone (Bordea, Bordea, 1987). It has also been identified in the Metaliferi Mountains (Mantea, Tomescu, 1986, Pl. V, Fig. 5 and Pl. XIV, Fig. 1, as *Lituonella* sp.).

From the three foraminifers, *Parurgonia caelinensis* is, no doubt, the most frequent form in the Albioara Limestone. Taking into account the presence of this species especially in the lower part of the limestone concerned, as well as its association with the other two foraminifers, the Portlandian age of the Albioara Limestone could be admitted, considering the possibility that in its lower third part it should include a part of the Kimmeridgian.

3.4. Bauxites

Bauxites represent the first term of the Lower Cretaceous of the area concerned. Their relations with the underlying limestones in whose karst they have been accumulated, have been shown in detail by Patrulius (in Ianovici et al., 1976). Being non-carbonate rocks, they have not been of interest for this study.

3.5. Characea and Gastropod Limestones

The bauxite deposits accumulated during the Post-Jurassic exondation are overlain by the Characea Limestones. They could not be separated from the overlying Gastropod ones, especially in boreholes. That is why, when characterizing them in point of microfacies, we shall consider the two lithostratigraphic entities as a whole. They are made up of micrites/biomicrites containing Characeae, ostracods and small-sized gastropods (Pl. VII, Fig. 6, 7). From the Characeae, Dragastan et al. (1966) have identified *Atopochara trivolvis* PECK, *Clavator harrisi* PECK and *Porochara* sp. In the thin sections cut by us we have found *Porochara* sp. *gyrogonites* (Pl. I, Fig. 3–7) and numerous sections through various Charophyte stems and branches (Pl. I, Fig. 8).

At the upper part of the Characea–Gastropod Limestones or immediately over them, we have identified a micropaleontological assemblage consisting of: *Salpingoporella* sp., *Pseudotriloculina* sp., *Valvulineria* sp. and *Pseudotextulariella* sp. Besides the marine character of this assemblage, we also notice that the *Pseudotextulariella* forms resemble very much specimens of the same type identified by one of us (I. Bucur) in the Hauterivian in the Reşiţa-Moldova Nouă Zone.

3.6. Lower Pachyodont Limestone

That is the formation from which most of the analyzed samples have been collected. The Barremian limestones include a large range of structures, the few main types mentioned below being possibly separated, following the analysis of quite many sections:

a) *Mud supported limestones:* biomicrites with tiny detritus of Dasycladaceae, micrites/microsparites, micrites/pelmicrites with strain lamination (Pl. VII, Fig. 8), pelmicrites containing foraminifers, pelmicrites/pelsparites with strain lamination, intraclast biopelmicrites/biopelsparites containing miliolids.

b) *Grain supported limestones:* biopelsparites containing lamellibranch and foraminifer fragments (Pl. IX, Figs. 2, 8), biosparites containing lamellibranch fragments and echinoderm plates, intrasparites and intrapelsparites with strain lamination.

c) *Fenestral limestones:* micrites/pelmicrites with desiccation cracks, laminitic fenestral micrites (Pl. VIII, Figs. 7, 8, 10), micrites with Stromatactis type fenestrae (Pl. VIII, Fig. 9), fenestral intraclastic pelmicrites/pelsparites, fenestral pelsparites/biopelsparites (Pl. IX, Figs. 3, 4, 7).

d) *Oolitic limestones:* oopelmicrites/oopelsparites (Pl. IX, Figs. 5, 6), oomicrites, slightly fenestral oolitic biopelmicrites, oolitic pelmicrites/pelsparites, slightly oolitic pelsparites (Pl. IX, Fig. 1), oolitic micrites/pelmicrites.

e) *Bacinella-bearing biolithites.* As can be inferred from Figure 2, fenestral facies are the prevailing microfacies. Within them, two subtypes can be separated: laminitic fenestral limestones (Pl. VIII, Figs. 7, 8) and fenestral limestones of Stromatactis type (Pl. VIII, Fig. 9). A particular type is that of granular limestones in which fenestral voids can be distinguished from the other intergranular spaces only because of their big sizes and irregular shape (Pl. IX, Fig. 7).

The participation of oolitic limestones is relatively reduced, as far as the quantity is concerned. Certain poorly sorted varieties are interesting, in which rare pisoids also occur (Pl. IX, Fig. 5). With the exception of normal ooids, varying in size, there are certain peculiar varieties, such as broken and regenerated ooids and distorted ooids (distorted oolites, Carozzi, 1961) (Pl. IX, Fig. 6).

In the sequence under examination, pelsparitic varieties are usually associated with oolitic and intraclastic ones, in the same way in which micritic varieties (pelmicrites) are associated, as far as sequences are concerned, with fenestral limestones. Biolithites have local development.

It is very difficult to establish a sequence of the various types of microfacies, which could apply to a larger interval. Nevertheless, the following virtual sequence can be identified, from bottom to top: a) micritic, sometimes fenestral limestones; b) peletoidal/oolitic limestones; c) fenestral, generally non-laminitic limestones; d) pelsparitic/oosparitic limestones; e) fenestral, prevailingly laminitic limestones; f) poorly sorted oolitic limestones and g) micritic/microsparitic limestones.

At certain levels, in the sequence of Barremian limestones there occur microsparites quite similar, especially at macroscopic level, with the Albioara microsparites (? dedolomicrosparites) as well as biopelsparites bearing lamellibranch fragments and echinoderm plates, that can be taken for certain sequences of the Farcu Limestone, which leads to confusions as far as the boundary between Jurassic and Cretaceous deposits is concerned, in certain exploration boreholes for bauxite. In this connexion we should also point out certain intraformational discontinuities that can be noticed both at macroscopic and at microscopic level, showing striking resemblences with the discontinuities in certain sequences at the boundary between Jurassic and Cretaceous limestones (micritic/microsparitic limestone, resembling the Albioara one, followed by a peletoidal sediment, supporting, in its turn, a micrite containing gastropods fragments, quite similar with certain sequences of the Characeae-Gastropod Limestones (Fig. 3).

The Lower Pachyodont Limestone has yielded a relatively rich assemblage of foraminifers and calcareous algae: *Earlandia conradi* ARNAUD-VANNEAU (Pl. V, Fig. 1), *Pseudolituonella gavonensis* FOURY, *Bolivinopsis* cf. *labeosa* ARNAUD-VANNEAU, *Novalesia* cf. *producta* (MAGNIEZ) (Pl. V, Figs. 3, 6), *Vercorsella scarsellai* (DE CASTRO) (Pl. V, Figs. 2, 5), *Sabaudia minuta* (HOFKER) (Pl. V, Fig. 7), *Orbitolinposis* cf. *capuensis* (DE CASTRO) (Pl. VI, Figs. 3–6), *Urgonina* cf. *alpillensis* FOURY (Pl. VI, Fig. 8), *Paracoskinolina maynci* (CHEVALIER) (Pl. VI, Fig. 9), *Paracoskinolina? jourdanensis* (FOURY et MOULLADE) (Pl. VI, Figs. 1, 2, 7), *Charentia cuvillieri* NEWMANN (Pl. V, Figs. 14, 15), *Pfenderina globosa* FOURY (Pl. V, Figs. 8, 9, 11, 13), *Arenobulimina* aff. *corniculum* ARNAUD-VANNEAU (Pl. V, Fig. 12), *Arenobulimina* sp. (Pl. V, Fig. 16), *Maycina* cf. *termieri* HOTTINGER (Pl. V, Fig. 10), *Neotrocholina* cf. *infragranulata* (NOTH), *Derventina filipescui* NEAGU, *Rumanoloculina ponticuli* NEAGU (Pl. VI, Fig. 10), *R. robusta* (NEAGU), *Istriloculina alimanensis* NEAGU, *I. emiliae* NEAGU, *I. eliptica* (IOVTCHEVA) (Pl. VI, Fig. 12), *Schytiloculina* cf. *confusa* NEAGU, *Moesiloculina danubiana* NEAGU (Pl. VI, Fig. 13), *M. scithica* (NEAGU), *Belorusiella* sp., *Glomospirella* sp., *Meandrospira* sp., *Ammobaculites* sp., *Textularia* sp., *Do-*

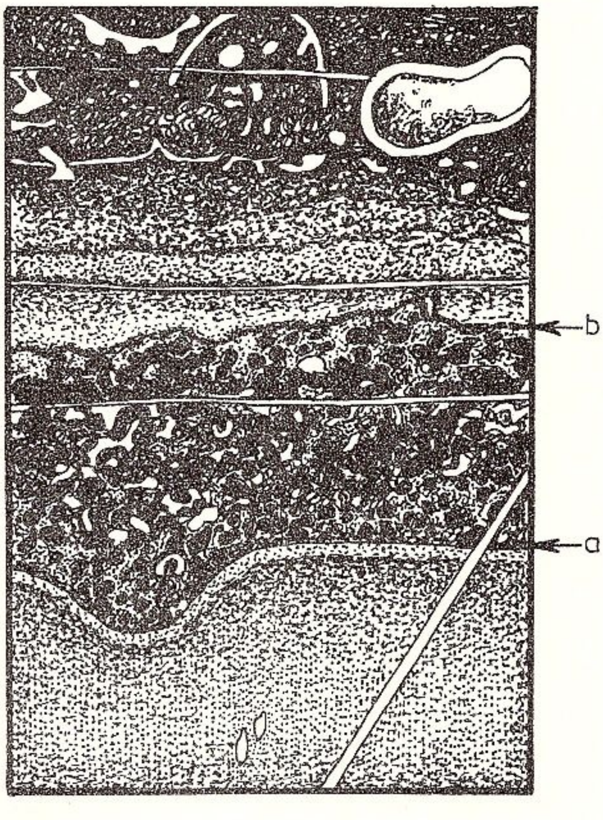

Fig. 3 - Intraformational discontinuities (a, b) in the Lower Pachyodont Limestone (drawn after a microphoto); x 17.

brogelina sp., *Valvulineria* sp., *Trocholina* sp. (Pl. V, Fig. 4), *Decussoloculina* sp. (Pl. VI, Figs. 11, 14), *Pseudotriloculina* sp. (Pl. VI, Figs. 16–21), *Salpingoporella muehlbergii* (LORENZ) (Pl. I, Figs. 9, 11–14), *S. melitae* RADOIČIĆ (Pl. I, Fig. 16), *S.* gr. *cemi* (RADOIČIĆ) (Pl. II, Fig. 3), *S.* cf. *pygmaea* (GÜMBEL), *S. katzeri* CONRAD et RADOIČIĆ (Pl. I, Fig. 15; Pl. II, Figs. 1, 2), *S.* gr. *genevensis* (CONRAD), *S. steinhauseri* CONRAD, PRATURLON et RADOIČIĆ (Pl. I, Fig. 10), *Apinella hispanica* (CONRAD et GRABNER), *Cylindroporella elliptica* BAKALOVA (Pl. I, Fig. 17), *Clypeina maslovi* (PRATURLON), *Heteroporella* (?) *paucicalcarea* CONRAD (Pl. II, Fig. 4), *Praturlonella danilovae* (RADOIČIĆ), *Pseudoactinoporella silvaeregis* BUCUR, *Rivularia fructiculosa* (JOHNSON et KASKA), *Rivularia theodori* DRAGASTAN (Pl. II, Fig. 9), *Rivularia piae* (FROLLO) (Pl. II, Fig. 6), *Scotlandella alpina* DRAGASTAN, *Parortonella getica* DRAGASTAN, *Garwoodia fluegeli* DRAGASTAN, *Hedstroemia moldavica* DRAGASTAN, *Bevocastria* cf. *elliotti* (DRAGASTAN), *Carpathocodium anae* (DRAGASTAN), *Bacinella irregularis* RADOIČIĆ.

The assemblage is typical of carbonate platform facies of Barremian-Aptian age. But a few foraminifers, especially certain orbitolinids, show, no doubt, the Barremian, i.e. *Orbitolinopsis* cf. *capuensis* (DE CASTRO), *Urgonina* cf. *alpillensis* (FOURY), *Paracoskinolina maynci* (CHEVALIER) and particularly *Paracoskinolina? jourdanensis* (FOURY et MOULLADE), to which *Pfenderina globosa* (FOURY), another Barremian form, is added.

We should mention that each microfossil in the pile of Barremian limestones has a different frequence. Anyhow, the *Rivulariaceae* algae prevail, strongly related to the prevalence of intertidal facies. The Dasycladaceae, relatively rare, are a little bit more abundant in the Lazuri Valley-Măgura Summit area. From the foraminifers, it is miliolids and pfenderinids that are more frequent.

Besides algae and foraminifers, Barremian limestones also contain lamellibranch fragments (especially pachiodonts), echinoderm plates (relatively rarely), crustacean coprolites (*Favreina* sp. very rarely), as well as rare tubes of annelid worms.

4. Considerations on the Diagenesis and the Paleoenvironment in Which Various Types of Limestones Are Formed

In the lithification of Upper Jurassic–Lower Cretaceous sediments in the region of our study, several diagenetic processes played an important part. Their action was obvious in time and it started with the genesis of the sediment (micritic borders, for instance) up to its complete lithification. Neomorphism, cementation, pressure solution, silicification and probably dolomitization/dedolomitization are worth noticing.

Neomorphic sparite (microsparite) frequently appears both in the Albioara Limestone and in Barremian limestones, either affecting the whole rock mass, or forming spheroidal-ovoidal aggregates (Pl. VIII, Fig. 2). The overgrowths on the echinoderm skeletal plates, occurring especially in the Farcu Limestone are equally neomorphic in nature.

Cementation could occur in sub-marine environment, in subsurface or subaerial ones, consisting in one or several cement generations. Pressure solution, evidenced by stilolithic fissures, has an important part to play in the final phase of cementation, by the redeposition of calcium carbonate, which this process releases (Purdy, 1968; Bathurst, 1975).

As mentioned, it is probable that some of the microsparites of the final sequence of the Albioara Limestone should be dedolomicrosparites. In this case, a penecontemporaneous dolomitization is involved. Deposited in a protected, probably lagoonal environment, the carbonate muds could have been subjected to restrictive conditions, in certain moments, by the advanced isolation of lagoons in respect of the rest of the basin. Under these conditions, the water from the sediment could reach "brine" concentrations, an important increase in the Mg/Ca ratio taking place, leading to dolomite rhombohedra formation (Folk, 1974; Folk, Land, 1975). A subsequent dedolomitization process could move off magnesium, maintaining the rhombohedral habitus of the crystals. On a smaller scale, this phenomenon has probably taken place also during the Barremian.

Silicification represents a late diagenetic process. In fact, this phenomenon is not at all frequent in the rocks under study. But, in exchange, a relatively frequent phenomenon is "strain deformation". It occurs both in Jurassic limestones and in Barremian ones, bestowing a laminitic aspect on the rock. Strain lamination is sometimes superimposed on the original sedimentary lamination, which proves that the lamination mentioned above is not a sedimentary one (as in case of laminitic fenestral micrites) as well as the deformation of bioclasts (Pl. VII, Fig. 8).

Referring to the paleoenvironment in which the studied limestones are formed, with the exception of the Vad Limestone (having the character of a hemipelagic deposit) and of the Characeae–Gastropod Limestones (fresh water, brackish water deposit respectively), the other types of limestones were formed at the expense of carbonate platform deposits.

The Farcu Limestones mainly correspond to an external infralittoral realm, corresponding to a reef barrier and to its adjacent zones. The Albioara Limestone corresponds to a protected internal infralittoral realm (probably of lagoonal type), isolated from the basin by a reef or oolitic barrier. The Albioara Limestone no doubt corresponds to the innermost realm of the Upper Jurassic carbonate platform, whose external realm is superimposed on the Cornet Limestone, the Aştileu one respectively. The protected character of the environment in which carbonate sediments were deposited is proved, on the one hand, by the relatively poor paleontologic contents, and, on the other hand, by the abounding algal oncoids, that usually live in restrictive environments (Masse, 1979), and possibly by the presence of dedolomicrosparites.

The Lower Pachyodont Limestone also corresponds, in this area, to an internal environment. It is, this time, a prevailingly mediolittoral (intertidal) environment, which has been proved by the abundance of fenestral, often laminitic structures. The micrite of these fenestral laminites is, no doubt, algal in nature. We have to suppose the existence in this zone, during the Barremian, of large intertidal surfaces of algal mats. They are, most of them, schizophytoids (sensu Masse, 1979) in which *Rivulariceae*, very frequent in Barremian limestones, can also be included. The algal origin of certain peletoids in the peletoid-fenestral limestones is proved by the existence of "bridges" uniting them (Pl. IX, Fig. 3, 4), resulted from the micritization of the mucilage uniting algal aggregates. The frequence of schizophytoidic structures (with the exception of Rivulariaceae), as well as the frequence of the miliolid *Pseudotriculina* in these types of rocks are other arguments referring to the restrictive environment they were formed in (Masse, 1979; Arnaud-Vanneau, 1980). Even the peletoidal or oolitic varieties are generally poor in microfauna and indicate an internal infralittoral or mediolittoral (subtial or intertidal) realm. Consequently, pachiodont-bearing bioherms are not largely developed in this area either, and therefore they appear rather rarely in the Barremian sequence.

5. Conclusions

The study carried out on the Upper Jurassic-Lower Cretaceous carbonate deposits in the south of the Pădurea Craiului Mountains (Sclavu Pleş-Albioara-Măgura area) has pointed to the existence of micropaleontological assemblages, interesting in point of biostratigraphy, especially in the Albioara Limestone and in the Lower Pachyodont Limestone. Thus, the *Parurgonina caelinensis, Kurnubia palastiniensis* and *Kilianina rahovensis* assemblage suggests a Portlandian age for the Albioara Limestone, with the possibility that it should contain, at its lower part, equally some part of the Kimmeridgian. On the other hand, *Urgonina* cf. *alpillensis, Paracoskinolina maynci* and *Paracoskinolina? jourdanensis* offer arguments for the Barremian age of the Lower Pachyodont Limestones.

The microfacial and the diagenetic features make it possible for the paleoenvironment to be established, in which these limestones were formed. A protected (probably lagoonal) environment, corresponding to the internal infralittoral realm, relatively often isolated from the rest of the basin, characterizes the Albioara Limestone. The Lower Pachyodont Limestone was formed prevailingly in the mediolittoral realm, as shown by the richness in fenestral-laminitic structures, in schizophytoids and in the miliolid *Pseudotriloculina*.

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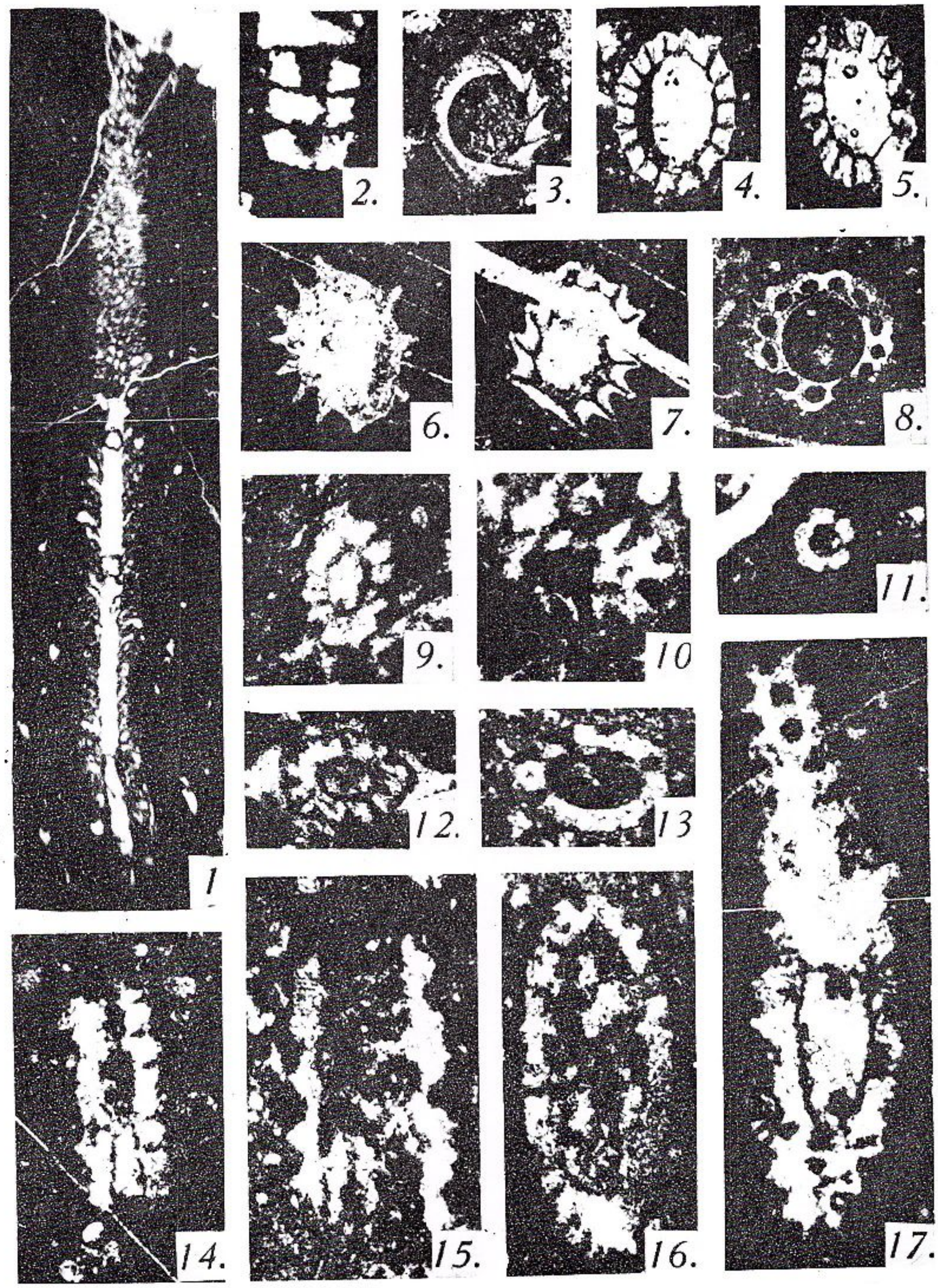

Plate I

Fig. 1 – *Macroporella lazuriensis* BUCUR, Albioara Limestone, sample 220, Lazuri Valley outcrop.
Fig. 2 – *Salpingoporella annulata* CAROZZI. Albioara Limestone, sample 658, F 5020.
Figs. 3–7 – Sections through *Porochara* div. sp. *gyrogonites* Characeae-bearing limestone. 3, 4 – sample 861, F 5011; 5, sample 1094, F 658A; 6, 7 – sample 1095, F 658 A.
Fig. 8 – Section through Characeae stems. Characeae-bearing limestone sample 370, F 1618.
Figs. 9, 11–14 – *Salpingoporella muehlbergii* (LORENZ). Lower Pachyodont Limestone. 9, sample 84, F 2276, 11 – sample 259, F 5022; 12, sample 619, F 2276; 13, sample 231, Lazuri Valley outcrop; 14 – sample 485, F 5027.
Fig. 10 – *Salpingoporella steinhauseri* CONRAD, PRATURLON et RADOIČIĆ. Lower Pachyodont Limestone, sample 440, F 5025.
Fig. 15 – *Salpingoporella* gr. *katzeri* CONRAD et RADOIČIĆ. Lower Pachyodont Limestone, sample 1996, F 1025 C.
Fig. 16 – *Salpingoporella melitae* RADOIČIĆ. Lower Pachyodont Limestone sample 590, F 5028.
Fig. 17 – *Cylindroporella elliptica* BAKALOVA. Lower Pachyodont Limestone, sample 1024, F 1017.

Magnifyings: 1 x 13, 2–17 x 45.

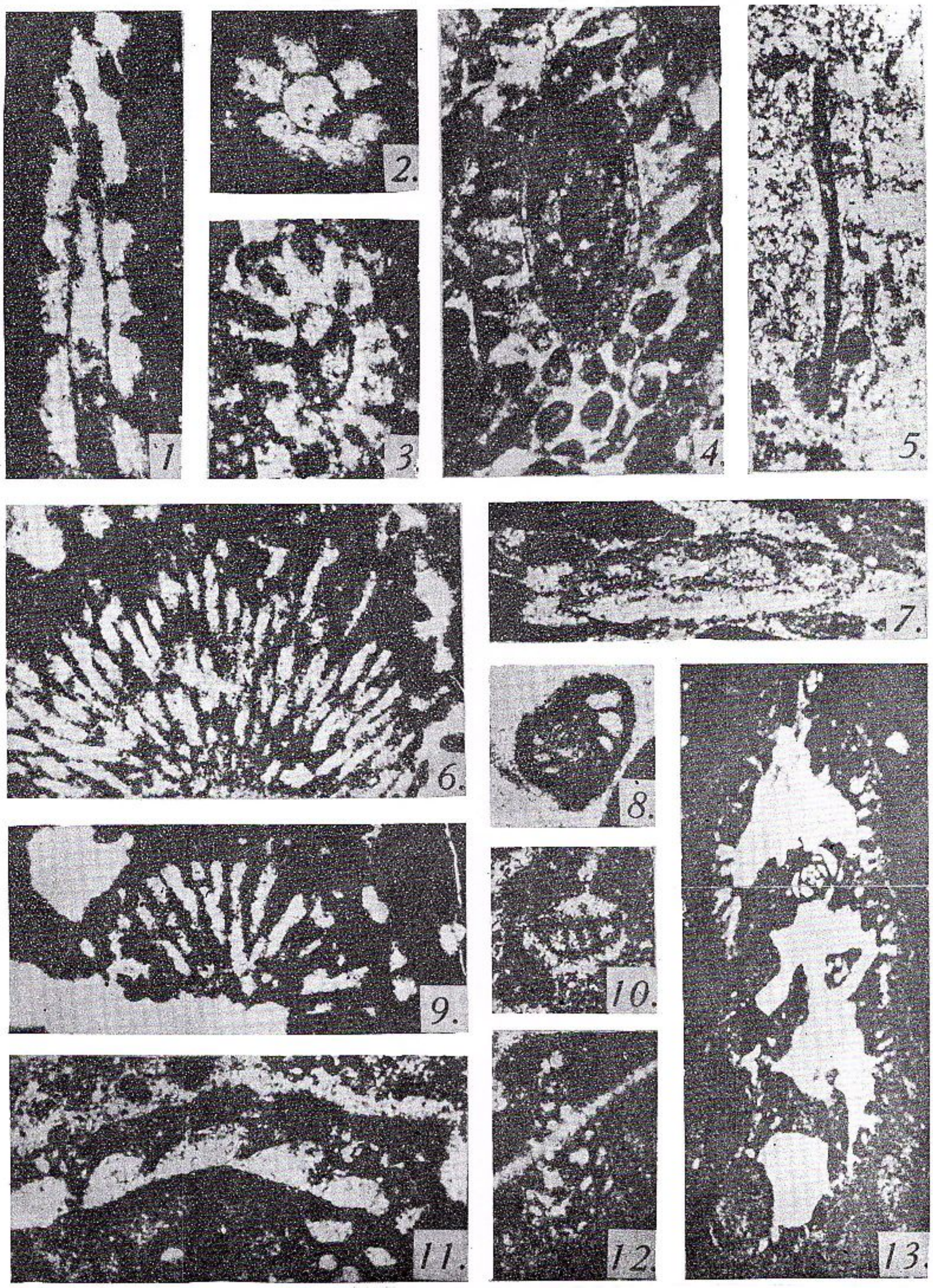

Plate II

Figs. 1, 2 – *Salpingoporella katzeri* CONRAD et RADOIČIĆ. Lower Pachyodont Limestone, sample 588, F 5028.
Fig. 3 – *Salpingoporella* gr. *cemi* (RADOIČIĆ). Lower Pachyodont Limestone, sample 250, Lazuri Valley outcrop.
Fig. 4 – *Heteroporella (?) paucicalcarea* CONRAD. Lower Pachyodont Limestone, sample 231, Lazuri Valley outcrop.
Fig. 5 – *Cornuloculina* sp. Farcu Limestone, sample 870, F 205.
Fig. 6 – *Rivularia piae* (FROLLO). Lower Pachyodont Limestone, sample 817, F 1619 C.
Fig. 7 – *Conicospirillina basiliensis* MOHLER, Farcu Limestone, sample 219, Lazuri Valley outcrop.
Figs. 8, 10 – *Protopeneroplis striata* WEINSCHENK. 8, uppermost part of the Farcu Limestone, sample 2015, Lazuri Valley outcrop; 10, Farcu Limestone, sample 871, F 205.
Fig. 9 – *Rivularia theodori* DRAGASTAN. Lower Pachyodont Limestone, sample 633, F 5020.
Fig. 11 – *Acruliammina* cf. *neocomiana* BARTENSTEIN. Farcu Limestone, sample 870, F 205.
Fig. 12 – *Kurnubia* cf. *variabilis* REDMOND. Albioara Limestone, sample 117, Valea Roşici outcrop.
Fig. 13 – *Rectocyclammina* sp. Albioara Limestone, sample 220, Lazuri Valley outcrop.

Magnifyings: 1–12 x 45; 13 x 30.

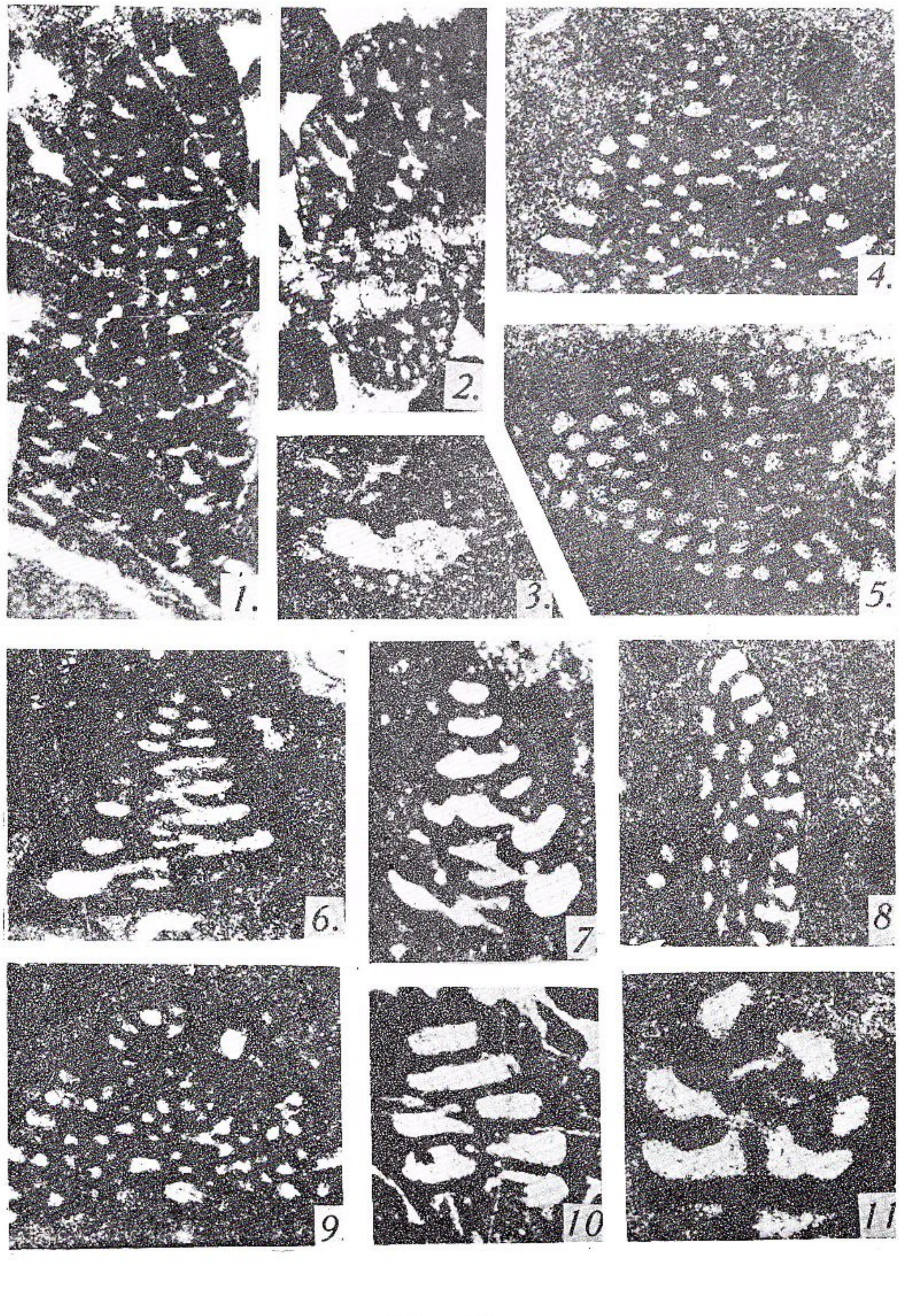

Plate III

Figs. 1–3 – *Kurnubia palastiniensis* HENSON. Albioara Limestone; 1, 2, sample 95, F 1887; 3, sample 2070, Lazuri Valley outcrop.

Figs. 4, 5, 8, 9 – *Kilianina rahonensis* FOURY et VINCENT, Albioară Limestone; 4 – sample 2031, Lazuri Valley outcrop; 5, sample 220 A, Lazuri Valley outcrop; 8, 9 – sample 2060, F 690/3.

Figs. 6, 7, 10, 11 – *Valvulina lugeoni* SEPTFONTAINE, Albioara Limestone; 6 – sample 2019, Lazuri Valley outcrop; 7, sample 2059, F 690/3; 10, sample 95, F 1887, 11 – sample 116, Lazuri Valley outcrop.

Magnifyings: 1–6, 8, 9, 11 x 45; 7, 10 x 30.

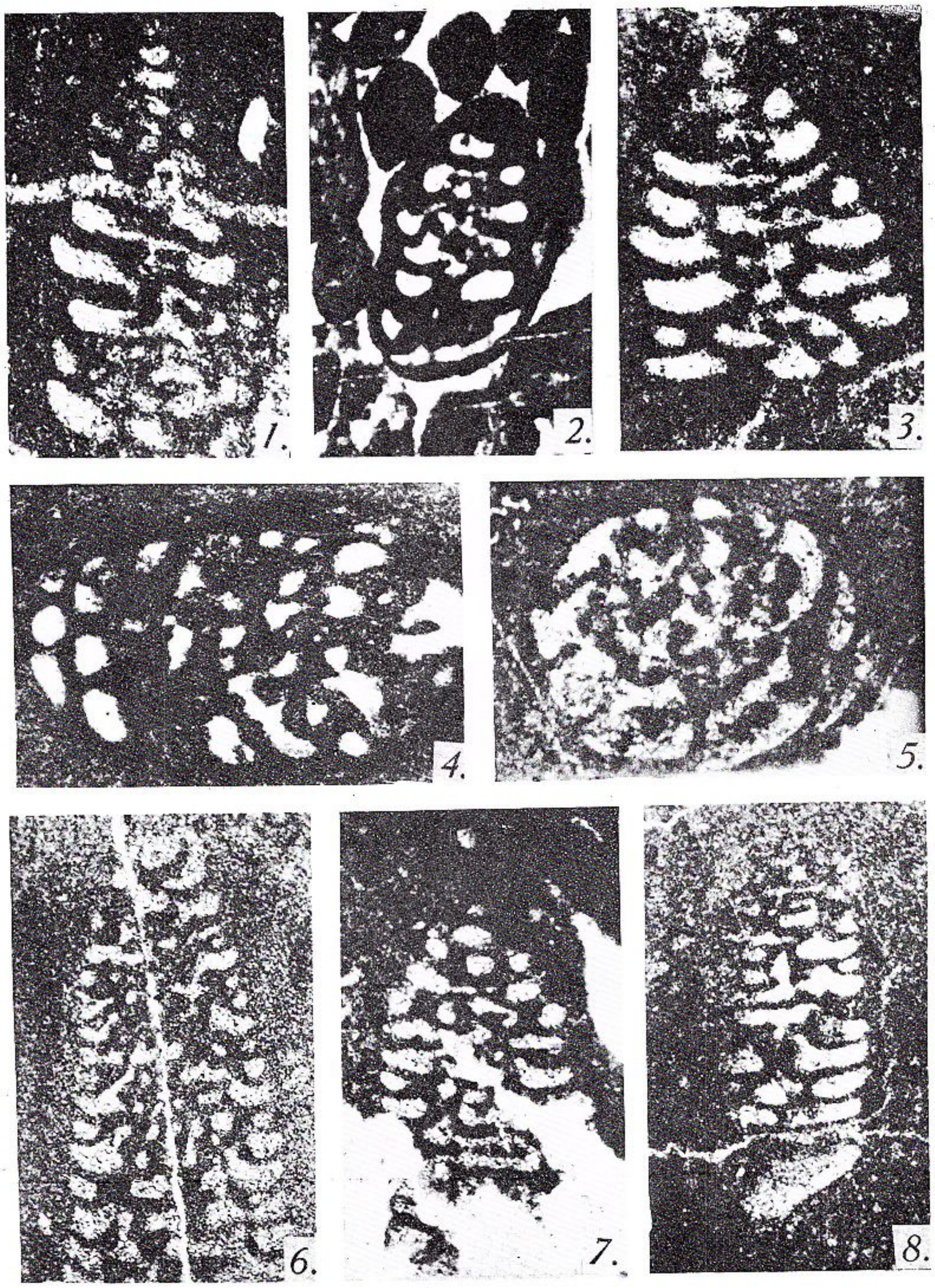

Plate IV

Figs. 1–8 – *Parurgonina caelinensis* CUVILLIER, FOURY et PIGNATTI-MORANO. Albioara Limestone; 1, sample 2009, Lazuri Valley outcrop; 2, sample 95, F 1887; 3, 5, 7 – sample 116, Valea Roşici outcrop; 4, sample 220, Lazuri Valley outcrop; 6, sample 2059, F 690/3; 8, sample 2020, Lazuri Valley outcrop.

Magnifyings: 1, 3, 4, 6–8 x 45; 2, 3 x 30.

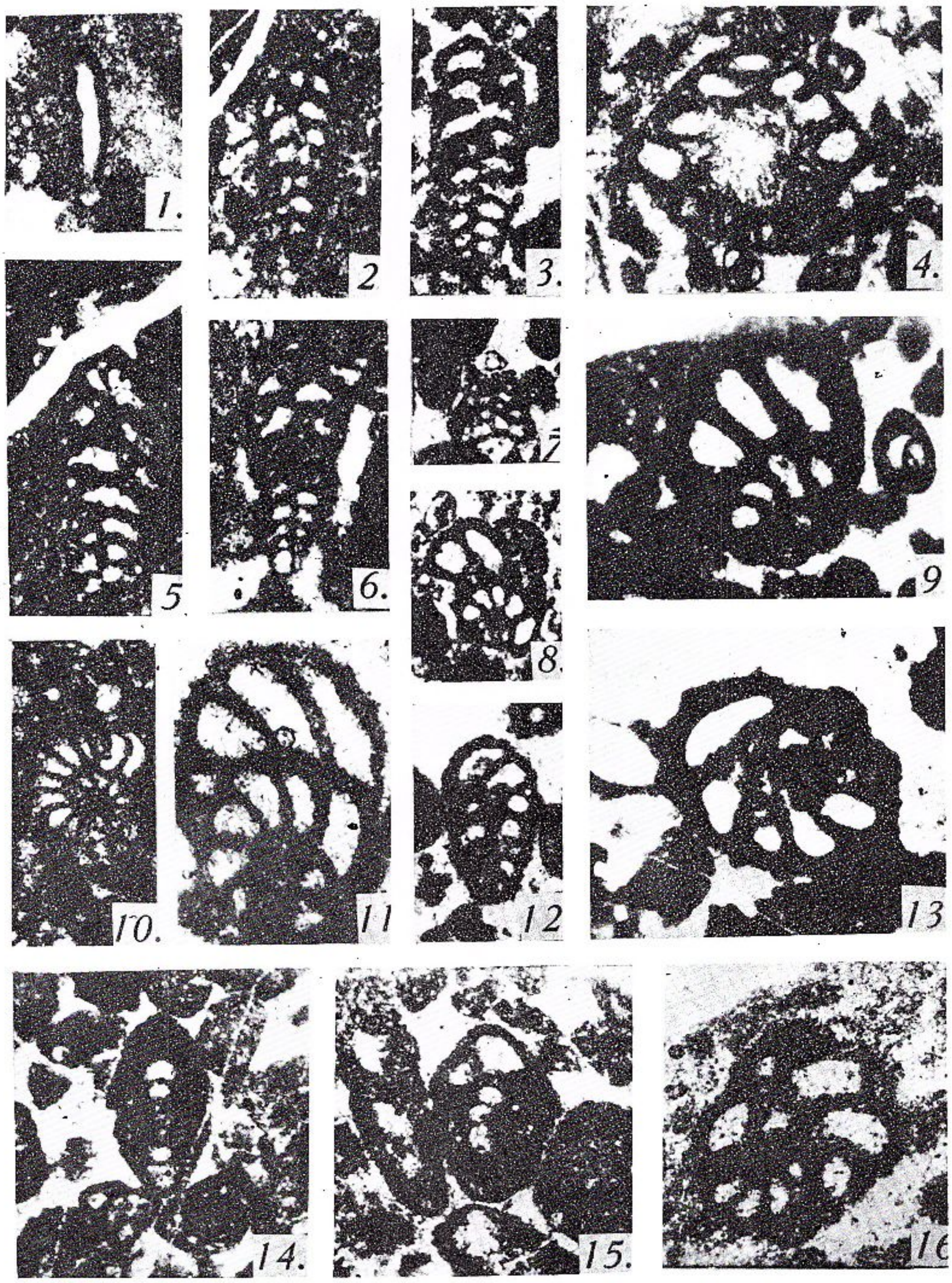

Plate V

Lower Pachyodont Limestone

Fig. 1 – *Earlandia conradi* ARNAUD-VANNEAU, sample 810, F 1619 C.
Figs. 2, 5 – *Vercorsella scarsellai* (DE CASTRO). Sample 494, F 5027.
Figs. 3, 6 – *Novalesia* cf. *producta* (MAGNIEZ). 3, sample 820, F 1619 C; 6, sample 802, F 1619 C.
Fig. 4 – *Trocholina* sp. Sample 84, F 1887.
Fig. 7 – *Sabaudia minuta* (HOFKER). Sample 783, F 5011.
Fig. 8, 9, 11, 13 – *Pfenderina globosa* FOURY. 8, sample 240, Lazuri Valley outcrop; 9, sample 526, F 5027/1; 11, samplé 1060, F 1017; 13, sample 633, F 5020.
Fig. 10 – *Mayncina* cf. *termier* HOTTINGER. Sample 76, F 1887.
Fig. 12 – *Arenobulimina* aff. *corniculum* ARNAUD-VANNEAU. Sample 429, F 5025.
Figs. 14, 15 – *Charentia cuvillieri* NEUMANN, 14, sample 254, Lazuri Valley outcrop.
Fig. 16 – *Arenobulimina* sp. Sample 436, F 5025.

Magnifying: 1 x 45.

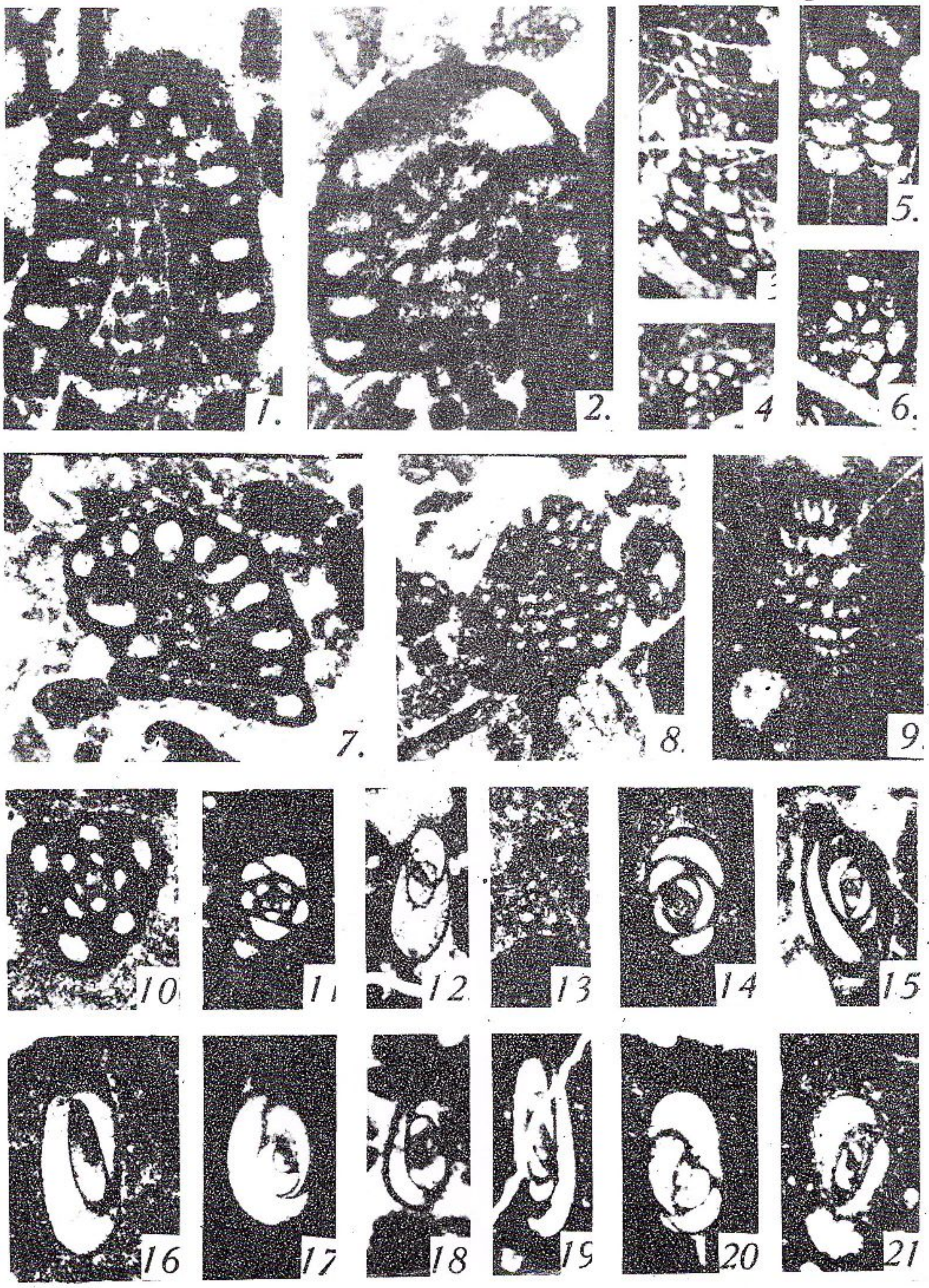

Plate VI

Lower Pachyodont Limestone

Figs. 1, 2, 7 – *Paracoskinolina (?) jourdanensis* FOURY et MOULADE. 1, sample 606, F 2276; 2, sample 224, Lazuri Valley outcrop; 7, sample 226, Lazuri Valley outcrop.
Figs. 3–6 – *Orbitolinopsis* cf. *capuensis* (DE CASTRO). 3–5, sample 783, F 5011; 6, sample 517, F 5027/1.
Fig. 8 – *Urgonina* cf. *alpillensis* (FOURY). Sample 224, Lazuri Valley outcrop.
Fig. 9 – *Paracoskinolina maynci* (CHEVALIER). Sample 260, F 5022.
Figs. 10, 15 – *Rumanoloculina ponticuli* NEAGU, 10, sample 845, F 5027/1, 15, sample 429, F 5025.
Figs. 11, 14 – *Decussoloculina* sp. 11, sample 366, F 5023; 14, sample 363, F 5023.
Fig. 12 – *Istriloculina eliptica* (IOVTCHEVA). Sample 1017, F 5012.
Fig. 13 – *Moesiloculina danubiana* (NEAGU). Sample 266, F 5023.
Figs. 16–21 – *Pseudotriloculina* div. sp. 16, sample 88, F 1887; 17, sample 635, F 5020; 18, sample 764, F 5011; 19, sample 633, F 5020; 20, sample 681, F 5016; 20, sample 822, F 1619.

Magnifyings: 1, 2, 5 – 21 x 45; 3, 4 x 30.

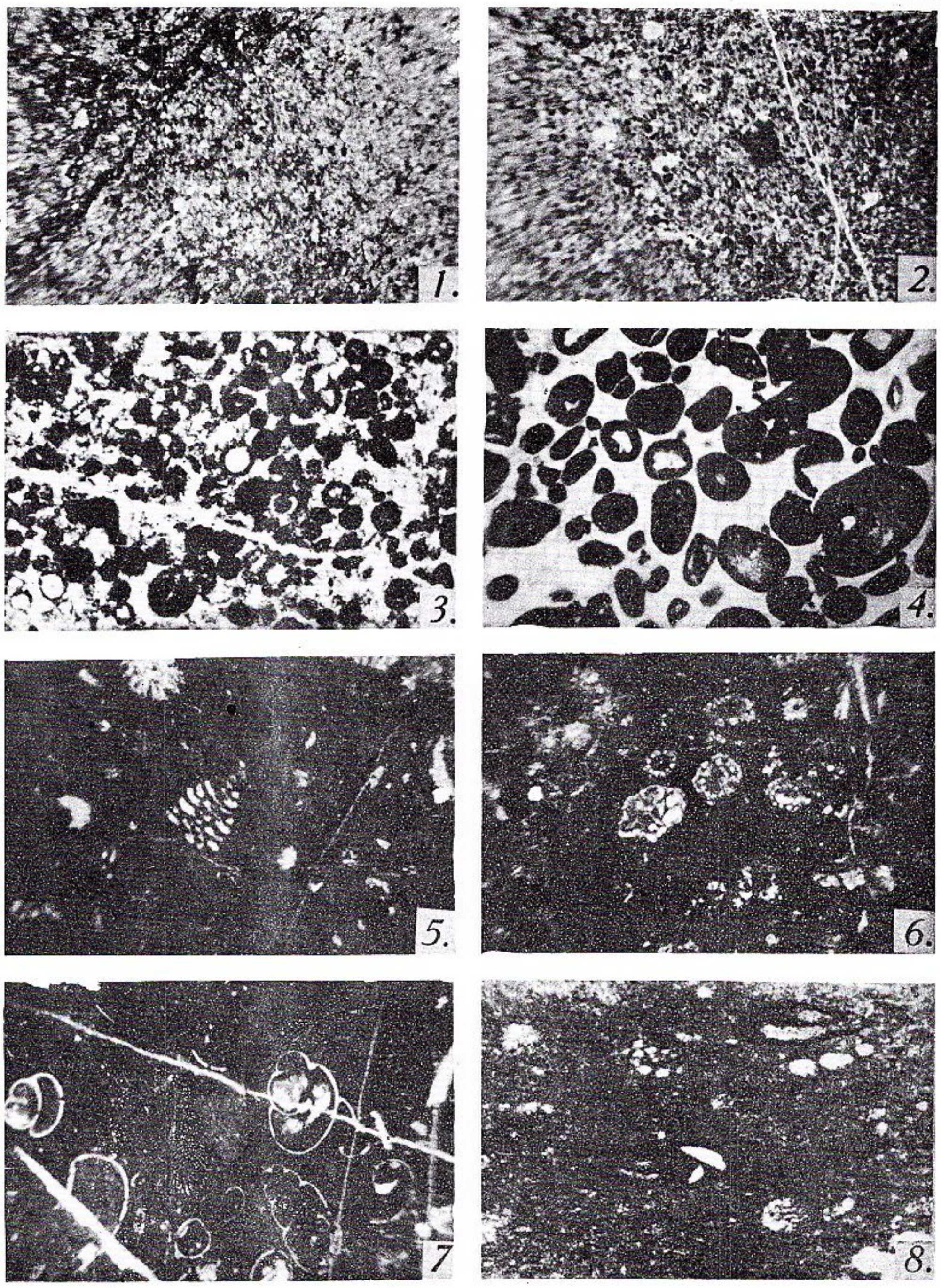

Plate VII

Fig. 1 – (Dolomitic ?) pseudosparite with terrigenous matter, pigmented with Fe oxides-hydroxides. Vad Limestone, sample 106, F 1887.

Fig. 2 – Tiny pelsparite containing Crinoideae plates and *Globochaete* sp. Vad Limestone, sample 105, F 1887.

Fig. 3 – same as Figure 2. Detail.

Fig. 4 – Oosparite. Level of passage from the Farcu Limestone to the Albioara Limestone. Sample 2015, Lazuri Valley outcrop.

Fig. 5 – Peletoidal micrite/microsparite containing *Parurgonina caelinensis* CUVILLIER, FOURY et PIGNATTI-MORANO and *Macroporella lazurensis* BUCUR. Albioara Limestone. Sample 116, Valea Roşiei outcrop.

Fig. 6 – Characeae-bearing biomicrite. Sample 1096, F 658/1.

Fig. 7 – Microgastropods-bearing biomicrite. Characeae Limestone. Sample 112, Valea Roşiei outcrop.

Fig. 8 – Laminitic biopelmicrite; strain lamination evidenced by the abnormally long habitus of the bioclasts. Lower Pachyodont Limestone. Sample 492, F 5027/1.

Magnifyings: 1, 2, 4, 7 x 13; 3, 8 x 45.

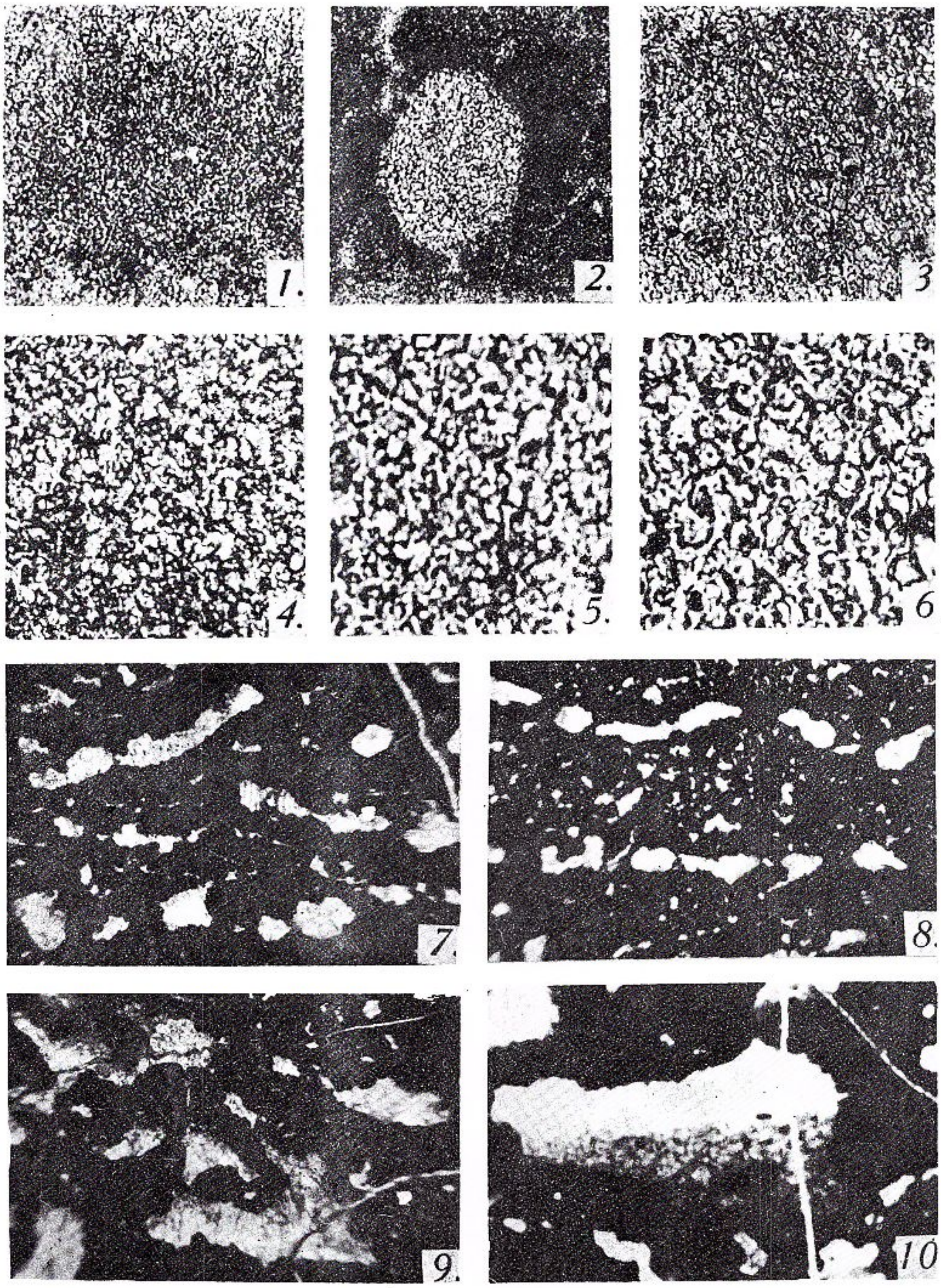

Plate VIII

Figs. 1, 3–6 – Microsparite (dedolomicrosparites ?) Albioara Limestone. 1, 3, 5, 6 – sample 1129; 4, sample 564, F 5024.
Fig. 2 – Microsparitization in zones spheroidal in shape ("boule de neige"). Albioara Limestone. Sample 2018, Lazuri Valley outcrop.
Fig. 7 – Laminitic fenestral micrite with dessication cracks. Lower Pachyodont Limestone. Sample 761, F 5011.
Fig. 8 – Laminitic fenestral micrite of schizophytoidic type. Lower Pachyodont Limestone. Sample 514, F 5027/1.
Fig. 9 – Fenestral micrite of *Stromatactis* type. Lower Pachyodont Limestone. Sample 338, F 5024.
Fig. 10 – Fenestral space filled with geopet sediment (ford silt ?). Lower Pachyodont Limestone. Sample 767, F 5011.

Magnifyings: 1 x45; 2,3 x 90; 4–6 x 180; 7–9 x 13; 10 x 30.

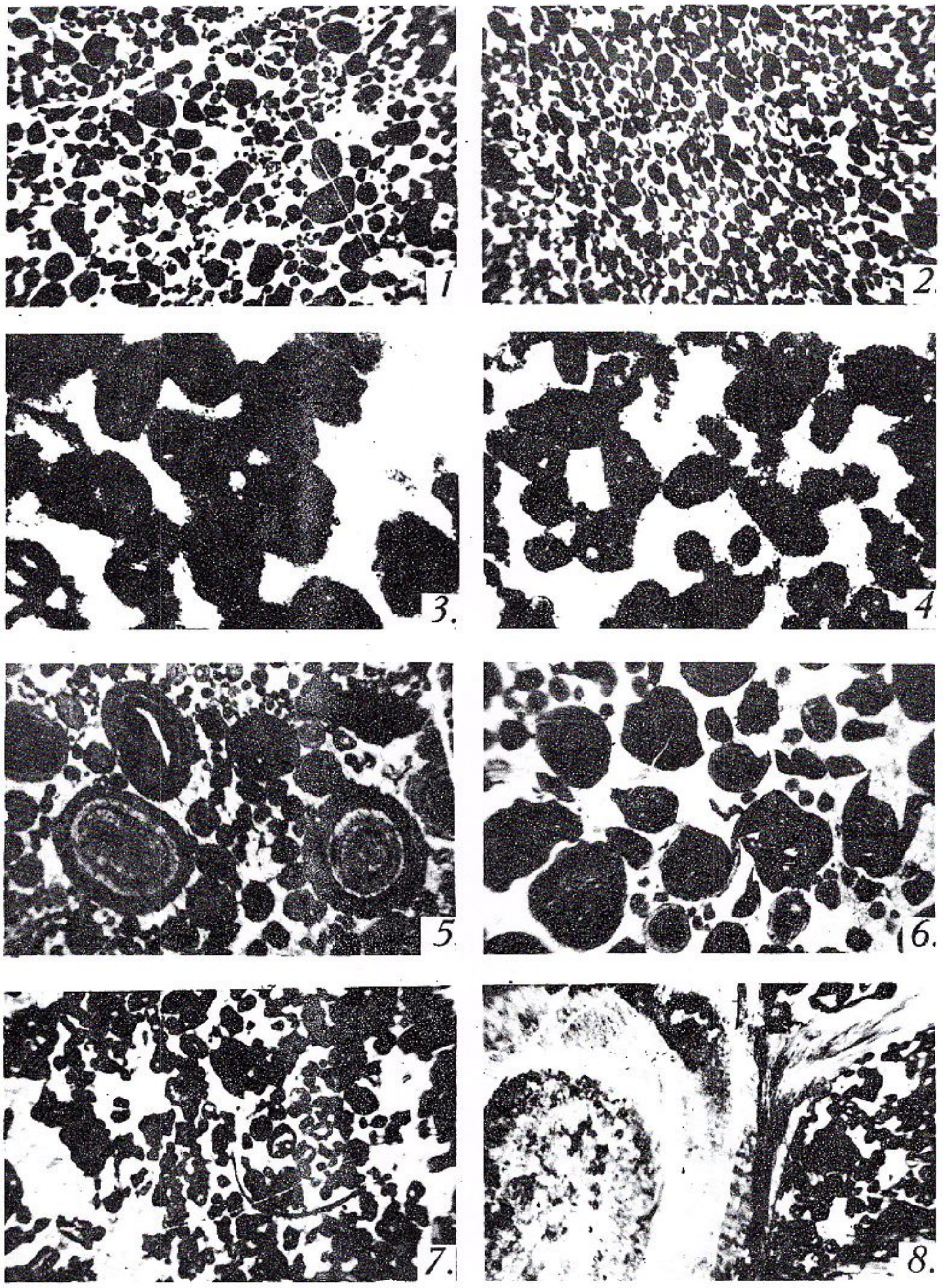

Plate IX

Lower Pachyodont Limestone

Fig. 1 – Slightly oolitic pelsparite. Sample 736, F 5016.
Fig. 2 – Compact pelsparite containing rare foraminifers. Sample 2034. F 1025 C.
Figs. 3–4 – Fenestral pelsparites of schizophytoidic type. The schizophytoidic nature of the micritic pelitoids is indicated by the presence of micritic bridges between them, probably formed by the cementation of the mucilages linking the cyanobacteria aggregates. These bridges look like the "meniscus" type cement. 3, sample 806, F 1619 C; 4, sample 679, F 5016.
Fig. 5 – Poorly sorted oopelsparite. Sample 341, F 5024.
Fig. 6 – Oosparite with distorted ooids ("distorted oolites", Carozzi, 1961). Sample 766, F 5011.
Fig. 7 – Fenestral biopelsparite. Sample 1068, F 1017.
Fig. 8 – Pachyodonts – containing biopelmicrite/biopelsparite. Sample 1027, F 5012.

Magnifyings: 1, 2, 5, 8 x 13; 3, 4 x 45.

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UPPER CRETACEOUS PLANKTONIC FORAMINIFERAL BIOSTRATIGRAPHY FROM THE CARPATHIANS AND NORTHERN DOBROGEA (ROMANIA) RELATED TO MACROPALEONTOLOGICAL ZONATION

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Project 262 – Tethyan Cretaceous Correlation

Key words: Biozonation. Ammonites. Inocerams. Planktonic Foraminifera. First occurrence. Extinction. South Carpathians – East Carpathians – North Dobrogea – Upper Cretaceous – Romania.

Abstract: *Upper Cretaceous Planktonic Foraminiferal Biostratigraphy from the Carpathians and Northern Dobrogea (Romania) Related to Macropaleontological Zonation.* The paper presents the distribution and zonation of the planktonic foraminifers for the Cenomanian–Coniacian of North Dobrogea (Babadag Basin) and South and East Carpathians (Romania). From the chronostratigraphic point of view these are controlled through the direct correlation with the distribution and zonation of the ammonite and inocerams faunas. For the Santonian–Maastrichtian the scale of the planktonic foraminiferal zones proposed by the author for the Romanian Carpathians is presented. Its chronostratigraphic assignment is mostly established by correlation with other such zonal scales from the Tethyan realm, the macrofaunal levels for calibration being very rare.

1. Introduction

The present paper puts forward our biostratigraphic data available according to planktonic foraminifers from the Upper Cretaceous in the Northern Dobrogea and the East and South Carpathians areas (pl. I, II). Special attention is paid to the Cenomanian-Coniacian biostratigraphy as the mentioned areas offer sequences relatively rich in both ammonites and inocerams as well as planktonic foraminifera, so that integrate biostratigraphy has been appealed to. As concerns the Santonian-Maastrichtian biostratigraphy only certain aspects from the Carpathians areas will be presented because in North Dobrogea only a reduced part of Santonian deposits are present, and they are less interesting from a biostratigraphical point of view. Most of the Santonian-Maastrichtian sequences preserved in the Carpathians area, mainly within Median Dacides post-tectogenetic cover, are relatively rich in planktonic foraminifers, while the macrofauna is sparse or insufficiently studied. In this case the absence of direct orthochronologic evidence of macrofauna impedes on assuming the evolution of planktonic foraminiferal assemblages of each chronostratigraphic sub-division. Our data will regard the evolution of assemblages within the biozoning scale proposed by us for the Romanian Carpathians. The relevance of each biozone has been mostly established by comparison with other similar biozonations and only sometimes by direct correlation with faunas yielded by the studied sequences.

Part of the data resulting from our studies, in view of integrate biostratigraphy (Szász, ammonites and inocerams; Ion, planktonic foraminifers), have already been published (Szász, Ion, 1984, 1988; Ion, Szász, 1989). Several previous studies concerning the Upper Cretaceous biostratigraphy based on planktonic foraminifers are worth mentioning: Săndulescu, 1967, 1969; Ion, 1975, 1976, 1978, 1982, 1983; Ion et al., 1987, unpubl. report; J. Ion et al. (in press). The papers concerning the ammonite and inoceram faunas worth noting are: Pop, Szász, 1973; Szász, 1981, 1982 (a, b, c), 1985, 1986 (a, b).

2. Cenomanian-Coniacian Biostratigraphy

The Cenomanian-Coniacian sequence in Northern Dobrogea is preserved in the Babadag Basin and represents the post-tectogenetic cover of the Northern Dobrogea Orogen (Fig. 1). It offers some type sections

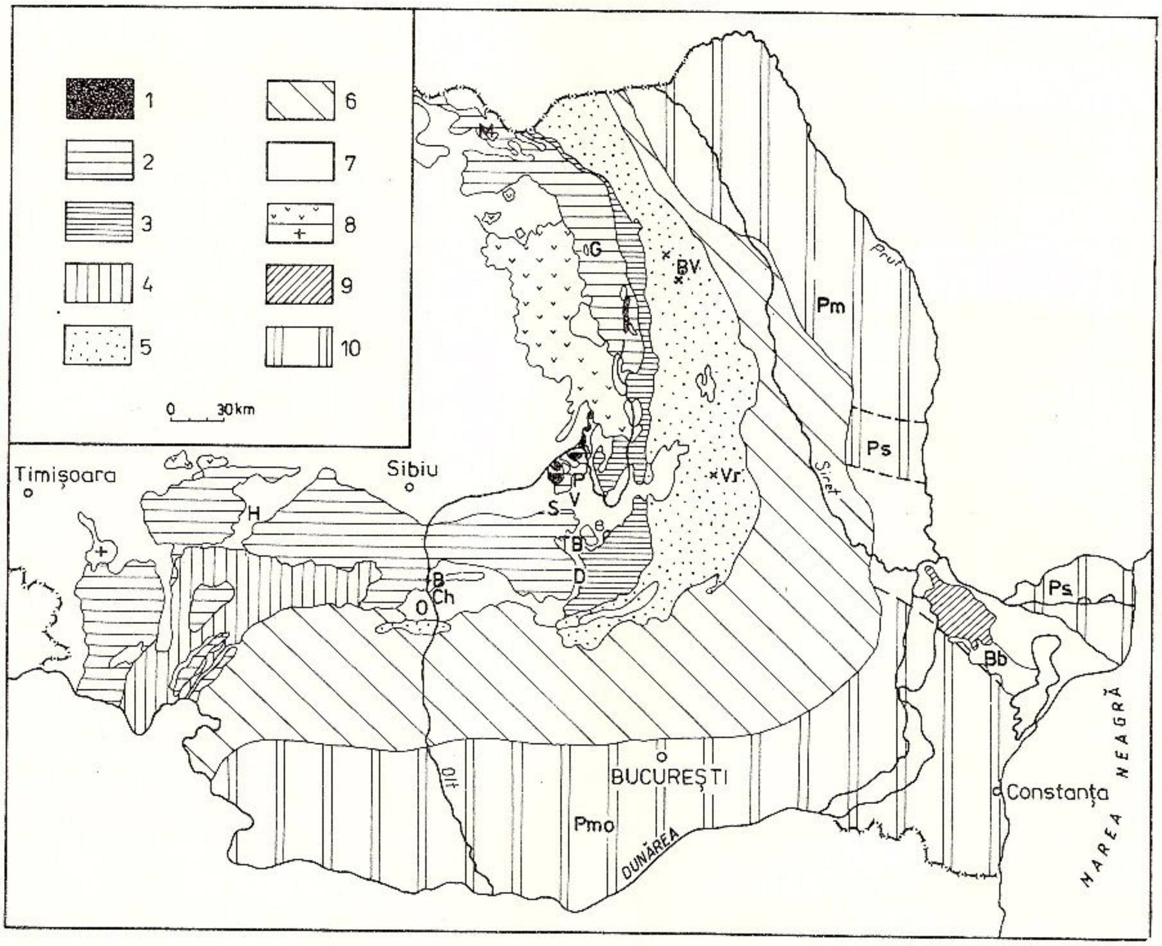

Fig. 1 – Simplified tectonic sketch of Romania (acc. to M. Săndulescu, 1984).
1, Transylvanides, Transylvanic Nappe: Perşani Nappe, Hăghimaş Nappe, Olt Nappe, Rarău Nappe; 2, Median Dacides: Bucovinian Nappe, Sub-bucovinian Nappe and Supragetic Nappe, Infrabucovinian Nappe and Getic Nappe; 3, Outer Dacides: Black Flysch Nappe and Baraolt Nappe, Ceahlău Nappe and Severin Nappe, Bobu Nappe; 4, Marginal Dacides (Danubian); 5, Moldavides: Curvicortical Flysch Nappe, Macla Nappe, Audia Nappe, Tarcău Nappe, Marginal Folds Nappe, Subcarpathian Nappe. Post-tectogenetic elements: 6, foredeep; 7, molasse depressions and post-tectogenetic covers; 8, subsequent magmatites (a, Neogene; b, Neocretaceous-Paleogene); 9, North Dobrogea Orogen; 10, Platforms (Pm = Moldavian Platforms, Ps = Scythian Platform, Pmo = Moesian Platform). East Carpathians: M, Maramureş region; G, Glodu Basin; BV, Bistriţa Valley region; P, Perşani Mts; V, Vlădeni Corridor; Vr, Lepşa Valley in Vrancea half-window. South Carpathians: H, Haţeg Basin; Vînturariţa-Cozia region (O, Olăneşti Basin; Ch, Cheia Valley; B, Brezoi Basin); D, Dîmbovicioara region; TB, Ţara Bîrsei Basin; S, Sinca region.

(proposed as hypostratotype sections, Szász, Ion, 1988) mainly of Coniacian biostratigraphy.

In the Southern and Eastern Carpathians, the Cenomanian-Coniacian deposits abounding in ammonites, inocerams, planktonic foraminifers occur mainly in the more internal areas represented by (Fig. 1): Haţeg Basin, Vinturariţa- Cozia area (Olăneşti Basin, Cheia Valley, Brezoi Basin, southern slope of the Cozia Massif), Dîmbovicioara-Ţara Bîrsei area (Braşov area), Şincai-Vlădeni Corridor, Perşani Mts., Glodu Basin, Maramureş area. These deposits are assigned to the Median Dacides post-tectonic cover. Of biostratigraphic interest are also some Middle Cretaceous sequences of the Outer Dacides in the Eastern Carpathians (Icoveşti-Bădeni in the Dîmboviţa Valley).

As part of integrate biostratigraphy, the orthochronology based on ammonites and inocerams allowed us to determine, for the studied area, firstly the regional range of several planktonic foraminifer species and mainly of the index ones, the correlation between planktonic foraminifer zone and the chronostratigraphic scale, to point out the similarities and differences with respect to other regional or global planktonic foraminifers biostratigraphy.

2.1. Cenomanian

In the Carpathian area (Icoveşti-Bădeni and Cheia Valley), *Thalmanninella brotzeni* and *Th. globotruncanoides* first appear in the Upper Vraconian, in the beds yielding the assemblage of Stoliczkaia dispar Zone, so that the Vraconian-Cenomanian boundary cannot be delimited based on planktonic foraminifers.

Lower Cenomanian. In the Carpathians most of the Lower Cenomanian, corresponding to the interval Mantelliceras mantelli Zone - lower part of Mantelliceras dixoni Zone, is characterized by the Thalmanninella brotzeni/ Th. globotruncanoides Interval Zone (acc. to def. Săndulescu, 1969) which comes from terminal part of the Upper Vraconian. The sequences of this age rarely yield (Tocila Valley and north of the Rîşnoava Valley, in Ţara Bîrsei area) index ammonite species but they overlie the Upper Vraconian deposits yielding ammonites of the Stoliczkaia dispar Zone (Icoveşti-Bădeni; Tocila Mică Valley) and underlie strata containing the *Mantelliceras couloni, M.* aff. *dixoni* (Icoveşti-Bădeni) and the first appearance of *Thalmanninella porthaulti* assigned to the *reicheli* spectrum. This spectrum is unanimously accepted as having the first occurrence at the upper part of the interval of Mantelliceras dixoni Zone.

Within the interval thus delimited persist from the Upper Vraconian the species *Th. brotzeni, Th. globotruncanoides, Th. appenninica, Th. balernaensis, Praeglobotruncana stephani, P. delrioensis, P. marginaculeata*; upwards the base of the Lower Cenomanian, *Th. greenhornensis* does appear, followed by *Th.* aff. *Th. appenninica, Th. micheli* and *Th. marchigiana.* Upwards, below the interval of Mantelliceras dixoni Zone, several other *Thalmanninella* species (*Th.* aff. *Th. brotzeni, Th. jaquesigali, Th. acuta, Th. oresti, Th. runcui, Th. caroni*) and *Rotalipora thomei* appear (i.e., Icoveşti-Bădeni, below the levels yielding *Mantelliceras* aff. *dixoni, M. couloni*; in the beds yielding *Turrilites dorsetensis dorsetensis, Mantelliceras* sp., *Inoceramus virgatus*, in the left-side tributary to the Rîşnoava Valley - Ţara Bîrsei area) and at the lower part of the M. dixoni Zone interval, *Rotalipora montsalvensis* is highly frequent.

In Northern Dobrogea the Lower Cenomanian foraminiferal assemblage assigned to Th. brotzeni/Th. globotruncanoides Zone is similar but less abundant.

At the upper part of the Lower Cenomanian, equivalent (based on *Mantelliceras couloni* and *M.* aff. *dixoni* beds at Icoveşti-Bădeni) with the upper part of Mantelliceras dixoni Zone, the first occurrence of *Thalmanninella porthaulti* is noted (in the Carpathians and Northern Dobrogea areas), a taxon initially assigned by us (Săndulescu, 1969) to "*Rotalipora (Th.) reicheli* not typical form". The appearance of this species marks the lower boundary of the Th. porthaulti Interval Subzone (acc. to def. Ion, 1981/1983 = subzone Th. reicheli non-typique, Ion 1977/1978); the upper boundary of this subzone is marked by the appearance of *Th. reicheli.* In our opinion the Cn 2b zone at the terminal part of the Lower Cenomanian in south-eastern France (Porthault, 1974) is the same with Th. porthaulti Subzone, its index being *R.* aff. *reicheli*, which, according to us, belong to the species *Th. porthaulti.* Now it is assumed that in both boreal and mesogean areas (i.e. Robaszynski et al., 1980; Robaszynski, M. Caron et Groupe Trav., 1979; Marks, 1984) *Th. reicheli* first appears at the top of the Lower Cenomanian, correlated to the terminal part of M. dixoni Zone. We do not know whether this distribution, below the Middle Cenomanian boundary, regards the typical specimens of *Th. reicheli.* The question arises with respect to the *Th. reicheli* or *Th. deeckei* specimens mentioned in other places (Libya, Barr, 1973; Tunisia, Salaj, 1980; Austria, Sturm, 1969) at the top of the Lower Cenomanian.

Middle Cenomanian. The planktonic foraminiferal assemblages of the Middle Cenomanian and at the lower part of the Upper Cenomanian from both the Carpathians and Northern Dobrogea area exhibit features similar to those from some Alpine areas in Central Europe, some regions in Northern Africa (i.e. Marocco), Northern Spain, Crimea and the Caucasus. In those areas, the rotaliporas of the group *cushmani-turonica* do not appear in the Middle Cenomanian - similarly to most of the West European areas - but in the Upper Cenomanian, while the Middle Cenomanian is characterized only by the appearance of *Thalmanninella reicheli* and *Th. deeckei.*

The lower part of the Middle Cenomanian is considered to be defined for both the Carpathian and Northern Dobrogea areas by the Thalmanninella reicheli Interval Subzone (acc. to def. Ion, 1977/1978). The Th. reicheli Subzone has this age as: - the beds with Acanthoceras rhotomagense Zone (Northern Dobrogea - Caraburun Hill) yield this microfauna (unfortunately the section of these beds does not contain outcrops of the lower boundary layers); - in the Ţara Bîrsei region (Tocila Mică Valley) the sequence bearing the assemblage of Th. reicheli Subzone yields *Inoceramus virgatus* from the base and *Scaphites obliquus* from the top; - the sequence bearing the assemblage of Th. reicheli Subzone is directly overlain (in Haţeg Basin - Ohaba Ponor Valley; Northern Dobrogea, at Ciucurova) by the beds yielding the assemblage of Acanthoceras jukesbrownei Zone (upper part of the Middle Cenomanian) and the first appearance of *Th. deeckei* (starting from the base), while it overlies the beds (lacking in macrofauna) bearing the assemblage of Th. porthaulti Subzone, assigned to the terminal part of the Lower Cenomanian. As one may infer from the data above, no orthochronologic proofs are available for the levels situated at the Lower Cenomanian–Middle Cenomanian boundary. *Th. reicheli* was considered to appear at this boundary by taking into account its absence from the lower beds yielding *Mantelliceras couloni, M.* aff. *dixoni* (Icoveşti-Bădeni) and *Th. porthaulti* of the Lower Cenomanian.

The upper part of the Middle Cenomanian, corresponding to the ammonite zone Acanthoceras jukesbrownei, is assigned to the Thalmanninella deeckei Interval Subzone (acc. to def. Ion, 1977/1978, 1982). Its lower boundary is marked by the appearance of *Th. deeckei*, followed by the appearance of the *Dicarinella algeriana, Marginotruncana* aff. *renzi* (first occurrence of a taxon assigned to the genus *Marginotruncana*), while around its upper boundary, which is also the boundary of the Upper Cenomanian, *Th. appenninica* disappears.

The fact that the Th. deeckei Subzone covers the same time interval as Acanthoceras jukesbrownei ammonite zone is well shown in the Haţeg Basin (Ohaba Ponor) where: – *Th. deeckei* appears from the base of the succession bearing the A. jukesbrownei Zone, marking it entirely; – the Upper Cenomanian deposits with *Eucalycoceras pentagonum*, deposits situated immediately above the beds with A. jukesbrownei Zones, yield from the first levels a new planktonic foraminifer assemblage containing *Rotalipora* gr. *cushmani-turonica*. The Th. deeckei Subzone covers the same time interval in Northern Dobrogea as: – *Th. deeckei* appears in the first levels overlying the beds which yield Acanthoceras rhotomagense Zone and Th. reicheli Subzone (Caraburun Hill): – the *Th. deeckei* assemblage is typical (at Ciucurova) of the entire sequence yielding an ammonite assemblage equivalent to A. jukesbrownei Zone; – the layers bearing the Th. deeckei Subzone assemblage underlie those yielding the *R.* gr. *cushmani- turonica* assemblage, which according to orthochronology appear for the first time in the Upper Cenomanian.

Upper Cenomanian. Both in the Carpathians and in Northern Dobrogea, just like all over the world, the planktonic foraminiferal assemblages are enriched during the Upper Cenomanian by the appearance of new genera and species. In our case, during this interval, the three first occurrences are important for microbiostratigraphy: the one of the rotaliporas assigned to *Rotalipora cushmani-turonica* group, situated at the base of the Upper Cenomanian (they appear in the first levels of *Eucalycoceras pentagonum* beds, in the Haţeg Basin), the first appearance of *Whiteinella paradubia* (within certain beds lacking in macrofauna, overlying the *E. pentagonum* beds, also in the Haţeg Basin) and finally the appearance of *Dicarinella imbricata* at the top of the Upper Cenomanian (in Northern Dobrogea it appears close to the base of the Lower Turonian beds yielding *Mytiloides* gr. *labiatus*). Three biozones are delimited within the Upper Cenomanian: Rotalipora gr. cushmani-turonica Interval Zone (acc. to def. Ion, 1977/1978), Whiteinella paradubia Interval Zone (acc. to def. Ion, 1977/1978, as subzone) and Dicarinella imbricata Interval Zone (acc. to def. Samuel et Salaj, 1966) which also reaches the base of the Lower Turonian.

The direct correlation between these three foraminiferal zones and the ammonite zones in Romania needs improving. It is only known that they succeed within an interval marked at the base by the upper boundary of Acanthoceras jukesbrownei Zone (Northern Dobrogea, Southern Carpathians – Haţeg Basin) and the first appearance of *Eucalycoceras pentagonum* (Haţeg Basin), and at the top by the first appearance of *Helvetoglobotruncana helvetica* in the beds with *Mytiloides* gr. *labiatus* (in Northern Dobrogea) or with *Neoptychites cephalotus* (Băniţa–Haţeg Basin). It is also to add that the assemblage of the last two foraminiferal zones are almost identical to those reported in the upper part of the Upper Cenomanian from south-eastern France. Therefore it has been admitted that: R. gr. cushmani-turonica Zone is reported to Eucalycoceras pentagonum Zone; W. paradubia Zone might have its base above the *E. pentagonum* beds and might last till the interval marked by Metoicoceras geslinianum Zone when *D. imbricata* and the entire suite of other *Dicarinella* species do appear; D. imbricata Zone starts from *M. geslinianum* beds and occurs as far as the lower part of the Mammites nodosoides s. l. ammonite Zone (lower part of the Watinoceras coloradoense Zone respectively), i.e. the lower part of the Lower Turonian.

At the base of R. gr. cushmani-turonica Zone, the appearance of the index species is associated with the one of the genus *Pseudorotalipora* (with the species *Ps. praemontsalvensis, Ps. altispira, Ps. lobata*): in the upper part of this zone there appear the genera *Whiteinella* (with the species *W. brittonensis, W. baltica, W. inornata), Helvetoglobotruncana* (with the species *H. praehelvetica*) and *Marginotruncana renzi* (convex forms); *Thalmanninella* taxa (*Th.* aff. *Th. micheli, Th.* aff. *Th. marchigiana*) with angularly-lobate contour appear at the top.

The assemblage of Whiteinella paradubia Zone abounds in species of the genus *Whiteinella*; within it the first *W. archaeocretacea* (only in Northern Dobrogea) and *Praeglobotruncana oraviensis oraviensis* (only in the Carpathians – Haţeg Basin, Ţara Bîrsei) appear.

The lower boundary of the Dicarinella imbricata Zone is usually marked by the almost simultaneous first appearance of *D. imbricata, D. indica (= hagni),* *D. canaliculata*, *D. biconvexa biconvexa*, *Archaeoglobigerina cretacea* as well as *Falsotruncana loeblichae*, *Marginotruncana elenae*, which appear at the base, also including *Praeglobotruncana* gr. *oraviensis* (its first appearance in Northern Dobrogea), and *Rotalipora s.l.* species. We should mention its similarity to "Tu 1 Zone" assemblage from south-eastern France (Porthault, 1974, 1978) where even the species *D. biconvexa* is present (in the *Fagesia superstes* beds), followed by rotaliporas s.l. as far as the Mammites nodosoides Zone; zone "Tu 1" does not contain *A. cretacea*. The D. imbricata Zone in the Western Carpathians, (in Czechoslovakia) and Tunisia is compared to the one reported from south-eastern France and our country, being considered a later one, appearing at the base of the Lower Turonian, succeeding the disappearance of rotaliporas s.l. (although in Tunisia rotaliporas s.l. are also cited from the base of *D. imbricata Zone*, see Salaj, 1980) as well as poorer (out of the species typical of this assemblage only *D. imbricata*, *A. cretacea*, and *W. paradubia* are reported from Tunisia; *D. indica = D. hagni* appears at the top similarly to south-eastern France).

It has been noted that in Romania, *W. archaeocretacea (= lehmanni = gigantea)* first appears in the Carpathian area (Ion, 1983) within the upper part (Lower Turonian in age) of the D. imbricata Zone, while in Northern Dobrogea area (Ion, in Szász, Ion, 1988) its appearance is earlier, within the W. paradubia Zone. The inconstant range of *W. archaeocretacea* and the contradictory extinction of rotaliporas s.l. are the reasons for not using the W. archaeocretacea Zone as Interval Zone or Partial Range Zone in the Romanian Carpathian area and Northern Dobrogea. The D. imbricata Interval Zone typical of the Carpathians and Crimea–Caucasus is kept.

We also mention: in Czechoslovakia, Tunisia, south-eastern France *P. oraviensis oraviensis*, *P. oraviensis trigona* appearing concomitantly with *Helvetoglobotruncana helvetica*; in Romania they appear from the terminal part of the Upper Cenomanian, earlier in the Carpathian area (from W. paradubia Zone) and later (from D. imbricata Zone) in Northern Dobrogea area. Therefore we cannot use them for Turonian biozoning as it was proposed by Salaj et Samuel (1966), Salaj, Bellier (1976), Salaj (1977–1980), Salaj, Gasparikova (1983) and others.

2.2. Turonian

In the studied areas the Cenomanian-Turonian boundary cannot be marked based on planktonic foraminifers, similarly to the situation known all over the world. The base of the Lower Turonian contains the microfauna extended from the late Cenomanian, assigned to Dicarinella imbricata Zone, including the same typical whiteinellas and dicarinellas as well as rotaliporas s.l.. The only difference is due to the appearance only in the Carpathians of *D. biconvexa gigantea* and *W. archaeocretacea*.

The deposits investigated in order to recognize the planktonic foraminifer assemblages at the Cenomanian– Turonian boundary are represented by sequences in which the only relevant macrofauna marker level of the Turonian base are the first *Mytiloides* gr. *labiatus* beds (in Northern Dobrogea, Eastern Carpathians - Perşani Mts, Postăvarul Mts). In Northern Dobrogea the lower boundary of the D. imbricata Zone, below the *Mytiloides* gr. *labiatus* beds, and its extension to their base till the appearance of *Helvetoglobotruncana helvetica* are best exposed. One may also notice that the beds yielding *M. gr. labiatus* and D. imbricata Zone as far as those containing *M.* gr. *labiatus*, *H. helvetica* also bear (*in situ?*) *Rotalipora s.l. (cushmani, turonica, reicheli)*.

Lower Turonian. This time interval is characterized by: its lower part (related to the lower part of Watinoceras coloradoense Zone) which contains the upper part of Dicarinella imbricata Interval Zone, while its upper part (related to the remaining part of W. coloradoense Zone and Mammites nodosoides Zone) corresponds to the Helvetoglobotruncana helvetica Interval Zone (without *Marginotruncana sigali* and *M. schneegansi*) and probably to the base of the Marginotruncana sigali Interval Zone.

As we have already mentioned, the upper part of the D. imbricata Zone has been identified in the first *Mytiloides* gr. *labiatus* beds (in Northern Dobrogea), corresponding to the base of the Lower Turonian. The second zone is typical of the upper part of the *M.* gr. *labiatus* beds (Northern Dobrogea, the Carpathians) or of *Mammites nodosoides* beds (the Carpathians – Haţeg Basin, at Mălăieşti) or of *Neoptychites cephalotus* levels (Haţeg Basin, at Băniţa) assigned to the upper part of the Lower Turonian. According to these data we concluded that the *D. imbricata* assemblage from the base of the Lower Turonian in Romania corresponds to *Whiteinella archaeocretacea* assemblage related to the lower part of the Watinoceras coloradoense ammonite Zone of the base of the Lower Turonian in France (Robaszynski et al., 1982; Robaszynski, 1983). Unfortunately, in the Carpathians there was not identified a conclusive foraminiferal assemblage in the equivalent beds (in the Maramureş area) to W. coloradoense Zone.

The Lower Turonian assemblage of D. imbricata Zone contains the first appearance (in the Carpathian area) of *D. biconvexa gigantea*, *W. archaeocretacea* and the extension from the late Cenomanian of *Whiteinella* species (*archaeocretacea* in Dobrogea; *brit-*

tonensis, aprica, baltica, paradubia, inornata), several *Dicarinella* species (*imbricata, indica, canaliculata, biconvexa biconvexa, algeriana*), numerous *Praeglobotruncana (gigantea, gibba, stephani, marginaculeata, delrioensis, aumalensis, carpathica)*; it also includes *Marginotruncana* aff. *renzi* (convex forms), *M. renzi* (convex forms), *M. elenae, Helvetoglobotruncana praehelvetica, Falsotruncana loeblichae, Archaeoglobigerina cretacea* as well as *Rotalipora s.l.* species (*cushmani, turonica, reicheli*) (*in situ ?*).

The assemblage of H. helvetica Interval Zone (acc. to def. Ion, 1983, for "*H.helvetica* sans *M. schneegansi*" Subzone) includes all the species of the preceding assemblage in addition to *H. helvetica, Dicarinella cachensis*, plus (only in North Dobrogea) *Marginotruncana pseudolinneiana* and *M. marginata*. The lower boundary of this zone is marked by the first appearance of the index species and only in the Carpathian area also by the first appearance of *D. cachensis. Rotaliporas* s.l. species persist (*in situ ?*) all over the zone in the Carpathians, namely in the levels containing the appearance of *M. schneegansi* and *M. sigali*, and in Northern Dobrogea above the level containing the appearance of *H. helvetica*. It is in Northern Dobrogea that the first appearances of *M. marginata* and *M. pseudolinneiana* are noted at the top of H. helvetica Zone (at Slava Rusă, a few meters below the strata yielding the ammonite *Romaniceras* cf. *kallesi* and the foraminifers *H. helvetica* and *M. sigali*).

It should be mentioned that in the Romanian Carpathians and Northern Dobrogea as well as in other parts of the Mesogean area (Crimea, Maslakova, 1971; south-eastern France, Porthault, 1974: Tunisia etc., fide Ion, 1983) the extinction (?) of *Rotalipora s.l.* took place later. As we have already mentioned, a comparison between the Eastern Carpathian area and Northern Dobrogea shows the same dissimilarity: as regards the former, the last rotaliporas s.l. disappear at the Lower Turonian/Middle Turonian boundary or at the top of the Lower Turonian, while from the latter they dissapear in the lower part of the Lower Turonian. These peculiar extinctions are probably due to boreal or mesogean influence controlled by a different evolution of the two areas. One may also think of reworkings, but there are instances in which the Lower Turonian deposits which still contain rotaliporas s.l. do not show resedimentation.

Lower Turonian-Middle Turonian boundary. There are not enough data available for describing the Lower Turonian-Middle Turonian boundary. This boundary might be marked by the appearance of *Marginotruncana sigali* and *M. schneegansi* in the Carpathian area and only by the appearance of *M. sigali* in Northern Dobrogea area (where *M. schneegansi* seems to appear firstly in the Middle Turonian).

In order to define this boundary the following data are available at present: – In Northern Dobrogea the terminal part of *M.* gr. *labiatus* beds, which according to the extinction age of this inoceram group might belong to the lower third of the Middle Turonian, yields the assemblage of the *Marginotruncana sigali* Interval Zone with the index species, *H. helvetica, M. pseudolinneiana, M.* cf. *marginata* and lacking in *Rotalipora s.l.*. Also to the terminal part of the Lower Turonian or the base of the Middle Turonian might belong the first deposits (devoid of macrofauna) bearing the *H. helvetica, M. schneegansi* and *M. sigali* assemblage in the Carpathian area. They also yield *Rotalipora s.l.* species which account in fact for their Lower Turonian age, the age of maximum extinction known in certain Tethyan regions for this group; – In Northern Dobrogea the *Romaniceras* cf. *kallesi* beds (at Slava Rusă), considered to mark approximately the base of Middle Turonian, contain in the first levels the species *M. sigali*, while *M. schneegansi* appears at approximately 4 m stratigraphic thickness above.

The species *M. sigali* may appear at the terminal part of the Lower Turonian both in Northern Dobrogea and in the Carpathians, and the species *M. schneegansi* in the Lower Turonian in the Carpathians and in the Middle Turonian in Northern Dobrogea. It is also possible that at least in Northern Dobrogea *M. sigali* first appears at the Lower Turonian-Middle Turonian boundary.

Middle Turonian. It is defined by the assemblages of two zones: Marginotruncana sigali Interval Zone, covering most of this substage interval, and the base of Marginotruncana coronata Interval Zone (extended in the Upper Turonian, too) covering its top. On the whole, similarly at world scale, the Middle Turonian marks the beginning of the *Marginotruncana* proliferation.

The M. sigali Zone corresponds to the subzone which was initially named by us the M. schneegansi Subzone (Ion, 1983) or M. schneegansi/M. sigali Subzone (Ion in Szász et Ion, 1988). We redefine it according to the name of the species *M. sigali*, as *M. schneegansi* appears at different moments in Northern Dobrogea and the Carpathians. The M. sigali Zone is delimited also as Interval Zone in France (the stratotype inclusively, Robaszynski et al., 1982; Robaszynski, 1983) where it covers the same stratigraphic interval as in the areas investigated by us.

The M. sigali Zone is characterized by the appearance of the index species, *M. schneegansi, Carpathoglobotruncana marianosi*; it also includes *Dicarinella turona* (in the Carpathian area) and *M. pseudolinneiana* (the first specimens in the Carpathians). The other component species come from H. helvetica Zone and the high frequency of *P. oraviensis oraviensis*.

and *P. oraviensis trigona* should be mentioned.

In Northern Dobrogea and the Carpathians, at the terminal part of the Middle Turonian, the assemblage including *Marginotruncana coronata, M. sinuosa, M. "renzi"* (plane-convex forms) appears as well as *M. angusticarinata, M. tohanensis, M. jekeliana, M.* aff. *elenae* (transition from *M. paraconcavata*) only in the Carpathians. It also contains *H. helvetica, H. praehelvetica (?), D. biconvexa biconvexa, M. pseudolinneiana* and (only in Dobrogea) *M. marginata*, plus all the species assigned to the Lower Turonian H. helvetica Zone. We note that in the stratotype area and in other regions of France *M. coronata* appears at the top of the Middle Turonian.

This Middle Turonian assemblage marked by the first appearance of the species mentioned above, with *M. coronata* as index species, is characteristic from *Collignoniceras carolinum* and *Inoceramus cuvieri* strata in Northern Dobrogea as well as from the deposits (lacking in macrofauna) underlying the *Subprionocyclus neptuni* beds in the Carpathian area (Haţeg Basin).

In Northern Dobrogea (Visterna quarries) there is a sequence including: Upper Turonian gritty limestones (bearing *Tongoboroyceras* cf. *rhodanicum* and several inocerams); ca 40 cm below *Collignoniceras carolinum* and ca 1 m below *Inoceramus cuvieri* a macrofauna assemblage lies, pointing to the Middle Turonian top. The whole late Middle Turonian–Upper Turonian sequence contains *M. coronata, M. sinuosa, M. "renzi"* (plane-convex forms).

In the Carpathian area (Haţeg Basin – Valea lui Ion) the Upper Turonian *Subprionocyclus neptuni* beds and the underlying sequence (ca 60 m thick), devoid of macrofauna, yield the same assemblage as the late Middle Turonian- Upper Turonian in Northern Dobrogea (at Visterna) as well as *M. angusticarinata, M. tohanensis, M. jekeliana, M.* aff. *elenae*. From the Carpathians and Northern Dobrogea only the species from the terminal part of the Middle Turonian, not known in the subjacent M. sigali zone, have been cited.

Middle Turonian-Upper Turonian boundary. Both in the Carpathians and in Northern Dobrogea, the mass appearance of several *Marginotruncana* species assigned to "grandes *Rosalines* plates" and to the index species of M. coronata Zone is at the terminal part of Middle Turonian age. Therefore this biohorizon cannot be appropriate for tracing the Middle/Upper Turonian boundary. Otherwise, both in the Carpathians and in Northern Dobrogea, *Marginotruncana undulata undulata* appears in the base of *Subprionocyclus neptuni* beds. Its appearance might mark the beginning of the Upper Turonian, which should be still examined.

We cannot apply the extinction of the species *H. helvetica* for marking the Middle/Upper Turonian boundary as it is done in several other places in the world. In Romania this species has been encountered as far as the Lower Coniacian (reaching the beds bearing Yabeiceras zone assemblage in Northern Dobrogea and in the Carpathians). In other Circum–Mediterranean and Pacific-Californian regions, *H. helvetica* is mentioned as far as the Lower Coniacian (fide J. Ion, 1983).

Upper Turonian. This time interval is characterized by the Marginotruncana coronata Interval Zone (acc. to def. Ion, 1979/1982, as subzone) extended from the top of the Middle Turonian and occurring till the appearance of *M. tarfayensis*, that is the Lower Coniacian boundary. The Upper Turonian assemblage contains all the characteristic species since the Middle Turonian, i.e. *M. coronata, M. sinuosa, M. "renzi"* (plane-convex forms). In the Carpathians it includes the first *M. marginata* (which in Northern Dobrogea is associated with Lower Turonian H. helvetica Zone) and *M. undulata undulata* both in the Carpathians and Northern Dobrogea; only in the Carpathians *M. pseudolinneiana* specimens with very involute chambers on the spiral part, *M. pseudolinneiana* – type *obliqua, M. pseudolinneiana* with thorny contour (transition to *G. spinea* ?), *Marginotruncana* n. sp. (1), *M. pileoliformis, M.* aff. *Carpathoglobotruncana marianosi, Praeglobotruncana hilalensis* do occur. The assemblage also contains all the species reported from the Middle Turonian *M. coronata* assemblage. We note the presence of *H. helvetica*. The Upper Turonian is marked by the extended speciation and proliferation of the genus *Marginotruncana*.

The Upper Turonian assemblage of M. coronata Zone in the Carpathians and Northern Dobrogea is characteristic of strata bearing the ammonite fauna of Subprionocyclus neptuni Zone and the inocerams assemblage typical of this interval. The *M. coronata* assemblages at the base of the Upper Turonian have been well pointed out both in Northern Dobrogea (at Visterna, where the Middle Turonian-Upper Turonian boundary is well delimited) and in the Carpathians (Haţeg Basin Valea lui Ion) within the ammonite beds assigned to S. neptuni Zone; those at the Upper Turonian top, at the Coniacian boundary, are well represented in Northern Dobrogea (at Visterna, Coşaru Mic quarries) and in the Carpathians (Perşani Mts – Valea Cărbunelui).

2.3. Coniacian

The Upper Turonian-Coniacian boundary is marked both in the Carpathians and in Northern Dobrogea by the appearance of *Marginotruncana tarfayensis*. It is the index species of the M. tarfayensis Interval Zone (acc. to def. Ion, 1979/1982 as subzone) which is considered to mark the beginning of the Coniacian in the Romanian Carpathians and Northern Dobrogea.

In the Carpathians (Perşani Mts – Valea Cărbunelui) this species appears first in the *Inoceramus incertus* levels, followed by *I. carpathicus*, and characterizes the beds with *Forresteria petrocoriensis* and *I. schloenbachi*, as shown below; the subjacent Upper Turonian sequence yields the assemblage with *Subprionocyclus neptuni* and microfauna of Marginotruncana coronata Zone. In Northern Dobrogea *M. tarfayensis* appears just below (at 2 m stratigraphic thickness) the first levels yielding the ammonite assemblage of Forresteria petrocoriensis Zone typical of the lower part of the Lower Coniacian. In Romania it has been confirmed (Ion, 1979/1982) also by taking into account the data above, that the first appearance of the species *Marginotruncana tarfayensis* marks the lower boundary of the Coniacian. Porthault (1974) proposed for south-eastern France the appearance of *M. tarfayensis* and *M. sinuosa* to mark arbitrarily the lower boundary of the Coniacian. In this case the appearance level of these species is assigned to some deposits lacking in macrofauna, at ca 200 m stratigraphic thickness below those bearing characteristic Lower Coniacian macrofauna. Our studies have shown that *M. sinuosa* can not be used as marker of the Coniacian lower boundary, as far as it is also present in Upper Turonian strata (bearing the macrofauna of Subprionocyclus neptuni Zone) (Ion, 1979/1982) and appears, according to recent data (Ion in Szász et Ion, 1988) at the terminal part of the Middle Turonian (in the levels yielding *Collignoniceras carolinum* and *Inoceramus cuvieri*). The first appearance of *Marginotruncana "renzi"* (plane convex forms), *M. angusticarinata*, in other parts considered as markers of the Coniacian lower boundary, is not of the same importance for our case. The first appearance of *M. "renzi"*, both in Northern Dobrogea and in the Carpathians is in the strata with macrofauna assigned to late Middle Turonian, while the latter first appears in the Carpathians, in the Middle Turonian strata and at the Coniacian base in Northern Dobrogea only.

Lower Coniacian. In Northern Dobrogea most of the Lower Coniacian is characterized by the Marginotruncana tarfayensis Interval Zone. Its terminal part is marked by the base of the Dicarinella concavata Interval Zone (acc. to def. Sigal, 1955). M. tarfayensis Zone starts in the levels, 2 m thick, devoid of macrofauna, followed by beds containing ammonites of Forresteria petrocoriensis Zone (at Caucagia, including *Forresteria petrocoriensis, Barroisiceras haberfellneri*, the first *Didymotis* layers, as well as *Inoceramus rotundatus, I. carpathicus* etc.) and then in the lower part of the beds with the Yabeiceras Zone (south of Bal-Bair Hill, including *Yabeiceras* aff. *orientale* as well as inocerams, *Inoceramus schloenbachi, I. deformis* etc.). The first appearance of the *Dicarinella concavata*, the base of the D. concavata Zone respectively, is characteristic of the upper part of the beds yielding the Yabeiceras Zone (north of Baia, with *Neocrioceras (Schlueterella) kossmati* etc., and *Inoceramus schloenbachi, I. deformis* etc.; south of Bal-Bair Hill, with *I. schloenbachi*, and the second *Didymotis* level, overlying the beds with *Yabeiceras* aff. *orientale* and with *Marginotruncana tarfayensis* assemblage). The appearance of *Dicarinella asymetrica* in the upper part of the beds with *Yabeiceras* is not sure, as it is identified only from thin sections. It should be mentioned that the first occurrence of the *D. concavata* followed of the first appearance of the *Globotruncana spinea*, of the extinction of *Falsotruncana* (confirming the extinction mentioned by Caron, 1985) as well as of the species *Helvetoglobotruncana helvetica* and several *Dicarinella, Whiteinella, Praeglobotruncana* and *Marginotruncana*.

In the Romanian Carpathians, the Lower Coniacian deposits studied in view of their planktonic foraminiferal content exhibit the following synthetic sequence of Ammonites and Inocerams faunas (acc. to the data reported from the Perşani Mts. and Vînturariţa-Cheia Valley area, Szász, 1986 b): they start with levels bearing *Inoceramus incertus*, followed by *I. carpathicus* (Perşani Mts.-Valea Cărbunelui); there followed the *Didymotis* strata (Perşani Mts.-Valea Racilor) or (Vînturariţa-Cheia Valley area) the deposits containing the assemblage of the upper part of Forresteria petrocoriensis Zone and an assemblage (including *Peroniceras* aff. *isamberti, Forresteria nicklesi* and *Inoceramus schloenbachi*) belonging to the interval of Yabeiceras Zone; the final deposits contain (Perşani Mts.-Valea Satului) *Inoceramus schloenbachi, Scaphites geinitzi* and *Tissotioides haplophyllus* asigned to the top of the Yabeiceras Zone. The Lower Coniacian sequence with the mentioned fauna yields Marginotruncana tarfayensis Zone microfauna exhibiting the same main features as in Northern Dobrogea. At the top, within the deposits yielding macrofauna of Yabeiceras Zone, *Falsotruncana* and *H. helvetica* disappear, specimens of *Globotruncana spinea* type appear, similarly to Northern Dobrogea – but *Dicarinella concavata* was not identified here. It is to mention that in Northern Dobrogea the disappearance of *Falsotruncana* and *H. helvetica* precede the appearance of the *D. concavata* (and ? of the *D. asymetrica*, too), the base of the Dicarinella concavata Zone respectively. As regards the studied area of the Carpathian Realm, the foraminiferal assemblage of the uppermost Lower Coniacian, corresponding to the top of Yabeiceras Zone – which seems to correspond to the base of D. concavata Zone from Northern Dobrogea – is similar to the rest of the Lower Coniacian excepting the absence of *Fal-*

sotruncana, H. helvetica and the presence of *G. spinea* type. So far in the Carpathian Realm no *D. concavata* and *D. asymetrica* have been encountered by us in the Coniacian deposits yielding macrofauna. In this area the first occurrences of *D. concavata* and *D. asymetrica* are known from sequences with no orthochronology (i.e. Ţara Birsei, J. Ion, 1983; Vlădeni Corridor, Săndulescu, 1967, 1970 etc.), its chronostratigraphic assignment being based on geometric criteria and correlations.

On both sections presented, one for Northern Dobrogea and the other for the Carpathian area, the assemblage of M. tarfayensis Zone shows the same essential features. It starts with the appearance of *M. tarfayensis* and also *M. paraconcavata, M. schneegansi* - with two keels, followed by the appearance of *Dicarinella* aff. *P. oraviensis trigona, Marginotruncana sigali* - with two keels, *M.* gr. *tilici*. In the Carpathians the first Lower Coniacian levels also contain *Marginotruncana* n. sp. (1), *Praeglobotruncana alta, Hedbergella flandrini, Marginotruncana undulata lehmanni, Archaeoglobigerina blowi*, specimens of *M. pseudolinneiana* - type *obliqua, M. pseudolinneiana* - to *G. spinea (?), M. pseudolinneiana* - with very involute chambers. Here too, in the *F. petrocoriensis* layers there appear type specimens which should be assigned to *Globotruncana rosetta* and *G. marici*. It should be mentioned that in Northern Dobrogea these levels also contain the first *M. angusticarinata*, a later appearance than in the Carpathians, where it is known since the late Middle Turonian. The assemblage of M. tarfayensis Zone abounds in species also reported from the Upper Turonian. We mention the occurrence of *H. helvetica* and *F. loeblichae*.

Middle-Upper Coniacian. If the lower boundary of the Coniacian is marked by planktonic foraminifers, the other boundaries of Coniacian sub-division as well as the boundary to the Santonian do not benefit from similar evidence. It might be possible that in Northern Dobrogea and in the Carpathians the first appearance of *Contusotruncana fornicata* be typical of the Middle Coniacian and the first occurrence of *Globotruncana lapparenti* and *G. tricarinata* mark the Upper Coniacian. Both in the Carpathians and Northern Dobrogea, the whole interval ranging from the terminal part of the Lower Coniacian to the lower part of the Lower Santonian is represented at present by the D. concavata Zone.

In Northern Dobrogea, the Middle and Upper Coniacian zoning is based on ammonites and inoceramus, so that the presence of Dicarinella concavata Zone within this interval is controlled by orthochronology. Thus the beds bearing the fauna of Peroniceras tridorsatum Zone and/or Inoceramus mantelli Zone (quarries north of Baia and quarry next to Ceamurlia de Jos) yield *Dicarinella concavata, D. asymetrica*, the first appearance of *Contusotruncana fornicata*, also *Contusotruncana plummerae* (?), *Marginotruncana "renzi"* (plane-convex forms), *M. paraconcavata, M. pseudolinneiana, Whiteinella paradubia*. The *Gauthiericeras roquei* strata (at Jurilovca) of Upper Coniacian age contain the same microfauna assigned to Dicarinella concavata Zone, with the index species, *D. asymetrica, Contusotruncana fornicata, D. biconvexa biconvexa, D. canaliculata, Praeglobotruncana gibba, P. oraviensis, Marginotruncana undulata lehmanni, M. jekeliana, M. "renzi"* (plane-convex), *M. pseudolinneiana* etc.

In the Carpathians, as we have already mentioned, D. concavata Zone has been reported so far from sequences lacking in typical macrofaunas. However, we know that these deposits overlie the Lower Coniacian ones yielding the M. tarfayensis Zone and underlie the uppermost Lower Santonian deposits with *Globotruncana bulloides* and (at Fizeşti-Haţeg Basin) *Texanites collignoni*. It is also known that the Lower Coniacian sequences (excluding its terminal part) with the fauna of Forresteria petrocoriensis Zone, of the lower part of the Yabeiceras Zone, and with *I. schloenbachi* assemblage, yield the foraminiferal M. tarfayensis Zone. Thus delimited, the development interval of the Dicarinella concavata Zone is assigned to the Lower Coniacian pro parte-Lower Santonian pro parte. As regards its content, we note the occurrence of *Contusotruncana fornicata* at its middle part, similarly to Northern Dobrogea, and still the presence of *M. paraconcavata*. Its species occur in Northern Dobrogea, too, to which other species also reported from the Upper Turonian or Lower Coniacian assemblages assigned to the Carpathian Realm and already mentioned by us are added.

In the Carpathian Realm, among the Coniacian sections with characteristic macrofauna and planktonic foraminifers, excepting the already described Lower Coniacian one, it is to note a sequence of deposits (Brezoi Basin - left slope of the Lotru Valley) yielding macrofauna of Peroniceras tridorsatum Zone (in the Lotru Valley including *Peroniceras (Zuluiceras) bajuwaricum, Scaphites meslei, Gaudryceras glaneggense, Inoceramus mantelli mantelli* etc.) typical of Middle Coniacian and a sequence (yielding *Protexanites* cf. *strozzi, Scaphites* ex gr. *ventricosus, I.* ex. gr. *undabundus*) which might be assigned to Gauthiericeras margae Zone, i.e. Upper Coniacian. A comparison between this planktonic foraminiferal assemblage and the already described assemblage reported from the top of the *Yabeiceras* Beds, i.e. the top of the Lower Coniacian, shows that the former also includes the appearance of *Contusotruncana fornicata* (during the Middle Coniacian). By comparison with the Middle-Upper Coniacian foraminiferal assemblage yielded by some deposit sequences containing macrofauna from Northern Dobrogea and by others lacking in macrofauna of the Carpathian Realm, this foraminiferal assemblage does not contain *D. concavata* and *D. asymetrica*.

3. Santonian-Maastrichtian Biostratigraphy

This chapter regards only the Santonian-Maastrichtian sequences of the Carpathian area (Pl. II). The biozoning presented has been mainly inferred from the study of sequences in pelagic facies assigned to the post-tectogenetic cover of the Median Dacides in the Ţara Bîrsei region and Şinca area, the Vlădeni Corridor, the Perşani Mts, Brezoi and Haţeg basins. Interesting data have been supplied by the Moldavides flysch sequences (Moldova-Bistriţa valleys sector; Lepşa Valley sector in the Vrancea Mts) (Fig. 1).

3.1. Santonian

The Dicarinella concavata Interval Zone extends also in the basal Lower Santonian. The Lower Santonian assemblage of the D. concavata Zone resembles the Upper Coniacian one for its first part. Towards the boundary of the Globotruncana bulloides Zone the Lower Santonian assemblage includes *Contusotruncana manaurensis, C. nothi, Globotruncana obliqua*, while *D. biconvexa biconvexa* has its last occurrence.

The Globotruncana bulloides Interval Zone (acc. to def. Ion, 1983) (Lower Santonian pro parte–Upper Santonian) represents the interval starting with the first appearance of the index species till the appearance of *Globotruncanita elevata*. The Lower Santonian assemblage of the basal part of this zone contains the index species and probably *Contusotruncana scutilla* (below the *Texanites collignoni* beds at Fizeşti, in the Haţeg Basin), *Globotruncana austinensis* and in the uppermost part (near *T. collignoni* beds) probably the last occurrence of *Marginotruncana paraconcavata*. This assemblage also includes *G. lapparenti* (which at least in Northern Dobrogea occurs starting with the Upper Coniacian), *G. linneiana* and the species belonging to the Upper Coniacian assemblage of the Dicarinella concavata Zone, such as *D. concavata, D. asymetrica*, specimens of *Globotruncana spinea* and *G. rosetta*. The Upper Santonian assemblage of this zone probably lacks in *M. paraconcavata* and includes the first appearance of *Rugoglobigerina pilula*.

3.2. Campanian

According to certain data, the first appearance of *Globotruncana arca*, followed by the first occurrence of *Globotruncana orientalis* might be asigned below the Santonian/Campanian boundary and prior to the appearance of *Globotruncanita elevata*. Therefore in the beds (in Brezoi Basin, with *Inoceramus lingua - patootensiformis, I. lingua -angustus, I. (Cardiceramus) bueltensis -arnoldi, Pachydiscus* cf. *cayeuxi, Hauericeras* ex. gr. *gardeni*) considered to belong to the base of the Lower Campanian or uppermost Santonian, the foraminiferal assemblage does not contain *Gt. elevata* yet, but only *G. orientalis* and *G. arca*. It also contains *Marginotruncana pseudolinneiana, M. sinuosa, M. "renzi"* (plane convex forms), *M. renzi* (convex forms), *M. coronata, Globotruncana bulloides, G. linneiana, G. lapparenti*. Moreover, the underlying foraminiferal assemblage, identified in the deposits with 20 m thickness already, contains *G. arca* and next to *I. lingua–I. angustus* beds, *G. orientalis* does occur. The data presented seem to demonstrate that probably the lower boundary of the Gt. elevata Zone does not coincide with the Santonian/Campanian boundary, being located upwards.

The Globotruncanita elevata Interval Zone (acc. to def. Ion, 1983) (Lower Campanian) represents the interval starting with the appearance of the index species till the appearance of *Globotruncana rugosa*. Its lower boundary is marked also by the last occurrence of *Dicarinella asymetrica, D. biconvexa gigantea* and probably by the last *Whiteinella brittonensis*. The basal section contains the last appearance of several *Praeglobotruncana* species (*P. oraviensis oraviensis* – lens-shaped with very convex dorsal part, *P. alta, P. aumalensis*), *Dicarinella concavata, Marginotruncana sinuosa, Carpathoglobotruncana filipescui, Rugoglobigerina pilula (=bulbosa), Whiteinella archaeocretacea*. In this zone there appear *Globotruncanita stuartiformis*, then at the top (over the *Inoceramus monticulli* beds, at Tohan, Ţara Bîrsei) *Globotruncanella havanensis, Globotruncana verrucosa, G. morozovae, G. mariai* have the first occurrence and *Dicarinella turonica, D. difformis, Praeglobotruncana prahovae, P. oraviensis oraviensis, P. oraviensis trigona* their last occurrence.

The Globotruncana rugosa Interval Zone (acc. to def. Ion, in this paper) (Upper Campanian pro parte) represents the interval starting with the appearance of the index species till the appearance of *Globotruncanita calcarata*. At the base, the index species is associated with *Globotruncana ventricosa, Gt. atlantica* (at Tohan, Ţara Bîrsei, it appears above the strata containing *Inoceramus regularis, I. alaeformis, I. sagensis, I. balticus*), *Rugotruncana subcircumnodifer, Contusotruncana patelliformis, Globotruncanita subspinosa* and *Globotruncana hilli*. The other species extend from the Gt. elevata Zone.

The Globotruncanita calcarata Total Range Zone (acc. to def. Herm, 1962) (Upper Campanian pro parte) represents the entire range of the index species. This zone also includes the first *Globotruncana falsostuarti* and *Globotruncanita stuarti*, while the other species are also known from G. rugosa Zone.

3.3. Maastrichtian

The Globotruncanita stuarti/Globotruncana falsostuarti Partial Range Zone (acc. to def. Sigal, 1977) (lower part of the Lower Maastrichtian) represents the interval starting with the last occurrence of *Globotruncanita calcarata* till the appearance of *Gansserina gansseri*. Starting from the base the assemblage of this zone contains *Plummerita hantkeninoides* and the species *G. falsostuarti, Gt. stuarti*, extending from the last zone. By the middle part of this zone *Globotruncana austinensis* disappears but there appear *G. aegyptiaca, Globotruncanella petaloidea*, besides *G. insignis*, and at its terminal part *Whiteinella paradubia* disappears. The assemblage containing *Globotruncana falsostuarti* and *Globotruncanita stuarti* has been identified (in the Haţeg Basin) from beds bearing the macrofauna assemblage with *Pseudoxybeloceras riosi, Ventilabrilla bayei, Hauericeras* ex gr. *gardeni, Inoceramus balticus.*

The Gansserina gansseri Interval Zone (acc. to def. Bronnimann, 1952) (uppermost Lower Maastrichtian-lowermost Upper Maastrichtian) represents the interval from the first appearance of the index species till the first appearance of *Abathomphalus mayaroensis.* The base of this zone also includes the first occurrence of *Contusotruncana contusa, Ct. plicata* and probably of the *Gt. conica, Globotruncanella pschadae* and *Rugoglobigerina hexacamerata.* In the upper part *Abathomphalus intermedius* appears.

The Abathomphalus mayaroensis Interval Zone (uppermost Upper Maastrichtian) is in agreement with the definition by Bellier et al. (1983), representing the interval from the first appearance of the index species till the appearance of the first lowermost Danian "petites globigerines". In the Moldavides flysch deposits, Eastern Carpathians, it has been thus delimited (in Moldova-Bistriţa valleys sector of the Tarcău Nappe, Ion in Ion et al., 1988, in print; in the Lepşa Valley sector of the Vrancea half-window, Marginal Folds Unit, Ion, in Ion et al., 1987, unpubl. report) with its two subzones: the Abathomphalus mayaroensis Total Range Subzone (acc. to def. Bronnimann, 1952, as zone) and the "non nommée" Interval Subzone (Bellier et al., 1983).

The A. mayaroensis Subzone is well represented in the Carpathian area. It covers the entire range of the index species, including the appearance of *Globotruncanita falsocalcarata.*

The "non nommée" Subzone corresponds to Smit's (1977, 1979) Guembelitria cretacea Zone. We have suggested (Ion in Ion et al., 1988, in print) to name it the Plummerita hantkeninoides Subzone, this new index species being one of the most representative taxon occurring more frequently than *Guembelitria cretacea* and easier to notice.

The "non nommée" Subzone or the Plummerita hantkeninoides Subzone has its lower boundary marked by the extinction of *A. mayaroensis* and the upper one is not marked in the Moldavides flysch by the appearance of the first "petites globigerines" from the Danian base, namely from G. fringa Zone. The first levels of these "petites globigerines" noticed so far (in the Vrancea half-window, Ion-Săndulescu, 1975) belong to the Globigerina eugubina Zone, representing the second "petites globigerines" Zone. In our case, the assemblage contains the species *"Globigerina" eugubina, "G."* cf. *minutula, "G." fringa, "G." trifolia, Guembelitria*, heterohelicede specimens. Usually in the Carpathian area, the first Danian planktonic zone, easily recognized, is Morozovella pseudobulloides Zone, while within the interval of "G." fringa Zone and "G." eugubina Zone it is represented (Ion in Ion et al., 1988, in press) by the Haplophragmoides miatliukae Zone or the first part of the Valvulineria alpina Zone.

P. hantkeninoides Subzone or "non nommée" Subzone, contains two types of assemblages. The lower part assemblage consists of *Plummerita hantkeninoides, Globotruncanita falsocalcarata, Gt. stuarti, Gt. stuartiformis, Gansserina gansseri, Globotruncana falsostuarti, G. arca, G. aegyptiaca, G. rugosa, Contusotruncana fornicata, Rugotruncana subcircumnodifer, Rugoglobigerina reicheli, R. macrocephala.* The upper part assemblage is lacking in *Gt. falsocalcarata* but usually includes *P. hantkeninoides, Gt. stuartiformis, Gt. stuarti, G. arca.*

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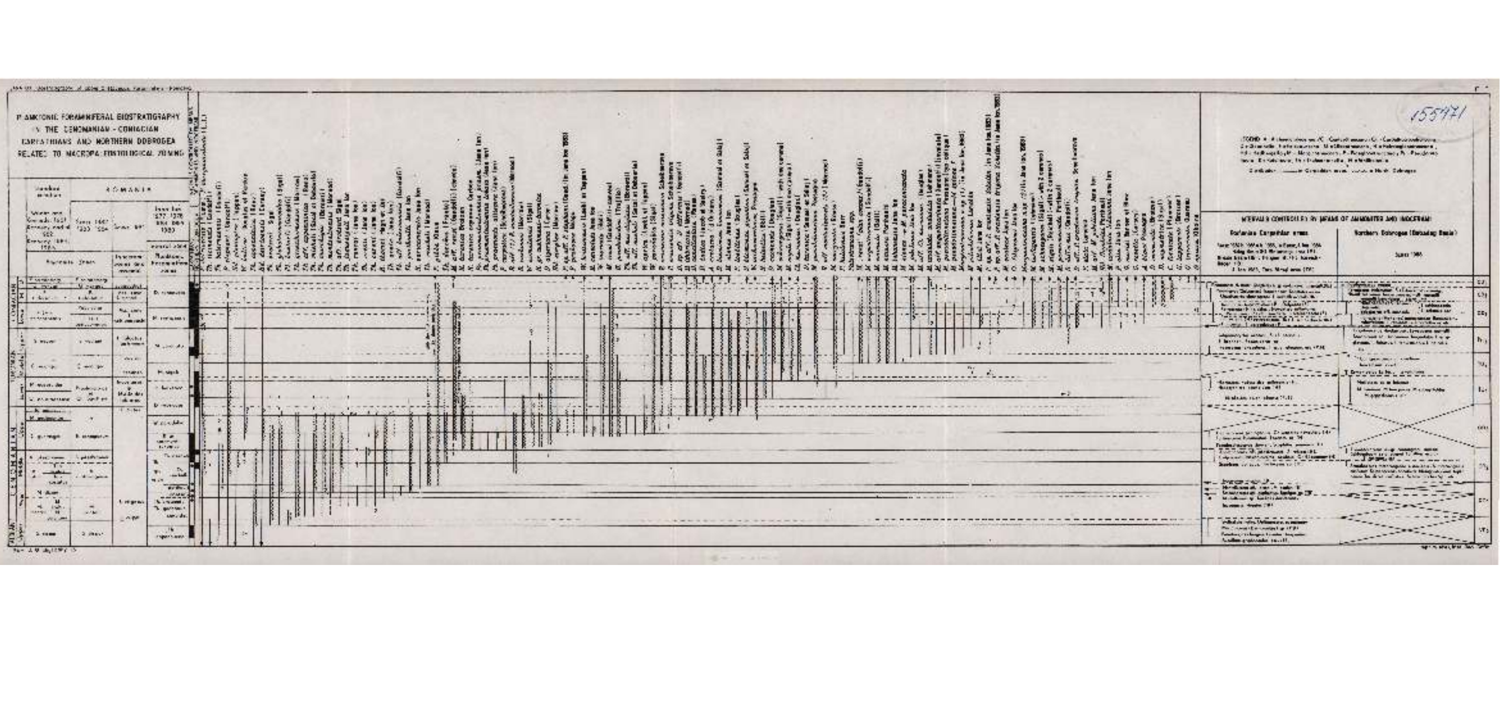
PLANKTONIC FORAMINIFERAL BIOSTRATIGRAPHY
IN THE CENOMANIAN - CONIACIAN
CARPATHIANS AND NORTHERN DOBROGEA
RELATED TO MACROPALEONTOLOGICAL ZONING
ROMANIA

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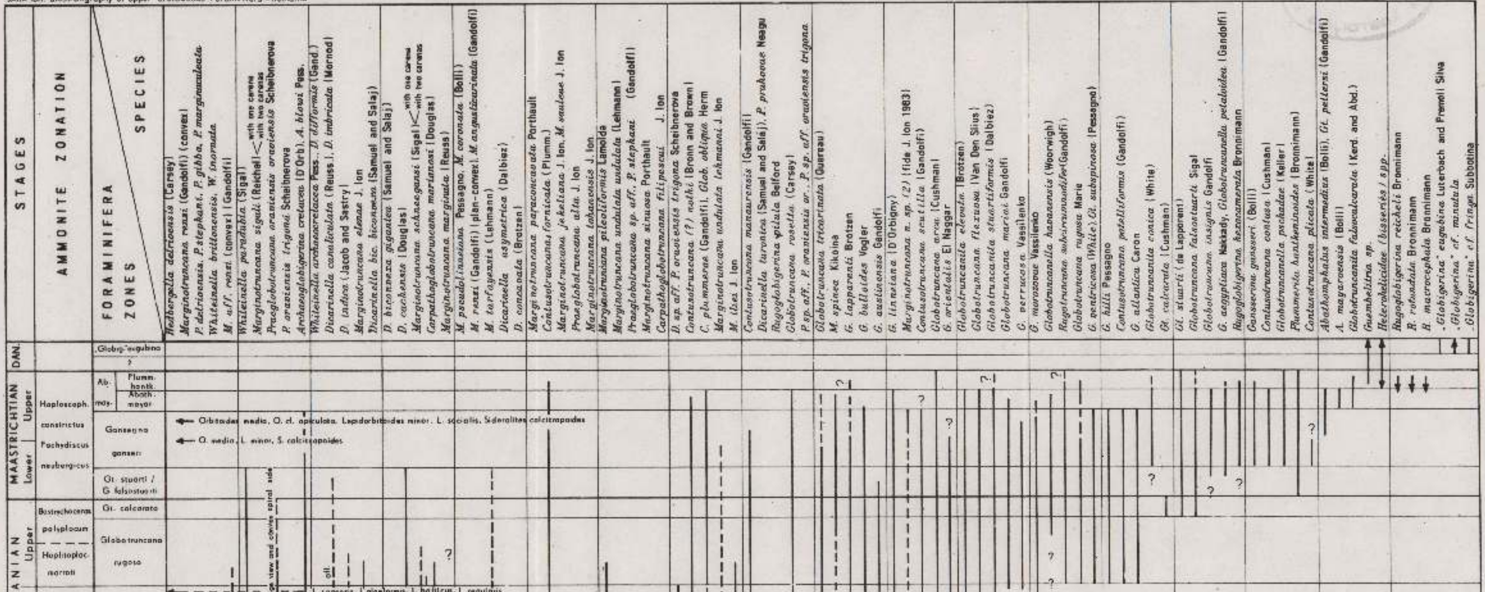
PLANKTONIC FORAMINIFERA DISTRIBUTION AND ZONATION IN THE SANTONIAN - MAASTRICHTIAN - CARPATHIANS AREA



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PRÉCISIONS CONCERNANT LA STRATIGRAPHIE DES FORMATIONS SÉDIMENTAIRES DU BASSIN DE GLODU (CARPATHES ORIENTALES)

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Key words: Upper Cretaceous-Eocene. Sedimentary basin. Sedimentary cycle. Stratigraphic units. Biostratigraphy. East Carpathians. Bistriţa Mountains. Median Dacides.

Abstract: *Considerations on the stratigraphy of the sedimentary formations in the Glodu basin (East Carpathians).* By the new researches for mapping, of revising the macrofaunas identified by our predecessors and the microbiostratigraphic (foraminifera) studies, new points of view have been considered concerning certain aspects of mapping of the Upper Cretaceous-Eocene in the Glodu Basin (Median Dacides area of the East Carpathians) but especially the stratigraphy, the ages of the lithostratigraphic subdivisions, the discontinuities. Four lithostratigraphic subdivisions has been established in this deposits: (a) the Formation with Exogyra columba, transgressively, conglomeratic in the first part, Lower-Middle Cenomanian in age, with glauconitic sandstones next, Upper Cenomanian possibly also the upper part of the Middle Cenomanian in age (according to the revised macrofauna); (b) the marly Păstinăreşti Formation, deposited after a slight stratigraphic break, is of Middle Turonian-Coniacian age (according to the revised macrofauna and after the microfauna); (c) the clayey-marly Glodu Formation follows after an important sedimentary break and is of Upper Maastrichtian-Thanetian age (according to the microfauna); (d) the transgressive "sandy- conglomeratic complex" of Ypresian(?)-Lutetian age. Each lithostratigraphic entity represents a cycle of sedimentation, separated by sedimentation and/or erosion breaks that occupy smaller or bigger time intervals. As regards the geological mapping, a new point of view is that the two synclines, where there lie the sedimentary deposits of the Glodu Basin, are separated by a strike fault, and the western limb the western syncline is normal.

1. Introduction

Les dépôts sedimentaires des bassin de Glodu représentent un reste non-érodé de la couverture sédimentaire post-autrichienne appartenant aux Dacides Médianes des Carpathes Orientales. Il sont conservés dans un pli synclinal des schistes cristallins et appartiennent au Crétacé supérieur et au Paléogène. Ici on peut délimiter plusieurs entités lithostratigraphiques représentant de divers cycles de sédimentation, séparés par des lacunes de sédimentation et/or d'érosion plus ou moins importantes.

Nos recherches, qui consistent surtout des travaux pour les cartes géologiques de Şarul Dornei et Bilbor (Carte géologique de la Roumanie, échelle 1:50.000) et des études micropaléontologiques (Ion), plus l'analyse (Szász) des associations de macrofaune connues jusqu'à présent, nous ont permis de faire quelques précisions concernant l'image cartographique du sédimentaire du bassin de Glodu (fig. 1), mais surtout concernant l'âge de diverses entités lithostratigraphiques et les lacunes stratigraphiques qui les séparent (pl. I–IV). A l'occasion de la présente étude nous introduisons les dénominations, la formation à *Exogyra columba*, la Formation de Păstinăreşti, la formation de Glodu, pour trois des entités lithostratigraphiques séparées dans le bassin de Glodu. Pour la quatrième, encore incomplètement étudiée, nous employons la dénomination de "complexe gréseux-conglomératique".

Dès le dernier siècle, on connait des données concernant la constitution géologique du bassin de Glodu (Ştefănescu, 1888; Uhlig, 1897; Athanasiu, 1898, 1899). Le dernier des auteurs cités considère, à base des arguments macropaléontologiques, que dans ce bassin on trouve le Cénomanien, le Turonien et au moins une partie du Sénonien. Les premières remar-

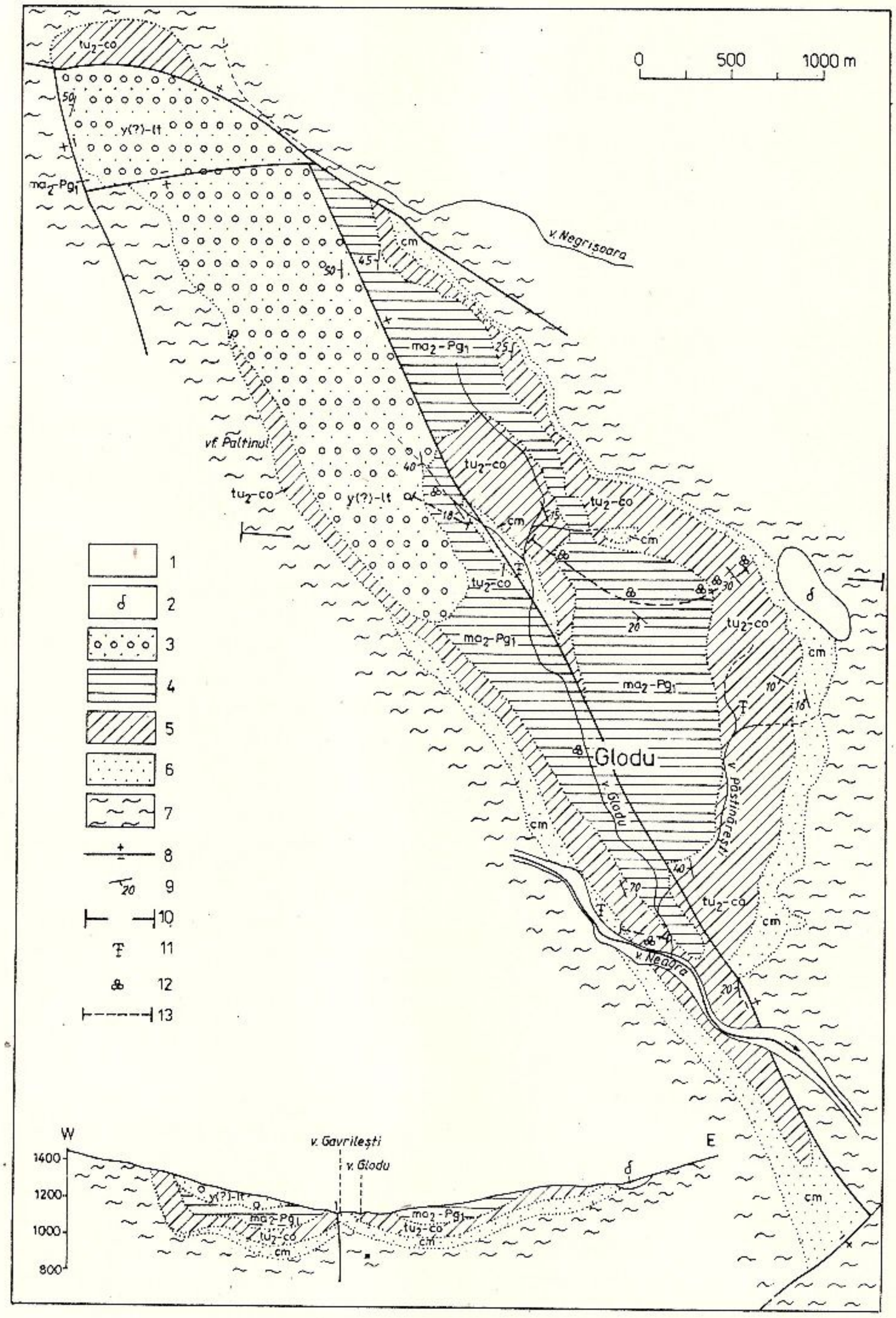

Fig. 1 — Carte géologique du bassin de Glodu (selon Szász). 1, Quaternaire: alluvions; 2, Malvensien: dacites du type de Drăgoiasa; 3, Lutétien-Yprésien (?): alternance de grès et d'argiles grises, conglomérats calcaires; 4, Paléocène-Maestrichtien supérieur: alternance d'argiles rouges et de grès; 5, Coniacien-Turonien moyen: marnes et marno-calcaires rouges et grises à ammonites et inocérames; 6, Cénomanien: conglomérats polymictiques, conglomérats quartzitiques, grès et siltites glauconique à mollusques; 7, Soubassement pré-cénomanien: schistes cristallins de divers types; 8, Failles; 9, Position des couches; 10, Tracé du coupe géologique; 11, Gisement fossilifère - macrofaune; 12, Microfaune (foraminifères); 13, Profil micropaléontologique.

ques sur l'existence de la microfaune du bassin de Glodu appartiennent à Kräutner (1937), qui cite du complexe marneux *"Rosalina linnei"*.

Une étude de détail sur le sédimentaire du bassin de Glodu a été faite par Mutihac (1959) qui, à base de la macrofaune connue antérieurement, ou à celle collectée par soi-même, tout comme de la microfaune (les déterminations appartiennent à Tocorjescu et Mihăilescu), met en évidence l'existence du Cénomanien, du Turonien, du Sénonien et du Paléogène (qu'il attribue à l'Eocène inférieur, Mutihac, 1959; plus tard, Mutihac et Ionesi, 1974, l'attribuent au Danien-Paléocène, sans une argumentation supplementaire).

Plus récemment, Alexandrescu et Brustur (1975) reprennent l'étude des dépots de Glodu, en mettant en évidence la présence du microfaciès à *Pithonella ovalis*. Par rapport aux autres régions carpathiques ou de l'avant-pays, les auteurs arrivent à la conclusion que le microfaciès à *Pithonella ovalis* est cantonn dans l'intervalle stratigraphique Turonien-Coniacien, représenté par des marnes rouges et grises. Pour l'âge d'autres complexes lithologiques il n'y a pas d'arguments nouveaux.

La plus récente étude qui a comme sujet la géologie du bassin de Glodu appartient à Preda, Todiriţă et Pauliuc (1980). Sur la carte qui accompagne l'étude on observe certaines modifications concernant l'image cartographique du bassin de Glodu par rapport à la carte présentée par Mutihac (1959). Les séparations lithologiques sont pourtant identiques, mais avec quelques modifications concernant leurs ages. A ceux-ci on va revenir.

2. Données lithostratigraphiques et biostratigraphiques

2.1. Formation à Exogyra columba (Cénomanien)

C'est le premier terme de la couverture sédimentaire post-autrichienne conservé à présent dans diverses régions seulement sous forme de lambeaux plus ou moins étendus.

La formation à *Exogyra columba* correspond (fig. 2) à ce qu'on a séparé par les recherches antérieures (Mutihac, 1959) dans le bassin de Glodu comme l'horizon des conglomérats basaux (Cénomanien inférieur, Mutihac, 1959; Vraconien, Preda et al., 1980) et l'horizon des grès glauconiques (Cénomanien supérieur, Mutihac, 1959; Cénomanien, Preda et al., 1980).

Dans le bassin de Glodu cette formation contient en base des conglomérats polymictiques ou des conglomérats surtout quartzitiques décrits par les chercheurs antérieurs comme un horizon à part. De ce niveau on passe d'habitude graduellement aux grès plus ou moins glauconiques, plus ou moins calcaires et riches en restes d'organismes. Il contient surtout des mollusques, dont l'élément caractéristique pour le segment orientale des Dacides Medianes est *Exogyra (Rhynchostreon) columba*, à laquelle s'ajoutent des espèces du genre *Pinna, Neithea, Pycnodonte*, puis des diverses espèces d'ammonites, gastéropodes, coelentérés, serpulides etc. Au dehors du périmètre étudié (au nord des monts Bîrgău, au bassin de Ţibău, Maramureş etc.) la succession est complétée par des grès calcaires fins et des siltites argileuses verdâtres avec un riche contenu d'inocérames et d'ammonites d'âge turonien inférieur (fide Szász, 1982, 1986).

L'âge de cette formation est établi à partir de la faune d'ammonites et de mollusques. Ainsi, Mutihac (1959) cite de Glodu (de la collection de S. Athanasiu) *Acanthoceras rhotomagense*, et Preda et al. (1980) cite *Mantelliceras mantelli, Calycoceras* sp. etc. En jugeant selon cette faune, à Glodu serait présent (cf. Preda et al., 1980) tout le Cénomanien dans l'horizon gréseux-glauconique; dans ce cas les conglomérats sous-jacents seraient plus anciens, peut-être vraconiens. En étudiant tout de même la faune citée, on a constaté que le spécimen figuré par Mutihac (1959, pl. 1, fig. 1) appartient en réalité au genre *Eucalycoceras* (*E. spathi* cf. Szász, 1983, pl. 20, fig. 2a, b) d'age cénomanien supérieur. Quant aux spécimens de *Mantelliceras mantelli*, cités par Preda et al. (1980, pl. 4, fig. 7; pl. 5, fig. 2) on constate qu'il s'agit des exemplaires très déformés qui, selon leur ornementation, appartiennent plutôt au genre *Eucalycoceras* ou *Pseudocalycoceras*. Pour cette interprétation plaide aussi le niveau stratigraphique indiqué par le reste de la faune (*Eucalycoceras, Calycoceras*), tout comme les spécimens trouvés dans d'autres régions (*Pseudocalycoceras lattense* - figuré par Szász, 1983, sous le nom de *Pseudocalycoceras* sp., pl. 18, fig. 3a, b). Généralement, on peut admettre que le niveau riche en faune du Cénomanien du bassin de Glodu, tout comme des autres régions des Dacides Medianes des Carpathes Orientales, représente le Cénomanien supérieur, tout au plus la partie terminale du Cénomanien moyen et les niveaux basaux conglomératiques, généralement non-fossilifères ne sont pas plus anciens que le Cénomanien inférieur (fig. 3).

2.2. Formation de Păstinăreşti (Turonien moyen-Coniacien)

Cette formation repose, après une légère lacune stratigraphique, sur les dépots cénomaniens et dans certaines régions (le flanc occidental du synclinorium) elle les dépasse et repose directement sur le soubassement cristallin. La meme situation est observée aussi dans l'extrémité septentrionale du bassin, à Păltiniş

La Formation de Păstinăreşti (fig. 2) correspond ap-

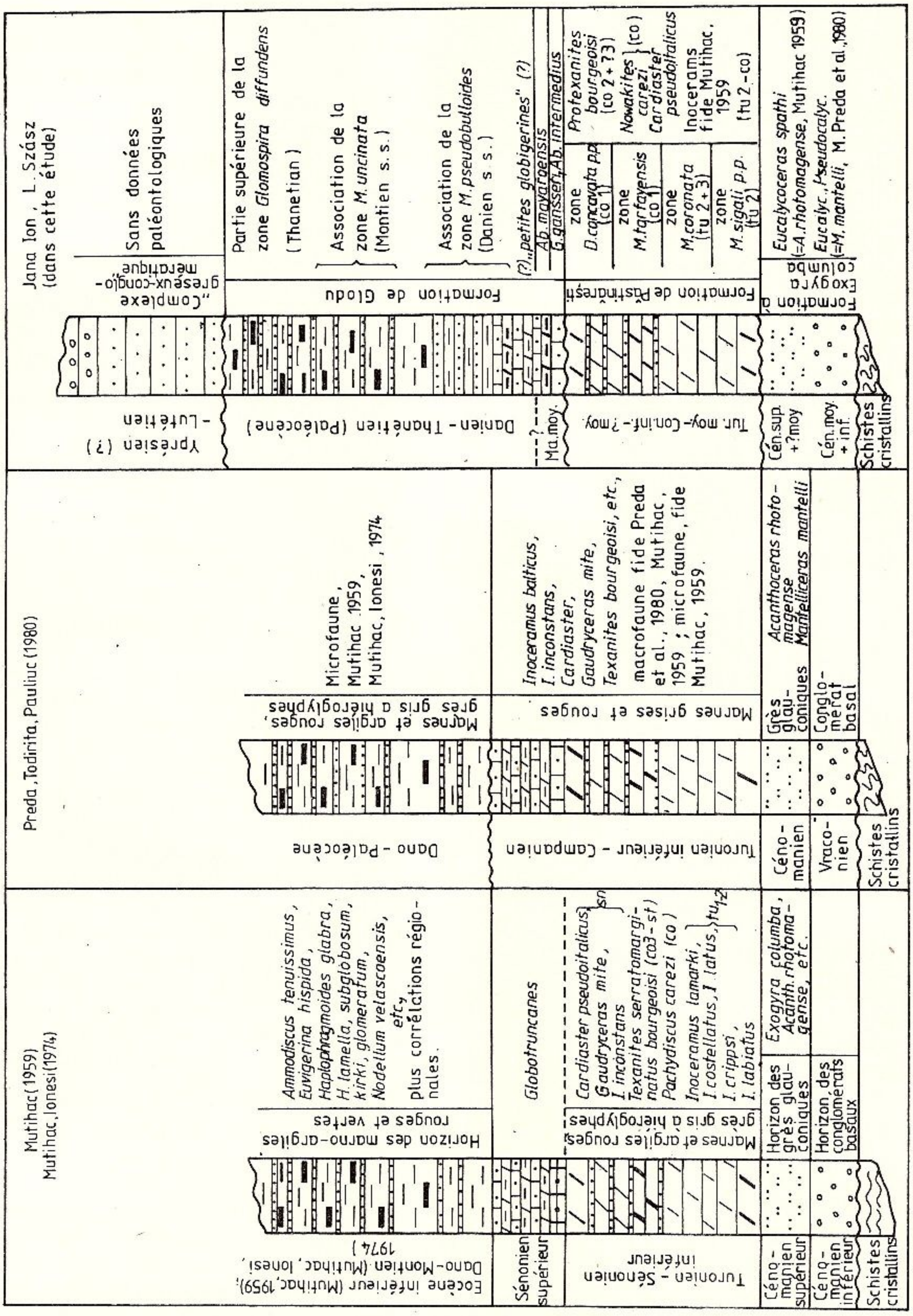

Fig. 2 – Etapes dans la connaissance de la stratigraphie du sédimentaire du bassin de Glodu: tu1 = Turonien inférieur; tu2 = Turonien moyen; tu3 = Turonien supérieur; co = Coniacien; co 1 = Coniacien inférieur; co 2 = Coniacien moyen; co 3 = Coniacien supérieur; st = Santonien; sn = Senonien.

proximativement à la partie inférieure de l'horizon des marnes rouges et des grès gris, séparé par Mutihac (1959), attribuée (macrofaune et microfaune) au Turonien-Sénonien inférieur. Preda et al. (1980) considèrent qu'elle commence dès le Turonien inférieur.

Du point de vue lithologique, la Formation de Păstinăreşti est formée de paquets de marnocalcaires en couches bien individualisées, surtout rouges à la partie inférieure de la formation, plus rarement noirâtres vers la partie supérieure, où il y a aussi des intercalations centimétriques jusqu'à décimétriques de grès fins et durs. L'épaisseur de toute la succession peut avoir 100 m. Des affleurements bien individualisés dans ces dépôts se trouvent sur la rive gauche du ruisseau de Neagra Broştenilor (en amont de la confluence avec le ruisseau de Glodu), dans la vallée de Păstinăreşti, sur la colline de Surduc, sur la route située au nord de la vallée de Păstinăreşti etc.

L'âge que nous avons attribué à la Formation de Păstinăreşti a comme arguments la faune d'inocérames et d'ammonites citée par les chercheurs antérieurs et revu par l'un de nous (Szász), tout comme les associations de foraminifères planctoniques mises en évidence par l'un de nous (Ion).

La plupart des espèces d'inocérames figurées par Mutihac (1959) et par Preda et al. (1980) plaident pour l'âge turonien moyen-coniacien, la faune étant assez similaire à celle connue des monts Perşani. Dans toutes les deux études on cite la présence de l'espèce *Inoceramus labiatus*. Le spécimen figuré sous ce nom par Mutihac (pl. 1, fig. 5) fait partie, selon notre opinion, du groupe *I. striatoconcentricus* et celui cité par Preda et al. n'est pas figuré. Nous avons des doutes sur la présence des inocerames du groupe *Mytiloides labiatus* dans le complèxe marneux, car dans d'autres régions les espèces de ce groupe sont présentes dans les siltites grises au-dessous du complexe marneux, des dépôts qui sont absents dans le bassin de Glodu.

Quant aux ammonites, les espèces citées par Mutihac (1959), à savoir *Nowakites carezi* et *Protexanites bourgeoisi*, sont coniaciennes, tout comme l'échinide *Cardiaster pseudoitalicus*, très fréquent dans le Coniacien de Perşani. Les arguments macropaléontologiques pour les âges plus nouveaux que le Coniacien pour la Formation de Păstinăreşti, nous les considérons non-concluants, car il s'agit de déterminations spécifiques erronées.

Les recherches micropaléontologiques (foraminifères) récentes (Ion in Peltz et al., 1987, 1988, rapports nonpubl.), à l'occasion de l'élaboration de la carte Bilbor, ont montré que la Formation de Păstinăreşti est formée des dépots du Turonien moyen-Coniacien inférieur, pour les suivantes raisons (pl. I–IV):

(a) – La partie inférieure de la formation (10–15 m d'épaisseur; marnes et marnocalcaires gris et rouges) contient tant sur le profil de la route située au nord de la vallée de Păstinăreşti (le flanc oriental du synclinorium de Glodu), que sur le profil de la vallée de Neagra Broştenilor (le flanc occidental du même synclinorium), l'association de foraminifères planctoniques, caractéristique pour le Turonien moyen pro parte, qui appartient à la zone à Marginotruncana sigali (cf. def. Robaszynski, 1983). L'association est formée de: *Helvetoglobotruncana helvetica, Marginotruncana sigali, M. schneegansi, M. elenae, M. renzi* (convexe), *M.* aff. *renzi* (convexe), *M. pileoliformis, Dicarinella turonica, D. imbricata, D. cachensis, Carpatoglobotruncana marianosi, Praeglobotruncana oraviensis oraviensis, P. oraviensis trigona, P. gibba, P. stephani, P. prahovae, P.* aff. *stephani, Archaeoglobigerina cretacea, Whiteinella paradubia, Falsotruncana* spp.; on trouve aussi *Uvigerinammina jankoi* et des ostracodes.

(b) – A la partie médiane (10–15 m d'épaisseur: marnes et marnocalcaires gris) sur le profil de la route située au nord de la vallée de Păstinăreşti, se trouve l'association de la zone à Marginotruncana coronata (cf. def. Ion, 1979, fide 1982 – mais comme une sous zone caractéristique pour la partie terminale du Turonien moyen-Turonien supérieur. L'association est formée de: *M. coronata, M. "renzi"* (plan-convexe), *M. pseudolinneiana, M. sinuosa, M. angusticarinata, M. sigali, M. schneegansi, M. renzi* (convexe), *M. pileoliformis, Dicarinella imbricata, D. canaliculata, Praeglobotruncana gibba, P. stephani*, plus *Pithonella* spp.

(c) – La partie supérieure (15 m d'épaisseur; marnes et marnocalcaires gris et rouges à intercalations de grès) est caractérisée sur les deux flancs du bassin (les profils mentionnés) par l'association de foraminifères planctoniques de la zone à Marginotruncana tarfayensis (cf. def. Ion, 1979, fide 1982 – mais comme une sous-zone.), qui représente la partie inférieure du Coniacien inférieur. L'association est formée de: *M. tarfayensis, M. paraconcavata, M. undulata undulata, M. n.* sp. (2), *M. "renzi"* (plan-convexe), *M. renzi* (convexe), *M. sinuosa, M. pseudolinneiana, M. coronata, M. marginata, M. angusticarinata, M. schneegansi, M. sigali, M. elenae, Dicarinella* aff. *primitiva, Carpathoglobotruncana marianosi, Dicarinella cachensis, D. canaliculata, D.* aff. *imbricata, D. imbricata, Archaeoglobigerina cretacea, Praeglobotruncana oraviensis trigona, Falsotruncana* spp.; elle contient aussi *Pithonella* spp.

(d) – La plus jeune association de foraminifères de la partie supérieure de la Formation de Păstinăreşti serait celle à *Dicarinella concavata, D.* cf. *asymetrica, Marginotruncana undulata undulata, M. schneegansi, M. sinuosa, M. coronata, M. tarfayensis, M. pseudolinneiana, M. marginata, M. renzi* (convexe), *M. angusticarinata* etc., identifiée sur le profil de la vallée de Neagra Broştenilor. Elle appartient à la

zone à Dicarinella concavata (cf. def. Sigal, 1955, emend. Ion, 1983), spécifique pour l'intervalle Coniacien inférieur (pro-parte) – Santonien inférieur. Mais le fait que (l'association est mise en évidence en lames minces) *D. asymetrica* n'est pas représentée par un exemplaire bien conservé et *Contusotruncana fornicata* est absente, nous avons des doutes en ce qui concerne l'extension de l'âge du complexe marneux plus haut que le Coniacien inférieur. La présence du Coniacien moyen (éventuellement le Coniacien supérieur) serait suggérée encore par l'espèce *Protexanites bourgeoisi* citée (Mutihac, 1959) dans ces dépots, mais que, malheureusement, nous ne l'avons pas retrouvée dans la collection. Par conséquent, pour l'extension de la Formation de Păstinăreşti en haut du Coniacien inférieur, les données ne sont pas encore certaines et la question reste ouverte.

2.3. Formation de Glodu (Maestrichtien supérieur-Paléocène)

Cette formation est toujours transgressive, en reposant sur les dépots plus anciens après une lacune stratigraphique qui contient peut-être au moins la plus grande partie du Santonien, plus le Campanien et le Maestrichtien inférieur. Il faut mentionner que dans les études précédentes on n'a pas clarifié les relations entre les dépots de cette formation et les dépots crétacés sous-jacents.

La Formation de Glodu correspond (fig. 2), dans la succession stratigraphique donnée par Mutihac (1959), à la partie approximativement supérieure (à grès fins blanchatres) de l'horizon des marnes rouges et des grès gris, attribuée au Sénonien supérieur à base d'une microfaune (diagénisée, non-déterminée spécifiquement) plus l'horizon des marno-argiles vertes et des grès à hiéroglyphes attribuées premièrement à l'Eocène inférieur, puis (1974) au Danien-Paléocène (à base de la meme microfaune). Elle correspond à une partie des marnes grises et rouges, turonien inférieur-campaniennes, appartenant au premier cycle de sédimentation, plus les dépots (marnes, argiles rouges, grès à hiéroglyphes) du deuxième cycle paléogène, de la succession stratigraphique du bassin de Glodu présentée par Preda et al. (1980).

La formation de Glodu est formée d'argiles (parfois de marnes aussi) rouge-bordeaux où sont intercalés des grès, généralement à épaisseurs centimétriques, parfois décimétriques, gris-verdatre, à hiéroglyphes, diaclasés, parfois courbicorticaux. Dans certains secteurs (surtout sur le flanc oriental du synclinorium) près de la base de la formation, il y a un paquet à couches centimétriques de grès calcaires ou de calcarénites très riches en globotruncanidés. L'épaisseur de toute la succession est de jusqu'à 100 m.

L'âge de la Formation de Glodu a été établi exclusivement sur des critères micropaéontologiques (Ion in Peltz et al., 1987, 1988, données nonpubl.). Il est Maestrichtien supérieur-Paléocène, car (pl. I–IV):

(a) – Sa partie basale (jusqu'à 10 m d'épaisseur; marno-argiles rouges à intercalations de grès calcaires et/or calcaires gréseux) contient premièrement l'association à *Plummerita hantkeninoides* ou à *Gansserina gansseri* et des exemplaires de passage d'*Abathomphalus intermedius* à *A. mayaroensis* (le flanc oriental du bassin sur la route du nord de la vallée de Păstinăreşti, tout comme près de la faille qui affecte le centre du bassin), suivie de l'association à *A. mayaroensis* (le versant gauche de la vallée du Glodu, la route du nord de la vallée de Păstinăreşti/vallée de Neagra Broştenilor).

La première association indique la partie inférieure du Maestrichtien supérieur (appartenant à la partie supérieure de la zone à Gansserina gansseri, cf. def. Bronnimann, 1952). Elle contient, *in situ*, *Gansserina gansseri*, *A. intermedius*, *Plummerita hantkeninoides*, *Globotruncana insignis*, *G. dupeublei*, *Contusotruncana contusa*, des exemplaires de passage d'*A. intermedius* à *A. mayaroensis*. Ici on trouve remaniées beaucoup d'espèces de foraminifères planctoniques (*H. helvetica*, *P. oraviensis*, *M. pileoliformis*, *M. sigali*, *M. schneegansi*, *M. tarfayensis*, *M.* "*renzi*" – plan convexe –, *M. renzi* – convexe), appartenant aux associations du Turonien moyen-Coniacien, donc à la formation sous-jacente.

La deuxième association, à *A. mayaroensis*, est caractéristique pour la partie supérieure du Maestrichtien supérieur (la zone à A. mayaroensis, cf. def. Bellier et al., 1983) et contient dans notre cas l'espèce index plus: *Gansserina gansseri*, *Plummerita hantkeninoides*, *Globotruncana arca*, *G. linneiana*, *G. lapparenti*, *Globotruncanita conica*, *Gt. stuartiformis*, *Gt. stuarti*, *Globotruncanella havanensis*, dans le versant gauche de la vallée du Glodu, près de la faille qui affecte le centre du bassin: *A. intermedius*, *P. hantkeninoides*, *Contusotruncana walfischensis*, *C. plummerae*, *C. contusa*, *Globotruncanita falsocalcarata*, *Rugoglobigerina rugosa*, *R. macrocephala*, sur la route du nord de la vallée de Păstinăreşti.

(b) – Vers la partie supérieure du paquet basal à calcaires gréseux (dans la vallée de Păstinăreşti), les grès contiennent, en sections minces, dans certaines agglomérations de microfaune planctonique non-déterminable généralement, des exemplaires très petits de *Siderolites*, beaucoup de *Rugoglobigerina*, *Globotruncana aegyptiaca*, *Globotruncanella petaloides*, *Globotruncanita* cf. *stuarti* et de très petits exemplaires (sous 0.12 mm) à trois loges qui pourraient appartenir aux "globigérines petites" et donc seulement celles-ci seraient in situ (les premiers niveaux du

Danien ?).

(c) – Au-dessus ou dans les premiers niveaux à grès micafères moux qui suivent au-dessus du paquet à grès et gréso-calcaires, les pélites comportent (le profil de la route du nord de la vallée de Păstinăreşti et sur la rive gauche de la vallée du Glodu) de petits exemplaires du type *Subbotina triloculinoides* et *Globoconusa daubjergensis*, plus microfaune remaniée (*H. helvetica, H. praehelvetica, P. stephani* etc) des dépots crétacés pré-maestrichtiens. En échange, les grès contiennent comme remaniée une microfaune maestrichtien supérieure (*Siderolites, Gt. stuarti, Gt. conica, Gt. subspinosa, C. contusa, G. aegyptiaca, P.hantkeninoides, A. mayaroensis, Rugoglobigerina* spp.), plus in situ *Globoconusa daubjergensis, Subbotina triloculinoides.* Nous avons identifié aussi l'association à *G. daubjergensis* et *S. triloculinoides* dans le synclinal occidental (vallée de Gavrileşti) du bassin de Glodu.

L'association à *Subbotina triloculinoides, Morozovella pseudobulloides, Globoconusa daubjergensis* (respectivement la zone à Morozovella pseudobulloides, def. Leonov et Alimarina, 1961, emend. Bolli, 1966) est caractéristique pour la base du Danien. C'est l'association la plus répandue qui indique d'habitude la présence de la base du Danien. Les premières associations du Danien sont en réalité celle à *"Globigerina" fringa* suivie par celle à *"Globigerina" eugubina* (la zone à G. fringa cf. Herm et al., 1981 et la zone à G. eugubina, def. Luterbacher et Premoli Silva, 1964 - fide Toumarkine et Luterbacher, 1985); l'association à *M. pseudobulloides* est la troisième qui apparait. Les premières deux association ou zones sont difficilement à séparer à cause des "globigérines petites" (sous 0,10 mm) et car elles occupent de petites épaisseurs de dépots (de quelques centimètres jusqu'à maximum quelques mètres) selon les données connues du monde entier. Dans notre pays, pratiquement dans les séries de flysch avec une continuité de sédimentation, le commencement du Danien est repéré par l'apparition de l'association à *M. pseudobulloides*. Seulement dans la Nappe des plis marginaux, à la partie terminale des couches de Lepşa de la demi-fenêtre de Vrancea, on connait (Ion-Săndulescu, 1975) la zone *"G." eugubina*.

Quant à la microfaune crétacée remaniée qui accompagne celle danienne du bassin de Glodu, nous sommes d'avis que ces remaniements sont souvent rencontrés dans les niveaux paléocènes du flysch des Carpathes Orientales (par exemple, dans les couches avec la microfaune de la zone à "G." eugubina ou avec celle de la zone à M. pseudobulloides, ou de la zone à M. uncinata, fide Ion in Antonescu et al., 1988, données nonpubl.).

(d) – Vers la partie médiane de la formation de Glodu, à l'aspect flyschoïde, on a identifié (d'une part et de l'autre de l'axe du synclinal oriental), quelques associations à exemplaires du type *Morozovella uncinata, M. inconstans, M. pseudobulloides, Planorotalites compressa, Subbotina triloculinoides, Globoconusa daubjergensis* (type *koslowskii*), qui appartient à la partie inférieure de la zone à M. uncinata (def. Bolli, 1957, emend. Bolli, 1966), donc au Montien s.s.

(e) – Dans le reste du paquet flyschoïde (25–50 m) on a identifié jusqu'à présent surtout des associations de foraminifères benthoniques, surtout d'agglutinants, avec beaucoup d'espèces de *Recurvoides, Trochamminoides, Kalamopsis grzybowski, Plectina coniformis, P. lenis, Reophax splendidus, R. pilulifera, Dendrophrya* spp. , *Rhabdammina* spp. etc., communes au moins pour le Sénonien-Paléocène.

Y font exception, étant significatives du point de vue chronostratigraphique, deux associations dont une identifié (sur la route du nord de la vallée de Păstinăreşti) dans les argiles variolées de l'axe du synclinal oriental, donc évidemment à la partie supérieure de la Formation de Glodu. Elle contient: *Rzehakina complanata, Haplophragmoides miatlukae, Nodellum velascoense, Hormosina ovulum ovulum, Nuttallides truempyi, Plectina coniformis.* La deuxième association provient de la rive droite de la vallée du Glodu, du synclinal occidental et contient: *Glomospira diffundens, Nodellum velascoense, Hormosina ovulum ovulum, Plectina coniformis, P. conversa, Recurvoides* spp., *Dendrophrya* spp.

Toutes les deux associations sont caractéristiques pour le Thanétien carpathique. Elles appartiennent à la partie supérieure de la zone à Glomospira diffundens (conf. Ion: Săndulescu, 1973, p. 20, tab. V; Ion in Ion et al., 1985, p. 121; emend. Ion in Antonescu et al., 1988, données nonpubl. et in Ion et al., 1988, sous presse) ou à la sous-zone à Spiroplectammina spectabilis (conf. def. Ion in Antonescu et al., 1988, données nonpubl. et in Ion et al., 1988, sous presse). Dans les niveaux de cet age sont frequents les remaniements de microfaune d'age crétacé moyen.

2.4. *"Complexe gréseux-conglomératique" (Yprésien ? - Lutétien)*

Le dernier paquet de la succession de dépots sédimentaires du bassin de Glodu est formé de bancs de grès jusqu'à épaisseurs métriques, avec des intercalations d'argiles grises. Vers la partie supérieure du paquet il y a des microconglomérats et des conglomérats calcaires. Dans ces dépôts on n'a pas trouvé de macrofaune ou de microfaune. Leur age éocène a été apprécié par corrélation avec les dépots de l'est des monts Bîrgău, similaires du point de vue lithostratigraphique. L'épaisseur de ce complexe peut être appréciée à plus de 100 m, à épaisseurs maximum dans la moitié septentrionale et occidentale du bassin de Glodu.

3. Structure du bassin de Glodu

Du point de vue structural, dans le bassin de Glodu il y a deux synclinaux qui viennent en contact au long d'une faille longitudinale. Le synclinal oriental est plus large et plus court. Il faut souligner que les dépôts cénomaniens de la vallée du Glodu sont à la base du flanc occidental faillé du synclinal oriental et pas comme une boutonnière dans l'axe d'un anticlinal central, selon les données antérieures. Le synclinal occidental est 2 km environ plus long que celui oriental, et il est plus écrasé, avec le flanc occidental redressé à la verticale. Sur ce flanc le contact des séries sédimentaires avec le soubassement cristallin est normal et pas faillé, selon Preda et al., 1980. On remarque aussi que nous n'avons pas réussi à mettre en évidence les deux failles transversales figurées sur la carte des auteurs cités.

4. Conclusions

Le sédimentaire du bassin de Glodu est formé de quatre formations distinctes qui correspondent à quatre cycles de sédimentation séparés par des lacunes plus ou moins importantes.

– Le premier cycle est représenté par la formation à *Exogyra columba*, conglomératique-gréseuse, cénomanienne. Son âge est établi à base de la macrofaune qui contient des espèces caractéristiques pour le Cénomanien supérieur et la partie supérieure du Cénomanien moyen. Cette formation contient des conglomérats polymictiques ou quartzitiques, suivis par des grès glauconiques à mollusques. Les dépôts de ce type sont très répandus dans les Dacides Medianes des Carpathes Orientales, et peuvent inclure aussi des grès fins et des siltites argileuses à macrofaune turonien inférieure. Donc, sur l'ensemble des Dacides Medianes des Carpathes Orientales, la durée de ce cycle est jusque dans le Turonien inférieur.

– Le deuxième cycle de sédimentation est représenté par les dépôts de la Formation de Păstinăreşti, marneuse, d'âge turonien moyen-coniacien (éventuellement seulement coniacien inférieure). Des arguments pour cet âge sont offerts tant par la macrofaune, que par la microfaune. Des successions semblables à la Formation de Păstinăreşti ont une large distribution dans les Dacides Medianes des Carpathes Orientales, mais jusqu'à présent les dépôts marneux de Glodu sont les plus riches en macrofaune, en formant un étalon pour les successions marneuses de cette zone.

– Le troisième cycle de sédimentation est représenté par la Formation de Glodu, argileuse-gréseuse, à argiles rouges à intercalations de grès, qui appartient à l'intervalle Maestrichtien supérieur-Paléocène. Elle suit après une importante lacune stratigraphique et débute par des dépôts très condensés du point de vue stratigraphique. Ces dépôts ont été attribués par des chercheurs antérieurs soit à l'Eocène inférieur, soit au Dano-Paléocène, sans arguments certains. Dans la présente étude on montre, par la microfaune, que la Formation de Glodu commence dès le Maestrichtien supérieur, elle contient aussi le Paléocène, et qu'entre le Maestrichtien et le Paléocène il y a une continuité de sédimentation.

– Finalement, le quatrième cycle de sédimentation, mis en évidence dans cette étude, est représenté par le "complexe gréseux-conglomératique" peut-être d'âge éocène. Il est évidemment transgressifet contient des grès à intercalations d'argiles grises, microconglomérats et conglomérats. Jusqu'à présent, nous n'avons pas d'arguments paléontologiques directs pour l'âge de ces dépôts.

– Du point de vue structural, le bassin de Glodu représente un synclinorium formé de deux synclinaux (celui oriental plus court et plus large, celui occidental plus écrasé et plus long), qui viennent en contact au long d'une faille longitudinale.

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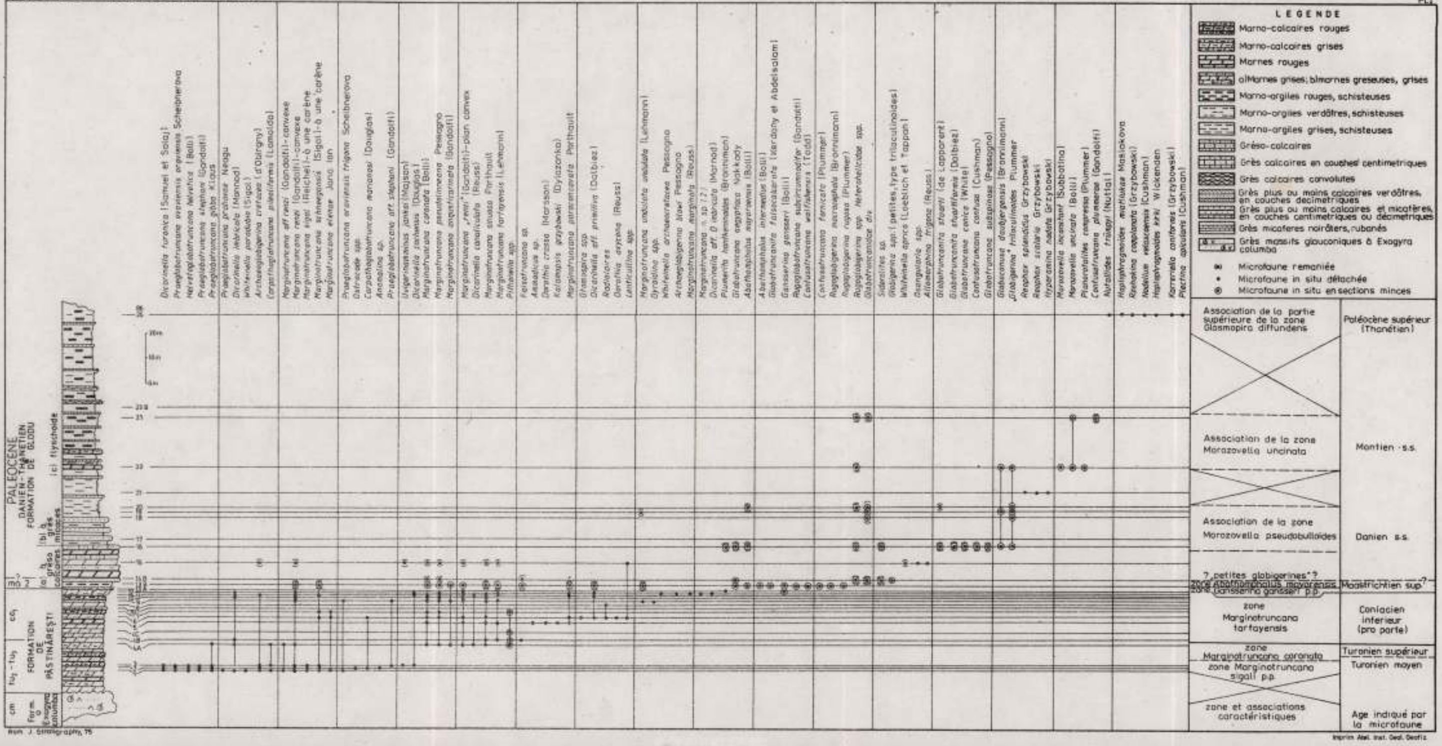
LE TURONIEN-PALÉOCÈNE DU BASSIN DE GLODU, LE PROFIL DE LA ROUTE SITUÉE AU NORD DE LA VALLÉE DE PĂSTINĂREŞTI (FLANC ORIENTAL DU SYNCLINAL D'EST)

PL. I



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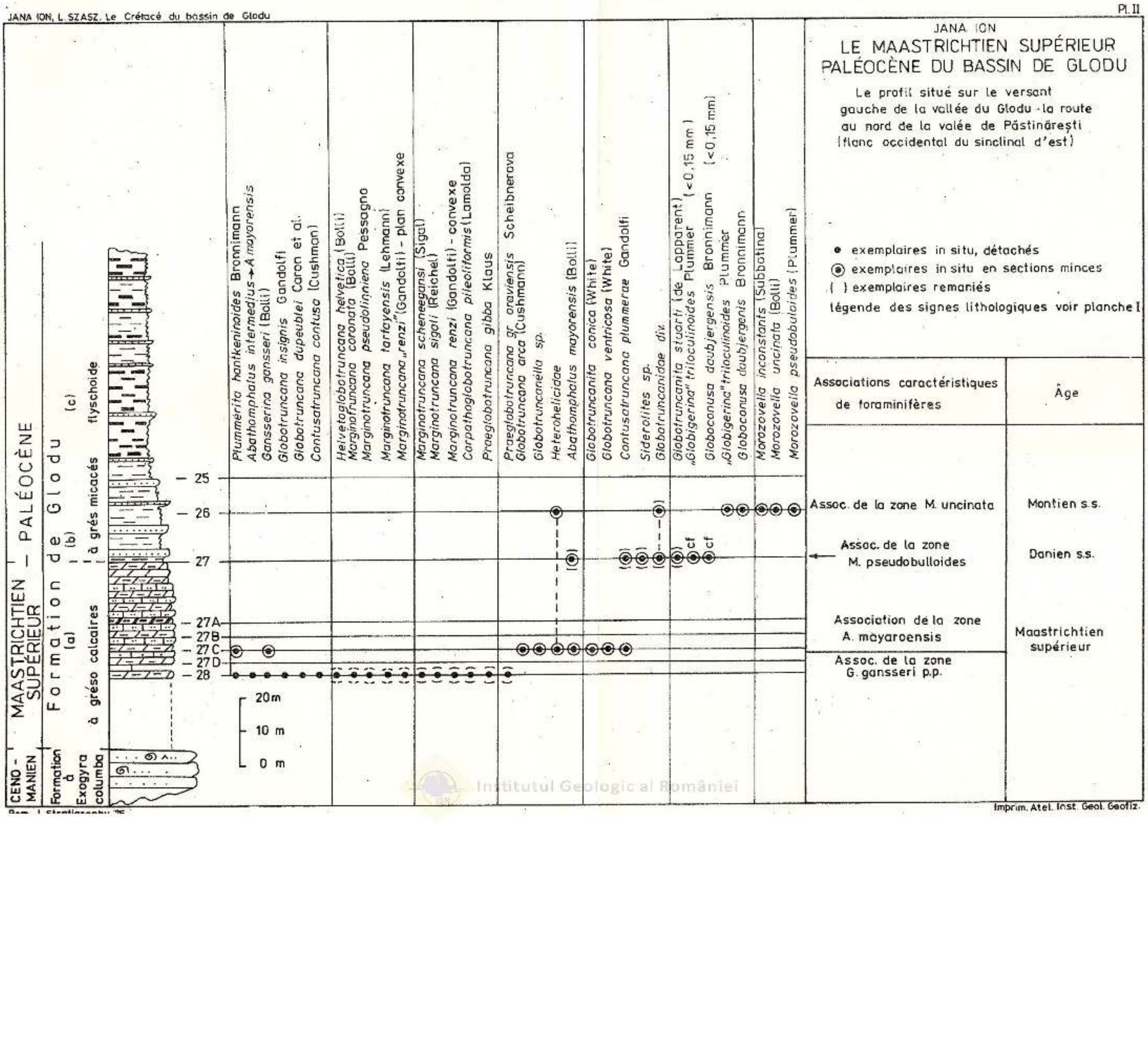

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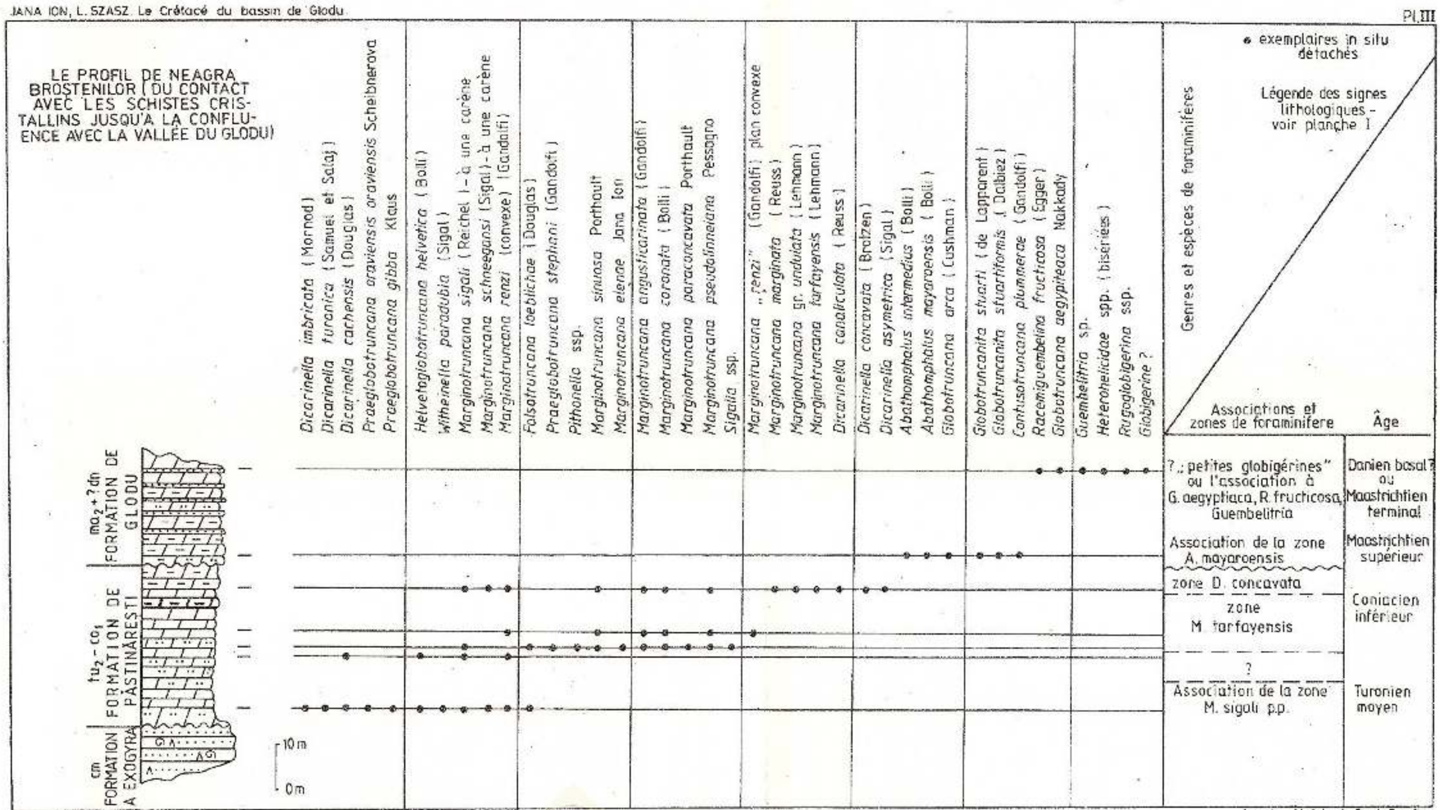

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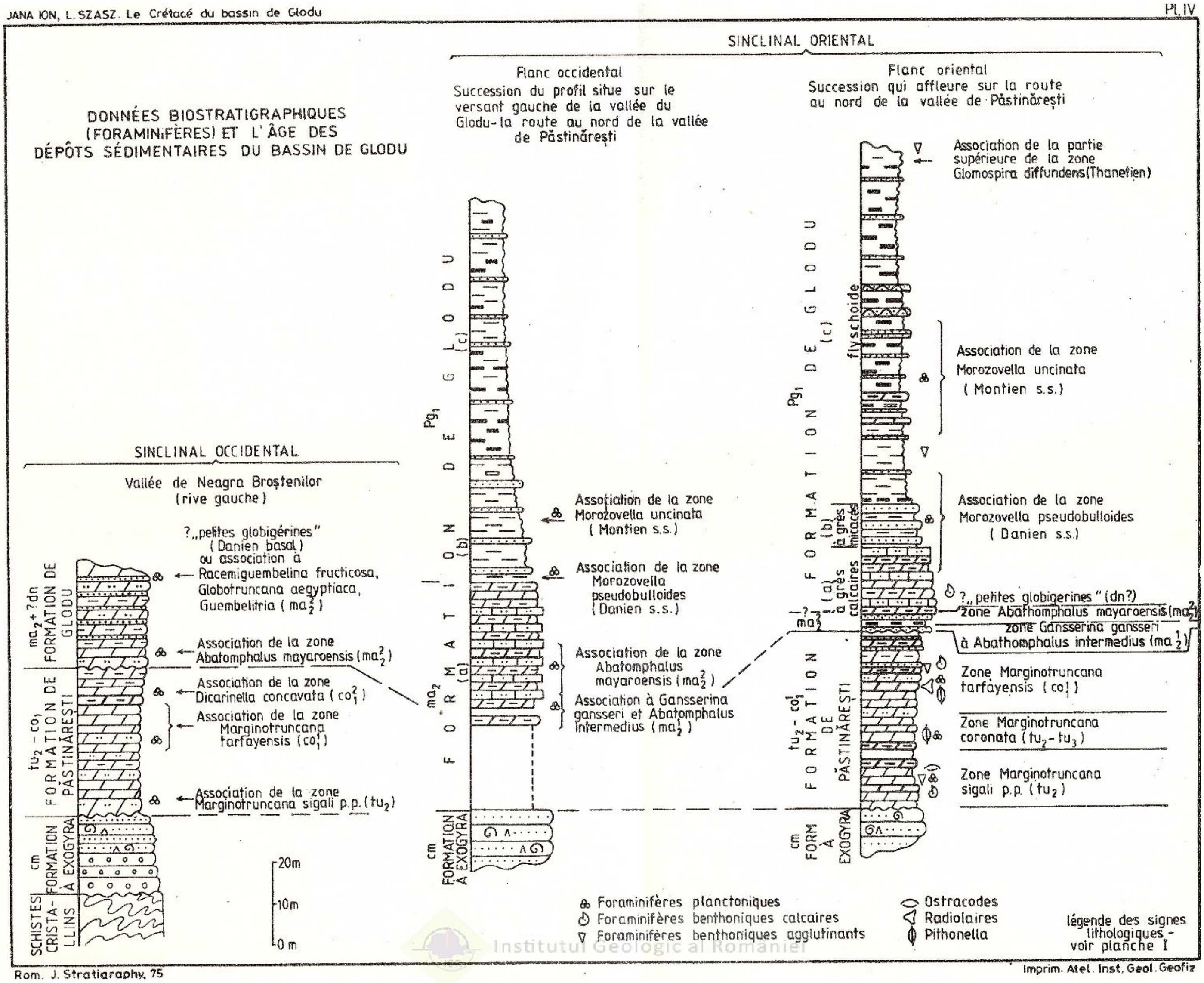
DONNÉES BIOSTRATIGRAPHIQUES (FORAMINIFÈRES) ET L'ÂGE DES DÉPÔTS SÉDIMENTAIRES DU BASSIN DE GLODU
SINCLINAL ORIENTAL
Flanc occidental
Succession du profil situe sur le versant gauche de la vallée du Glodu-la route au nord de la vallée de Păstinăreşti
Flanc oriental
Succession qui affleure sur la route au nord de la vallée de Păstinăreşti
SINCLINAL OCCIDENTAL
Vallée de Neagra Broştenilor (rive gauche)
?„petites globigérines" (Danien basal) ou association à Racemiguembelina fructicosa, Globotruncana aegyptiaca, Guembelitria (ma$_2^2$)
Association de la zone Abatomphalus mayaroensis (ma$_2^2$)
Association de la zone Dicarinella concavata (co$_1^1$)
Association de la zone Marginotruncana tarfayensis (co$_1^1$)
Association de la zone Marginotruncana sigali p.p. (tu$_2$)
ma$_2$+?dn FORMATION DE GLODU
tu$_2$ - co$_1$ FORMATION DE PĂSTINĂREŞTI
cm FORMATION À EXOGYRA
SCHISTES CRISTALLINS
20m
10m
0 m
Pg$_1$ FORMATION DE GLODU (c) (b)
ma$_2$ (a)
Association de la zone Morozovella uncinata (Montien s.s.)
Association de la zone Morozovella pseudobulloides (Danien s.s.)
Association de la zone Abatomphalus mayaroensis (ma$_2^2$)
Association à Gansserina gansseri et Abatomphalus intermedius (ma$_2^1$)
cm FORMATION À EXOGYRA
Association de la partie supérieure de la zone Glomospira diffundens (Thanetien)
Pg$_1$ FORMATION DE GLODU (c) flyschoide (b) à grès micacés (a) à grès calcaires
Association de la zone Morozovella uncinata (Montien s.s.)
Association de la zone Morozovella pseudobulloides (Danien s.s.)
?„petites globigerines" (dn?)
?ma$_2^2$
Zone Abathomphalus mayaroensis (ma$_2^2$)
zone Gansserina gansseri à Abathomphalus intermedius (ma$_2^1$)
tu$_2$-co$_1^1$ FORMATION DE PĂSTINĂREŞTI
Zone Marginotruncana tarfayensis (co$_1^1$)
Zone Marginotruncana coronata (tu$_2$-tu$_3$)
Zone Marginotruncana sigali p.p. (tu$_2$)
cm FORM. À EXOGYRA
Foraminifères planctoniques
Foraminifères benthoniques calcaires
Foraminifères benthoniques agglutinants
Ostracodes
Radiolaires
Pithonella
légende des signes lithologiques - voir planche I

Rom. J. Stratigraphy, 1993, 75, p. 65–69

CONTRIBUTIONS TO THE GEOLOGY OF THE GURA PUTNEI HALF-WINDOW

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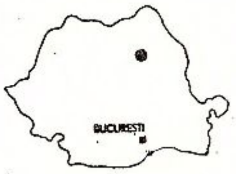

Key words: Paleogene-Lower Miocene. Lithostratigraphy. Regional tectonics (structural data). East Carpathians Flysch Zone. Marginal Folds Nappe (Gura Putnei Half-Window). Obcinele Bucovinei Mountains.

Abstract: The paper brings new lithostratigraphic and structural contributions to the knowledge of the studied area. The Jgheabu Mare Beds, the Salt Formation and the Conglomerate-Gritty Formation are described and correlated with equivalent deposits of the other regions. Several new structural data are also presented and commented upon.

Field studies carried out in 1987 in the Suceava Basin in view of elaborating the Geological Map of Romania, scale 1:50 000, Putna Sheet (Micu, Constantin, 1988) have given us the opportunity to obtain new stratigraphic and structural data concerning the Marginal Folds Nappe that crops out in the Romanian part of the Gura Putnei Half-Window and which we present in what follows.

The region under study has made the object of detail geological researches between 1950–1963, and of a few papers, those by Joja (1950, 1954, 1957), Filimon and Albu (1957, unpubl. report) and by Joja et al. (1963) being worth mentioning.

Stratigraphy

Jgheabu Mare Beds (Ypresian-Lutetian). Under this name we have separated the oldest deposits cropping out in the Gura Putnei Half-Window, south of the frontier with the Ukraine. Lithologically, they are represented by an alternance of well cemented, subdecimetrical finely grained green quartzose sandstones and slightly silicified green pelites, that are associated with coarse quartzose sandstone, microconglomerates with green schists elements and grey-greenish siltstones in decimetric beds. These deposits, with visible thickness of 15–35 m, lie under the Doamna Limestones and crop out in the axis of the Vărăria and Poiana Mare anticlines, south of the Suceava Valley. The lithological affinities with the Jgheabu Mare Beds as well as their identical stratigraphic position have made us call them the same as the synchronous deposits in the Bistriţa Half-Window. Filimon and Albu (1957, unpubl. report) have, in fact, a similar point of view, but, on the map, they assign these deposits to the Pasieczna (Doamna) Limestones.

Doamna Limestones (Lutetian) have been evidenced in the axial zone of the above mentioned anticlines, where they are represented by decimetric beds of micritic white-greyish or slightly greenish micritic limestones that admit thin green pelite interbeds as well as subdecimetric or decimetric beds of medium-grained oligomictic quartzose sandstones, white in colour. They differ from the Pasieczna Limestones, their stratigraphic equivalent of the Tarcău Nappe, first of all by the existence of accumulations irregular in shape, more or less tabular, of quartzose white sandstones in the limestone mass, probably due to the action of lithophagous worms as well as to the slight development of siliceous accidents in the middle part of the limestone beds. The thickness of the Doamna Limestones in the Gura Putnei Half-Window is of 25–40 m.

Red and green clays (Lutetian) overlie, with no sedimentation break, the Doamna Limestones, being represented by a flysch sequence of the "hieroglyph beds" type, in which green quartzose sandstones, with numerous bioglyphs, alternate with red and green clays. The thickness of these deposits in the region ranges between 5 and 15 m.

Unlike Joja (1960) and Joja et al. (1963), who use for these deposits the name of Strujinoasa Beds, we have preferred to use the term of red and green clays, previously introduced by Olteanu (1953) for synchronous deposits in the Bistriţa Half-Window. The Strujinoasa Beds (Joja, 1960), initially used as a biostratigraphic entity in the Tarcău Nappe, subsequently also separated in the Gura Putnei Half-Window, represent, in our opinion, only a peculiarity of the Plopu Beds, i.e. their sequence that includes red clay interbeds. In this connexion, we consider that the use of the name of the Upper Plopu Beds (Joja, 1960) for the succession overlying the deposits with red clay interbeds, of Viţeu Beds respectively (Joja et al., 1968), for those between the Doamna/Pasieczna Limestones and the red clay interbeds do not correspond with the initial definition of the Plopu Beds (Atanasiu, 1943) and should therefore be abandoned.

Concerning the red and green clays, they should not be referred to the base of the Bisericani Beds (Filimon, Albu, 1957, unpubl. report) as, in the description of the stratotype of the Bisericani Beds, Athanasiu et al. (1927) do not mention the presence of red clays in the lower part of these deposits.

Bisericani Beds (Priabonian) have not a very large distribution in the Gura Putnei Half-Window. They are lying with no sedimentary break over the red and green clays and support the Globigerina-bearing Marls and the Fierăstrău Sandstone. These deposits are made up of clays and grey-greenish siltstones, accidentally accompanied by thin interbeds of hieroglyph sandstones. The thickness of the Bisericani Beds in the Gura Putnei Half-Window ranges between 200 m and 300 m.

Globigerina-bearing Marls and Lucăceşti Sandstone (Priabonian-Rupelian) lie over the Bisericani Beds, being represented by a sequence, 10–15 m thick, made up of grey-whitish marls sometimes interbedded with centimetric turbiditic sandstones as well as whitish-greenish quartzose, sometimes slightly micaceous sandstones, in decimetric beds (Lucăceşti Sandstone).

These deposits had been previously referred to the Priabonian, based on Globigerina assemblage (Joja et al., 1963). More recent researches in other regions (Micu et al., 1980, unpubl. report; Micu, Gheţa, 1986) evidenced the presence at the upper part of the Globigerina-bearing Marls of calcareous nannoplankton assemblages as well as of Globigerinae that could make it possible for these deposits to be assigned to the Priabonian-Rupelian.

Slaty Bituminous Shales and Fierăstrău Sandstone (Oligocene) have been separated by us for the first time in the Gura Putnei Half-Window, where they range between 5 and 30 m thick. These deposits are made up of bituminous shales in decimetric or centimetric beds with decimetric interbeds, rarelier up to 1.5–2 m thick, of quartzose sandstones of Kliwa type. The sandstone beds, usually quite brittle and medium-coarse, are sometimes well cemented, finely-grained, made up of quartz grains, very rarely accompanied by glauconite grains. The small thickness of the deposits concerned has made us render them on the map together with the Lower Menilites and the Brown Marls overlying them.

Lower Menilites and Brown Bituminous Marls (Oligocene) represent a very characteristic lithological entity in the sequence of Oligocene deposits. At its lower part, on a thickness of 3–8 m, menilites appear in centimetric beds, followed, on 8–20 m thickness, by Brown Bituminous Marls, whose weathered surfaces are grey-whitish in colour.

Lower Dysodilic Shales (Oligocene) range between 30 and 50 m in thickness, and are made up mainly of bituminous shales with rare interbeds of centimetric quartzose sandstones of Kliwa type.

Kliwa Sandstone (Oligocene) crops out on a rather limited area, constituting the filling of a syncline situated between the Gura Putnei locality and the Suceava River, that is further continued in the left bank of this river, in the Deluţul Summit. West of this syncline the occurrences of these deposits are very scarce. They only crop out on limited areas, west of the Ziminel Creek and along a series of tributaries of the Suceava River, south of Straja.

A characteristic feature of the Kliwa Sandstone of the Gura Putnei Half-Window is the absence of green schists elements along all the sections under study. We have only found one block of Kliwa sandstone, about 10 cm in diameter, containing green schists fragments, nearby the railway enbankment, north of the Turanu Hill, but it probably originates somewhere else, as we have not found such rocks in the Turanu Hill-Hutei Hill Syncline.

Upper Dysodilic Shales and Menilites (Oligocene-Miocene) only crop out on a series of torrents that flow down from the Măgura Mică Summit to Bivolăria. There, overlying the Kliwa Sandstone, on a thickness of a few meters, schistose dysodilic shales as well as diatomites occur. Towards the upper part of the sequence, menilites also appear, in centimetric beds.

Gura Şoimului Beds (Lower Miocene) have the same outcropping area as the underlying deposits, which they overlie without any sedimentary break. They consist of silty grey-greenish marls, slightly marly green clays, fine greenish platy sandstones and very thin interbeds of microconglomerates with green schists elements. Sometimes in the mass of the clays, that can be up to 3–4 m thick, well rolled pebbles of green schists, 1–3 cm in size, are disseminated.

The passage from the underlying deposits is gradual. In the base of the Gura Şoimului Beds there appear thin beds of slightly bituminous black clays.

Salt Formation (Lower Miocene). The deposits belonging to this formation crop out in the south-western corner of the Gura Putnei Half-Window, in the Glodu Creek Basin, where they consist of silty marls and grey-greenish, sometimes slightly reddish clays, with rare interbeds of thin grey-yellowish sandstones and subdecimetric microconglomerates with green schists elements. A single outcrop mentioned by previous researches in the region (Joja, 1957; Filimon, Albu, 1957, unpubl. report) is covered now by the Glodu Stream alluvia, but in that very place there is a source with a high NaCl concentration. Such a salty source also appears in the outcropping area of the Salt Formation, quite near borehole 12 Putna, in Poiana Glodu.

The above mentioned deposits were assigned by Filimon and Albu (1957) to the Helvetian, based on a microfaunal assemblage, where besides reworked species from the Cretaceous and the Paleogene, several species assigned to the Middle Miocene are mentioned, the latter without being figured. These deposits were considered by the respective authors to appear in a small window, south of the Gura Putnei Half-Window.

Joja (1957) also mentions the presence of these deposits on the Glodu Creek and assigns them to the Miocene, only according to lithologic criteria, without specifying the relations between them and the Bisericani Beds, which the author figures on the map on the Glodu Stream and which he considers to belong to the Gura Putnei Half-Window.

To what we have already mentioned we should add that borehole 12 Putna on the Glodu Creek has intercepted three levels of rock salt between 128 and 560 m, 1066 and 1142 m and 1244 and 1429 m respectively, which proves without doubt the presence of the Salt Formation in the Glodu Stream Basin. The existence of the Salt Formation to the south has also been proved by boreholes 2, 11 and 15 Putna.

The drilling data in the region point out the existence within the Salt Formation of some conglomeratic-gritty episodes with green schists elements, a similar situation being known from other areas of the Marginal Folds Nappe (Vrancea Half-Window, Bistriţa Half-Window).

Conglomeratic Gritty Formation (Lower Miocene) mainly develops in the western part of the Gura Putnei Half-Window, being caught under the overthrust of the Cretaceous and Paleogene deposits of the Tarcău Nappe. This formation is quite varied lithologically. Conglomerates with green schists elements prevail, with a poorly cemented, gritty or silty matrix, grey in colour. These conglomerates generally are of "matrix-supported" type, but there are equally "clast-supported" conglomerates or various intermediate varieties. The sizes of the pebbles, usually well rounded, are centimetric. Rarely, angular-shaped pebbles, 0.1–0.15 m in size, are also present.

Coarse, slightly micaceous sandstones also occur, associated with conglomerates. Decimetric graded-bedded sandstones, sometimes obliquely laminated at their upper part have centimetric interbeds of grey-greenish clays or green siltstones.

As concerns the relations of the Conglomeratic-Gritty Formation with the underlying deposits, our researches have proved that it has a clearly unconformable position. That is very clear in the zone of the Bostan Stream springs, where the sequence concerned unconformably overlies various terms of Oligocene deposits and even the Bisericani Beds, Priabonian in age. On numerous sections, especially in the right bank of the Suceava River, in the very base of the Conglomeratic-Gritty Formation there are blocks of various rocks, reworked from the underlying deposits (Bituminous Brown Marls, Menilites, Fierăstrău Sandstones), which proves the sedimentation of this formation on a surface previously subject to erosion. The data from the boreholes drilled in the region suggest that erosion would post-date the deposition of the Salt Formation, the relations between the latter and the Conglomeratic-Gritty Formation being also one of unconformity.

We have equally assigned to the Conglomeratic-Gritty Formation the deposits cropping out in the railwaycut, right of the Putna Creek, nearby the confluence with the Gura Putnei Creek. Over here, without evidencing the direct relations with the underlying rocks, there appear green-grey silty marls, with interbeds of green clays and a bed of microconglomerates with green schists, 0.1–0.15 m thick. The above described deposits are lithologically identical with those cropping out in the upper course of the Bostan Stream. This outcrop has been referred by Filimon and Albu (1957) to the Oligocene, although they mention the presence of Cibicides species, probably Miocene in age. The deposits concerned were previously considered by Joja (1954, 1955) to represent a Miocene that appears in a small window under the Oligocene of "the lower unit of the outer flysch" (= Marginal Folds Nappe), the same author subsequently referring them to the Tortonian (Joja et al., 1968).

It is quite difficult to follow the relations between the Salt Formation and the Conglomeratic-Gritty Formation. The only place where someone could examine these relations would be the saddle representing the dividing line between the Glodu and Gura Putnei Creek basins. From there northwards, there are only conglomerates and sandstones that crop out, unconformably lying over deposits older than the Salt Formation. Southwards, in the Glodu Creek Basin, there is a depression zone, a thin overlier of Quaternary deposits, under which boreholes have proved the presence of the Salt Formation. To all appearances, the Conglomeratic-Gritty Formation is overlying the Salt Formation. Considering data we have already presented we are of the opinion that the relations between the two formations are of unconformity.

The Conglomeratic-Gritty Formation is considered by Joja (1955, 1957) to represent a conglomeratic facies locally substituting different terms of Oligocene deposits, starting with the Lower Dysodilic Shales.

Filimon and Albu (1957, unpubl. report) also consider the deposits under discussion to be of Oligocene age in a marginal, conglomeratic facies, conformably overlying the Lower Dysodilic Shales or the Kliwa Sandstone. Although they point out the presence, in the base of the conglomerates, of reworked blocks of bituminous marls and menilites, neither of author raises the matter of possible unconformable relations between the conglomerates concerned and the deposits underlying them.

As shown above, our researches have clearly demonstrated the unconformable position of the Conglomeratic-Gritty Formation both in the Gura Putnei Creek Basin and on certain tributaries of the Suceava River (Slatina, Nistor, Scorbura, Andronic Rivulets). In exchange, the cores and the electric logs of borehole 12 Putna of the Glodu Creek have shown that equally in the Salt Formation there are numerous interbeds of sandstones and conglomerates with green schists elements. On the basis of regional correlation we are of the opinion that the Conglomeratic-Gritty Formation of the Gura Putnei Half-Window is Burdigalian in age, just like the Salt Formation underlying it.

The conglomeratic deposits of this type certainly have a larger outcropping area north of the frontier, because, on all the left tributaries of the Suceava River (Ziminel, Balmus, Pîrîul Rău) as well as on the summits between them, we have frequently found numerous elements of green schists from the Conglomeratic-Gritty Formation that unconformably overlies older deposits.

As regards regional correlations, the Conglomeratic-Gritty Formation of the Gura Putnei Half-Window represents the equivalent of the Almaşu Conglomerates of the Bistriţa Half-Window. North of the frontier, the equivalent of this formation could be the Sloboda Conglomerates, the latter lying also unconformably over older formations (Glusko and Kruglov, 1971; Burov et al., 1974) etc.

Tectonics

The formations of the Marginal Folds Nappe of the Gura Putnei Half-Window crop out from under the overthrust of the Tarcău Nappe deposits, in the Putna upper course and in the Suceava Valley. The erosion contour of the overthrust plane of the Tarcău Nappe approximately trends north-south between the Straja locality and the Glodu Stream. Southwards its direction becomes approximately south-west-north-east, being rather sinuous, which suggests a quite small overthrust angle and relatively small thicknesses of the Tarcău Nappe in the Putna Basin.

The dominant structural elements of the Marginal Folds Nappe are the two anticlines, Vărăria to the west and Poiana Mare to the east, in whose axis we have mentioned the presence of the Jgheabu Mare Beds. The western limb of these anticlines is laminated, the eastern ones are affected by normal faults. Both structures suddenly sink north of the Suceava River, where they contain only Priabonian deposits (Bisericani Beds) in their axis.

West of the Vărăria Syncline a series of secondary anticlines and synclines are developed, whose Upper Eocene and Oligocene deposits are unconformably overlain by the Conglomeratic-Gritty Formation.

East of the above mentioned anticlines representing the central zone of the Gura Putnei Half-Window a series of large synclines is developed, separated by narrow anticlines with the eastern limb affected by reverse faults. In the axis of these anticlines, Bisericani Beds crop out. We should also mention that the data figured on the Geological Map of Romania, 1:500 000, sheet 2a (1957), referring to the presence of Oligocene deposits in the synclines north of the Suceava River, reflect the reality in the field better than subsequent maps, on which only the Bisericani Beds are figured.

Approximately along the Suceava River, we suppose the existence of a transverse fault trending WSW-ENE, whose northern compartment is sunk. That is attested by the sudden disappearance, north of the Suceava River, of the Lower and Middle Eocene formations in the axis of the Vărăraia and Poiana Mare anticlines.

Following the erosion contour line of the Tarcău Nappe overthrust we could notice that north of the above mentioned fault, it is almost rectilinear, which proves a rather big thrust angle. In exchange, south of the fault, this angle of the overthrust surface is, comparatively, smaller.

The overthrust of the Marginal Folds Nappe over the Subcarpathian Nappe is covered in the region of study by the Suceava and Putna Rivers alluvia, the boundary traced between the two units being in this area more or less conventional. In any case, the contact is made east of the Laura Stream, left tributary of the Suceava River, wherefrom it continues SSE in the north-eastern slope of the Măgura Mică Hill. From here southwards, the Marginal Folds Nappe is totally overpassed by the overthrust of the Tarcău Nappe that directly overlies Miocene formations of the Subcarpathian Nappe.

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BIOSTRATIGRAPHIC STUDY OF THE EOCENE-OLIGOCENE BOUNDARY IN THE TYPE SECTION OF THE BREBI MARLS (TRANSYLVANIA, ROMANIA)

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Project 174 – Terminal Eocene Events

Key words: Eocene. Oligocene. Stratigraphic boundary. Molluscs. Ostracods. Foraminifers. Nannofossils. Palynomorphs. Biostratigraphy. Apuseni Mts – Paleogene Zone – Transylvania.

Abstract: The paper presents the results of the researches on molluscs, ostracods, benthic foraminifers, calcareous nannoplankton and dinoflagellates from the type section of the Brebi Marls, considered as one of the most useful in the world for the study of the Eocene-Oligocene boundary. The malacologic assemblage is of Priabonian type in the whole section, the typically Oligocene species occurring in the suprajacent member. The changes in ostracod and benthic foraminifer faunas are gradual, without marking an obvious threshold in tracing the boundary. Two local zones of ostracods have been separated, the Sphaenocytheridea parvogracilis Zone and the Hazelina indigena Zone. The distribution of calcareous nannoplankton points to the presence of the NP 21 and NP 22 Zones, the Eocene-Oligocene boundary being placed in the first zone within the interval of the Ericsonia obruta acme. It corresponds to the boundary between the two ostracod zones. Dinoflagellates are referred to the Wetzeliella (Rhombodinium) perforatum Zone, the index form getting up in Transylvania to the top of the NP 21 Zone.
The Brebi Marls were deposited at a depth corresponding to the outer shelf in normal marine waters. Transylvania was part of the Mesogean Realm, but nannoplankton also indicates the influence of cool waters around the Eocene-Oligocene boundary.

Introduction

The investigations carried out all over the world, within the framework of Project no. 174 of IGCP "Geological Events at the Eocene-Oligocene Boundary", for selecting sedimentary sequences that would cross the Eocene-Oligocene Boundary, have emphasized the importance of the Brebi Marls of Transylvania (Bombiţă, Rusu, 1981).

Better known as Bryozoan Marls/Beds, due to their richness in bryozoa in the Cluj-Napoca–Huedin Area – Hauer and Stache (1863, "Die Laevigata und Bryozoen Mergel"), Pavay (1873, "Die Bryozoen – Tegel und Mergel") and Koch (1894, "Bryozoenschichten") – these deposits have been included in modern lithostratigraphic nomenclature as Brebi Marls, a term introduced by Hofmann (1879, "Breder Mergel") for the Jibou Region. We mention that in the present acception the Brebi Marls include in the base equally what Hofmann separated as "Intermedia Marls", a sequence of passage from the Cluj Limestone, which represents in fact a biohorizon (Nummulites fabianii Level) and not a lithostratigraphic unit.

Hofmann (1879) and Koch (1894) were the first to identify a rich and varied fossil content, consisting in molluscs, foraminifers, ostracods, bryozoans, echinoids etc. These authors, who considered the formation under study to be the equivalent of the Bryozoan Marls in the Budapest region, place it at the upper part of the Bartonian stage (last stage of the Eocene). In this way, the Eocene-Oligocene boundary has been placed in Transylvania, as early as the first researches, between the Brebi Marls and the Hoia Beds.

Accepted in this position for a long time, the boundary under discussion comes to be placed either within the Mera Beds, above the last nummulites-bearing level (Bombiţă, 1957), or at the base of the Brebi Marls, as it can result from the paper by Martini and Moisescu (1974), in which the nannoplankton Ericsonia subdisticha Zone (NP 21) is assigned, as a whole, to the Oligocene. It could be already inferred from the paper by Popescu and Gheţa (1972) that the NP 21 Zone starts at the base of the Brebi Marls, that is with the acme of the typical *Nummulites fabianii*, a form characteristic of the Mesogean Priabonian.

The results obtained by Martini and Moisescu (1974) were reinterpreted by Rusu (1977), who proposed that the Eocene-Oligocene boundary be placed between the NP 21 and the NP 22 standard zones, a level situated between the Pycnodonte gigantica marker-level and the top of the Brebi Marls. In agreement with the opinion expressed by several authors, we consider the Lattorfian and the Lower Tongrian (with elements of the NP 21 Zone) as last terms of the northern Upper Eocene, equivalent to the last part of the Mesogean Priabonian.

Based on the study of planktonic foraminifers from some geological sections in the Meseş Area (Brebi section included), Iva and Rusu (1982) established the Eocene-Oligocene boundary at the level of the extinction of the *"Globorotalia" cerroazulensis* group and the appearance of the species *Globigerina ampliapertura* BOLLI, a level situated in the upper third of the Brebi Marls, at the base of the Pycnodonte gigantica biohorizon. Traced in this way, the boundary was subsequently argued on the basis of larger forams by Bombiţă (1984) and nannoplankton by Gheţa (1984). Micu and Gheţa (1986), and presented as such in the papers for Project no. 174 (Rusu, 1985; Bombiţă, 1986).

Taking into account that the Brebi section, selected among the best sections in the world for reconstructing the sequence of events at the Eocene-Oligocene boundary (see the preface to the volume "Terminal Eocene Events" edited by Pomerol and Premoli-Silva, 1986), has been studied biostratigraphically only from the point of view of planktonic foraminifers, we have intended to analyze its contents equally as concerns other groups of organisms.

For this paper the field researches and the study of molluscs have been carried out by Anatol Rusu, the ostracods have been studied by Franz Wanek, the benthic foraminifers by Despina Brotea, the nannoplankton by András Nagymarosy, and the dinoflagellates and palynomorphs by Ana Ionescu.

Geographic and geologic setting

The Brebi locality (Sălaj district) lies in the northwestern extremity of the Transylvanian Plateau, on the regional road 191 C, midway between Zalău and Jibou (about 26 km).

As regards the geology, the region belongs to the so called Paleogene Zone of the Apuseni Mountains, representing the post-tectogenetic cover of the Inner Dacids. Within this area, characterized by an alternation of continental formations with marine neritic formations, three sedimentation areas have been identified: Preluca Area, in the north, Meseş Area, in the centre, and Gilău Area, in the south.

In the Brebi village region, lying in the northern part of the Meseş Area, the following lithostratigraphic units crop out (see the geological sketch, Fig. 1):

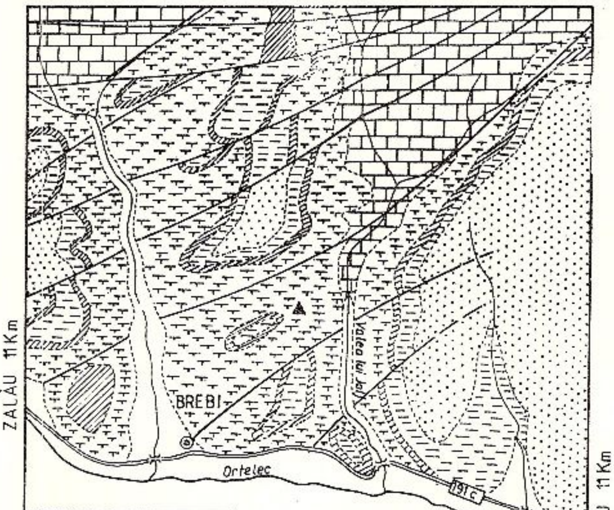


Fig. 1 Geological sketch of the Brebi region: 1, Moigrad Beds; 2, Curtuiuş Beds; 3, Hoia Beds; 4, Brebi Marls; 5, Cluj Limestone; 6, Setting of the Brebi Marls type section.

– Cluj Limestone, represented by bioclastic calcarenites and nodular calcilutites (about 30 m thick), rich in fossils, including nannoplankton elements of the NP 20 Zone (Gheţa, 1984);

– Brebi Marls, a marly formation that seems to reach its known maximum thickness here (about 75 m), developing gradually from the subjacent member through a clayey limestone pile, a few meters thick;

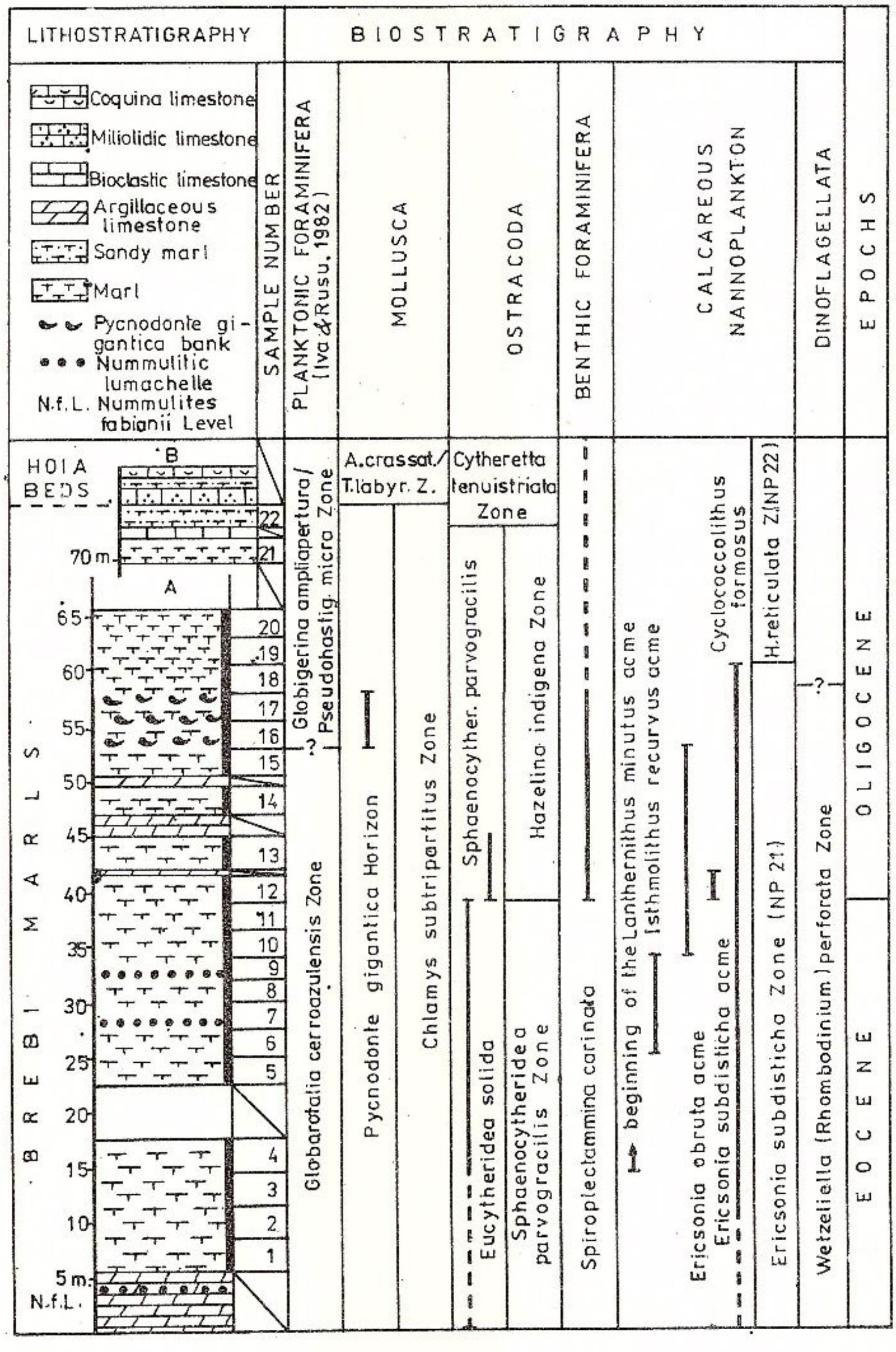


Fig. 2 – Type section of the Brebi Marls on the Jolj Valley-Brebi village (a) and complementary section at Strîmtura-Ortelec village (b) with the main biostratigraphic events.

– Hoia Beds, a 2 m thick member, consisting of two banks of skeletal limestones, separated by a marly interbed, all of them very rich in fossils, containing, in the Cluj region, the nannoplankton of the NP 22 Zone (Martini, Moisescu, 1974);

– Curtuiuş Beds, consisting of grey clays and marls (12–15 m thick) with brackish water fauna, comprised in the Preluca Area between deposits with an assemblage of the NP 22 Zone (Mészáros et al., 1977);

– Moigrad Beds, made up of silty red clays, sands and gravels, all of them non-fossiliferous deposits of continental origin (more than 100 m thick).

The outcrop, proposed by Iva and Rusu (1982) as type section of the Brebi Marls and sampled for this paper lies in the right slope of the Jolj Valley at Brebi, at about 700 m upstream the Zalău-Jibou road (Fig.

1, Pl. I, Fig. 1). The lithological column of the Brebi Marls here (Fig. 2) starts with a pile of 5.5 m of miliolidic clayey limestones containing bivalves, gastropods, echinoids, solitary corals etc. and rare specimens of Nummulites fabianii (PREV.), marking the N. fabianii Level (with lumachelic development in nearby sections). About 70 m of whitish grey fine marls follow up (the last 10–12 m of which are not exposed) by three interbeds of more calcareous marls (0.5–2 m thick) in the upper half of the outcrop. As paleontologic markers two thin levels of lumachels containing small nummulites appear in the lower half of the section, and a level with many shells of a big-sized gryphaeid *Pycnodonte gigantica* (SOL.) (Pl. I, Fig. 2), developed on 6 m thickness, in the upper third.

As the uppermost part of the Brebi Marls does not crop out in the type section a complementary section was selected, situated about 6 km westward, in the left slope of the Ortelec Valley, at "Strîmtură" (downstream the Ortelec locality). Over here, in the excavation of the Brebi-Zalău road, one can notice (Fig. 2): 3 m of grey-whitish marls, 60 cm of slightly gritty limestone with small nummulites and rare bivalves, 2 m of grey-yellowish, fossiliferous sandy marls, with biodetrital levels, 50 cm nodular clayey limestones with bivalve moulds, overlain by the Hoia Beds (miliolidic limestones bearing *Ampullinopsis crassatina* (LMK.), *Tympanotonos labyrinthum* (NYST), *Polymesoda convexa* (BRGT.), grey sandy marly clays and shelly limestones, amounting to 3 m thickness.

The Brebi Marls on the Jolj Valley were investigated for clay mineralogy, the clay fraction being 70–85% smectitic (Chamley, 1986, Fig. 2). The assemblage of clayey minerals in the lithological column shows a slight increase of illite at the expense of smectite, which is due to the increase in time of the amount of rock-derived (equally illustrated by the increase in quartz and feldspar) instead of soil-derived minerals. These transformations parallel to those registered in the World Ocean, that took place in a stable tectonic setting, suggested, according to the cited author, climate conditions that have become colder and slightly wetter.

The terrigenous-carbonatic sediments of the Brebi Formation have been deposited on the shelf in northwest Transylvania, in a large trough-like (approximately trending E-W), with a slight slope in the south (where the terrigenous material originated from) and with a steep slope towards the high Ticău-Preluca carbonate platform (Popescu, 1976), with an exclusively carbonatic sedimentation, in the north.

The Brebi Marls correspond to the maximum extension of the Priabonian Sea, followed by the rapid regression during the Early Oligocene, whose start is also well marked in the "Strîmtură" section.

Biostratigraphy

Five groups of marine organisms, three of which are benthic (molluscs, ostracods and benthic foraminifers) and two planktonic (calcareous nannoplankton and dinoflagellates) as well as remains of continental vegetation (represented by spores and pollen) have been examined for a more complete description of the Eocene-Oligocene passage interval in the type section of the Brebi Marls.

Mollusca

Molluscs make up the best represented megafaunal group of the Transylvanian epicontinental Paleogene on which the first regional stratigraphic studies are based.

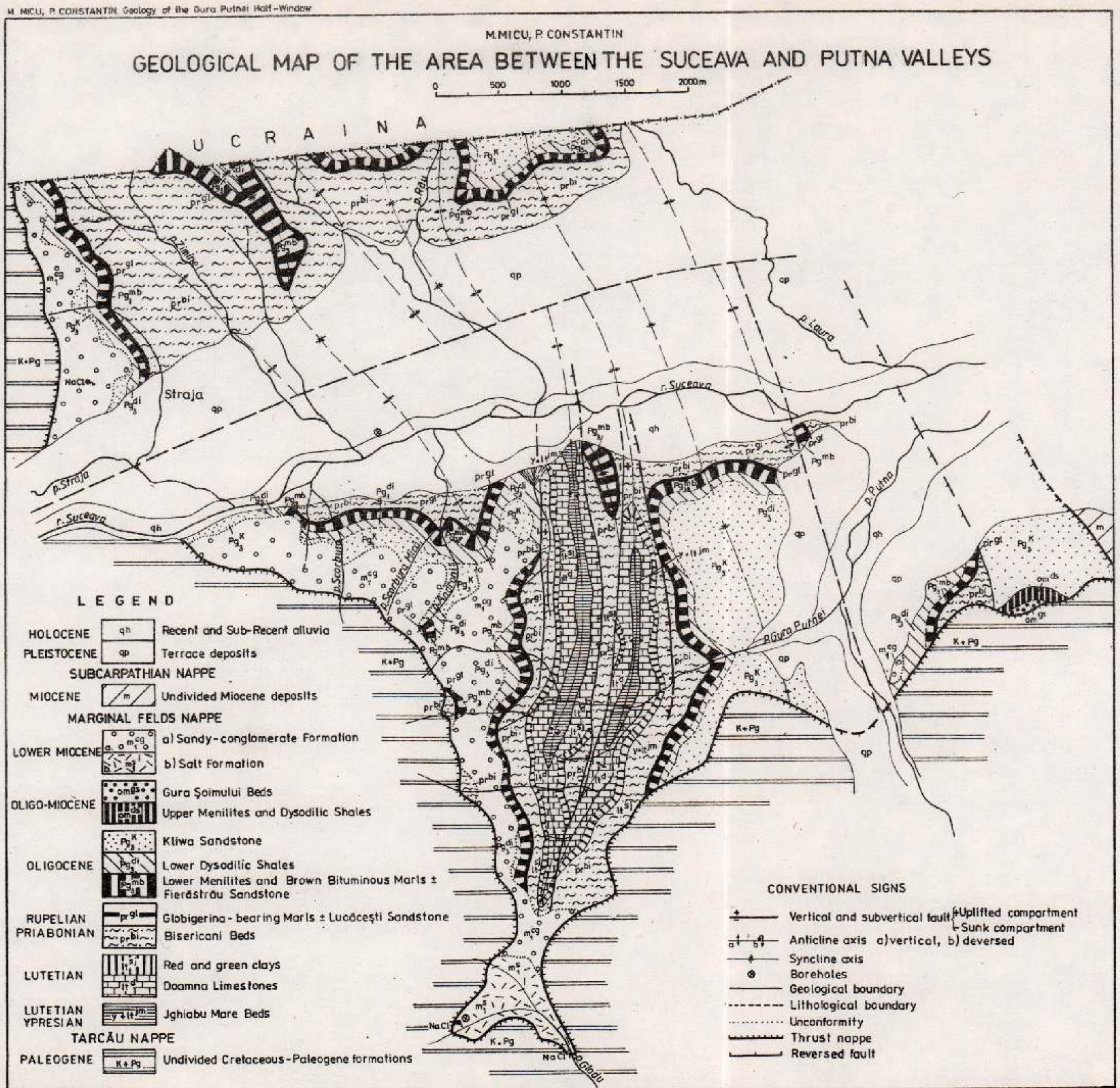
From the Brebi Marls, Hofmann (1879) and Koch (1894) identified a varied bivalve and gastropod fauna, later studied and enriched by Mészáros (1957) and Mészáros et al. (1963).

The molluscs identified by us for this paper have been collected from the marly interval of the Brebi Member, in the region of the Brebi locality, without including the basal clayey limestone sequence (Nummulites fabianii Level), that is richly fossiliferous but of no interest in the discussion of the Eocene-Oligocene boundary.

The mollusc fauna has been preserved as inner or outer moulds, with the exception of pectinids, ostreids, spondylids and chamids, that conserve their shells.

From the forms of bivalves identified, *Pycnodonte gigantica* (Pl. I, Fig. 2) occurs with greater frequence, forming a marker level (Fig. 2) of regional importance, followed by *Chlamys* ex. gr. *biarritzensis* and *Lentipecten corneus*, occurring with variable frequency in the whole section. The other bivalves and especially gastropods have a low frequency. Besides the 25 taxa identified at the level of species or subspecies (see Table 1), other specimens found have been assigned to the genera *Anadara, Pteria, Arctica, Nemocardium, Angulus, Cuspidaria* etc.

From the stratigraphic distribution of the molluscs in Table 1, it results that most of them range along several stages and only three taxa (*Chama granulosa, Panope allonsensis* and *Diastoma costellatum alpinum*) are strictly located at the level of the Priabonian. The concurrence interval of all the species identified is indeed the Upper Eocene. There are equally forms such as *Pecten arcuatus* and *Cardita laurae* whose evolution starts in the Priabonian, but are characteristic of the Oligocene. The Brebi Marls malacofauna, referred to the *Chlamys subtripartitus* local zone (Mészáros et al., 1989), is of Priabonian type and gives no elements for tracing the Eocene-Oligocene boundary. The greater evolutionary stability makes changes in the mollusc fauna and generally in benthic



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organisms to appear later on, often closely dependent on local paleoecological conditions. In Transylvania this change takes place at the level of the Hoia and the Mera Beds, at the same time with the appearance in the basin of typically Oligocene species, such as *Ampullinopsis crassatina* (LMK.), *Turritella biarritzensis* (BOUSS.), *T. clumancensis* (BOUSS.), *Pitar verneuili* (D'ARCH.), *Polymesoda convexa* (BRGT.) etc., belonging to the Ampullinopsis cressatina/Tympanotonos labyrinthum Zone (Mészáros et al., 1989).

bathimetric range, being developed in the shelf area, from the shore up to the continental slope. As the sedimentological features and the other groups of organisms plead for depths comparable with the outer shelf, it results that certain typically littoral forms like *Cubitostrea* or *Chama* have been transported here. The high frequence forms we have referred to and especially Pycnodonte gigantica proliferate in off-shore zones at middle depths of 100 m, in waters of lower turbidity.

Table 1
Stratigraphic distribution of the Mollusc species yielded by the Brebi Marls in the Brebi Region

SPECIES \ DISTRIBUTION	MIDDLE EOCENE	LATE EOCENE	RUPELIAN	CHATTIAN	MESOGEAN REALM	NORTHERN REALM
Chlamys biarritzensis subtripartitus (d'Arch.)	x	x			•	
Cubitostre cubitus (Desh.)	x	x			•	•
Corbula pixidicula (Desh.)	x	x				•
Velates perversus Gmelin	x	x			•	•
Mesalia fasciata Lmk.	x	x			•	•
Barbatia appendiculata (Sow.)	x	x			•	•
Spondylus buchi Phil	x	x	x		•	•
Pycnodonte gigantica (Sol.)	x	x	x		•	•
Plicatula martinsi (d'Arch.)	x	x	x		•	
Tibia goniophora (Bell.)	x	x	x		•	
Rimella fissurella (Linné)	x	x	x		•	•
Lentipecten corneus (Sow.)	x	x	x	x	•	•
Spondylus podopsideus Lmk.	x	x	x	x	•	
Pelecyora incrassata (Sow.)	x	x	x	x	•	•
Chama granulosa d'Arch.		x			•	
Panope allonsensis Boussac		x			•	
Diastoma costellatum alpinum (Tourn.)		x			•	
Chlamys biarritzensis biarritzensis (d'Arch.)		x	x		•	
Chlamys biarritzensis thorenti (d'Arch.)		x	x		•	
Chlamys bellicostata (Wood)		x	x		•	•
Megaxinus rectangulatus (Hofm.)		x	x		•	
Cardita laurae (Brgn.)		x	x		•	
Pecten arcuatus (Brocc.)		x	x	x	•	
Glossus subtransversus (d'Orb.)		x	x	x	•	•
Pholadomya puschi Goldf.		x	x	x	•	•

The typically marine Brebi Marls assemblage, clearly is of the Mesogean Realm (see Table 1), which is in agreement with the arguments offered by the presence in the basin of nummulites, echinoids or colonial corals. Most of the mollusc taxa identified have a large

Ostracods

It was Pávay (1873) who first spoke of the ostracod fauna from the Bryozoan Marls. This research is later resumed by Barbu (1956, 1963). The studies carried out by Olteanu (see Olteanu, Popescu, 1973; Bombiţă et al., 1975; Olteanu, 1977 a, 1977 b, 1980) contribute with much more important data than the ones mentioned before. He separated two biozones within the Brebi Marls: the Phlyctocythere eocaenica Zone and the Paijemborchella tricostata Zone (Olteanu, 1977 a; Popescu et al., 1978). Recently Wanek (1989) resumed the research on ostracods for establishing the Eocene-Oligocene boundary.

Analyzing a large amount of material from the Brebi section we have obtained the image of the variation in percentages of species, for every sample – rendered in Table 2. The identified assemblages are dominated by facies forms with a large stratigraphic range, without biostratigraphic value. The species of biostratigraphic value have moderate or low frequencies.

Eocene ostracod assemblages were replaced by Oligocene ones in a slow continuous process. Thus, in the first four samples from the base of the Brebi Marls sequence, from the 45 taxa present, species characteristic of the Eocene participate with 40 % (33 % being strictly Upper Eocene species), Eocene-Oligocene transition elements represent 38 %, and the strictly Oligocene elements are represented only by one species (2 %), in a single sample and with negligible frequency. The other elements included (20 %) cannot be stratigraphically evaluated at the present level of knowledge.

Following gradual changes at the uppermost level from which samples have been collected – that are in the very proximity of the Hoia Limestone – Eocene species are present in percentages of only 7 %, the transition one of 47 %, the strictly Oligocene ones of 26 % and those with still uncertain stratigraphic value of 26 %.

Our studies on the ostracod fauna in the Brebi Marls (see also Wanek, 1989) do not confirm the zonation introduced by Olteanu (1977 a). So, the nominal species for the Phlyctocythere eocaenica Zone (species no. 15 in Table 2) has been found only in one sample (sample 11) represented by two valves. If we take into account that from the stratigraphic interval assigned to this zone we have separated almost 16,000 ostracod valves (and from sample 11 about 800), this occurrence loses its significance for biozonation. For the upper biozone, the ones above the Pycnodonte gigantica Level, none of the about 8,800 ostracod valves belongs to the genus *Paijemborchella* (or *Eopaijemborchella*).

It does not mean that an ostracologic biozonation is out of question, but it is only possible based on studies concerning a multitude and a variety of sections with adequate details. For the time being, as there are not enough data for comparison, we can only point to the local character of the biozonations proposed in the present paper.

In the lower half of the succession (samples 1–11) we separate the Sphaenocytheridea parvogracilis Zone (the frequence of the species is low but relatively constant), considering that the index species seems to range within a rather limited stratigraphic interval (Olteanu, 1977 b) and to be widespread in the epicontinental realm of NW Transylvania. We point to the presence in this biozone of a series of species with isolated occurrences (rendered in Table 2 with ringed numbers): *Cytherura corniculata* SHEREMETA, 1969-(1), *C. bamburgensis* KEIJ, 1957-(2), *Uroleberis olteanui* WANEK, 1987-(3), *Cytheromorpha* cf. *bulla* HASKINS, 1971-(4), *Tuberocytheridea tuberosa* (LIENENKLAUS, 1900)-(5), *Haplocytheridea heizelini* KEIJ, 1957-(6), *Aulocytheridea faboides* (BOSQUET, 1852)-(7), *Phalcocythere horrescens* (BOSQUET, 1852)-(8), *Cushmanidea scorbiculata* (LIENENKLAUS, 1894)-(9), *Diebelina koeneni* (MOOS, 1969)-(16) (Pl. II, III).

A distinct element of the 11-21 sample sequence is given by the species *Hazelina indigena* MOOS (that occurs at the top of the preceding zone) and *Pterygocythereis fimbriata* (MÜNSTER) (with a larger stratigraphic continuity), characteristic of the Oligocene. That confirms our previous assertion (Wanek, 1989) that in the Brebi Marls sequence the ostracod faunas already have Oligocene elements under the Pycnodonte gigantica Level.

Based on the above data, we propose the Hazelina indigena Zone, defined in the base by the disappearance of the form *Sphaenocytheridea parvogracilis* OLTEANU and the first appearance of the species *Haplocytheridea helvetica* (LIENENKLAUS). This biozone is characterized by the occurrence of species with relatively high frequencies, of very short (? local) ranges, such as: *Eucytheridea solida* (LIENENKLAUS), *Cyamocytheridea buendensis* (LIENENKLAUS), *Alteratrachyleberis* aff. *lienenklausi* (OERTLI), *Cuneocythere praelevis* PIETRZENIUK, *Costa tricostata* (REUSS), *Pterygocythereis ceratoptera* (BOSQUET) etc. The species with isolated occurrences from this interval are: *Idiocythereis bartoniana* HASKINS, 1971-(10), *Occultocythereis rupelica* MONOSTORI, 1979-(11), *Acanthocythereis spinosa* (LIENENKLAUS, 1900)-(12), *Aequacytheridea kochi* (REUSS, 1855)-(13).

The uppermost level of the Brebi Marls in the sequence studied (sample 22) shows an important facies change and can be characterized by the explosive occurrence of the species *Cytherelloidea gantensis* MONOSTORI and *Haplocytheridea helvetica* (LIENENKLAUS). It is worth mentioning at this level the occurrence of the species *Cytheretta tenuistriata* (REUSS, 1853) (species 14 in Table 2), chosen by Olteanu (1977 a) as an index form for the upper biozone.

We also mention that a good possibility of biozonation is suggested by the sequence in time of the species belonging to the genus *Pterygocythereis*. They have short, partially concurrent stratigraphic ranges. Their size and peculiar shape make them quite appropriate for being used as biozone markers.

The associations found in this section (characterized by the dominance of *Krithinae* and *Cytherellidae*, followed by *Argilloecia* and *Cytheropteron*) prove normal marine conditions, with depths corresponding to the outer shelf towards bathyal zone (Peypouquet, 1980), similar with those in the Upper Eocene of the Aquitaine Basin (Ducasse, 1975) and of Hungary (Monostori, 1986). The paleoecological significance of the changes that took place at the uppermost level investigated (sample 22) consists in the setting in of lower salinity environments, proved by the disappearance of stenohaline marine forms and the explosion of high euryhaline ones. At this level we can also appreciate a decrease in the waters depth.

Benthic foraminifers

The first list of foraminifers from the "Bryozoan Beds" was published by Martonfy in 1880 (fide Koch, 1894). More detailed micropaleontological researches were carried out by Barbu (1956, 1961, 1963) who identified in the Jibou area more than 160 foraminifer species, almost all of them benthic. Later on, Iva and Rusu (1982) identified 40 planktonic foraminifer taxa, establishing in the Brebi Marls two biozones: Globorotalia cerroazulensis Zone (under the Pycnodonte gigantica Level) and Globigerina ampliapertura/Pseudohastigerina micra Zone (above this level), placing the Eocene-Oligocene boundary between them.

The analyses carried out by us intended to point to the distribution of benthic foraminifers in the type section of the Brebi Marls and to their stratigraphic and paleoecologic importance.

A first thing worth mentioning is that planktonic foraminifers prevail, are of small size and comparatively not so well conserved as the benthic ones.

Benthic foraminifers, diversified both generically and specifically, occur with variable frequencies (see Table 3). The material in samples 1–4 situated above the Nummulites fabianii Level was subject to a more advanced diagenesis that destroyed most of the test of calcareous foraminifers. In these samples, planktonic foraminifers are almost absent, and the assemblage of benthic ones is poor and contains badly preserved specimens. The situation is the same for the samples 21 and 22 at the uppermost part of the Brebi Marls.

The assemblage of benthic foraminifers is dominated by Miliolids that represent about 20 % from all the taxa identified. In what concerns the number of specimens, the best represented are certain species of the genera *Quinqueloculina, Triloculina, Cibicides, Pararotalia, Heterolepa, Guttulina* and *Globulina* (Pl. IV). There equally are forms that occur sporadically, usually represented by 1–2 specimens. *Sphaerogypsina peruviana*, that appears only in sample 15, but with high frequency, is an exception. Certain species with continuous presence become abundant at certain levels (*Eponides toulmini* in sample 18 and *Textularia budensis* in sample 19).

We are of the opinion that the larger foraminifer[1] lumachels (*Nummulites budensis, N. chavannesi* and *Operculina alpina*) from samples 7 and 9 represent allochthonous hydrodynamical accumulations. Examining the distribution of benthic foraminifers in the Brebi section, it is noticed that until under the Pycnodonte gigantica Level the number of species is reduced to a half, without creating an obvious threshold (Table 3). Sztrákos (1979) makes the remark, that can also be checked up in our case, that in the proximity of the Eocene-Oligocene boundary, the Priabonian species disappear gradually (some of them considered to be typically Eocene continue to lie also in the Early Oligocene) and that most foraminifers characterizing the Oligocene also appear in the Priabonian. It is the case of *Eponides toulmini* and *Textularia budensis*, Eocene forms that pass equally in the Oligocene part of our sequence where they can even proliferate (sample 18 and 19, respectively) or, of *Melonis pompilioides* that, according to Morkhoven et al. (1986), would occur as early as the Oligocene (Zone P 18) but which we could identify in surely Eocene levels (sample 4). *Spiroplectammina carinata* appears to have considerable significance for the Eocene-Oligocene boundary. It is a species known in the Oligocene and the Miocene that first occurs beginning with sample 12, therefore at the boundary between the two ostracod zones. The first part of the section (up to the Pycnodonte gigantica Level) is characterized by the presence of Eocene species of the genus *Pararotalia (P. inermis, P. curry, P. armata* etc.). The forms common for this interval are *Heterolepa dutemplei, Cancris auriculus, Melonis pompilioides* and *M. affinis*. The final part of the section is characterized to a slighter extent by benthic foraminifers. We should also mention the appearance of *Cibicides dampelae transilvanicum* described by its authors (Iva et al., 1967) in the Ciocmani Beds and considered to be strictly Oligocene – and the presence as common forms of *Pararotalia spinigera* and *P. byramensis* (at the top level).

In the benthic foraminifer fauna two assemblages are noticed: one characteristic of the inner shelf represented by miliolids, polymorphinids, discorbids, textulariids (Le Calvez, 1970; Poignant, 1972; Moulnier, 1988) and another one belonging to the outer shelf-upper bathyal zone. This last association includes *Spiroplectamina carinata* s.s., the genus *Cancris* (dom-

[1] Larger foraminifera have been determined by Bombiţă to whom we are very much obliged.

inant in the outer central shelf zone), the genera *Cibicides* and *Heterolepa* (proliferating at the border of the continental shelf) and representatives of the genera *Baggina* and *Giroidina*, forms of some greater depth – upper bathyal zone (Smith, 1971; Morkhoven et al., 1986). We are of the opinion that the mixing up of the two assemblages is due to a secondary transportation from the inner shelf towards the outer shelf. The abundance in planktonic foraminifers indicate a sedimentation at a certain distance from the shore of an open sea, with normal marine salinity.

Calcareous nannoplankton

The study of the nannoplankton assemblage of the Brebi Marls began with the paper by Popescu and Gheţa (1972). Their determination is in concordance with the results obtained by Martini and Moisescu (1974) who separated the standard zones of calcareous nannoplankton NP 21 and NP 22. Mészáros et al. (1973) and Mészáros and Moga (1987) identified in the Brebi Marls the zones NP 19 and NP 20, but the nannofossil lists better suggest a younger age. Gheţa (1984) placed the whole formation in zones NP 21-22 (see also Micu, Gheţa, 1986). Two important biohorizons, the Nummulites fabianii Level and the Pycnodonte gigantica Level, could be correlated with the NP 21 nannoplankton zone.

The most detailed data on the nannoplankton distribution published by Popescu and Gheţa (1972) and Martini and Moisescu (1974) were carried out on sections in the Cluj region. According to these results the lower part of the Brebi Marls can be characterized by the high abundance in *Isthmolithus recurvus* DEFLANDRE, *Coranulus germanicus* STRADNER and *Lithostromation perdurum* DEFLANDRE. The upper part contains the local acme of the species *Transversopontis zigzag* ROTH and HAY/= *T. obliquipons* DEFL. (HAY, MOHLER and WADE)/, *T. obliquipons* s.s. and *T. pulcheroides* (SULLIVAN) BÁLDI-BEKE.

The nannoplankton from the Brebi Marls stratotype is not very abundant and diverse, but, usually, a moderate quantity of nannofossils occurs in the samples, with the exception of samples 1–4 and 21, 22, where the number of specimens is rather low. The nannofossils are well preserved, however a strong overcalcifying makes the determination of the species difficult, mainly in the lower portion of the Jolj section (samples 1–3).

The nannoplankton assemblage identified in the type section (Table 4) is characterized by the great quantity of placoliths, the remarkable absence of helicosphaerids and the small or medium quantity of discoliths. The ratio of holococcoliths in the population is rather high. It can be due to the relative nearshore character of the marls having been deposited in the shelf realm. The high abundance of *Lanternithus minutus*, *Zygrhablithus bijugatus* and *Isthmolithus recurvus* indicates the influence of cool sea waters. The amount of tropical-subtropical elements (i.e. sphenoliths and discoaster) is very low. Nevertheless, the great quantity of the species *Reticulofenestra bisecta* in this section is difficult to explain, since this species is very typical and common in the nannoplankton assemblages of moderate and cool climate belts. A remarkable temperature difference appears between this biotop and the rest of the Eocene in Transylvania.

The stratigraphic position of the Brebi Marls stratotype corresponds to the standard nannoplankton zones NP 21–NP 22 (a few characteristic nannofossils are figured on Plate V). This ranging is supported by the following facts:

– absence of the *Discoaster saipanensis* and *D. barbadiensis* throughout the section;

– presence of *Reticulofenestra umbilica* and *R. hillae* occurs from the bottom to the top of the section;

– occurrence of *Cyclococcolithus formosus* only up to the upper third of the section.

The most important biostratigraphic events in the section under study are (see Table 4):

– beginning of the acme of *Lanternithus minutus* and *Zygrhablithus bijugatus* (from the sample 4 and 5, respectively);

– acme of the *Ericsonia obruta* + *E. subdisticha* (between samples 10–15);

– last occurrence of *Cyclococcolithus formosus* (sample 18) marking the boundary between zones NP 21 and NP 22.

The latest data concerning the position of the most important micropaleontological events in the proximity of the Eocene-Oligocene boundary, calibrated with the scale of magnetic polarity, have shown that these events are not synchronous everywhere in the world (see vol. TEE, Pomerol, Premoli-Silva, ed. 1986). Nevertheless a sequence of events could be established on the basis of several deep sea drillings in the Atlantic (La Brecque et al., 1983; Hsü et al., 1984; Jenkins, 1986) and land sections from Italy (Monechi, 1986; Nocchi et al., 1986 and "The ad-hoc meeting on the Eocene-Oligocene boundary at Ancona, Italy, 1987). The following events from old to new have been established:

– beginning of the *Lanternithus minutus* acme;

– last occurrence of the *Turborotalia cerroazulensis* group;

– last occurrence of *Hantkeninas* and *Cribrohanthenínas*, a moment considered to mark the Eocene-Oligocene boundary;

– acme of *Ericsonia obruta*;

– last occurrence of *Cyclococcolithus formosus*.

Unfortunately, in the Carpatho-Balkan region, the Hantkeninas are missing from the Upper Priabonian and the nearest event in the nannoplankton assemblage that should approximate this boundary would be the acme of the *Ericsonia obruta*. There exist data that show that in the Contessa Quarry section in Umbraria (Italy) this acme would start immediately under the last occurrence of the Hantkeninas.

According to what has been shown, on the basis of nannoplankton, the Eocene-Oligocene boundary would be placed in the Brebi Marls within NP 21 Zone, somewhere under the Pycnodonte gigantica Level (within the interval of samples 11–12). Placed between samples 11 and 12, this boundary would be situated towards the upper part of Zone CP 16 a (that ends with the extinction of the *Ericsonia subdisticha* acme) and would correspond to the boundary between the two local zones of ostracods. That fact appears to be in disagreement with the results based on planktonic foraminifera (Iva, Rusu, 1982), according to which this boundary would be placed at an upper level. It is therefore necessary to resume the study of planktonic foraminifers on several sections in North-West Transylvania to solve this problem.

Dinoflagellates, spores and pollens

Previous palynological studies on the Paleogene deposits in Transylvania have evidenced, in the Brebi Marls of the Gilău Area, a poor spore-pollen assemblage and the following dinoflagellates: *Deflandrea phosphoritica, Deflandrea* sp. and *Tytthodiscus* (Popescu et al.; Pl. XXIII, XXIV).

The analysed samples from the Brebi Section have yielded relatively varied dinoflagellates and spore-pollen assemblages, wherein marine elements (dinoflagellates) are dominant.

The degree of preservation of dinocysts is good, even very good with certain species. In the deposits studied the following dinoflagellates have been registered, with variable frequence: *Tytthodiscus* sp., *Deflandrea phosphoritica, Kisselovia* sp., *Wetzeliella (Rhombodinium) perforata, Batiacasphaera* cf. *compta, Cordosphaeridium funiculatum, C. gracilis, Cribroperidinium* cf. *giuseppei, Wetzeliella clathrata, Wetzeliella (Rhombodinium) draco, Pentadinium laticinctum, Palaeocystodinium golzowense* etc. (see Table 5, Pl. VI).

Of them, *Deflandrea phosphoritica, Pentadinium laticinctum, Palaeocystodinium golzowense* are known since the Lower Eocene-Middle Miocene, *Wetzeliella (R.) draco* since the Middle Eocene-Oligocene, *Cordosphaeridium funiculatum, Batiacasphaera* cf. *compta, Wetzeliella (R.) perforata* since the Middle Eocene-Upper Eocene, *Wetzeliella clathrata* since the Upper Eocene-Lower Oligocene. It can be noticed that the zone of concurrence of the cited species is part of the Upper Eocene interval.

The range of the species *W. (R.) perforata* is referred, in the zonation of Costa and Downie (1976), to the W. (R.) perforata Zone, considered the equivalent of the calcareous nannoplankton zones NP 17–NP 21 (?). That is the zonation we have used in this paper.

According to Chateauneuf and Cavagnetto (1978), the range of the species *W. (R.) perforata* covers the standard zones W 12 (Rhombodinium perforatum Zone) and W 13 (Kisselovia clathrata angulosa Zone) in their zonation.

Later on, Bujak (1979, fide Williams, Bujak, 1985) redefined the Rhombodinium perforatum Zone, restricting it, as compared to the initial definition, to the equivalent of zones NP 18–NP 20 (Upper Eocene). We can specify that the species *Wetzeliella (R.) perforata* goes up to the upper part of zone NP 21.

In Romania, *W. (R.) perforata* was recorded in the Plopu Beds (Upper Eocene) on the Trotuş Valley (Ionescu, Alexandrescu, 1987) and subsequently in other formations of the same age in the East Carpathians (Ionescu, unpubl. data).

The microplanktonical assemblage identified does not yield elements for establishing the Eocene-Oligocene boundary in the Brebi Section.

The upper part of the section have not yielded conclusive elements from biostratigraphic point of view, either for the Eocene or for the Oligocene. The situation is somehow comparable to that in the Umbria region (Italy), where, at the Eocene-Oligocene boundary, the very absence of dinoflagellates is noticed (Nocchi et al., 1986).

The presence of the genera *Tytthodiscus* and *Botriococcus* would indicate that the deposits under study were sedimented not very far from the shore.

The palynological assemblage is poorly represented both by spores and by pollen, with the exception of conifers, represented by the species *Pityosporites alatus* and *P. microalatus* (that at certain levels are abundent – see Table 5).

The taxa *Verrucatosporites favus, Trilites multivallatus, Cicatricosisporites* sp., *Triatriopollenites engelhardthoides, Inaperturopollenites hiatus* are sparse.

The identified land microflora has no biostratigraphic value. It can only possibly give climate indications.

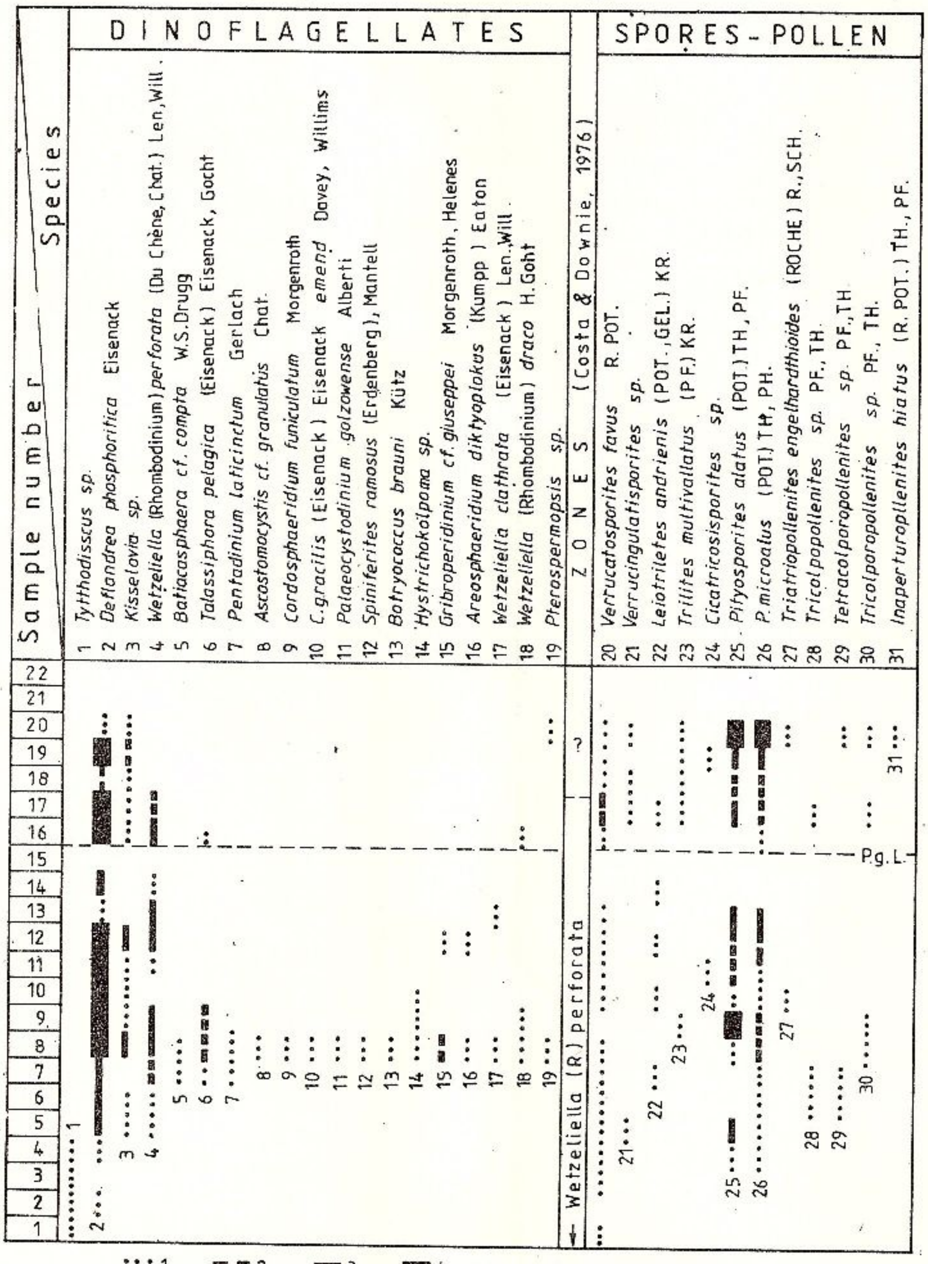
Within the microfloral assemblage mentioned above no reworked elements seem to occur.

Conclusions

The researches made on certain groups of fossils from the type section of the Brebi Marls (Jolj Valley) and the complementary section of "Strîmtură" (Ortelec) bring up a series of new data for the biostratigraphic picture of the Eocene-Oligocene boundary in Transylvania (fig. 2) and for evaluating the environmental conditions in which the terrigenous-carbonatic deposits in the region were sedimented.

Table 5

Dinoflagellate and spore-pollen distribution in the type section of the Brebi Marls.



1, < 10 specimens; 2, 10-30 specimens; 3, 31-100 specimens; 4, >100 specimens; 5, Pycnodonte gigantica Level.

The mollusc fauna from the marls belongs to the Mediterranean bioprovince and is of Priabonian type, the elements characteristic of the Oligocene appearing at the level of the Hoia Beds, at the same time with the paleogeographical changes, due to the regression during the Early Oligocene. The biohorizon bearing the big gryphaeid *Pycnodonte gigantica* is situated in the upper third of the stratotype and has been considered in the field the Eocene-Oligocene boundary, established on the basis of planktonic foraminifers (Iva, Rusu, 1982).

The study of ostracod assemblages shows a gradual passage from the lower part of the section, with the clear prevalence of Eocene forms, at the upper part, where the Oligocene ones begin to prevail. Based on stratigraphic distribution and frequence, two local ostracod biozones have been separated, Sphaenocytheridea parvogracilis Zone and Hazelina indigene

Zone, between which the Eocene-Oligocene boundary would be placed (Fig. 2).

The distribution of benthic foraminifers from the sections under study shows a reduction to half the number of species up to under the Pycnodonte gigantica Level, without giving birth to an obvious threshold. It is thought that the first occurrence of the species *Spiroplectammina carinata* could have a special significance for drawing up the Eocene-Oligocene boundary at the level evaluated on the basis of calcareous nannoplankton.

The study of calcareous nannoplankton confirms the presence of zones NP 21 and NP 22 and evidences a series of events that, from older to younger are (Fig. 2):

– beginning of the acme *Lanternithus minutus* and *Zygrhablithus bijugatus*;

– acme of *Ericsonia obruta + E. subdisticha*, and

– last occurrence of *Cyclococcolithus formosus* (marking the boundary between NP 21 and NP 22 Zones).

The sequence of events in the nannoplankton evolution appears to be in agreement with the latest data obtained in the world (see vol. TEE, ed. Pomerol, Premoli Silva, 1986). In the absence of the Hantkeninas, whose last occurrence was established as the Eocene-Oligocene boundary, this boundary can be approximated by the acme of *Ericsonia obruta*. Thus, being situated rather low in the Brebi Section (approximately in the middle of the formation) and in disagreement with what was known from the study of planktonic foraminifera, it is necessary to resume researches of these foraminifers on several sections in Transylvania.

Palynological studies have evidenced for the first time a relatively diverse assemblage of dinoflagellates, establishing the presence of the Wetzeliella (Rhombodinium) perforata Zone and the range of the index form (that has been known for sure only up to the level of nannoplankton NP 20 Zone). In Transylvania *Wetzeliella (R.) perforata* is found up to the upper part of NP 21 Zone.

The Brebi Marls were deposited south of the Ţicău-Preluca carbonate platform in an asymmetrical, large E-W trending trough-like, bounded to the south by land. The groups of organisms studied from the Brebi Section indicate for this region a shelf zone, covered by normal marine waters, with depths corresponding to the outer shelf. At the end of the Priabonian, Transylvania was part of the Mesogean Realm. The abundance of certain nannoplankton species, such as *Lanternithus minutus, Zygrhablithus bijugatus* and *Isthmolithus recurvus* around the Eocene-Oligocene boundary, indicates also the influence of cool waters.

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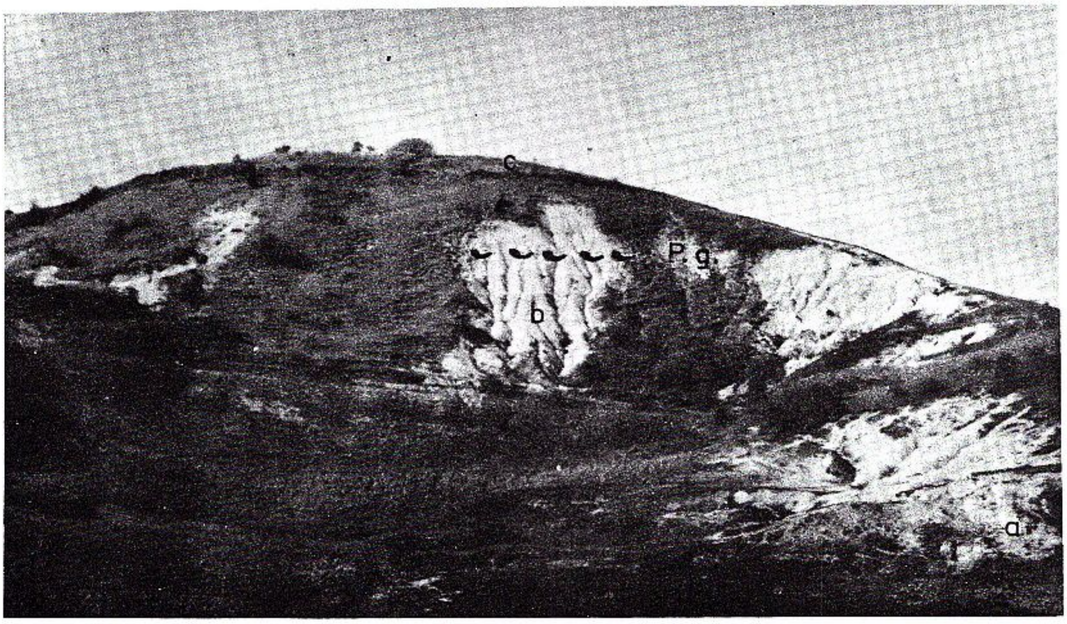

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Plate I

Fig. 1 – Type-section of the Brebi Marls in the right slope of Jolj Valley. a, Cluj Limestone; b, Brebi Marls (P.g. – Pycnodonte gigantica Level); c, Hoia Beds.

Fig. 2 – *Pycnodonte gigantica* (SOL.). Left valve, x 1, Bodia.

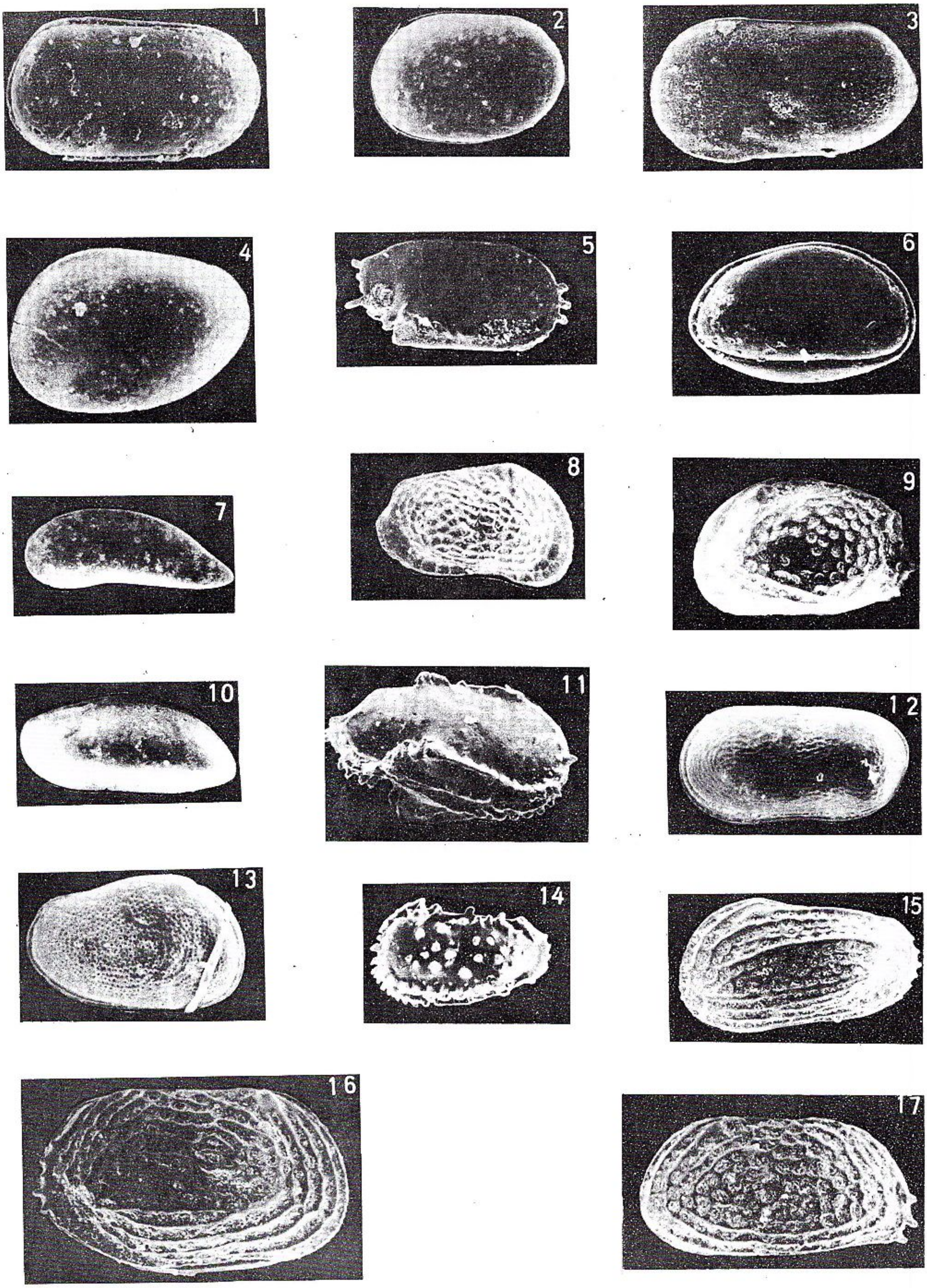
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Plate II

Fig. 1 – *Cytherelloidea gantensis* MONOSTORI, x 82.
Fig. 2 – *Cytherella dentifera* MÉHES, x 49.
Fig. 3 – *Cytherella beyrichi* (REUSS), x 74.
Fig. 4 – *Cytherella consueta* DELTEL, x 85.
Fig. 5 – *Pterygocythereis cornuta* (ROEMER), x 45.
Fig. 6 – *Cardobairdia hungarica* MONOSTORI, x 97.
Fig. 7 – *Paracypris prima* MANDELSTAM, x 38.
Fig. 8 – *Echinocythereis dadayana* (MÉHES), x 46.
Fig. 9 – *Pocornyella limbata* (BOSQUET), x 61.
Fig. 10 – *Argilloecia ampulloidea* MANDELSTAM, x 72.
Fig. 11 – *Pterygocythereis aquitanica* DUCASSE, x 55.
Fig. 12 – *Cytherella strangulata* DUCASSE, x 64.
Fig. 13 – *Loxoconcha* aff. *kuiperi* KEIJ, x 79.
Fig. 14 – *Pterygocythereis fimbriata* (MÜNSTER), x 42.
Fig. 15 – *Cytherella tricostata* OLTEANU, x 53.
Fig. 16 – *Alteratrachyleberis striatopunctata* (ROEMER), x 55.
Fig. 17 – *Alteratrachyleberis* sp. aff. *lienenklausi* (OERTLI), x 42.

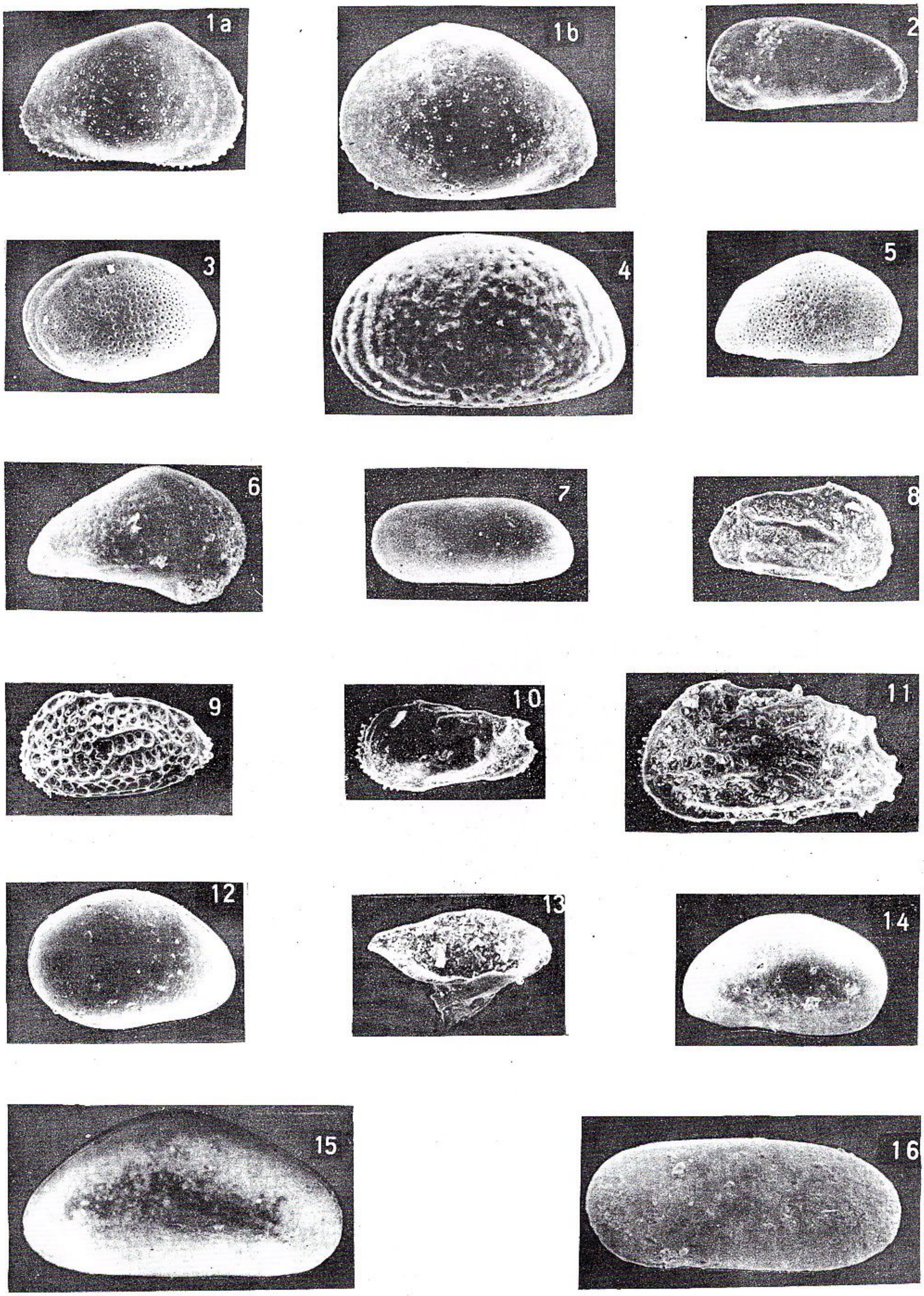
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Plate III

Fig. 1 – *Aequacytheridea perforata* (ROEMER), 1 a – x 61, 1 b – x 66.
Fig. 2 – *Sphaenocytheridea parvogracilis* OLTEANU, x 89.
Fig. 3 – *Cuneocythere proplevis* PIETRZENIUK, x 62.
Fig. 4 – *Haplocytheridea helvetica* (LIENENKLAUS), x 68.
Fig. 5 – *Eucytheridea solida* (LIENENKLAUS), x 52.
Fig. 6 – *Phlyctocythere cocaenica* KEIJ, x 103.
Fig. 7 – *Parakrithe* aff. *cristallina* (REUSS), x 65.
Fig. 8 – *Hazelina indigena* MOOS, x 44.
Fig. 9 – *Costa tricostata* (REUSS), x 57.
Fig. 10 – *Occultocythereis costalis* HASKINS, x 65.
Fig. 11 – *Quadracythere* aff. *vohrenkampi* MOOS, x 68.
Fig. 12 – *Xestoleberis gantensis* MONOSTORI, x 65.
Fig. 13 – *Cytheropteron gulincki* KEIJ, x 67.
Fig. 14 – *Xestoleberis* sp. 2, x 62.
Fig. 15 – *Aglaiocypris arcuata* (MÜNSTER), x 58.
Fig. 16 – *Turmaekrithe fragilis* PIETRZENIUK, x 115.

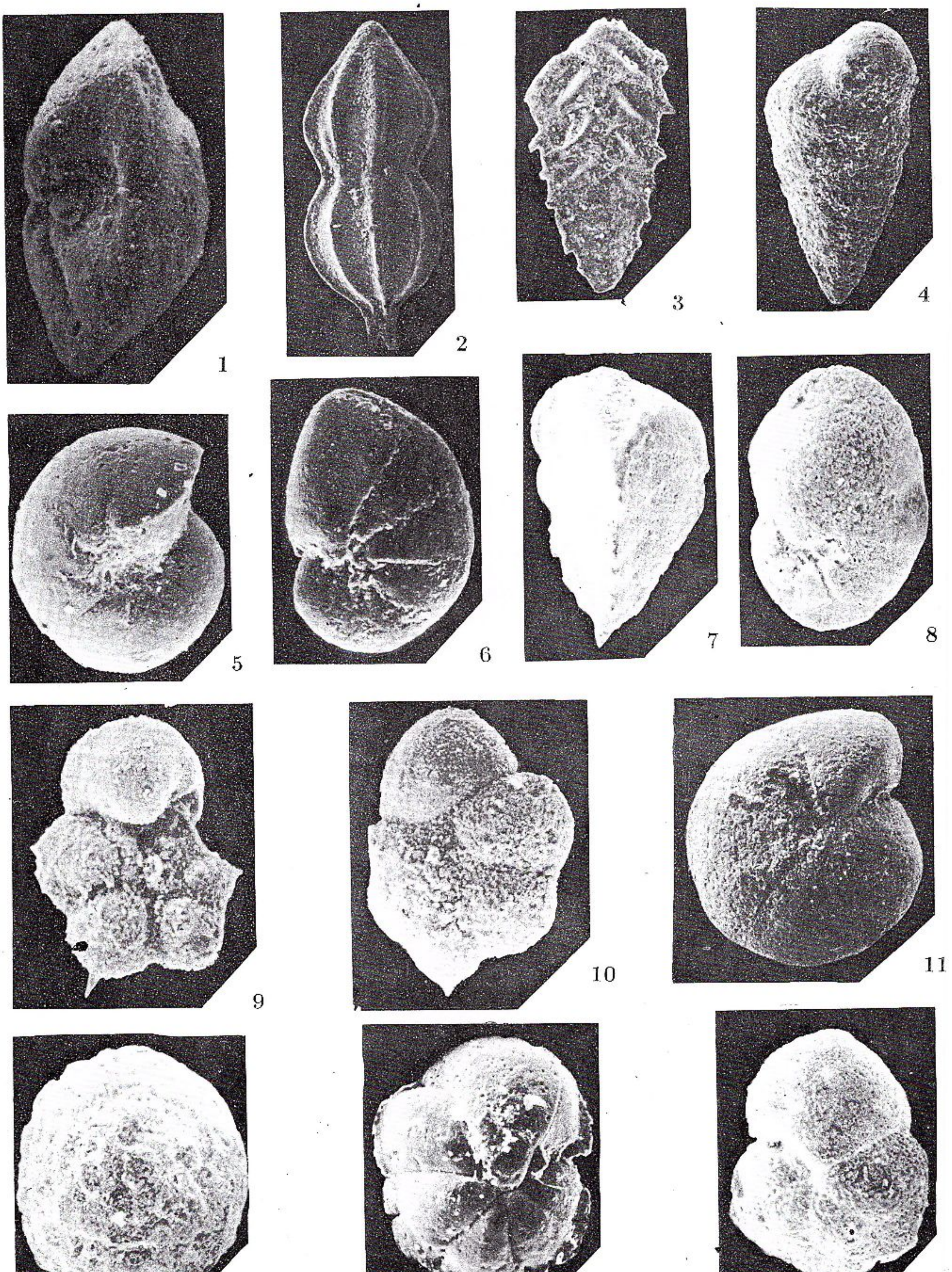
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Plate IV

Fig. 1 – *Eponides toulmini* BROTZEN, x 62.
Fig. 2 – *Nodosaria budensis* HANTKEN, x 24.
Fig. 3 – *Spiroplectammina carinata* (D'ORBIGNY), x 19.
Fig. 4 – *Textularia budensis* HANTKEN, x 19.
Fig. 5 – *Gyroidina soldanii* D'ORBIGNY, x 42.
Fig. 6 – *Florilus elongatus* D'ORBIGNY, x 51.
Fig. 7 – *Reussella spinulosa* (REUSS), x 29.
Fig. 8 – *Cancris auriculus* (FICHTEL & MOLL), x 45.
Figs. 9, 10 – *Pararotalia spinigera* (LE CALVEZ), 9–x 58, 10–spiral side – x 50.
Fig. 11 – *Heterolepa dutemplei* (D'ORBIGNY), x 33.
Fig. 12 – *Sphaerogypsina peruviana* BERRY, x 27.
Fig. 13 – *Cibicides dampelae intermedium* IVA, GHEORGHIAN, GHEORGHIAN, x 38.
Fig. 14 – *Cibicides dampelae transilvanicum* IVA, GHEORGHIAN, GHEORGHIAN, x 17.

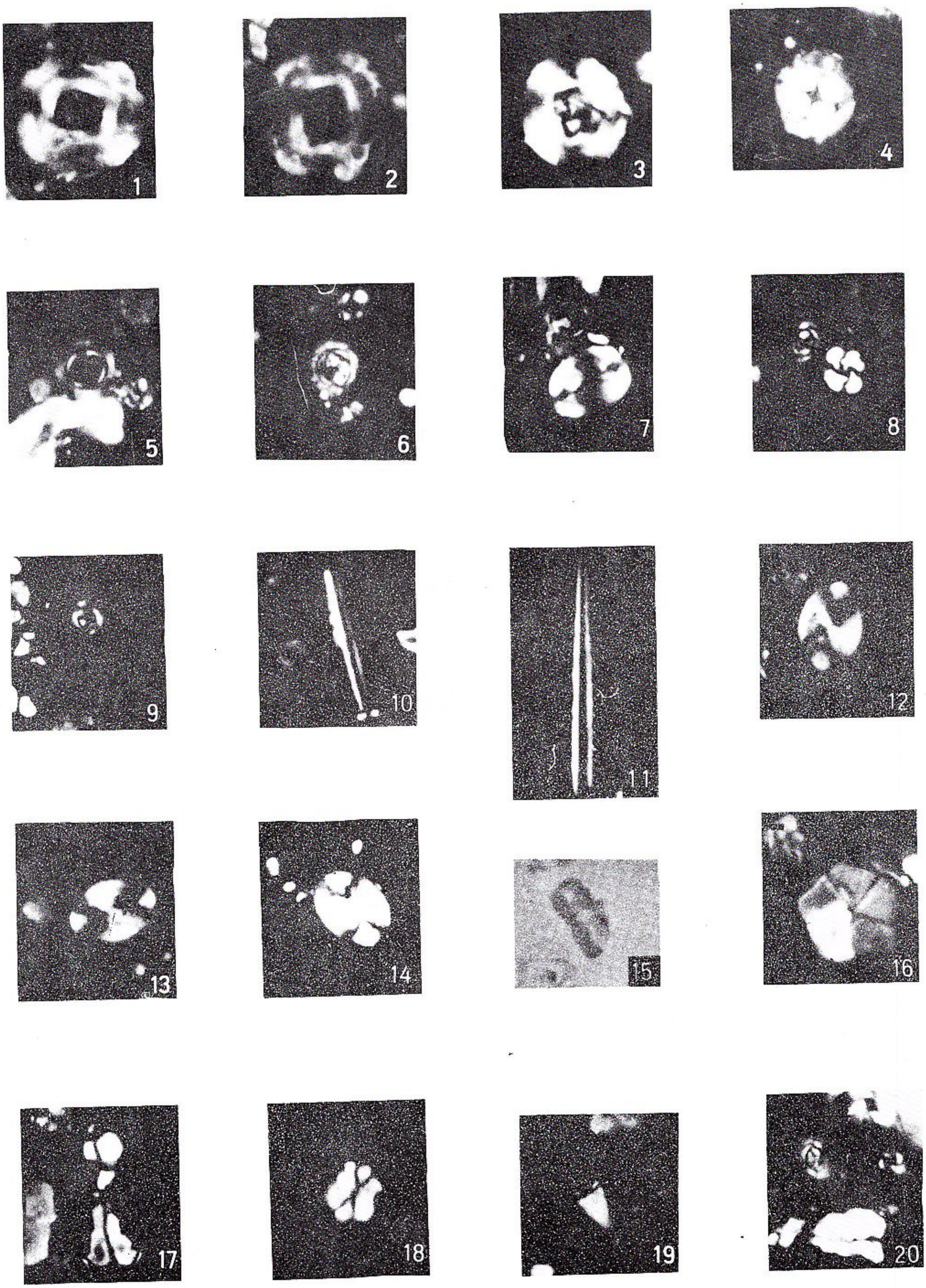
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Plate V

Figs. 1, 2 – *Reticulofenestra umbilica* (LEVIN), MARTINI & RITZKOWSKI, sample 6.
Fig. 3 – *Reticulofenestra hillae* BUKRY & PERCIVAL, sample 6.
Fig. 4 – *Cyclococolithus formosus* (KAMPTNER) ROTH, sample 6.
Figs. 5, 6 – *Cyclococolithina kingii* ROTH, sample 6.
Fig. 7 – *Cyclococolithina kingii* ROTH, and overcalcified specimen of *Pontosphaera multipora* KAMPTNER, sample 6.
Fig. 8 – *Ericsonia subdisticha* (ROTH & HAY) ROTH and a small placolith, sample 10.
Fig. 9 – *Ericsonia subdisticha* (ROTH & HAY) ROTH, sample 14.
Fig. 10 – *Rhabdosphaera perlonga* DEFLANDRE, sample 10.
Fig. 11 – *Rhabdosphaera spinula* LEVIN, sample 6.
Figs. 12, 13 – *Transversopontis pulcheroides* (SULLIVAN) BÁLDY-BEKE, sample 6.
Fig. 14 – *Pontosphaera multipora* (KAMPTNER) ROTH, sample 6.
Fig. 15 – *Istmolithus recurvus* DEFLANDRE, sample 6.
Fig. 16 – *Braarudosphaera bigelowii* (GRAN & BRAARUD) DEFLANDRE, sample 6.
Fig. 17 – *Zygrhablithus bijugatus* (DEFLANDRE) DEFLANDRE, sample 14.
Fig. 18 – *Lanternitus minutus* STRADNER, sample 5.
Fig. 19 – *Sphenolithus predistentus* BRAMLETTE & WILCOXON, its axis at 45° to the nicols, sample 10.
Fig. 20 – *Zygrhablithus bijugatus* (DEFLANDRE) DEFLANDRE and *Ericsonia obruta* PERCH-NIELSEN, sample 14.

All fig. x 1200 and between crossed nicols except fig. 15 – between parallel nicols.

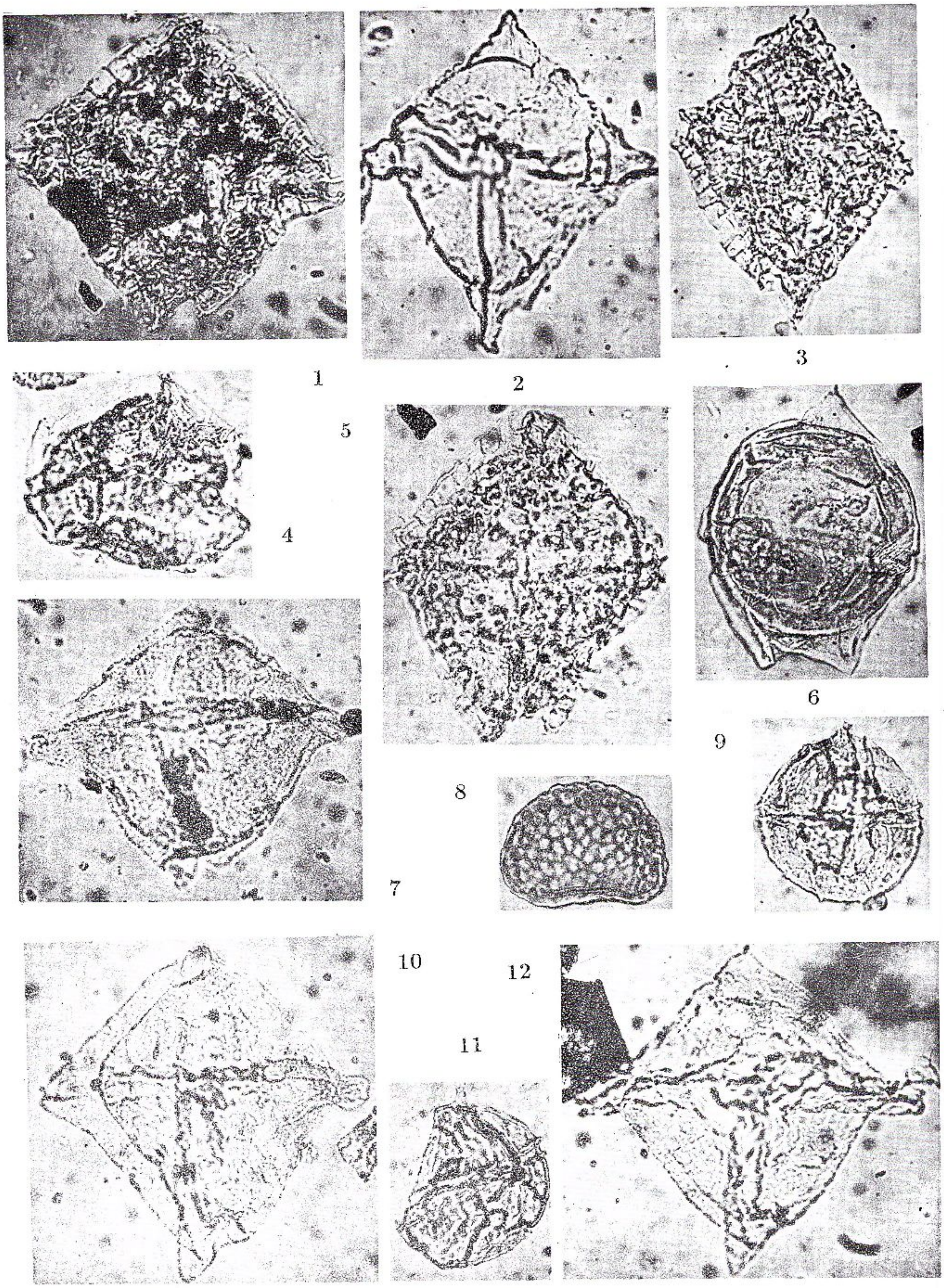
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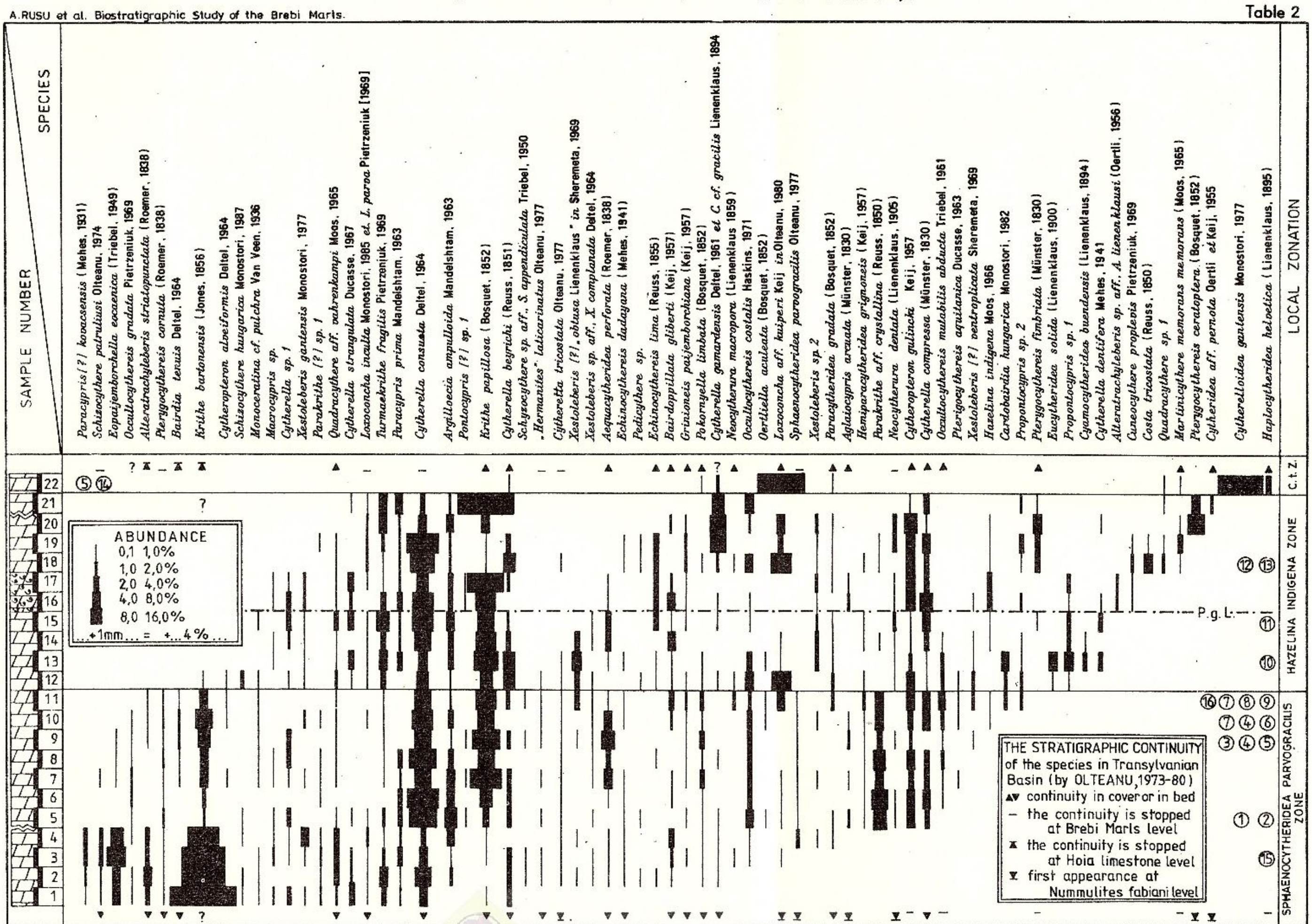
Plate VI

Fig. 1 – *Wetzeliella clathrata* (EISENACK) LEN., WILL, sample 8, 107μ/112.5μ.
Fig. 2 – *Wetzeliella (Rhombodinium) draco* (DU CHÊNE, CHAT.) LEN., WILL, sample 16, 117.5μ/105μ.
Fig. 3 – *Kisselovia* sp., sample 8, 110μ.
Fig. 4 – *Pentadinium laticinctum* GERLACH, sample 20, 75μ/80μ.
Fig. 5 – *Kisselovia* sp., sample 16, 120μ/102.5μ.
Fig. 6 – *Deflandrea phosphoritica* EISENACK, sample 18, 100μ/80μ.
Figs. 7, 10, 12 – *Wetzeliella (Rhombodinium) perforata* (DU CHÊNE, CHAT.) LEN., WILL., 7-sample 7, 107.5μ/125μ; 10-sample 8, 120μ; 12-sample 8, 112.5μ/137.5μ.
Fig. 8 – *Verrucatosporites favus* POT., sample 6, 52.5μ/40μ.
Figs. 9, 11 – *Cribroperidinium* cf. *giuseppei* MORGENROTH, HELENES, 9-sample 7, 67.5μ/60μ; 11-sample 8, 62.5μ/60μ.

OSTRACOD DISTRIBUTION IN THE TYPE-SECTION OF THE BREBI MARLS

(the ringed numbers correspond to species cited into the text only)

Table 2



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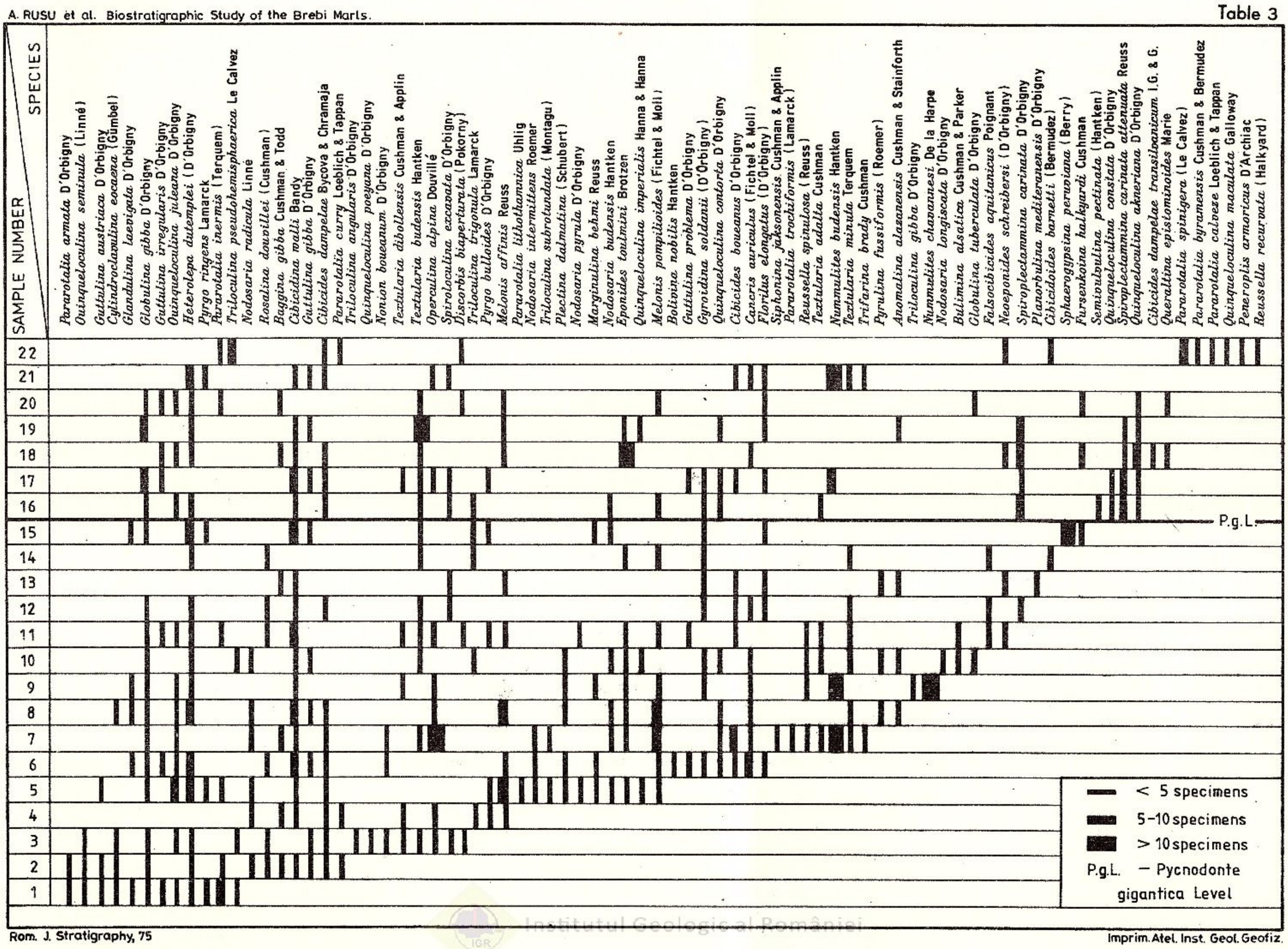
BENTHIC FORAMINIFERA DISTRIBUTION IN THE TYPE-SECTION OF THE BREBI MARLS

Table 3



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CALCAREOUS NANNOPLANKTON DISTRIBUTION IN THE TYPE-SECTION OF THE BREBI MARLS

Table 4

SAMPLE NUMBER \ SPECIES	*Transversopontis obliquipons* (Defl.) Hay, Mohler, Wade	*Sphenolithus moriformis* (Brönn & Strad.) Braml. & Wilcoxon	*S. predistentus* Bramlette & Wilcoxon	*Rhabdosphaera spinula* Levin	*Reticulofenestra bisecta* Hay, Mohler, Wade	*R. umbilica* (Levin) Martini & Ritzkowski	*R. hillae* Bukry & Percival	*R. minuta* Roth	*Coccolithus pelagicus* (Wallich) Schiller	*Cyclicargolithus floridanus* (Roth & Hay) Bukry	*Lanternithus minutus* Stradner	*Zygrhablithus bijugatus* (Deflandre) Deflandre	*Chiasmolithus oamaruensis* (Defl.) Hay, Mohler, Wade	*Cyclococcolithina kingii* Roth	*Cyclococcolithus formosus* Kamptner	*Rhabdosphaera perlonga* Deflandre	*Coccolithus eopelagicus* (Braml. & Ried) Braml. & Sullivan	*Ericsonia obruta* Perch-Nielsen	*Reticulofenestra ornata* Müller
22							1	F	R	F	F	R							2
21					2	1	1	C	F	C	C	F					R	F	1
20				1	R	F	F	C	F	C	C	C				F	R	R	
19					1	3		C	F	C	C	F						1	
18		R	2	R	R	4	1	A	F	A	A	C			1	2	F	R	
17		R	2		R	F	4	A	F	A	C	C			1	2	R	R	1
16		R	1	1	R	3		A	C	A	C	C		1			F	R	
15				R	C	4	3	A	C	A	A	A				2	R	F	
14		F	2	R	F	5	3	A	F	A	A	A			6	2	R	F	
13				R	R	3	3	C	F	C	C	F		2			R	F	
12	1	C	F	F	R	6	3	A	F	A	A	C			2	2	R	F	
11		F	4	F	R	3	2	A	F	A	A	A			1	R	R	F	
10		F	4	R	F	2	1	A	C	A	M	A		2	1	R	F	F	
9		C	5	R	C	F	F	A	F	A	M	A			1	R	C	2	
8		C	3	R	F	F	F	A	F	A	A	A			3	1	F	3	
7		C		R	R	F	F	A	F	A	A	A		2	2	R	R	1	
6	1	F	3	R	R	F	F	A	C	A	A	C		F	R	R	F	1	
5	1	F	3	R	R	F	F	A	A	A	A	A		R	5	R	F	1	
4		F	1	3	R	F	F	A	C	C	C	F			4	1			
3		R	2	R	2	F	5	C	C	C	F	R		1	2	1	3	2	2
2		3		R	3	F	3	A	F	F	R	R	1	1	1	1	3	1	1
1	1	1	2	3	3	F	2	A	F	F	F	F							

SAMPLE NUMBER \ SPECIES	*Braarudosphaera bigelowii* (Gran & Braarud) Defl.	*Pontosphaera plana* (Braml. & Sullivan) Haq	*Ericsonia subdisticha* (Roth & Hay) Roth	*Pontosphaera multipora* (Kamptner) Roth	*Isthmolithus recurvus* Deflandre	*Micrantholithus flos* Deflandre	*Coccolithus cf. crassipons* (Boucher)	*Micrantholithus vesper* Deflandre	*Transversopontis pulcheroides* (Sullivan) Báldi-Beke	*Helicopontosphaera bramlettei* Müller	*Discoaster tanii* Bramlette & Riedel	*Coronocyclus nitescens* (Kampt.) Braml. & Wilcoxon	*Pontosphaera latoculata* (Bukry & Percival) Perch.-Niels.	*Orthozygus aureus* (Stradner) Braml. & Wilcoxon	*Rhabdosphaera creber* Deflandre	*Reticulofenestra callida* (Perch.-Nielsen)	*R. cf. lokodensis* Báldi-Beke	*Reworking from the Paleogene*	*Reworking from the Cretaceous*	STANDARD ZONES (Martini 1971)	ZONES (Okada & Bukry 1980)
22	F			F					1											NP22	CP16c
21			R	R												1	1				
20	F		R	F				R								1					
19				F														1			
18	F			F	1			F	R							2		1		NP21	CP16b
17	R			F	R			R	F						1			1			
16	R			F				R	R												
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15	F		R	F	R				F												
14	R		R	F	F													1	1		
13			R	R	R				F												
12	R		F	F	3			R	C				1	R	1			1			CP16a
11	F		R	R	1		2	F	F			2	2								
10	F		R	C	R		F	R													
9	F			F	C		2	R													
8	C		1	F	C			R	3	1	2										
7	F			C	2			R													
6	F		2	C	C			1	3												
5	F		2	F	F	R	1	R	R										1		
4					1																
3	1	3	1	2														1			
2	1																				
1																		R			

LEGEND

3 = no. of specimens
R < 7 specimens
F = 7 – 50 sp.
C > 50 specim.
A = several hundreds of specim.
M = several thousands of specim.
P.g.L. = Pycnodonte gigantica Level

BIOSTRATIGRAPHIC EVENTS

beginning of the L. minutus acme
beginning of the Z. bijugatus acme
I. recurvus acme
E. obruta acme
E. subdisticha acme
last occurence of C. formosus

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ASPECTS OF THE POSSIBILITIES OF LITHOLOGICAL CORRELATION OF THE OLIGOCENE – LOWER MIOCENE DEPOSITS OF THE BUZĂU VALLEY

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Abstract: The paper is centered on the analysis of the geometrical position of the lithological markers represented by the Stănila Shales, the Tylawa Limestones, the Jaslo Shales, the Vineţişu, Mlăcile and Bătrîni Tuffs, within the Oligocene-Lower Miocene deposits as well as on the manner in which they can be used as synchronous correlation elements. The conclusion can be drawn that the Tylawa Limestones and the couple of the Vineţişu and the Mlăcile Tuffs are most important for correlations. The Jaslo Shales still raise correlation problems, from one scale to another, at each of their levels. The Bătrîni Tuffs also represent markers worth considering, especially in the zones where their presence is more constant. The correlations made on the basis of these lithological markers (Plate I) bring into relief the heterochronous character of the boundaries between the "horizons" separated within the Oligocene-Lower Miocene deposits.

I. Introduction and History

It is Stoica (1944), Grigoraş (1951), Popescu (1952) who established the lithological sequences in the Buzău Valley. The last two have proved that in the region there are two totally different facies, an inner one – of the Pucioasa Beds with Fusariu Sandstone – and an outer one – bituminous with Kliwa Sandstone, between which there are interfingering relations. In their interfingering area a mixed facies can be separated in which the lithological elements characteristic of the main facies are superposed in the same vertical cross-section. The detrital material of the facies of the Pucioasa Beds with Fusaru Sandstone comes from inner Carpathian sources, while that of the bituminous facies with Kliwa Sandstone comes from the outer part, from the Vorland.

Each of the mentioned facies (equally the mixed one) has its own sequences, that are quite constant on the longitude, but that transversally show varieties of thickness and composition, which have always led to correlation problems.

Grigoraş (1955) presents a clear picture of his view concerning the way of correlating the lithological separations in the Oligocene deposits between the Putna Valley and the Buzău Valley. Popescu et al. (1960, unpubl. report) modified this sketch for the part east of the Buzău Valley, proving that the Vineţişu beds and the Podu Morii Beds are not superposed but are partly synchronous. In fact this idea corresponds to that of Popescu (1952) and Pătruţ (1951) who had already drawn this conclusion for the region west of the Buzău Valley, using as a marker the tuffs that crop out both in the Vineţişu Beds and in the Podu Morii ones.

With certain detail modifications, the sequences established by the above cited authors have remained valid so far. Their correlation has also varied, within certain limits, especially for the intervals situated from the level Vineţişu-Podu Morii upwards. The sketch accepted today almost unanimously is that obtained on the occasion of the drawing up of the Covasna Sheet (1968) of the Geological Map of Romania, scale 1:200.000 (see the lithological columns).

The fact that researches have been resumed in a classic region like that of the Buzău Valley, where seemingly everything is known, has made us try and use another method for checking up the validity of the conclusions of previous researches. While separating already known lithological sequences we have taken the decision to follow as closely as possible – both transversally and longitudinally the lithological levels whose composition is in contrast with that of the detrital formations making up the main mass of the Oligocene-Lower Miocene deposits (more or less laminated) tuffs and carbonatic rocks respectively.

It was Anton (1944), Popescu (1952), Pătruţ (1951) who first asserted that certain tuff levels represent synchronous lithological markers, being used as such both by the last two authors mentioned and by prospectors who worked in this region during 1960, 1961 (Popescu et al., unpubl. report), 1969 (Vasilescu et al., unpubl. report).

Stoica (1944) first cited calcareous shales in the Colţi region overlying the lower menilites, therefore at the level of the Tylawa Limestones. In 1959 Wdowiarsz found calcareous shales ressembling the Jaslo ones in the Vineţişu Valley cross-section. Later on, calcareous shales were identified both overlying the lower menilites and in the Pucioasa Beds facies by Popescu et al. (1961), but both levels have been described as Jaslo Shales. In 1969 Jucha named the lower level of calcareous shales Tylawa Limestones. The coccolithic nature of these rocks has been demonstrated by Nowak (1965).

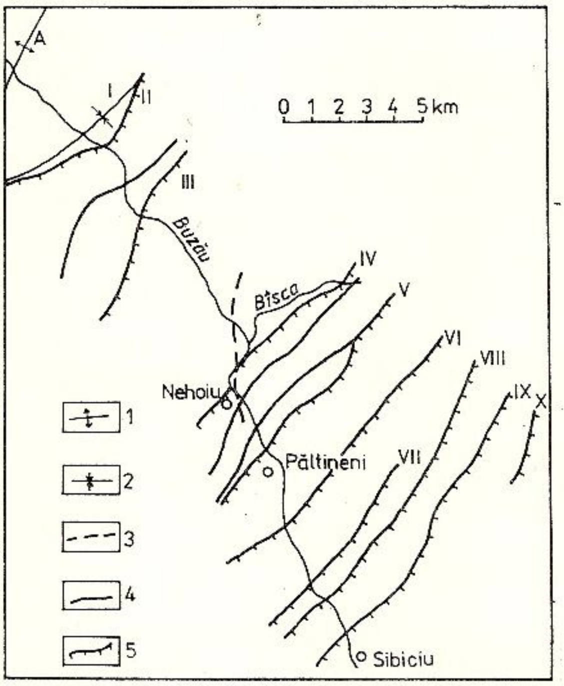
Models regarding the possibilities of correlation of calcareous shales have been achieved both for separate areas of the Carpathians-North Carpathians (Jucha, Kotlarczyk, Koszarski; Jytko, 1961; Haczewski, 1984); Ucrainian Carpathians (Sachin, 1958 from Jucha, Kotlarczyk, 1961); East Carpathians (Kotlarczyk, 1961; Alexandrescu, Brustur, 1985; Ionesi, 1986) and even for the whole area of the Carpathians (Visoţki, Grigoraş, 1964; Kotlarczyk, 1980; Haczewsky in Leszczynski, Vendorff, Haczewski, 1987).

The Buzău Valley and its tributaries offer an excellent cross-section (Figure) through all the facies of the Oligocene- Lower Miocene deposits of the Tarcău nappe, facies that are accompanied both by tuffs and by calcareous shales.

The present paper intends to present both the cineritic and the carbonatic levels and the way in which they can be used as synchronous levels of correlation. The detail columns for each structure are obtained by using first of all the data from the Buzău Valley to which data are added from neighbouring cross-sections, so that the sequences of plates should be as complete as possible.

II. Description of the levels of calcareous rocks and cinerites

As the classical lithological "Horizons" have already been analyzed in many papers, starting with those in which they were separated for the first time (papers cited in the previous chapter), we shall no longer refer to them. But we shall insist on the levels of tuffs and carbonatic rocks that have been mentioned either seldom or just never in the previous papers. They will be described in stratigraphic order, beginning with the older ones, irrespective of their petrographic nature, but mentioning in the same order the lithological horizons in which they lie.



Tectonic sketch of the Paleogene flysch in the Buzău Valley. A, Băile Siriu Syncline; I, Giurca Syncline; II, Monteoru Scale; III, Vineţişu Scale; IV, Leordeanu Scale; V, Păltineni Scale; VI, Mlăjet Scale; VII, Capu Dealului Scale; VIII, Valea Lupului Scale; IX, Malu Alb Scale; X, Găvanele Scale; 1, Anticline axis; 2, Syncline axis; 3, Transcurrent fault; 4, Secondary overthrust in the main scales; 5, Scale.

1. Stănila Shales

On almost the whole cross-section of the Buzău Valley, the Eocene deposits are overlain by a pile (Lingureţi Beds, Stoica, 1944) with mixed features, made up of rocks looking like the Eocene formations – green pelites, micaceous sandstones with laminated structure, thin marls looking like the Globigerina-containing ones, that alternate with rocks typical of Oligocene deposits – brown coloured marly-calcareous shales with scales and fragments of fossil fishes, shales that, when weathered, become white, resembling in this respect the Jaslo ones.

These shales are very well exposed near the Stănila locality, which is the reason for describing them under this name. Although it is not very easy to examine the Stănila Shales because of their small sizes and of the landslides frequently affecting the Lingureşti Beds, they have nevertheless been identified equally in the Valea Rea and the Giurca structures.

2. Tylawa Limestones

The first levels of stratiform calcareous rocks interbedded in the base of the deposits in bituminous facies were described in the North Carpathians as the Tylawa Limestones (Jucha, 1969).

It is under this name that we describe all the thin beds (0.5–12 cm) of white or white-yellowish limestones, compact or laminated in structure, that overlie the lower level of menilites in the lower horizon of menilites and of white bituminous marls. We mention that laminated limy rocks contain numerous remains of fossil fishes. Because of both their white colour and of their stripes, that are characteristic of them, they can be easily distinguished from the environing rock mass.

In well exposed cross-sections it can be noticed that the Tylawa Limestones are grouped in two distinct piles. The lower pile (generally under 1 m) is intercalated in the lower part of the white bituminous marls (with diatomites – Filipescu, 1934) and is made up both of compact limestones, about 1 cm thick, and of laminated limestones, about 1.2 cm thick. The upper level, much better represented than the lower one, reaches 1–2 m or only exceptionally 10–14 m. Calcareous shales are between 0.5 and 12 cm thick and compact limestones around 1 cm. The proportion between the thicknesses of the calcareous rocks and those of the pile containing them ranges between 4.5 percent and 35 percent. But we should mention that the total thickness of the interbeds of shales and Tylawa Limestones remains approximately constant, of 60–70 cm. It is therefore quite obvious that the thickness of these levels, as well as that of the distance between them (10–20 cm) depend on the quantity of white bituminous marls interbeds and on the sandstones separating them. The figures are given only for the development area typical of the bituminous facies with Kliwa Sandstone, as, west of it, Tylawa Limestones grow thinner up to their complete disappearance.

3. Jaslo Shales

From the special lithological terms of the Oligocene-Lower Miocene deposits in the Carpathians, it was the Jaslo Shales that were first identified (Uhlig, 1883).

In the Buzău Valley, the interbeds of Jaslo Calcareous Shales of the facies of the Pucioasa Beds with Fusaru Limestones were found out for the first time by Wdowiasz (1959).

In the pile of rocks assigned to the Jaslo Shales there are both white calcareous shales (1–10 cm thick) containing fossil fish fragments and compact white limestones (0.5–8 cm) or grey-yellowish marly limestones (up to 20 cm thick).

The Jaslo Shales appear interbedded both in the Pucioasa Beds with Fusaru Sandstone (in the horizon bearing the same name) and in the bituminous facies with Kliwa Sandstone (lower horizon of the Kliwa Sandstone).

Their development varies from one facies to another. According to the data in the region they are well developed in the mixed facies at the level of the Pucioasa Beds in the Vineţişu Scale, where they occur in three distinct levels and reach a thickness of about 95–100 cm, on a stratigraphic interval of 65–70 m. To the inner part they have been found up to the Giurca Syncline, where on another few meters they amount to 33 cm thickness. Out of the Vineţişu Scale, the Jaslo Beds are no longer so important, as they generally appear at only one level and have much smaller thicknesses. Thus, in the Leordeanu Scale they have about 23 cm, and more to the exterior, in the Păltineni Scale, on an interval about 6.5 m thick, the Jaslo Shales in the Buzău Valley cross-section (Topilele) amount to only 2.5 cm. 2.25 km to the north-east, following the direction of the same structure; the level mentioned above contains 25 cm of Jaslo Shales on a thickness of only 1.7 m.

Even more to the exterior, in the cross-section exposed on Valea Rea, in the Mlăjet Scale, the situation is somehow different. Thus, here appear two interbeds of Jaslo type. The lower one (3 m thick) is made up of coffee-coloured hard shales, only some levels of which (1–4 cm) are more calcareous, like the Jaslo ones. Overlying them, at about 150 m stratigraphic thickness, there is a laminated calcareous sandstone, made up only of elements of Jaslo Shales. The Jaslo Shales with a position similar to those of the lower level occur also in a more inner scale, i.e. the Păltineni one, on the northern slope of Vîrful Oii.

The outermost scale in which Jaslo Shales have been found is one named Valea Lupului-Colţii de Jos. In this scale, on a stratigraphic thickness of about 85 cm, Jaslo Shales and Limestones are interbedded, totalling 40 cm thickness.

From the field data at our disposal it becomes quite obvious that neither the number nor the total thickness of calcareous interbeds is constant, both of them depending on local accumulation conditions.

4. Valea Rea Marls

In the lower horizon of the Kliwa Sandstone, the piles of massive sandstones are separated on rhythmic levels with more or less micaceous turbiditic sandstones, alternating with grey and coffee-coloured pelites. In one of these piles, at about 450 m above the Tylawa Limestones, in the Valea Rea cross-section (right tributary of the Buzău Valley), on an interval of 25–30 cm, three beds less than 1 cm thick of calcareous marls are intercalated. The marls are whitish, yellowish when weathered, and contain many small-sized Globigerinae, which we call Valea Rea Marls. It is very difficult to watch these levels because of their small sizes. Nevertheless, when well exposed, they can be identified. Thus, they have also been identified in the Valea Fişici sequence, belonging to the outermost scale (Găvănele Scale).

5. Vineţişu Tuffs

In the Vineţişu Valley cross-section, in the beds bearing the same name, on 1.2 m, two beds (20 and 10 cm) are intercalated, of strongly bentonized olive tuff. The same type of rock, either fine or coarser-grained (when it contains also biotite) appears interbedded in all the well exposed cross-sections of the Vineţişu Beds and the Podu Morii Beds. Only in one case they appear in the prevailingly dysodilic pile (Topilele Beds) always placed between the above mentioned beds and the lower horizon of the Kliwa Sandstone. The thickness of the bentonite beds ranges between 10 and 30 cm. They are interbedded on an interval that is not thicker than 5–6 m. When well exposed, bentonites can be easily recognized due to their dark colours and to the depression-like shapes of their beds, caused by their brittle character.

6. Mlăcile Tuffs

The left slope of the Buzău Valley, in the area of the Mlăcile locality, offers good conditions of studying the Oligocene-Lower Miocene deposits. Thus, a little bit to the north of this locality, in the Podu Morii Beds (at their upper part) on 2.5 stratigraphic metres, five beds of white tuffs, tuffs which we call Mlăcile Tuffs, appear. The bed thickness ranges between 20 and 100 cm. Their grain-size also varies from one bed to another. The coarser varieties contain a relatively great amount of biotite. The thickest bed also has an obvious current lamination, which shows that the cineritic material was redistributed by bottom currents. These tuffs and especially the thick, biotite-containing one can be identified in all the good cross-sections in the Podu Morii Beds and even in those of the Vineţişu Beds (on the left slope of the Bisca Valley, a little bit downstream the Fetica locality). But the same type of tuff has also been discovered under the Podu Morii Beds, like in the case of the Buzău Valley cross-section, south of Păltineni (Topilele), where there are three beds of tuff, close to one another, the thickest of which reaches 1 m.

Since 1950, this type of tuff has been considered to belong the Dacitic Tuff category (Pătruţ, 1951).

7. Bătrîni Tuffs

The Podu Morii Beds are overlain by the upper horizon of the Kliwa Sandstone. The lower part of this horizon, on variable thicknesses, contains interbeds of light grey or whitish pelites, resembling the ones in the Podu Morii Beds. Unlike these, the pelites at the upper part of the same horizon are coffee-coloured, clayey, of dysodilic type. In this last part of the upper horizon of the Kliwa Sandstones, at several levels, thin beds are intercalated of grey, rust-coloured, or most frequently yellowish bentonitized tuffs; these bentonitized tuffs go up in the sequence beyond and even above the diatomites and the upper menilitic shales.

Tuffs and bentonites with the same characteristic features occur also west of the region studied by us in the Teleajen Valley (Filipescu, 1934; Pătruţ, 1951) as well as north of the Drajna Syncline, in the facies of the Pucioasa Beds with Fusaru Sandstones. In this last zone they are very well developed in the region of the Bătrîni locality, which has made us in fact describe them under this name. The thickness of the tuff beds ranges between 1 and 20 cm or even up to 1 m, the beds occurring on a thickness of a few hundred meters.

To end the description of the three groups of cinerites – Vineţişu, Mlăcile, Bătrîni – we mention that we have used the notion of tuffs for historical priority, as they are in fact mostly represented by bentonites.

8. Muncelu Cărămănesc Gypsum (lower gypsum)

The Oligocene-Lower Miocene deposits in the Tarcău Nappe are limited at the upper part by a pile of white gypsum of metric thickness with regional distribution, which represents an excellent stratigraphic marker.

Both the gypsum pile and its relations with its subjacent deposits are quite visible in a road cut section, north of Muncelu Cărămănesc, which is, in fact, why we have called it like that. It crops out equally along the Drajna Syncline.

III. Possibilities of correlation

In this chapter an analysis will be made of the way in which rock levels can be used for correlating the horizons of the Oligocene-Lower Miocene deposits and for pointing to the contribution these levels can have in the improvement of the correlation sketch.

The Stănila Shales – the oldest interbeds of calcareous shales – have a special importance for correlation as they overtake, towards the inner part, the area of distribution of both the lower menilites and the Tylawa Limestones. But for the time being, it is difficult to make more detailed remarks, as their extension is not known. In any case, in the Buzău Valley transverse, all the known occurrences occupy the same level and, consequently, can be considered to be practically synchronously accumulated.

The Tylawa Limestones occur in all the structures lying more to the exterior than the Giurca Syncline. They can be followed approximately on the whole length of the structures concerned both to the north-east (Săndulescu, Săndulescu, 1964), and to the west.

In addition, the great development of the Tylawa Limestones, starting from the North Carpathians, continuing in the East Carpathians, up to the eastern part of the South Carpathians, even if it has certain differences in thickness, represents an argument for using them as an element of regional correlating (Koltarczyk, 1981), an idea that is equally supported by the data obtained by us (Plate I).

Establishing in detail the position of the Tylawa Limestones and using them as a stratigraphic marker point out two quite important facts, i.e.: the unequal development of the white bituminous marls is due to their replacement by dysodilic shales, starting from the upper part to the lower part, so that the greatest lateral development is that of the white bituminous marls that lie immediately above the basal pile of menilites; the possibility of using as a synchronous marker for the area of the bituminous facies of the upper level of the thin shales, which occurs in the base of the lower horizon of dysodilic shales.

The Tylawa Limestones also represent a checking element that nevertheless points to the correctness of the interpretations given by Romanian geologists, who have considered for a long time that the lower menilite horizon (Lower level) represents a synchronous correlation pile on the basis of which the lower part of the Oligocene deposits has been established, either of various facies or of various units.

More delicate correlation problems begin to appear with the deposits overlying the horizon of the lower menilites and the white marls. This horizon is overlain by that of the lower dysodilic shales whose basal part sequence (6–12 m) is made up of real dysodiles, in the remaining part of the horizon shales alternating with thin beds of Kliwa Sandstones.

The sudden increase in the supply of terrigenous material both in the inner part and in the outer part has led to the accumulation of two horizons – Pucioasa Beds with Fusaru Sandstones in the inner part and the lower horizon of the Kliwa Sandstone in the outer part – with thicknesses that vary between 1000 and 600 m. The relative lithological monotony of these "Horizons" has made their more detailed dividing impossible. The only contrasting lithological elements are represented by the calcareous Jaslo Shales that are considered by Polish geologists (Jucha, Kotlarczyk, 1961; Kotlarczyk, 1980; Haczewski, 1984) to be a marker horizon, used even along the whole chain.

In the region under analysis the Jaslo – type calcareous shales occur interbedded along thicknesses ranging between 1.5 and 6.5–7 m. An exception is the Vineţişu Valley cross-section, where the main two levels are separated by 70 m of flysch deposits. To the outer part, in the Mlăjet Scale, there are two distinct levels: a lower one, at 160 m above the Tylawa Limestone and an upper one, represented by coarse sandstone with calcareous elements of Jaslo type)?, at about 350 m above the same Tylawa Limestones. The latter can be correlated with interbeds of the other scales situated between 450 and 550 m above the Tylawa Limestones. In exchange, the lower level of the Mlăjet Scale (Valea Rea) has its geometrical equivalent only in the scale immediately inner (of Păltineni), 7 km north-east of the Buzău Valley, on the northern slope of Vîrful Oii, where at about 200 m far from the Tylawa Limestones the same type of less calcareous shales crop out.

It is therefore clear that the Jaslo – type Shales should be correlated quite attentively, step by step, both on the direction of the beds and from one structure to another, so that the transition from one marker to another should be avoided as it would lead to wrong correlations.

For the time being there are only few possibilities of correlation based on the levels of Globigerina-bearing marls in Valea Rea, because of the limited area they have been identified on. But it can be easily noticed that the Valea Rea Marls appear approximately at the same level with the youngest interbeds of the Jaslo Shales and exactly in the scales where the shales are absent (Mlăjet, Găvanele). The situation presented leads to the hypothesis that the Valea Rea Marls could represent a stratigraphical equivalent of the upper level of the Jaslo Shales. This hypothesis seems to be almost supported by the fact that the Jaslo Shales also contain Globigerinae, that are so frequent in the Valea Rea Marls. Nevertheless, the hypothesis needs some checkings before becoming a certitude.

Both at the upper part of the horizon of the Pucioasa Beds with Fusaru Sandstone and at the lower level of the Kliwa Sandstone, there is a prevailingly pelitic level (dysodiles and grey marls + more or less micaceous sandstones), of variable thickness, which we call the Topilele Beds. In point of geometry always placed under the Vineţişu Beds and the Podu Morii ones, this pile corresponds to the lower part of what Teysseire (1911, fide Pătruţ, 1952) separated as the Podu Morii Beds as well as to the horizon of the upper dysodiles (Dumitrescu, 1952) in the Vrancea Half-Window. It clearly differs from the suprajacent deposits that, beginning with Popescu (1942, fide Popescu, 1952) have been dscribed as Podu Morii Beds (Izvoarele Beds, Pătruţ, 1951).

In the Vineţişu Beds and in the Podu Morii ones there are tuff interbeds that have been pointed out by the very first geological researches in the region (Izvoarele Beds, Pătruţ, 1951).

In the Vineţişu Beds and in the Podu Morii ones there are tuff interbeds that have been mentioned by the very first geological researches in the region. Their detail examination has evidenced the following data:

– the tuffs are grouped in two distinct levels, superposed on a distance of 25–50 m;

– each level is made up of more or less weathered cinerites, of various aspects, i.e. the lower level (Vineţişu Beds) of tuffs and bentonite having a dark green or olive colour, which contrasts with the white or slightly greenish colour of the tuffs and the bentonites in the upper level (Mlăcile Tuffs);

– this type of sequence – Vineţişu Tuffs, Mlăcile Tuffs – is constant on the whole length of the zone under study, of about 25 km, but it can also be identified along the strike of the structures, both north-east and south-west.

From the situation presented above but also considering their origin, it can be inferred that the two tuffs levels no doubt represent synchronous horizons that can be used as markers of lithological and stratigraphic correlation.

But the detail study of the sections has also pointed to an important aspect, i.e. that the position of the green-olive and white tuff couples is not constant as referred to the boundaries of the deposits in which they are intercalated. Thus, the white tuff with biotite (of Mlăcile) occurs in the following positions: in the Vineţişu Scale (at Fetica) at about 150–160 m above the lower boundary of the Podu Morii Beds; to the exterior, in the Valea Lupului Scale at 70–170 m far from the same boundary etc. To end with, the Păltineni Scale, where the same tuff occurs in an altogether special position, i.e. in the Topilele Beds, at about 7–8 m under the lower boundary of the Podu Morii Beds. We think that, from mentioning this situation it clearly results that the lower lithological boundary of the Vineţişu and the Podu Morii Beds is heterochronous. From the examination of the columns presented (Plate 1), it can be noted that this is also the situation at the upper part of the same beds. Consequently, even if the flysch deposits of the type Vineţişu-Podu Morii Beds represent a pile accumulated along the whole width of the transversal line of the Buzău Valley, this pile is heterochronous, (both the starting and the ending moment of its accumulation), showing variations, quite important at times.

In the region of study, the Vineţişu Beds are overlain by 10–15 m of deposits of the upper horizon of the dysodilic shales. South-west, along the direction of the Vineţişu Scale, this horizon becomes thicker and thicker, having a more important development in the area of the Bătrini locality, situated west of Starchiojd. In this region, in the upper horizon of the dysodilic shales, at 20–25 m far from its base, several levels of white or green tuffs are intercalated (see the medallion from the column of the Vineţişu Scale), partly bentonitized, partly silicified.

We come back to the region of study to point out that the first tuffs that overlie the Mlăcile Tuffs appear as early as the upper horizon of the Kliwa Sandstone in the Capul Dealului and Valea Lupului Scales. In the two most external scales, the first tuff interbeds appear immediately below the horizon of menilites (+ upper diatomites), tuffs which can be correlated as far as their position is concerned, with the Fălcău ones (Alexandrescu et al., 1984). Concerning the occurrence of the Bătrini Tuffs a lithological detail is also worth considering: as regards the outcropping conditions, the upper horizon of the Kliwa Sandstone has been separated as a unitary entity, although in well exposed cross sections it can be noticed that the pelites at the lower part are marls and grey clays of Podu Morii type, while those at its upper part are coffee-coloured clayey dysodilic shales. The Bătrini Tuffs always occur only in that part of the upper horizon of the Kliwa Sandstone that contains interbeds of dysodilic shales. That situation pleads for the synchronism between the tuffs interbedded in the upper horizon of the dysodilic shales lying over the Vineţişu Beds and those in the upper horizon of the Kliwa Sandstone. But the number and thickness of the Bătrini Tuffs vary from one section to another, which makes their correlation, bed by bed, at least for the time being, very difficult. Nevertheless the first tuffs representing the beginning of a new cycle of explosions can be considered synchronous along the whole width of the Tarcău Nappe on the Buzău Valley. Correlating first the cinerite episode and then the upper one in the order of their sequence, the image in Plate 2 is obtained, according to which the base of the upper menilite and diatomite "horizon" is situated at different thicknesses with respect to the median level of the Bătrini Tuffs. This can be easily explained by the lens-like (large scale) accumulation of the Kliwa Sandstone situated between the median tuff level and the base of the upper menilite and diatomite "horizon".

The dysodilic shales with Kliwa Sandstone interbeds overlying the upper menilite and diatomite "horizon" have been separated as the supramenilite "horizon". It does not differ at all from the Kliwa Sandstones with dysodiles separating the various menilite or diatomite levels. Its lower boundary is considered to be where the menilite intercalations disappear. The correlation of the menilite and/or diatomite levels very clearly indicates that the position of this boundary varies from one profile to another, having therefore a heterochronous character.

The real initial (prior to folding) relationships between the deposits discussed have been presented in the palynspastic sections from Plate 2. It should be mentioned here that the 25–26 km of the analysed section on the Buzău Valley represent about half of the initial width of 48–50 km over which the investigated Oligocene-Lower Miocene deposits accumulated.

IV. Conclusions

The analysis made in the previous chapters revealed some general aspects regarding the Oligocene-Miocene deposits, which will be further presented.

The Stănila shales extending towards the inner part of the investigated zone, beyond the lower menilite horizon, can be used as a correlation element for the base of the interval under consideration.

The Tylawa Limestones show an impressive longitudinal constancy from the Northern Carpathians to the eastern part of the South Carpathians. Transversely it is found that they disappear in the north-western structures, being present only in the inward extension zone of the bituminous facies. But it is worth mentioning that they also occur in the Oligocene deposits (Valea Caselor facies) from the post-tectonic cover of the units of Upper Cretaceous tectogenesis which extends up to the eastern part of the South Carpathians (Suslăneşti). Due to their impressive areal extension and constant position, the Tylawa Limestones constitute an excellent correlation element. The correlation made on their basis confirms the previous ones, which have considered as lithological markers the lower menilite and white marble "horizons".

The Jaslo shales have been also identified along the entire flysch zone, from the Northern Carpathians to the East Carpathians Bend Zone. The westernmost zone with calcareous shales in the Oligocene is situated in the South Carpathians, west of Suslăneşti (the Seţurile Limestones – Ştefănescu et al., 1983). In the Tarcău Nappe of the investigated zone the Jaslo shales are best represented both in the Mixed Facies and that of the Pucioasa Beds with Fusaru Sandstones. Outwards, in the bituminous facies with Kliwa Sandstone, both the number and thickness of the Jaslo shales become obviously smaller. The fact that they occur in place at rather great distances (70–150 m) and cannot be followed along strike (owing to outcropping conditions) makes their correlation difficult. However, by correlating the main, and in most cases the single level of occurrence, a lens-like accumulation of the Kliwa Sandstone can be noticed.

The Vineţişu and *the Mlăcile tuffs* couple can be identified from the basin of the Buzău Valley westwards throughout the outcropping zone of the Paleogene flysch. Several tuffs levels are quoted (Sikora, 1959) in the Northern Carpathians. Their correlation with those under discussion is very difficult. It is supposed that they correspond with the tuffs at Dynia W and Konskie (Wieser in Koszarski, 1985), which are the first cinerites overlying the Jaslo shales. Each tuff of this couple represents synchronous horizons, due to which it could be demonstrated that the lower boundary of the Vineţişu and Podu Morii type flysch is heterochronous.

The Bătrîni tuffs have also a great distribution in the East Carpathian Bend Zone and west of it, being identified (Ştefănescu in Dimitrescu et al., 1985) also in the upper horizon of the dysodilic shales overlying unconformably the crystalline schists of the Getic Domain. These tuffs seem to be equivalent to the tuffs at the base of the Gura Văii Beds from the Getic Depression. Due to their great distribution they represent a marker used on a large scale. The geometric correlation of the tuff intercalations from this group provides an image in which the upper horizon of the dysodilic shales is synchronous with the upper part of the upper "horizon" of the Kliwa Sandstone (which contains tuff and dysodilic shales intercalations) as well as the supramenilite "horizon". It is worth mentioning that only the lower part of the upper "horizon" of the Kliwa Sandstone, which contains grey pelite interbeds, should be considered an equivalent of the upper part of the Vineţişu Beds. Therefore, according to these correlations, most of the boundaries of the "horizons" of the Oligocene-Lower Miocene deposits should be regarded as lithological boundaries and not as synchronous stratigraphic boundaries.

In conclusion, of the three main levels in which calcareous shales were found, only the median one, that is the Tylawa Limestones, undoubtedly shows the characteristics of a synchronous horizon. Although the Jaslo shales show a considerable areal extension, they still raise transverse correlation problems from one scale to another, especially when they occur at several geometric levels. The most important cineritic rocks are those at Vineţişu and Mlăcile, which, being easy to identify and displaying a great regional constancy, can be surely used as synchronous markers. Finally it is worth noting that the utilization of the Bătrîni Tuffs as synchronous correlation markers points to the fact that both the menilites originating from diatomites and the Kliwa Sandstones separating them (the supramenilite horizon inclusively) show a lenticular development.

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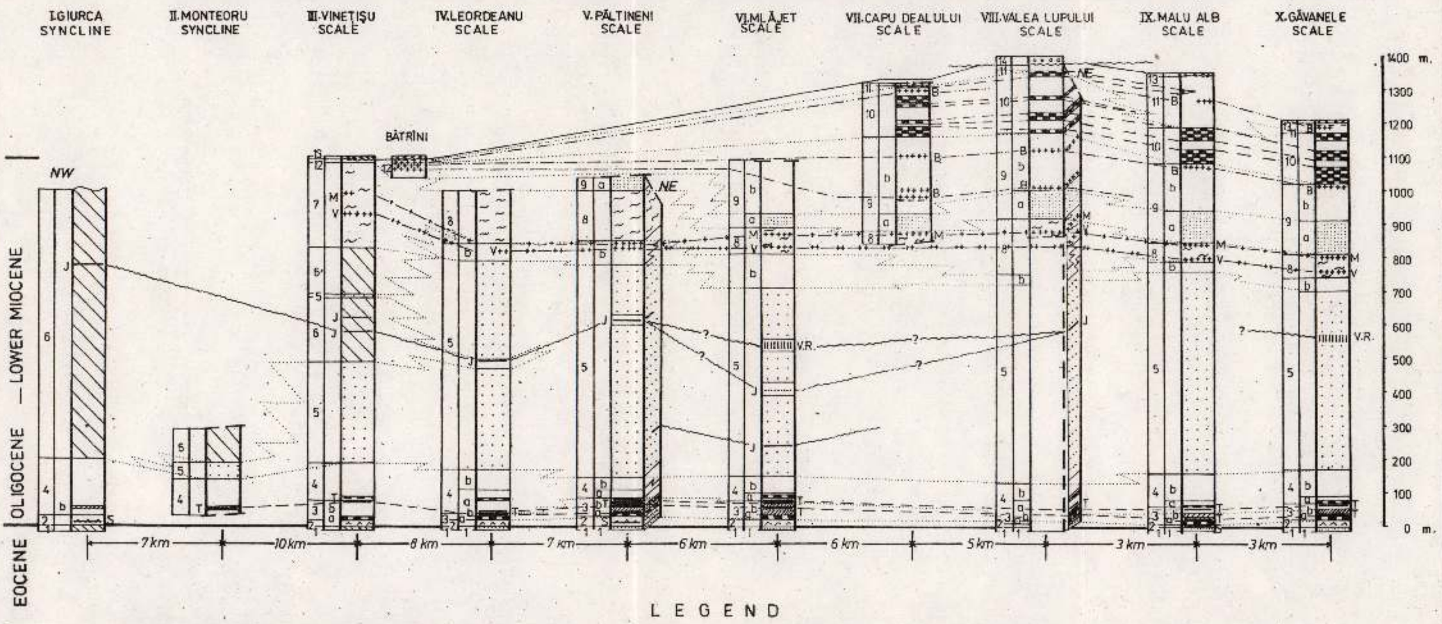
LITHOLOGICAL COLUMNS OF THE OLIGOCENE-LOWER MIOCENE DEPOSITS IN THE BUZĂU VALLEY



L E G E N D

THE FACIES OF THE PUCIOASA BEDS WITH FUSARU SANDSTONES

1. Horizon of the Globigerinaebearing marls
2. Lingurești Beds
4. Lower dysodilic shales horizon (b. white bituminous marls)
6. Horizon of the Pucioasa Beds with Fusaru Sandstone
7. Vinetișu Beds
12. Upper horizon of dysodilic shales

BITUMINOUS FACIES WITH KLIWA SANDSTONE

1. Horizon of the Globigerinaebearing marls

1.1. Lucăcești Sandstone

2. Lingurești Beds
3. Lower horizon of menilites (a) and of white bituminous marls (b)
4. Lower horizon of dysodilic shales without (a), or with sandstone interbeds (b)
5. Lower horizon of Kliwa Sandstone, b. Topilele Beds
8. Podu Morii Beds
9. Upper horizon of the Kliwa Sandstone with grey pelite interbeds (a), or of dysodilic shales (b)
10. Upper horizon of menilites and diatomites
11. Supramenilitic horizon

S = Stanila Shales
T = Tylawa Limestone
J = Jasło Shales
V.R = Valea Rea Marls
V = Vinetișu Tuffs
M = Mlăcile Tuffs
B = Bătrîni Tuffs
13. Muncelu Cărămănesc Gypsum
14. Doftana Molasse
7 km = polynspastic distance between columns

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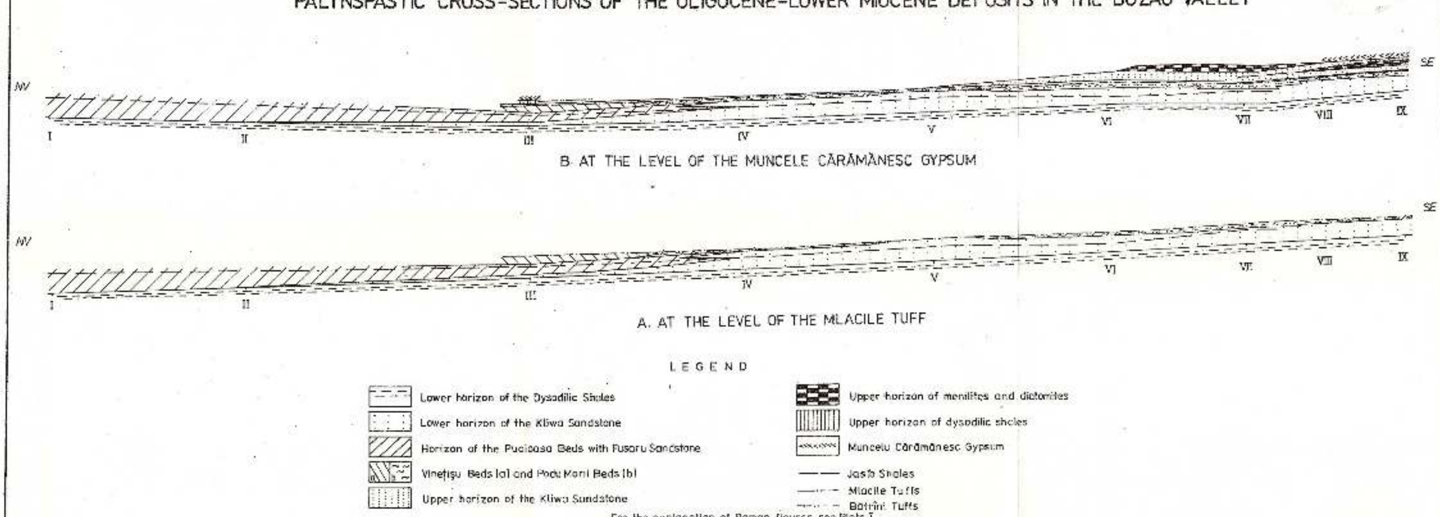
PALYNSPASTIC CROSS-SECTIONS OF THE OLIGOCENE–LOWER MIOCENE DEPOSITS IN THE BUZĂU VALLEY



For the explanation of Roman figures see Plate I

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Rom. J. Stratigraphy, 1993, 75, p. 91–96

CONTRIBUŢII LA STABILIREA LIMITEI OLIGOCEN-MIOCEN ÎN PÎNZA DE TARCĂU DIN BAZINUL VĂII BUZĂULUI, PE BAZA ASOCIAŢIILOR DE NANNOPLANCTON CALCAROS

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Key words: Oligocene. Miocene. Stratigraphic boundary. Flysch. Nannofossils. Plankton. East Carpathians – External Flysch Zone – Buzău Mountains.

Abstract: *Setting out of the Oligocene-Miocene Boundary in the Tarcău Nappe of the Buzău Valley Basin, Based on Calcareous Nannoplankton.* The purpose of the present paper is to establish the Oligocene-Miocene boundary, based on the study of calcareous nannoplankton, in an area of the outer flysch of the Eastern Carpathians, i.e. the Tarcău Nappe, in the Buzău Valley Basin. Based on the studied nannoflora, from the samples collected from three sections (the section on the Vineţişu Valley, the section of the Buzău Valley, upstream Nehoiu and the section at Gîrla Fişici), the Oligocen-Miocene Boundary has been traced at the level of the Vineţişu Tuff.

Introducere

Această lucrare are ca obiect precizarea limitei Oligocen-Miocen, respectiv Chattian-Aquitanian, pe baza studiului asociaţiilor de nannoplancton, într-un sector al flişului extern al Carpaţilor Orientali şi anume în pînza de Tarcău, din bazinul văii Buzăului.

Rezultatele obţinute de diverşi autori, pe baza studiului nannoplanctonului calcaros, în pînza de Tarcău, au dat rezultate diferite cu privire la această limită.

Astfel, Martini şi Lebenzon (1971), Lebenzon (1973), au fost primii care au studiat asociaţiile de nannoplancton din flişul extern al Carpaţilor Orientali. Ei au atras atenţia asupra faptului că limita Oligocen-Miocen este plasată în stratele de Vineţişu.

Asociaţiile de nannoplancton descrise de aceşti autori, din probele colectate de pe pîriul Răchitiş (bazinul Văii Tarcăului), din stratele de Vineţişu, au indicat ca vîrstă, pentru aceste depozite, Miocenul inferior (Aquitanian-Burdigalian).

Gheţa şi Dicea în Ştefănescu et al. (1979) studiind profilele de pe valea Lupei, valea Bizdidel, valea Vineţişu şi Gîrla Fişici, descriu asociaţii de nannoplancton, care plasează limita Oligocen-Miocen mai jos decît limita litologică dintre stratele de Vineţişu şi stratele de Pucioasa cu gresii de Fusaru, respectiv stratele de Podu Morii şi orizontul inferior al gresiei de Kliwa. Deci, orizontul stratelor de Pucioasa cu gresii de Fusaru şi orizontul inferior al gresiei de Kliwa sînt considerate Oligocene, partea superioară a acestor depozite fiind atribuită Miocenului inferior. Stratele de Vineţişu, stratele de Podu Morii, ca şi orizonturile suprajacente (orizontul superior al gresiei de Kliwa, orizontul superior al şisturilor disodilice) sînt considerate, de către aceşti autori, de vîrstă Aquitanian-Burdigalian (zonele de nannoplancton NN_2–NN_3).

Pentru stabilirea poziţiei limitei Oligocen-Miocen, în bazinul văii Buzăului, au fost studiate şi analizate, din punct de vedere al nannoplanctonului calcaros, trei profile: profilul de pe valea Vineţişu (fig. 1), cu depozite oligocene în faciesul stratelor de Pucioasa cu gresii de Fusaru, profilul de pe valea Buzăului, aval de localitatea Nehoiu (fig. 2) şi profilul de pe Gîrla Fişici (fig. 3), ultimele două profile cu depozite oligocene în faciesul bituminos al gresiei de Kliwa.

Conţinutul de nannoplancton

Colectarea şi analizarea probelor, pe baza asociaţiei de nannoplancton calcaros, a inceput cu orizontul inferior al gresiei de Kliwa.

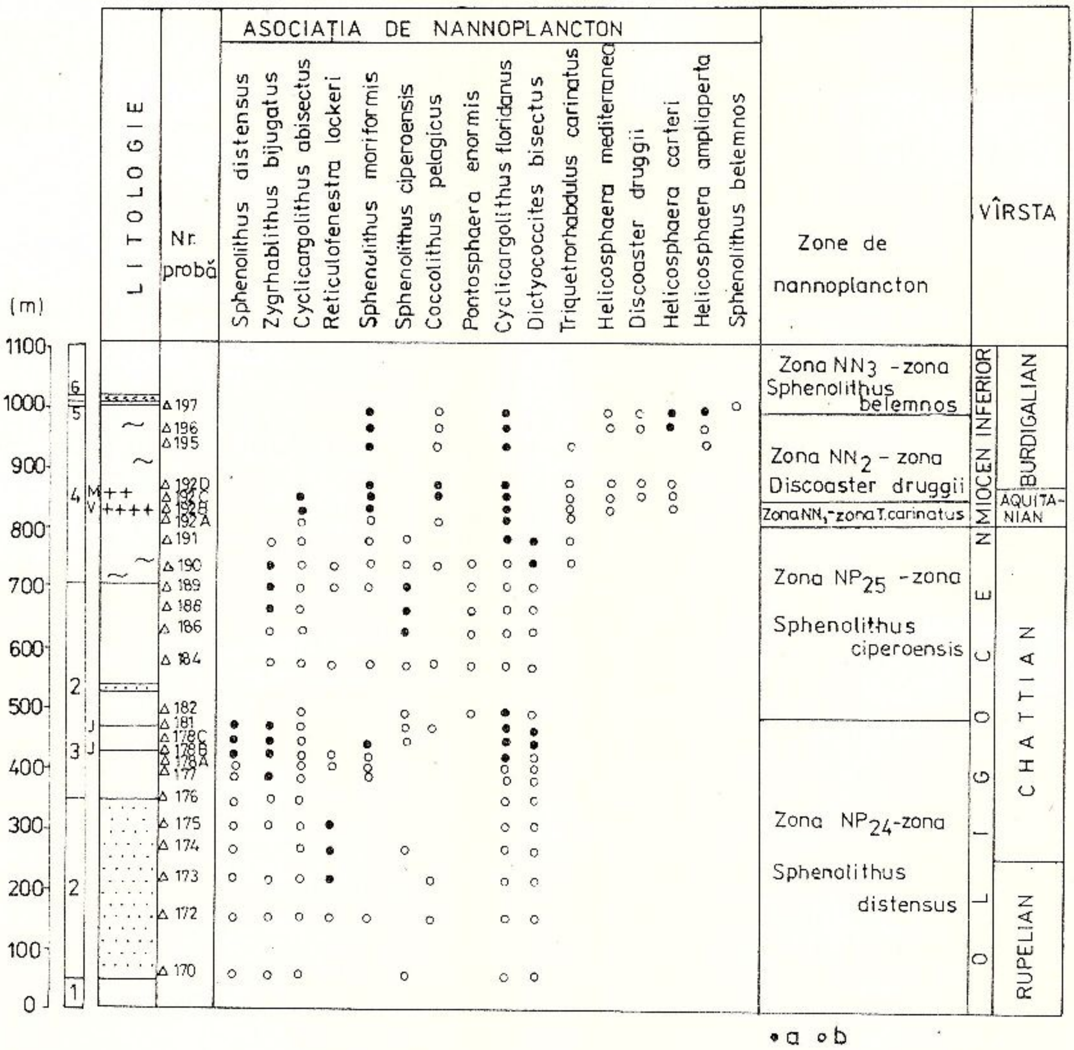

Fig. 1 – Formațiunile oligocen-miocen inferioare din profilul văii Vinețișu (după Ștefănescu et al., 1989) și asociația de nannoplancton corespunzătoare. 1, orizontul inferior al șisturilor disodilice; 2, orizontul inferior al gresiei de Kliwa; 3, stratele de Pucioasa cu gresii de Fusaru; 4, stratele de Vinețișu; 5, orizontul superior al șisturilor disodilice; 6, gipsul de Muncelu Cărămănesc; J, șisturi de Jaslo; V, tufuri de Vinețișu; M, tufuri de Mlăcile; a, frecvent; b, rar.

Pe profilele văilor Vinețișu și Buzăului, primele probe au fost prelevate din partea bazală a orizontului inferior al gresiei de Kliwa, chiar deasupra orizontului inferior al șisturilor disodilice, în timp ce, pe profilul Gîrla Fișici, probarea a început cu partea superioară a orizontului inferior al gresiei de Kliwa.

Primele probe recoltate de pe valea Vinețișu și valea Buzăului, au furnizat o asociație de nannoplancton caracteristică zonei NP_{24} – zona Sphenolithus distensus Bramlette et Wilcoxon (1967), zonă în care se plasează limita Rupelian-Chattian.

Asociația de nannoplancton determinată este formată din următorii taxoni:

Sphenolithus distensus (MARTINI) BRAMLETTE et WILCOXON

Zygrhablithus bijugatus (DEFLANDRE) DEFLANDRE

Cyclicargolithus floridanus (ROTH et HAY) BUKRY

Cyclicargolithus abisectus (MÜLLER) BUKRY

Reticulofenestra lockeri MÜLLER

Sphenolithus moriformis (BRONNIMANN et STRADNER) BRAMLETTE et WILCOXON

Sphenolithus ciperoensis BRAMLETTE et WILCOXON

Coccolithus pelagicus (WALLICH) SCHILLER

Dictyococcites bisectus (HAY, MOHLER et WADE) BUKRY et PERCIVAL

Prezența lui *Sphenolithus distensus* în asociație, distribuția speciei fiind cuprinsă în zonele de nannoplancton NP_{23}–NP_{24}, prezența speciilor *Sphenolithus*

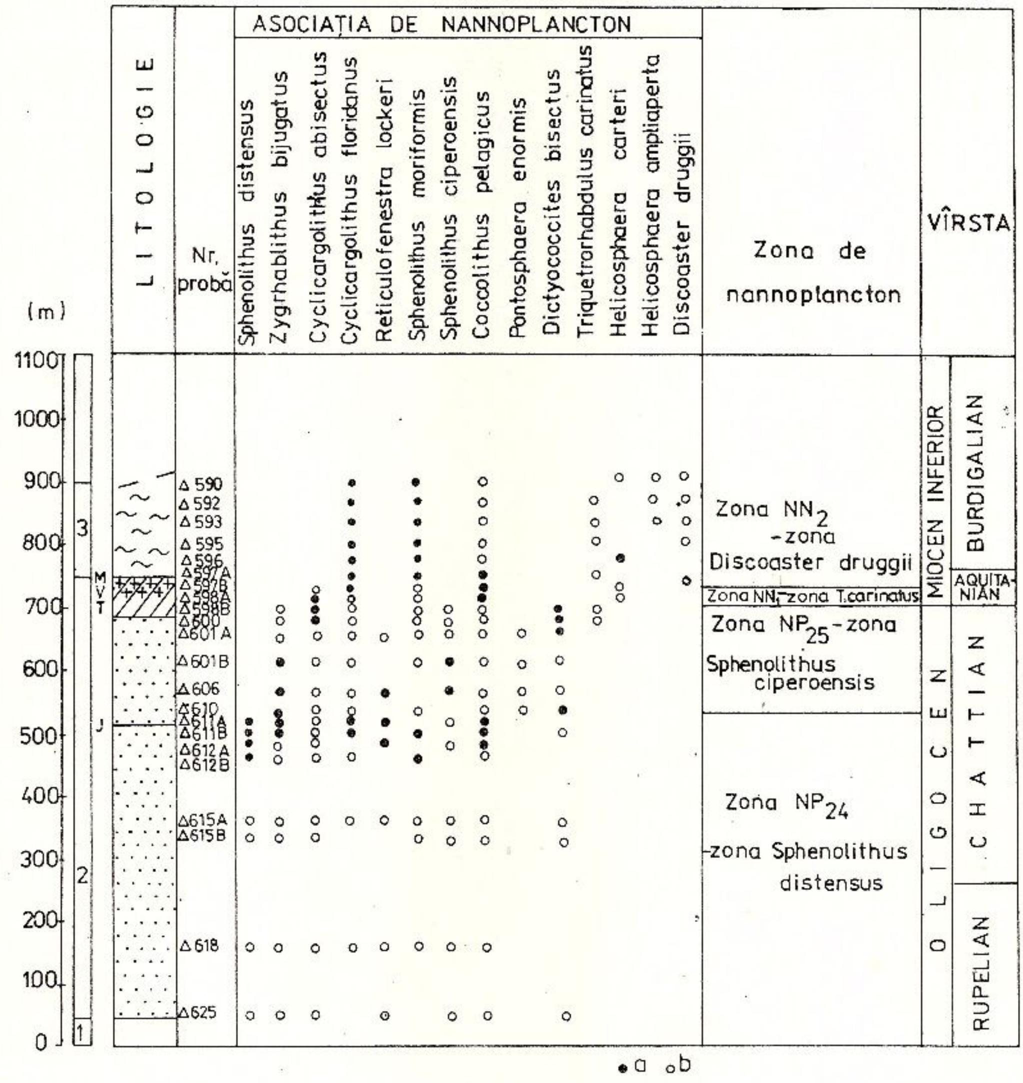

Fig. 2 – Formaţiunile oligocen-miocen inferioare din profilul văii Buzăului, aval de Nehoiu (după Ştefănescu et al., 1989) şi asociaţia de nannoplancton corespunzătoare. 1, orizontul inferior al şisturilor disodilice; 2, orizontul inferior al gresiei de Kliwa; 3, stratele de Podu Morii; J, şisturi de Jaslo; V, tufuri de Vineţişu; M, tufuri de Mlăcile; T, strate de Topile; a, frecvent; b, rar.

ciperoensis (cu distribuţie în zonele NP_{24} şi NP_{25}) şi *Cyclicargolithus abisectus* (cu distribuţie între NP_{24} şi NN_1 bazal) sînt argumente în favoarea plasării acestei asociaţii de nannoplancton în zona NP_{24}.

Probele colectate în continuare pe profilele văilor Vineţişu şi Buzău, din orizontul inferior al gresiei de Kliwa, au indicat faptul că ne aflăm tot la nivelul zonei NP_{24}, speciile de nannoplancton, menţionate anterior, menţinîndu-se pe tot acest interval.

Şisturile de Jaslo reprezintă, în general, un nivel caracteristic atît în orizontul inferior al gresiei de Kliwa (valea Buzăului), cît şi în stratele de Pucioasa cu gresii de Fusaru (valea Vineţişu).

Asociaţia de nannoplancton determinată din probele colectate din şisturile calcaroase de Jaslo este alcătuită din aceleaşi specii, care situează în zona NP_{24}.

Pe profilul Vineţişu, probe recoltate dintr-un pachet de pelite cafenii, la 70 m deasupra şisturilor de Jaslo, a furnizat o asociaţie de nannoplancton cu aceiaşi taxoni ca cei descrişi anterior, din care lipseşte însă *Sphenolithus distensus*. Faptul menţionat nu este izolat, acelaşi lucru constatîndu-se şi pe profilul Văii Buzăului, în probe recoltate la circa 30 m, deasupra şisturilor de Jaslo. În aceste probe se constată că *Sphenolithus ciperoensis* cu întreaga asociaţie de nannoplancton descrisă anterior se menţine; specia *Dictyococcites bisectus* are o frecvenţă foarte mare. În plus pe profilul Vineţişu apare specia *Pontosphaera*

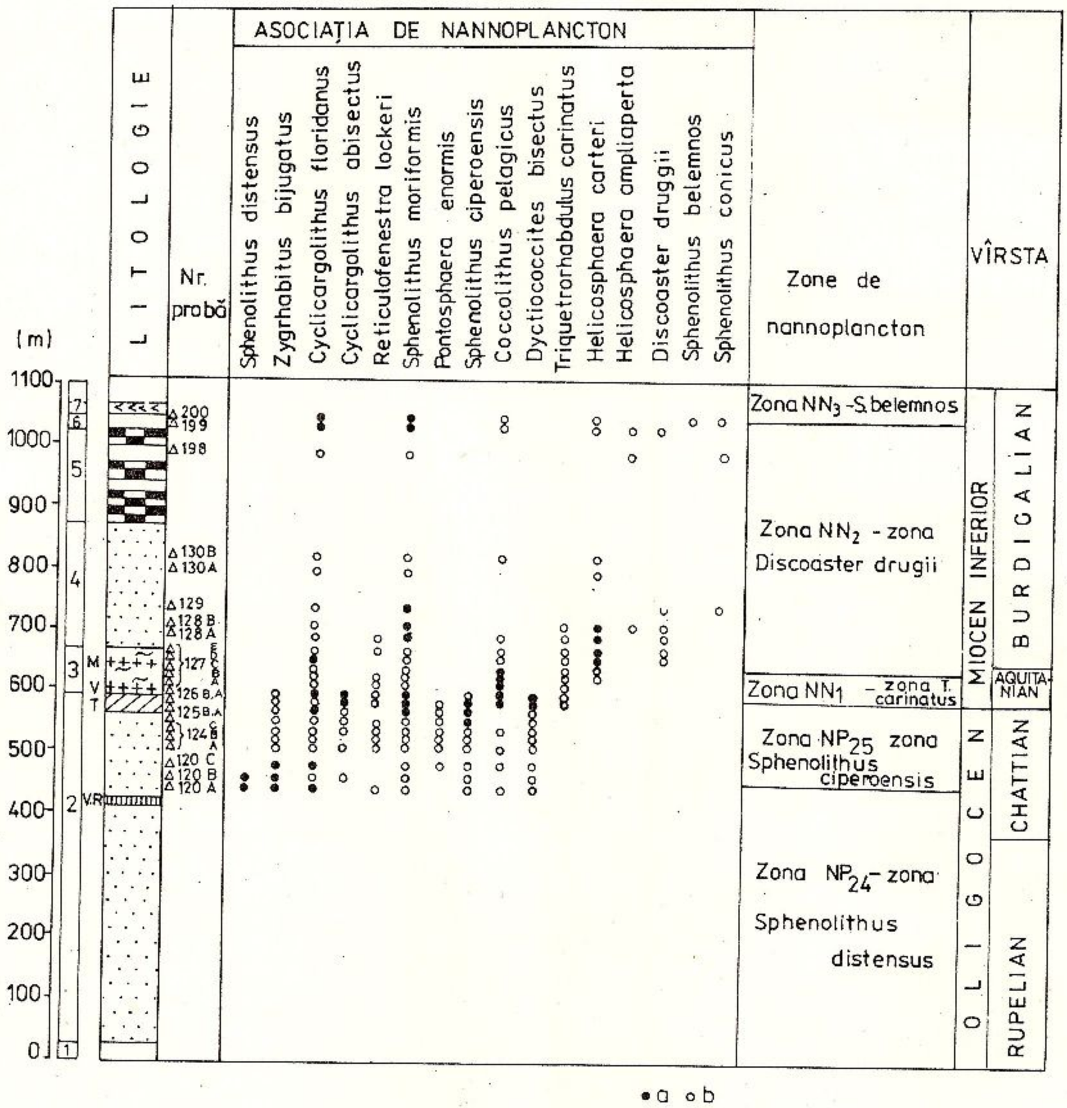

Fig. 3 – Coloana litologică a formațiunilor oligocen-miocen inferioare din profilul Gîrlei Fișici (după Ștefănescu et al., 1989) și asociația de nannoplancton corespunzătoare. 1, orizontul inferior al șisturilor disodilice; 2, orizontul inferior al gresiei de Kliwa; 3, stratele de Podu Morii; 4, orizontul superior al gresiei de Kliwa; 5, orizontul superior al menilitelor; 6, orizontul supramenilitelor; 7, gipsul de Muncelu Cărămănesc; V.R., marne de Valea Rea; V, tufuri de Vinețișu; M, tufuri de Mlăcile; T, strate de Topile; a, frecvent; b, rar.

enormis LOCKER, specifică zonei NP_{25}. Asociația de nannoplancton se poate încadra în zona NP_{25} – zona Sphenolithus ciperoensis Bramlette et Wilcoxon (1967) emend. Martini (1976), zonă corespunzătoare Chattianului.

În profilul Gîrla Fișici, orizontul reper al șisturilor de Jaslo nu apare. Aproximativ la același nivel stratigrafic, apar marnele de Valea Rea, similare cu cele de pe profilul Văii Rele. Probele colectate din aceste marne, pe Valea Rea, au indicat o asociație de nannoplancton cu *Sphenolithus distensus*, pe care o putem atribui zonei NP_{24}.

Pe profilul Gîrlei Fișici, probarea a început la partea superioară a orizontului inferior al gresiei de Kliwa, la circa 20 m deasupra marnelor de Valea Rea. Probele analizate au furnizat aceeași asociație cu *Sphenolithus distensus*, caracteristică zonei NP_{24}. La circa 50 m deasupra, probele conțin o asociație de nannoplancton cu: *Sphenolithus ciperoensis, Pontosphaera enormis, Cyclicargolithus abisectus, Sphenolithus moriformis, Dictyococcites bisectus, Cyclicargolithus floridanus, Zygrhablithus bijugatus, Coccolithus pelagicus*, asociație pe care o atribuim zonei NP_{25}.

Din probele colectate pe profilul văii Vinețișu, aproximativ la 100 m deasupra limitei litologice dintre stratele de Pucioasa cu gresii de Fusaru și stratele de Vinețișu a fost determinată specia *Triquetrorhabdu-*

lus carinatus MARTINI, alături de speciile determinate anterior.

Martini (1971) consideră că limita Oligocen-Miocen, în termenii nannoplanctonului calcaros, este legată de apariţia acestei specii. S-a constatat totuşi (Müller, 1981) că *T. carinatus* poate să apară frecvent în Oligocenul terminal. De aceea, considerăm că ne aflăm încă în zona NP_{25} (Chattian).

Probele recoltate de la nivelul pelitelor subiacente şi suprajacente, cît şi chiar din tuful de Vineţişu (ne referim la primul nivel de tufuri din stratele de Vineţişu, respectiv stratele de Podu Morii), pe cele trei profile studiate, conţin următoarea asociaţie de nannoplancton:

Cyclicargolithus floridanus (ROTH et HAY) BUKRY

Sphenolithus moriformis (BRÖNNIMAN et STRATNER) BRAMLETTE et WILCOXON

Cyclicargolithus abisectus (MÜLLER) BUKRY

Triquetrorhabdulus carinatus MARTINI

Coccolithus pelagicus (WALLICH) SCHILLER.

Se constată absenţa speciilor: *Sphenolithus ciperoensis, Dictyococcites bisectus* et *Zygrhablithus bijugatus*, specii prezente pînă acum. Presupunem că ne aflăm în zona de nannoplancton NN_1 – zona Triquetrorhabdulus carinatus (Bramlette et Wilcoxon, 1967, emend. Martini et Worsley, 1970), zonă ce corespunde Miocenului inferior (Aquitanian).

Limita Oligocen-Miocen este trasată, în termenii nannofosilelor calcaroase la limita zonelor NP_{25}–NN_1.

Müller (1981) propune ca limita Oligocen-Miocen să se traseze la nivelul extincţiei următoarelor specii: *Zygrhablithus bijugatus, Helicosphaera recta, Sphenolithus ciperoensis, Dictyococcites dictyodus, Ericsonia fenestrata.*

Biolzi et al. (1983) referindu-se la aceeaşi limită Oligocen-Miocen, utilizează aceiaşi taxoni, folosind în loc de *Dictyococcites dictyodus, Dictyococcites bisectus*; totodată se arată faptul că diversitatea speciilor de la limita Oligocen-Miocen este în general scăzută.

Se constată deci, în cazul trasării limitei dintre zonele NP_{25}–NN_1, utilizarea unor criterii de absenţă (Noël, 1972), deşi aceste criterii pot fi legate, ca şi la alte grupuri de fosile, de o simplă modificare de facies.

Într-adevăr, această limită nu este marcată de o schimbare distinctă în ansamblul de nannoplancton. De aceea, zona NN_1, se poate caracteriza mai mult ca o zonă de tranziţie, în care are loc descreşterea numărului de specii paleogene, fără prezenţa formelor tipic neogene. Aceste specii au prima ocurenţă în interiorul zonei NN_1.

În probele studiate se remarcă extincţia speciilor: *Sphenolithus ciperoensis, Dictyococcites bisectus* şi *Zygrhablithus bijugatus*, precum şi dezvoltarea în erupţie a speciei *Cyclicargolithus abisectus*.

Pe baza acestor date putem trasa limitele Oligocen-Miocen şi, respectiv, Chattian-Aquitanian la nivelul tufului de Vineţişu. Este de menţionat că în pelitele de sub tuful de Vineţişu de la Gîrla Fişici s-au găsit, alături de speciile mai sus menţionate şi alte specii, remaniate, din Cretacic, Eocen, Oligocen: *Watznaueria barnesae* BLACK, *Micula concava stauropora* GARNET, *Lucianorhabdus cayeuxii* DEFLANDRE, *Discoaster saipanensis* BRAMLETTE et RIEDEL, *Reticulofenestra umbilica* LEVIN, *Chiasmolithus oamaruensis* DEFLANDRE.

Remanierea acestor specii corespunde cu acumularea unora dintre nivelele de brecii sedimentare din faciesul de Slon (Popescu, 1958) şi poate fi deci pusă în seama masivelor alunecări submarine ce au dus la formarea acestor brecii.

Pe cele trei profile studiate, în pelitele suprajacente tufului de Vineţişu apare specia *Helicosphaera carteri* (WALLICH) KAMPTNER, iar pe profilul Vineţişu apare, în plus, specia *Helicosphaera mediterranea* MÜLLER, ambele specii debutînd în Miocenul inferior (Aquitanian).

Asociaţia de nannoplancton descrisă anterior se menţine, atît în stratele de Podu Morii, cît şi în stratele de Vineţişu, pînă la nivelul celui de-al doilea tuf (tuful de Mlăcile).

În probele colectate de la nivelul celui de-al doilea tuf, ca şi din pelitele suprajacente, apar faţă de speciile determinate anterior: *Discoaster druggii* BRAMLETTE et WILCOXON şi *Helicosphaera ampliaperta* BRAMLETTE et WILCOXON, specii a căror distribuţie începe din zona NN_2. Alături de aceste specii, asociaţia de nannoplancton cuprinde: *Sphenolithus moriformis, Coccolithus pelagicus, Triquetrorhabdulus carinatus, Helicosphaera carteri* şi *H. mediterranea.* Putem afirma că, de la nivelul celui de-al doilea tuf ne situăm în zona de nannoplancton NN_2 – zona Discoaster druggii (Martini et Worsley, 1970).

Probele recoltate din baza gipsului pe profilul de la Muncelu Cărămănesc au furnizat o asociaţie cu: *Sphenolithus belemnos, Helicosphaera ampliaperta, H. carteri, H. mediterranea, Discoaster druggii, Coccolithus pelagicus şi Cyclicargolithus floridanus.*

Apreciem că ne situăm, la acest nivel, în zona de nannoplancton NN_3.

Concluzii

Din probele studiate şi analizate pe cele trei profile, din bazinul văii Buzăului, se pot trage următoarele concluzii:

– Stratele de Pucioasa cu gresii de Fusaru, respectiv orizontul inferior al gresiei de Kliwa sînt de vîrstă rupelian-chattiană.

– Limita Rupelian-Chattian se plasează deasupra nivelului şisturilor de Jaslo.

– Nivelul tufurilor de Vineţişu şi de Mlăcile sînt repere corelabile pe toate cele trei profile. Limita Oligocen-Miocen se plasează la nivelul primului tuf (tuful de Vineţişu), din stratele de Vineţişu, respectiv spre baza stratelor de Podu Morii (pe profilul Gîrla Fişici). Pe profilul văii Buzăului, aval de Nehoiu, tuful de Vineţişu, la nivelul căruia se plasează limita Oligocen-Miocen, se situează peste partea superioară a orizontului inferior al gresiei de Kliwa, în stratele de Topilele. Limita Oligocen-Miocen este pe acest profil sub limita litologică a stratelor de Podu Morii.

– De la nivelul tufului de Vineţişu putem aprecia existenţa zonei de nannoplancton NN_1, zonă cu o extindere limitată, pînă la pelitele subjacente celui de-al doilea tuf.

– De la nivelul tufului de Mlăcile în sus există dovezi pentru separarea zonei de nannoplancton NN_2 (Aquitanian-Burdigalian). Această zonă se extinde pe profilul Vineţişu pînă la partea superioară a stratelor de Vineţişu. Pe celelalte două profile (valea Buzăului şi Gîrla Fişici) cuprinde întreg intervalul pînă sub gipsul inferior.

– Zona NN_3 începe odată cu primele pelite de sub gipsul inferior.

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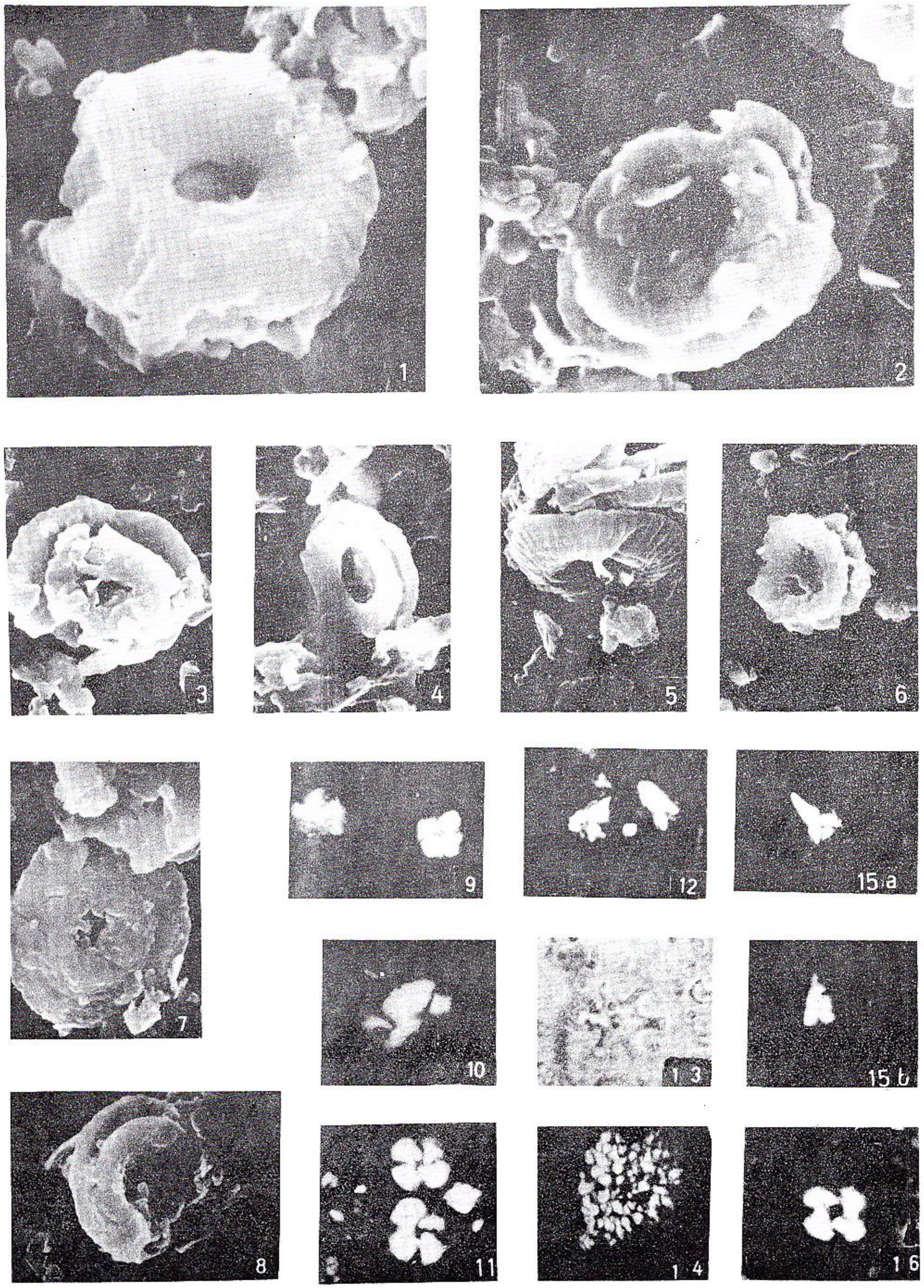
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Plate

Fig. 1 – *Cyclicargolithus abisectus* (MÜLLER) BUKRY, x 9500, SEM.
Fig. 2 – *Cyclicargolithus floridanus* (ROTH et HAY) BUKRY, x 9500, SEM.
Fig. 3 – *Dictyococcites bisectus* (HAY, MOHLER and WADE) BUKRY and PERCIVAL, x 5000 SEM.
Fig. 4 – *Cyclicargolithus floridanus* (ROTH and HAY) BUKRY, SEM x 5500.
Fig. 5 – *Coccolithus pelagicus* (WALLICH) SCHILLER, SEM x 6000.
Fig. 6 – *Reticulofenestra lockeri* MÜLLER, SEM x 5500.
Fig. 7 – *Dictyococcites bisectus* (HAY, MOHLER and WADE) BUKRY and PERCIVAL, SEM x 6500.
Fig. 8 – *Reticulofenestra lockeri* MÜLLER, SEM x 5500.
Fig. 9 – *Sphenolithus moriformis* (BRÖNNIMANN and STRADNER) BRAMLETTE and WILCOXON, LM cross-polarized light, x 2500.
Fig. 10 – *Helicosphaera carteri* (WALLICH) KAMPTNER, LM cross-polarized light, x 2500.
Fig. 11 – *Cyclicargolithus floridanus* (ROTH and HAY) BUKRY, LM cross-polarized light, x 2500.
Fig. 12 – *Sphenolithus conicus* DEFLANDRE; *Sphenolithus belemnos* BRAMLETTE and WILCOXON, LM cross-polarized light, x 2500.
Fig. 13 – *Discoaster* cf. *druggii* BRAMLETTE and WILCOXON, LM, phase-contrast light, x 2500.
Fig. 14 – *Thoracosphaera* sp., LM cross-polarized light, x 2500.
Fig. 15 – *Sphenolithus ciperoensis* MARTINI, LM cross-polarized light, x 2500, a-45°, b-0°.
Fig. 16 – *Cyclicargolithus floridanus* (ROTH and HAY) BUKRY, LM cross-polarized light, x 2500.

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PODU MORII BEDS IN THE TAZLĂUL SĂRAT BASIN

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Key words: Flysch. Oligocene-Miocene. Podu Morii Beds. Upper and Lower Kliwa Sandstone. Lithostratigraphy. Nannoplankton assemblage. Outer Flysch Zone. East Carpathians. Tazlău Mountains.

Abstract: The original contribution of this paper concerns the first separation of the Podu Morii Beds and of the Lower and Upper Kliwa Sandstone below and above them, respectively. An assemblage of calcareous nannoplankton belonging to the zone NN_2 - Discoaster druggi (Upper Aquitanian-Lower Burdigalian) within the middle part of the Podu Morii Beds indicates the Uppermost Oligocene-Lower Miocene age of these deposits.

The paper presents new litho- and biostratigraphic details of the bituminous facies with Kliwa Sandstone of the Zemeş "lambeau de rabotage", resulted from the researches carried out in the Tazlăul Sărat Basin in 1988.

The presence of numerous hydrocarbon reservoirs, that have been exploited since the second half of the past century, has led to intense geological researches. From the previous geological papers concerning the stratigraphy and the structure of the outer flysch zone of the Tazlăul Sărat Basin those by Băncilă (1940) and Hristescu (1953, unpubl. report – in Băncilă, 1958), followed by those by Turculeţ, Filimon (1956, unpubl report), Mirăuţă (1969), Matei et al. (1969, unpubl. report) and Polonic, Polonic (1969, unpubl. report) are worth mentioning. Detail geological mapping in the Zemeş region (Figure) have made it possible for us to identify in the Tazlăul Sărat talweg, upstream the bridge north of the confluence with the Zemeş Creek, a sequence of gritty-clayey flysch, about 145–150 m thick, intercalated between metric beds of Kliwa Sandstone. The deposits concerned are made up of decimetric beds of strongly convolute, slightly micaceous grey sandstones, alternating with marly-siliceous grey-greenish pelites, 0.1–0.4 m thick. Current marks and bioglyphs frequently occur in the sandstone base. Under the microscope, the sandstones – fine- or medium-grained – appear made up prevailingly of quartz grains, muscovite and feldspar being totally subordinate. The sandstone cement is made up of microgranular calcite.

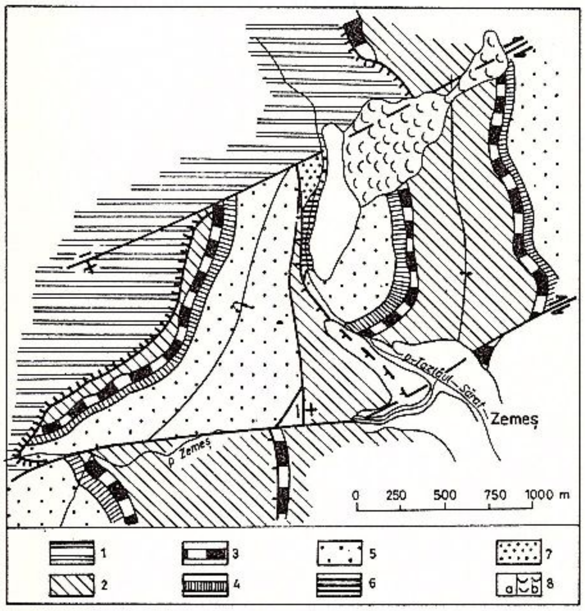

Fig. – Geological sketch-map of the Zemeş area. Tarcău Nappe: 1, Undivided Cretaceous-Paleogene flysch formations of the Tazlău Digitation. Zemeş "lambeau de rabotage"; 2, Undivided Eocene flysch formations; 3, Lower Menilites and Bituminous Brown Marls, with Slaty Shales and Fierăstrău Sandstone in the base (Oligocene); 4, Lower Dysodilic Shales (Oligocene); 5, Lower Kliwa Sandstone (Oligocene); 6, Podu Morii Beds (Oligocene–Lower Miocene); 7, Upper Kliwa Sandstone (Lower Miocene); 8, Quaternary deposits: a, terraces and alluvial plain; b, landslides.

For these deposits we use the name of Podu Morii Beds, as they have a lithology and a stratigraphic position similar with those of the homonymous formation in the bending zone of the East Carpathians. We mention that this type of deposits was previously described (Renz – in Drăghici, 1963; unpubl. report) in the Tîrgu Ocna area, in a series of "lambeaux de rabotage" with a structural position similar with that of the Zemeş one.

We should mention that the passage from the Podu Morii Beds to the deposits in the foot-wall and in the hanging-wall (Lower Kliwa Sandstone, Upper Kliwa Sandstone, respectively) is gradual, on a thickness of about 20-30 m, convolute sandstones and associated pelites occurring as interbeds at several levels between the thick beds of Kliwa Sandstone.

As the outcrops are discontinuous, we have not met so far, in the Podu Morii Beds at Zemeş, tuff interbeds, characteristic of these deposits in other regions.

From the pelitic interbeds of the middle part of the Podu Morii Beds at Zemeş, a micropaleontological sample contains the following calcareous nannoplankton assemblage: *Cyclicargolithus floridanus* (ROTH & HAY) BUKRY, *Sphenolithus moriformis* (BRÖNNIMANN & STRADNER) BRAMLETTE, *Thoracosphaera* sp., *Coccolithus pelagicus* (WALLICH) SCHILLER, *Triquetrorhabdulus carinatus* MARTINI, *Coronocyclus nitescens* (KAMPT.) BRAMLETTE & WILCOXON, *Helicosphaera mediterranea* MÜLLER, *H. ampliaperta* BRAM. & WILCOXON, *Reticulofenestra pseudoumbilica* (GARTNER) GARTNER, *Cyclococcolithus formosus* KAMPT., *Coccolithus abisectus* MÜLLER *Reticulofenestra bisecta* (HAY, MOHLER & WADE) ROTH, *Chiasmolithus altus* BUKRY & PERCIVAL, *Cyclococcolithus luminis* SULLIVAN, *Sphenolithus ciperoensis* BRAML. & WILCOX, *Transversopontis obliquipons* (DEFL.) HAY, MOHLER & WADE, *Reticulofenestra insignita* ROTH & HAY, *Helicosphaera carteri* (WALLICH) KAMPTNER.

From the above mentioned assemblage determined by Melinte, a series of taxa extend in range from the Eocene to the Oligocene up to the Miocene, being not specific to a certain nannoplankton zone, and other species are clearly reworked from older deposits, of Eocene or Oligocene age. In exchange, a series of taxa of this assemblage give important indications on the age of the deposits under study.

Thus, the species *Triquetrorhabdulus carinatus* MARTINI is characteristic of the interval NP_{25} Top Oligocene) – NN_2 (Upper Aquitanian-Lower Burdigalian), *Helicosphaera carteri* KAMPTNER ranges from the NP_{25} zone up of the Pleistocene, and *Helicosphaera mediterranea* MÜLLER appears as far as zone NN_1 up to the upper boundary of zone NN_4. The species *Helicosphaera ampliaperta* BRAMLETTE & WILCOXON occurs as far as zone NN_2, having the extinction at the upper part of zone NN_4 and *Reticulofenestra pseudoumbilica* (GARTNER) extends in the whole Miocene, sporadic occurrences being noticed as far as zone NN_1.

Taking into account the data presented above, the nannoplankton assemblage identified can be referred to zone NN_2-*Discoaster druggi* MARTINI and WORSLEY that belongs to the interval Upper Aquitanian-Lower Burdigalian. As the sample in which this association has been found is situated at the middle part of the Podu Morii Beds at Zemeş, we are of the opinion that these deposits can be assigned to the Upper Oligocene-Lower Miocene. The Oligocene-Miocene boundary can be placed within the lower part of the Podu Morii Beds, or, as Ştefănescu et al. (1979) considered, in the lower Kliwa Series. Correspondigly, the Upper Kliwas Sandstone cannot be older than Lower Burdigalian.

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CONTENTS IN CALCAREOUS NANNOPLANKTON OF THE MARINE LOWER AND MIDDLE MIOCENE BEDS, EAST AND NORTH-EAST OF ZALĂU

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Key words: Miocene. Nannofossils. Plankton. Apuseni Mountains – Paleogene zone of NW Transylvania.

Abstract: The paper is the result of the study of the nannoplankton contents of the Chechiş Beds and the Dej Tuff, developed east and north-east of Zalău. Based on the assemblage of calcareous nannofossils, the Chechiş Beds were assigned to the Upper-Aquitanian-Lower Burdigalian (Discoaster druggii Zone – NN_2, Martini, 1971) and the Dej Tuff to the Langhian (Sphenolithus heteromorphus Zone – NN_5, 1971). It is also admitted, at least for the Transylvanian Basin, the subdividing of the Discoaster druggii Zone – NN_2 in two subzones: Sphenolithus dissimilis subzone – NN_{2a} and the Helicosphaera kamptneri subzone – NN_{2b}.

The Miocene succession, developed east and north-east of Zalău, is represented by the Chechiş Beds and the Dej Tuff.

The Chechiş Beds, transgressively lying over various terms of the Paleogene or even directly over the Precambrian metamorphic rocks of the Mezeş Mts, consist of three lithostratigraphic sequences, whose thicknesses vary from one section to another, because of the frequent lithofacial changes along strike.

The lower sequence, well developed on the Rogoaze Stream, where it reaches a thickness of about 70 m, is made up of alternations of coarse slightly cemented sandstones and mediogranular up to coarse sands, with interbeds of microconglomerates or even mediogranular conglomerates (Fig. 1). In the base of this sequence, big blocks of eruptive and metamorphic rocks border the transgression boundary.

The middle sequence, well developed north of the Ortelec Stream, where it is about 100 m thick, is prevailingly pelitic, containing metric grey-greenish clays and centimetric conglomerates, hard sandstones or red siltstones (Fig. 1). Sometimes, at the lower part of this sequence, the clayey levels contain plant imprints and bivalve remains (Frumuşelu Stream, Răpaos Stream).

The upper conglomeratic sequence (Fig. 1) has seemingly small thicknesses on the Rogoazele Stream (about 15 m) and becomes thicker and thicker eastwards, because of the retirement northwards of the Langhian transgression boundary.

The sample collected from these deposits contain a rich nannoplankton assemblage i.e. *Coccolithus copelagicus* (BRAMLETTE & RIEDEL) BRAMLETTE & SULIVAN, *C. miopelagicus* BUKRY, *C. pelagicus* (WALLICH) SCHILLER, *Coronocyclus nitescens* (KRAMPTNER) BRAMLETTE & WILCOXON, *Helicosphaera ampliaperta* BRAMLETTE & WILCOXON, *H. intermedia* (MARTINI) HAY & MOHLER, *H. kamptneri* HAY & MOHLER, *Pontosphaera multipora* (KAMPTNER) ROTH, *Cyclicargolithus abisectus* (MÜLLER) WISE, *C. floridanus* (ROTH & HAY) BUKRY, *Reticulofenestra dictyoda* (DEFLANDRE) STRADNER, *R. lockeri* MÜLLER, *Reticulofenestrae, Rhabdosphaera* sp., *Sphenolithus dissimilis* BUKRY & PERCIVAL, *S. moriformis* (BRÖNNIMANN & STRADNER) BRAMLETTE & WILCOXON (Pl.).

From the species identified *Helicosphaera ampliaperta, H. kamptneri* and *Sphenolithus dissimilis* are of particular interest. The existence of *Helicosphaera ampliaperta* unaccompanied by *Sphenolithus belemnos*, within the context of the assemblage presented, could lead to referring the deposits that contain them to the stratigraphic interval of the Helicosphaera ampliaperta Zone-NN_4 (Martini, 1971). In the last period of time it is well known that the first occurrences of *Helicosphaera ampliaperta* and *H. kamptneri* are found in the interval occupied by the upper part of the Discoaster druggii Zone-NN_2, while that of *Sphenolithus belemnos* a little bit later, before the extinction of *Tri-*

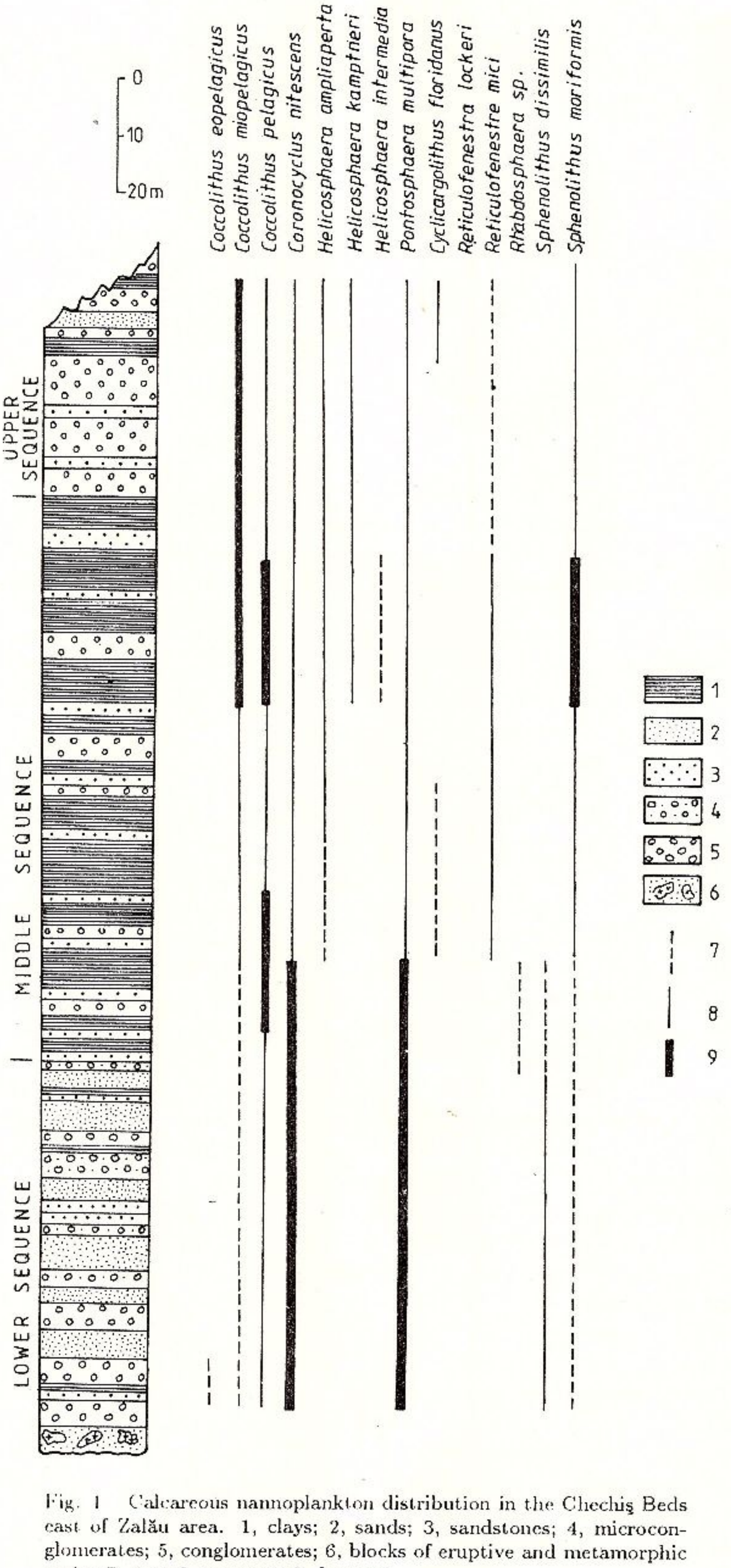

Fig. 1 Calcareous nannoplankton distribution in the Chechiş Beds east of Zalău area. 1, clays; 2, sands; 3, sandstones; 4, microconglomerates; 5, conglomerates; 6, blocks of eruptive and metamorphic rocks; 7, rare; 8, common; 9, frequent.

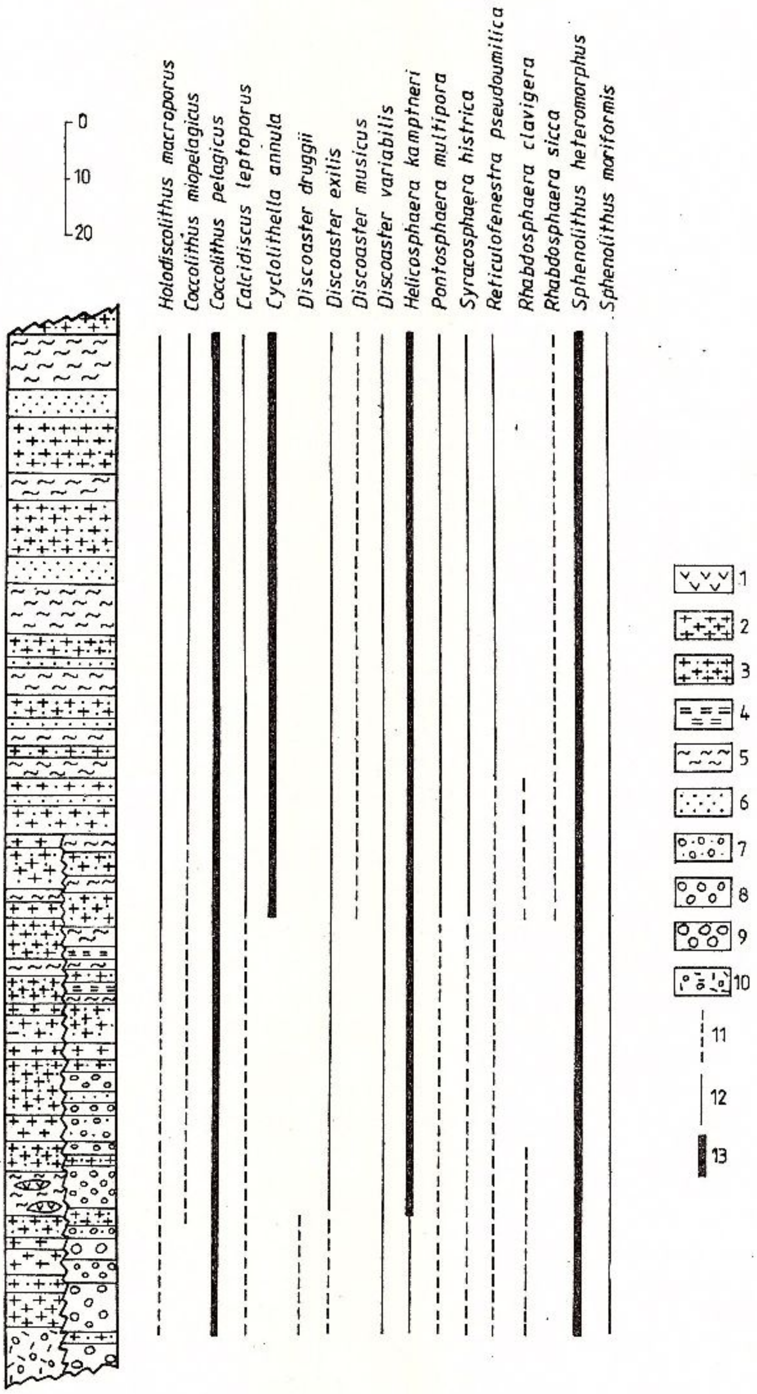

Fig. 2 - Calcareous nannoplankton distribution in the Dej Tuff east and north of Zalău area. 1, gypsum; 2, tuffs; 3, tuffites; 4, calcareous shales; 5, marls; 6, sandstones; 7, microconglomerates; 8, fine conglomerates; 9, conglomerates; 10, volcanic agglomerates; 11, rare; 12, common; 13, frequent.

quetrorhabdulus carinatus (that marks the boundary NN_2-NN_3) (Perch Nielsen in: Bolli et al., 1985). It is also known that *Sphenolithus dissimilis* becomes extinct towards the middle part of Zone NN_2.

From the distribution of the calcareous nannofossils (Fig. 1) it is noticed that the lower sequence and the base of the middle one are characterized by the existence of *Sphenolithus dissimilis*, while in the remaining part of the middle section and in the upper one there occur *Helicosphaera ampliaperta* and *H. kamptneri*.

From what has been shown, although index fossils are missing, we can assert that at least the assemblage of the middle and the upper sequences are equivalents of the upper part of the Discoaster druggii Zone – NN_2 (Martini, 1971). The nannoplankton contents of the lower sequence and of the base of the middle one are referred, in this case, to the lower part of the same zone. Consequently, the age of the Chechiş Beds, developed east and north-east of Zalău, is Lower Aquitanian-Lower Burdigalian (nannoplankton zone NN_2 – sensu Martini, 1971).

The sedimentary succession described was considered to be the equivalent of the Chechiş Beds (in the zone of Moigrad), taking into account the macrofaunal contents and the existence in the base of the formation of the "glauconitic level" characteristic of the Chechiş Beds (Rusu, 1977).

The nannoplankton association evidenced in the investigated area is almost identical with that of the typical Chechiş Marls, identified by Mészáros et al. (1976), with the exception of *Sphenolithus belemnos*.

In the Transylvanian Basin, considering the distribution of a few species, the Discoaster druggii Zone – NN_2 can be divided into two subzones:

– Sphenolithus dissimilis Subzone – NN_{2a}, corresponding to the interval between the first occurrence of *Discoaster druggii* and the last occurrence of *Sphenolithus dissimilis* (corresponding to the first occurrence of *Helicosphaera ampliaperta*);

– Helicosphaera kamptneri Subzone – NN_{2b}, delimited by the last occurrence of *Sphenolithus dissimilis* (or the first occurrence of *Helicosphaera ampliaperta*) and the first occurrence of *Sphenolithus belemnos* approximately corresponding to the last occurrence of *Triquetrorhabdulus carinatus*).

This subzoning is proved to be valid both in the southern part of the Transylvanian Basin, in the sedimentary deposits of "the conglomeratic breccious horizon", in the "marly horizon" and the "cinerite-bearing sequence" (described by Gheorghian, 1976; unpubl. report), developed along the Sebeşul de Sus, Sebeşul de Jos, Tălmăcel Valleys as well as in the outer part of the South Carpathians in the Muereasca Sandstones and the Gura Văii Beds on the Muereasca Valley (separated by Popescu et al., 1977).

The Dej Tuff, that crops out east north-east of Zalău, transgressively lies over the Lower Miocene and the Paleogene formations. The thickness of the deposits ranges between 40 m (at Bîrsa) and 250 m (between Mirşid and Firminiş), because of the Pannonian transgression. In the succession of the Dej Tuff, where it is better represented (Lazu Stream, Valea cu Rîpi Stream), two sequences can be identified, according to the prevailing lithologic character (Fig. 2).

The lower sequence, with thicknesses up to 150 m, laterally shows a great lithofacial variety. Thus, south of Mirşid, it is made up of decimetric alternances of tuffs and tuffites, with interbeds of volcanic agglomerates in the lower half and white tuffaceous marls with gypsum lenses in the upper half. From Mirşid eastwards, microconglomeratic interbeds begin to occur, and in the Barcău Hills and the Spinaţi Vineyards the sequence becomes prevailingly conglomeratic, with rare interbeds of tuffaceous marls and calcareous shales towards the upper part. Towards Bîrsa, where only the lower part of the sequence crops out in fact, the deposits become prevailingly tuffitic.

The upper sequence, that is not thicker than 100 m, is well developed on the Lazu Stream, where it is made up of decimetric alternances of tuffs and tuffites, with centimetric interbeds of white tuffaceous marls and hard grey-whitish sandstones, with hieroglyphs and calcite diaclases.

The contents in calcareous nannofossils are almost identical in the whole sequence of the Dej Tuff, being made up of *Holodiscolithus* cf. *H. macroporus* (DEFLANDRE) ROTH, *Coccolithus miopelagicus* BUKRY, *C. pelagicus* (WALLICH) SCHILLER, *Calcidiscus leptoporus* (MURRY & BLACKMAN) LOEBLICH & TAPPAN, *C. leptoporus centrovalis* (STRADNER & FUCHS) PERCH NIELSEN, *Cyclolithella annula* (COHEN), *Discoaster exilis* MARTINI & BRAMLETTE, *Helicosphaera kamptneri* HAY & MOHLER, *Pontosphaera multipora* (KAMPTNER) ROTH, *Syracosphaera histrica* KAMPTNER, *Reticulofenestra pseudoumbilica* (GARTNER) GARTNER, *Rhabdosphaera clavigera* MURRAY & BLACKMAN, *Rh. sicca* (STRADNER) MARTINI, *Sphenolithus heteromorphus* DEFLANDRE, *S. moriformis* (BRÖNNIMANN & STRADNER) BRAMLETTE & WILCOXON, *Pontosphaera* (?) sp. (Pl. 1).

The presence of *Sphenolithus heteromorphus* and of *Discoaster exilis*, as far as the base of the lower sequence, makes it possible for the nannofossils assemblage under study to be referred, without any doubt, to the Sphenolithus heteromorphus Zone – NN_5 (Martini, 1971). Consequently, the deposits of the Dej Tuff, developed east and north-east of Zalău are assigned to the Langhian (Lower Badenian). The inexistence of the base of the Langhian, i.e. the upper part of the Helicosphaera ampliaperta Zone – NN_4 (Martini, 1971), that we have found at Călan-Hunedoara, is probably due to a slight delay of the Langhian transgression in this zone.

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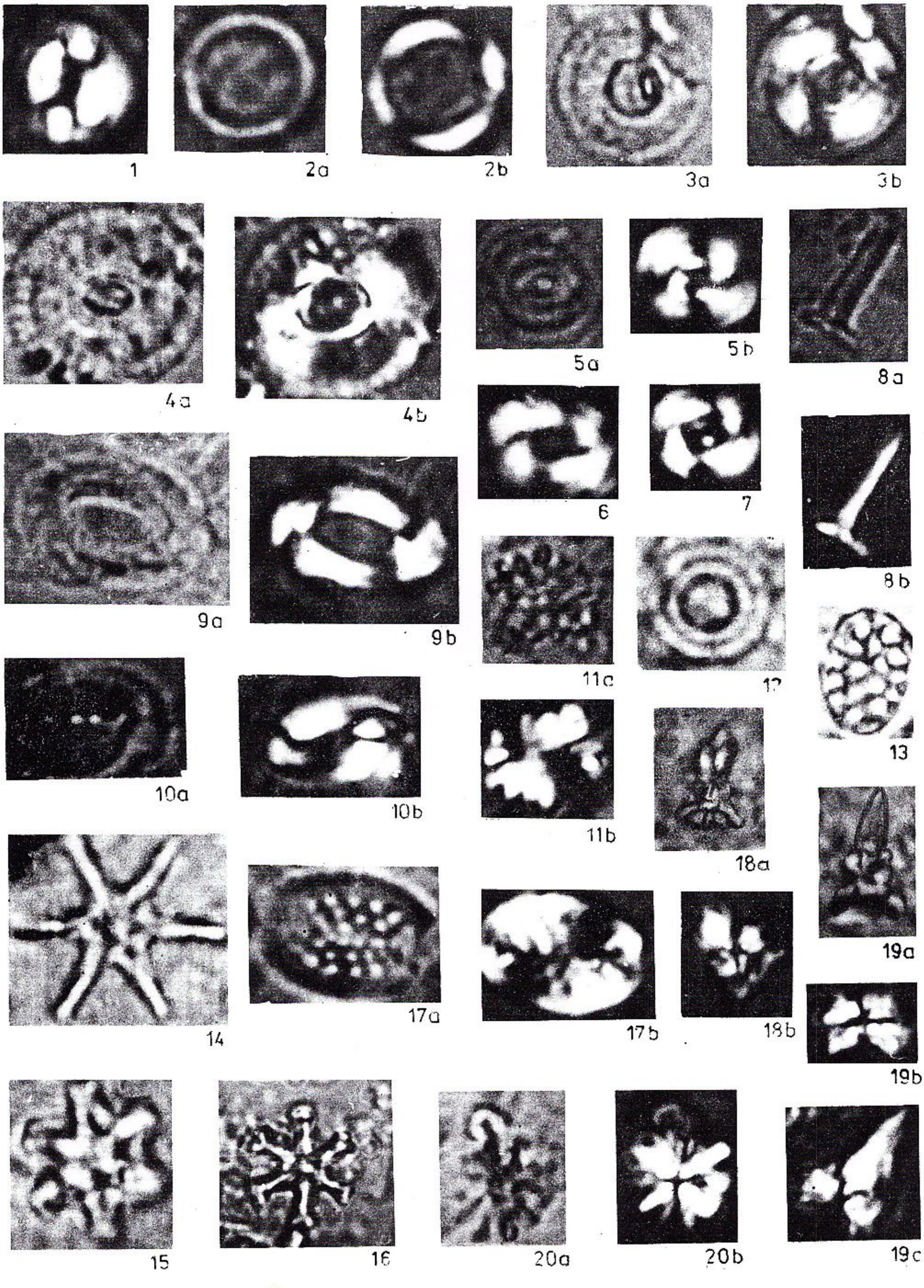
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18b
19b
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16
20a
20b
19c

Plate

Fig. 1 – *Coccolithus pelagicus* (WALLICH); N+; Chechiş Beds; Rogoaze Stream; x 2500.
Fig. 2 – *Coronocyclus nitescens* (KAMPTNER); 2a–N II; 2b–N+; Chechiş Beds; Rogoaze Stream; x 3000.
Fig. 3 – *Calcidiscus leptoporus* (MURRAY & BLACKMAN); 3a–N II; 3b–N+; Dej Tuff; Lazu Stream; x 2500.
Fig. 4 – *Calcidiscus leptoporus centrovalis* (STRADNER & FUCHS); 4a–N II; 4b–N+; Dej Tuff; Cuceu; x 3000.
Fig. 5 – *Cyclicargolithus abisectus* (MÜLLER); 5a–N II; 5b–N+; Chechiş Beds; Rogoaze Stream; x 2500.
Fig. 6 – *Reticulofenestra pseudoumbilica* (Gartner); N+; Dej Tuff; Lazu Stream; x 3300.
Fig. 7 – *Reticulofenestra lockeri* MÜLLER; N+; Chechiş Beds; Rogoaze Stream; x 3300.
Fig. 8 – *Rhabdosphaera sicca* (STRADNER); 8a–N II; 8b–N+; Dej Tuff; Cuceu; x 3000.
Fig. 9 – *Helicosphaera ampliaperta* BRAMLETTE & WILCOXON; 9a–N II; 9b–N+; Chechiş Beds; Rogoaze Stream; x 3500.
Fig. 10 – *Helicosphaera kamptneri* HAY & MOHLER; 10a–N II; 10b–N+; Chechiş Beds; Frumuşelu Stream; x 3300.
Fig. 11 – *Pontosphaera (?)* sp.; 11a–N II; 11b–N+; Dej Tuff; Otlău Stream; x 3500.
Fig. 12 – *Cyclolithella annula* (COHEN); N II; Dej Tuff; Otlău Stream; x 3000.
Fig. 13 – *Holodiscolithus macroporus* (DEFLANDRE); N II, Dej Tuff; Cuceu; x 3000.
Fig. 14 – *Discoaster exilis* MARTINI & BRAMLETTE; Dej Tuff; Otlău Stream; x 3300.
Fig. 15 – *Discoaster musicus* STRADNER; Dej Tuff; Otlău Stream; x 3300.
Fig. 16 – *Discoaster variabilis* MARTINI & BRAMLETTE; Dej Tuff; Cuceu; x 3000.
Fig. 17 – *Pontosphaera multipora* (KAMPTNER); 17a–N II; 17b–N+; Chechiş Beds; Rogoaze Stream; x 3700.
Fig. 18 – *Sphenolithus dissimilis* BUKRY & PERCIVAL; 18a–N II; 18b–N+ at 45^0; Chechiş Beds; Rogoaze Stream; x 3300.
Fig. 19 – *Sphenolithus heteromorphus* DEFLANDRE; 19a–N II; 19b–N+ at 0^0; 19c–N+ at 45^0; Dej Tuff; Otlău Stream; x 3000.
Fig. 20 – *Sphenolithus moriformis* (BRÖNNIMANN & STRADNER); 20a–N II; 20b–N+; Dej Tuff; Otlău Stream; x 3300.

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SUR LES FORMATIONS NÉOGÈNES DE LA ZONE DE MARGINA – LĂPUGIU

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Key words: Neogene. Badenian. Sarmatian. Pontian. Pannonian Depression – Adjacent Depressions – Lipova Depression.

Abstract: *On the Neogene Formations on the Margina-Lăpugiu Sheet.* The authors have approached four problems: a) the agglomerates partly support Badenian sedimentary deposits and partly are interbedded with them; b) the authors have only identified Lower Badenian (=Langhian) deposits, at whose upper part they have noticed gravels with fresh water fauna (unionids); c) they have paleontologically proved that the Sarmatian formations overlie a much more restricted area; d) it is only the quartzose gravels that they have assigned to the (only Lower) Pontian, the other gravels being considered to represent older, equally detrital, Miocene formations that have been surficially weathered.

La carte "Margina-Lăpugiu" couvre une région délimitée à l'ouest et à l'est par ces deux localités, et par la rivière Mureş au nord et nord-est et approximativement la rivière Bega au sud. Plus de la moitié de ce territoire est couvert par des terrains néogènes, appartenant à ce que nous avons nommé "le golfe de Făget" (Marinescu et al., 1977). C'est une zone qui se développe à l'est de Lugoj, ayant en axe les localités de Făget, Margina et Lăpugiu. Récemment, on a creusé ici plusieurs forages peu profonds, qui ont fourni des informations supplementaires, particulièrement utiles, en confirmant ou en complétant les données déjà connues des affleurements. Ces forages ont fait l'objet de deux études (Mihăilescu et al., 1987; Papaianopol et al., 1987; non publ.) à conclusions plus ou moins contradictoires. Au dernier nous avons participé nous mêmes, en étudiant les forages encore conservés à l'époque, desquels quelques uns avaient été exécutés après la fin du premier.

Par rapport à ce qu'on connaissait déjà des études antérieures, les dépôts néogènes de ce périmètre soulevaient une série de problèmes, dont certains concernaient aussi les secteurs voisins:

1. – les relations entre les agglomérats et les formations sédimentaires, un problème concernant aussi, au moins partiellement, l'âge des éruptions;

2. – l'existence du Badénien supérieur;

3. – la surface couverte par les formations sarmatiennes;

4. – la surface couverte par les formations pontiennes.

Les problèmes sont aussi le résultat du nombre réduit de coupes bien ouvertes et de bons affleurements, la région étant bien recouverte par la végétation; ensuite, une série de notes publiées contredisaient certaines conclusions plus anciennes, mais n'étant pas confirmées par les recherches ultérieures. Notre intention est justement d'élucider ces problèmes.

1. *La question des relations entre les agglomérats et les formations sédimentaires.* Des études plus anciennes, ces relations ne semblaient pas être assez claires; certains auteurs étaient d'avis qu'ils reposaient sur les dépôts badéniens, tandis que les autres remarquaient le rapport invers. Ultérieurement, on a élucidé certaines choses: a) dans un secteur voisin, à l'est, on a établi que l'âge des dépôts attribués au Miocène moyen qui supportaient les agglomérats était crétacé; b) au nord, vers Alba Iulia, des agglomérats qui semblaient faire partie de la même formations que ceux de Lăpugiu, supportent des dépôts éocène supérieurs; c) les datations radiométriques de quelques échantillons provenant du sud du Mureş ont indiqué un âge d'approximativement 65 m.a.; d) un forage de Lăpugiu de Jos, après avoir traversé des dépôts badéniens, à une profondeur de 66 m, a rencontré les agglomérats qu'il a traversé 34 m.

Il en résulterait que ces agglomérats sont antérieurs au Badénien. D'ailleurs, même le contour de la limite cartographique montre que les dépôts badéniens de la zone méridionale de la feuille (à Pietroasa-Crivina de Sus), de l'ouest de Lăpugiu, tout comme ceux du nord (Coştei, Fintoag), reposent toujours sur les agglomérats. Exceptant les situations déjà mentionnées, dans la zone septentrionale, les relations entre les deux formations discutées sont souvent tectoniques. Cependant, on a rencontré aussi un autre type de relations: près de l'embouchure de la Valea lui Popa (Şesului), au sud d'Ohaba, même dans le premier affleurement, au rive droite de cette vallée, en aval d'une bergerie, au-dessous des agglomérats, suivent des tufs et ensuite des argiles siltiques, muscovitiques, violacé-noirâtre. Ces argiles, assez tectonisées, contiennent d'aileurs un banc de tuffites aussi.

On connait déjà que les agglomérats comportent plusieurs intercalations cinéritiques; parfois elles sont melangées de lapilli, de bombes et d'éléments de grandes dimensions d'eruptif. D'autres fois il y a seulement des tufs blanchâtres, ou légèrement verdâtres. En poursuivant le plus haut banc de tuf, au long de la région d'Ohaba-Baştea-Coşeviţa, on a constaté que d'abord celui-ci est présent, encadré par les agglomérats, sur la crête de l'ouest de la Valea lui Popa (Şesului), d'où il descend vers l'ouest, dans Valea Rea, encadré toujours par les agglomérats; on le retrouve, cette fois-ci accompagné par des lapilli et seulement à peu d'éléments d'agglomérats, dans la vallée immédiatement suivante vers l'ouest, la vallée de la Mohira, en supportant ici des graviers centimétriques ou plus fins, mais aussi avec des éléments de 5–10 cm, en alternance avec des sables (Fig. 1).

Plus vers l'ouest, à l'extrémité de l'interfleuve entre la vallée de la Mohira et la vallée de Groşi, on retrouve la même couche de tuf, accompagnée toujours par un banc de lapilli, intercalé entre des sables dans le lit et des graviers au toit. Toujours encadrée par des sables et des graviers, la couche de tuf est présente aussi au bout d'une colline de l'ancien village de Baştea, tout comme au nord de Coşeviţa, ou au sud-ouest de ce village, dans la vallée de la Homoşdia; ici elle dépasse 2 m d'épaisseur, accompagnée toujours par le banc de lapilli. Dans ce dernier endroit le tuf et les lapilli sont intercalés entre des argiles siltiques fossilifères, à nombreux mollusques badéniens (la même faune qu'à Lăpugiu, donc Badénien inférieure) (Fig. 2).

La situation décrite indique nettement qu'en même temps avec l'éruption qui a généré une importante quantité d'agglomérats à l'est (Lăpugiu), pas plus loin vers l'ouest (Baştea, Coşeviţa) s'accumulaient lentement les sédiments argileux et sableux, fossilifères, où, comme résultat d'une explosion qui a produit aussi un materiel plus fin, de la cendre et même des lapilli, s'est déposé aussi le banc de tuf mentionné. Cela ne contredit point les éléments énumérés:

– On ne peut pas soutenir que toute la masse d'agglomérats, d'une si vaste région, au sud des Monts Apuseni, se soit accumulée brusquement. Elle peut être le résultat des éruptions successives, à des intervalles de temps, donc d'âges différents. Même les centres d'émission de ces matériels éruptifs nous sommes convaincus d'être différents.

– L'éruption mentionnée représente seulement un court épisode par rapport à la durée du Langhien. Ainsi on peut expliquer qu'au dessus des dépôts éruptifs respectifs les dépôts langhiens se sont accumulés continuellement, en les recouvrant "en transgression".

2. *L'existence du Badénien supérieur.* De ce qu'on connait déjà jusqu'à présent, les dépôts badéniens de Lăpugiu et Coştei représentent le Badénien inférieur, étant souvent comparés au Langhien. Tous les forages que nous avons étudiés dans la région n'ont traversé que des dépôts de cet âge, particulièrement fossilifères, surtout des foraminifères (à l'exception du forage 14 Ohaba, qui a traversé des argiles, des siltites et des sables en faciès continental-lacustre, sans faune).

Dans une étude non-publiée sur le secteur de Măgura-Fintoag, Mihailescu et al. (1985, non publ.) mentionnent plusieurs fois la présence du Badénien supérieur (Kossovien): le cours supérieur de la vallée de la Năndreasca, dans la vallée de l'Icui, dans Valea Mare au nord de Fintoag. Les arguments sont quelques mollusques marins indifférents (*Pitaria italica, P. chione et Venus* sp.), tout comme deux espèces de *Velapertina*, dans une seule site. Nous n'avons pas retrouvé ces arguments; de plus, *Velapertina* a été parfois confondue avec *Candorbulina*. En même temps, il faut souligner la ressemblance lithologique des dépôts attribués au Kossovien avec ceux langhiens. C'est la raison pour laquelle c'était impossible de les reconnaître et de les cartographier.

Il faut faire encore une précision: dans l'étude mentionnée (Mihăilescu et al., 1985; non publ.), il s'agit du "faciès des calcaires de Leitha" du Kossovien. On est devenue une habitude de parler de calcaires de Leitha pour toutes les intercalations des calcaires, surtout récifaux. De plus, on ajoute que c'est un calcaire vacuolaire, friable. C'est un heritage incorrect, qui doit être éliminé du langage courant. En réalité, le calcaire de Leitha est un calcaire compact, particulièrement dur, d'origine récifale, c'est vrai, à nombreux mollusques caractéristiques pour ce faciès, du Badénien inférieur. L'aspect des calcaires mentionnés, qui se développent à l'est de Coştei, ne ressemble pas celui des calcaires dont on a parlé, bien que par rapport à l'âge (Badenien inférieur), ils peuvent être synchrones. Dans ce cas, il s'agit d'une double incorrectitude, concernant tant l'âge, que l'aspect. On devrait éviter de telles comparaisons et références à distances si grandes, sans une vérification documentée. Ni même l'expression "du type"... ne peut être adéquate, au moins pour ce cas.

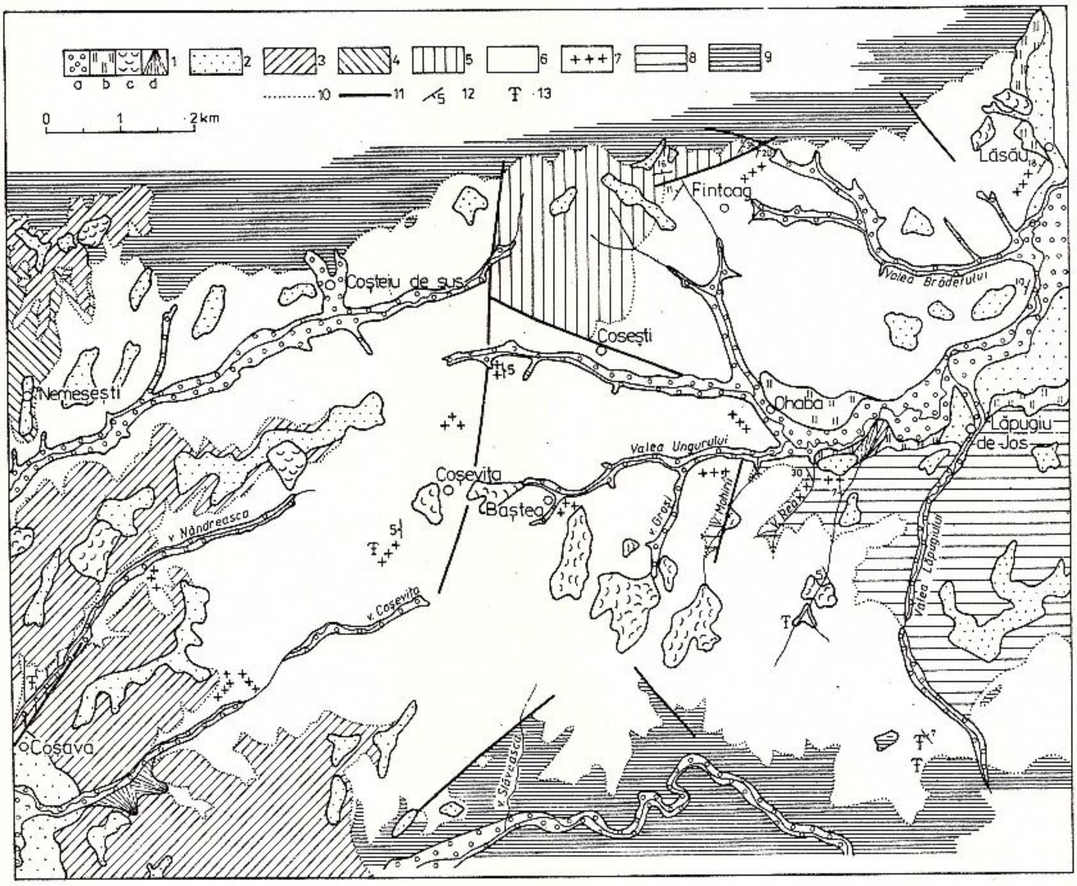

Fig. 1 Formations néogènes de la zone de Margina-Lăpugiu. Holocène: 1, alluvions (a), dépôts déluviaux (b), dépôts colluviaux (terrains glissés) (c), cône de déjection (d). Pléistocène: 2, terrasses. Pontien: 3, graviers, argiles. Malvensien: 4, argiles, sables. Sarmatien (Volhinien): 5, sables, argiles sableuses. Badénien: 6, sables, argiles sableuses; 7, tufs volcanics; 8, aglomérats volcanics; 9, formations de bordure; 10, limite géologique; 11, faille; 12, position des couches; 13, site fossilifère.

Nous n'avons pas assez d'arguments pour nier l'affirmation de nos collègues devanciers (d'ailleurs c'est la situation de toutes assertions), surtout qu'à l'ouest le Badénien est bien mis en évidence, tant du point de vue lithologique, que paléontologique, dans les forages au sud de Caransebeş; en échange, nous n'avons pas pu le détecter dans la région. De plus, en nous trouvant dans une région de développement classique du Badénien inférieur (Lăpugiu, Coştei), bien représenté, exclusivement dans tous les forages (Gh. Popescu, in Papaianopol et al., 1987; non publ.), sur la carte nous n'avons pas pu figurer un "Badénien non-divisé", car celui-ci n'aurait pas correspondu à la réalité, le Badénien inférieur étant partout très argumenté.

Ayant en vue que le Badénien supérieur est présent tant à l'ouest de la région (Caransebeş), qu'à l'est (Buituri), on pourrait supposer qu'il avait couvert ce territoire aussi, étant érodé ultérieurement, pendant le Sarmatien inférieur. Cette hypothèse semble tout de même un peu forcée; plus plausible nous semble son absence par la non-accumulation, après l'interruption de la sédimentation pendant le Langhien supérieur. Alors, en Transylvanie et dans toute l'aire carpathique, a eu lieu un élèvement, avec l'isolement d'une large région lagunaire, y s'accumulant les évaporites. C'est normal que dans les secteurs immédiatement voisins aient eu lieu des exodations. Cette opinion est soutenue aussi par des graviers à zones conglomératiques, à l'aspect fluviatile, qui finissent la sédimentation badénienne et dans lesquels on a trouvé ici quelques valves d'Unio, malheureusement dans un mauvais état de conservation. Cette formation se trouve entre Lăsău et Lăpugiu de Jos, au sud de l'embouchure du ruisseau du Brădet, au long d'un chemin forestier récemment coupé. Ces données, corroborées avec l'aspect continental-fluviatile des dépots traversés par le forage 14 Ohaba (approximativement dans l'axe de la zone), nous confirment la contemporanéité de ces dépôts avec les évaporites de

Transylvanie (Badénien moyen=Wiélicien=Langhien supérieur), accumulées dans une phase d'élèvement et donc de colmatage de ce fond de golfe. Les eaux sont revenues ici à peine pendant le Sarmatien inférieur et même pas dès le commencement de celui-ci.

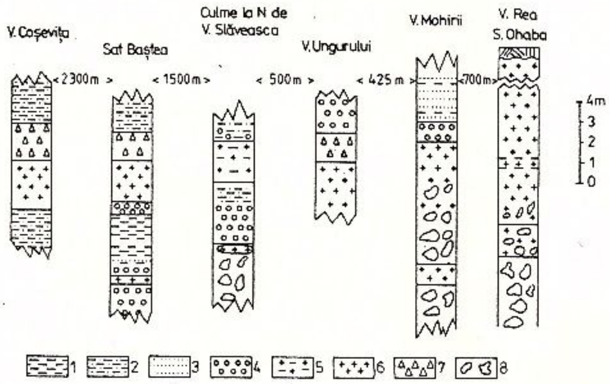

Fig. 2 – L'horison de tufs badéniens du secteur Lăpugiu-Coşeviţa (Banat). 1, argiles; 2, siltits; 3, sables; 4, graviers; 5, bentonites; 6, tufs volcanics; 7, lapilli; 8, aglomerats volcanics.

3. *L'extension des formations sarmatiennes.* Les dépôts sarmatiens ont été découverts par Huică et al. (1968, nonpubl.) et Huică et Dragu (1970). Des trois secteurs où ils ont été signalés, le seul avec une faune riche et représentative, est celui de la vallée de Năndreasca, au bout en amont du village de Coşava. Là-bas il y a des sables avec des graviers en base, suivis d'argiles sableuses, à nombreux mollusques, dont on a déterminé plusieurs espèces caractéristiques pour le Volhynien (Orăşanu et al., 1971).

Un autre gisement fossilifère, que nous avons retrouvé aussi, se trouve sur un affluent de droite de la vallée du Fintoag, au bout en amont du village avec le même nom; il s'agit d'argiles noirâtres charbonneuses comportant un banc lumachellique de 50–60 cm d'épaisseur environ, qui repose sur une couche, sous 1 m, de tuf benthonitisé, blanchâtre-verdâtre. La lumachelle est formée d'ostréidés (*Griphaea gingensis*), avec quelques autres mollusques qui, seulement en forçant les choses, peuvent être considérés caractéristiques pour le Sarmatien. En réalité, ils indiquent un faciès saumâtre du Miocène moyen en général. La position littorale de la site et l'existence de certains dépôts charbonneux justifient le degré plus réduit de salinité des dépôts badéniens. La couche de tuf mentionnée permet la corrélation avec les dépôts du sud, même la précision de sa position dans les successions stratigraphiques à faune badénienne (de Coşeviţa, Coştei, Lăpugiu, Lăsău).

Les autres sites, attribuées au Sarmatien, décrites de la vallée du Şeş, à Ohaba, comportent certainement des dépôts badéniens. De plus, on peut mentionner un autre secteur, au nord de Coseşti, encadré par des failles et traversé, approximativement à son centre, par le forage 6 Coseşti. Exceptant la microfaune, dans ce forage à 175 m de profondeur, on a rencontré une lumachelle épaisse de 15 cm, formée exclusivement par des cérites (*C. pictum, C. mitrale, C. rubiginosum*). La base du Sarmatien doit être tracée donc au moins au-dessous de ce banc (Papaianopol et al., 1987, non publ.).

On peut coclure que, bien que les dépôts sarmatiens aient recouvert auparavant une surface beaucoup plus étendue, en faisant peut-être la liaison entre ceux de Banat et ceux de Haţeg et de la Transylvanie, ils ont été ultérieurement fortement érodés, ne se conservant certainement que seulement deux sites: celle de la vallée de la Năndreasca, à Coşava et au nord de Coseşti.

4. *L'extension des formations pontiennes.* L'intervalle stratigraphique qui suit le Sarmatien sur la feuille de Margina appartient au Malvensien supérieur (zone E du Pannonien); il est bien daté du point de vue paléontologique, surtout par les gisements fossilifères de Groşi et de Sinteşti et Zorani (Orăşanu et al., 1971; Marinescu et al., 1977). Y suivent des sables et des graviers blancs quartzeux, dépourvus de faune; il s'agit de la même formation que celle exploitée dans les mines de Gladna et de Jupâneşti (au dehors des limites de la feuille). Au-dessus de ceux-ci il y a un horizon d'argiles réfractaires, qui ne se retrouve pas sur la feuille de Margina. L'âge de ces dépôts a été attribué au Pontien inférieur (Odessien), par leur position stratigraphique, étant encadrés entre le Malvensien supérieur et le Portaferrien.

Ces graviers forment les versants d'une série de collines au nord et à l'est de la localité de Margina; entre Coşava et Coşeviţa, il y a plusieurs années, il y avait aussi quelques petites mines dont les traces sont encore visibles au long de la route, sur la coupe de la colline entre les vallées de la Năndreasca (au nord) et de la Homoşdia (au sud).

A partir de ces situations, en extrapolant, tous les graviers des coupes des collines entre les vallées de la Bega et du Mureş (vers Gurasada) ont été interprétés comme représentant le Pontien inférieur. Ainsi, a été délimitée sur de diverses cartes, une plaque horizontale de graviers polygènes, dépourvus de faune, en couvrant toute sorte de formations; la solution semblait être commode, solution pour une série de lacunes d'information. On omettait pourtant deux choses: a – tant les formations sarmatiennes, que celles badéniennes contiennent des dépôts détritiques, des sables et toute sorte de graviers (mais en aucun cas de sables blancs quartzeux); b – les dépôts pléistocènes recouvrent vraiment certaines surfaces, les plus hautes, ou des bouts de collines, étant représentés tant par des argiles diluviales, à concrétions féromanganeuses ("bobovines", bohnerz), que des sables et des graviers alluviaux. La distinction dans ce dernier cas est très difficile et, s'ils ne recouvrent pas effectivement des terrasses, cette distinction est même hasardée.

En suivant le passage des dépôts détritiques miocènes des coupes de certaines vallées, vers ces graviers couvrant les collines, nous avons constaté que, le plus souvent, ces-derniers ne sont que des dépôts badéniens, ou sarmatiens, altérés vers la partie supérieure des coupes, ou sur les coupes des collines. Comme les dépôts badéniens ne sont, eux-aussi, consolidés que par le tassement, l'action d'érosion est très facile. Pour ces raisons nous sommes d'avis que pour les dépôts pontiens (odessiens) il faut garder seulement les graviers et les sables quartzeux, en les considérant comme appartenant aux dépôts miocènes plus anciens.

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LE NÉOGÈNE DANS LES FORAGES DE LA DÉPRESSION PANNONIQUE

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S. C. "Prospecţiuni" S. A. Str. Caransebeş 1, 78344 Bucureşti 32.

Key words: Neogene. Stratigraphic units. Mollusks. Ostracods. Biostratigraphy. Cores. Romanian Pannonian Basin. Pannonian. Depression.

Abstract: *Neogene in the Boreholes of the Pannonian Depression.* The authors present the biostratigraphic study of the material provided by 40 drillings carried out in the Pannonian Depression by the I.F.L.G.S., I.P.E.G.-Banat and I.P.E.G.-Cluj. The Sînnicolau de Munte Formation (Langhian-Kossovian) and the Sînnicolau Mare-Tomnatec Formation (Pannonian s.str. - Malvensian, Pontian, Dacian, Romanian) have been pointed out. The last deposits ending the sedimentation in the Pannonian Depression belong to the Pleistocene. The Neogene formations, especially the Pontian ones, exceed 3000 m in thickness.

La Dépression Pannonique, unité structurale majeure située dans la partie occidentale de notre pays, à suscité un grand intéret surtout en ce qui concerne les substances énergetiques y présentes (hydrocarbures, charbons et eaux thermales).

Sa masse (superposée aux éléments plissés et à la couverture post-tectogénétique du soubassement plissée) appartient au Néogène, étant representée par des dépôts badéniens, sarmatiens, pannoniens, pontiens et quaternaires.

Il y a un riche matériel de spécialité concernant la géologie de cette unité: Paucă (1935, 1976), Liteanu et al. (1965), Ichim et al. (1967), Paucă et al. (1968), Drăgulescu et al. (1968), Mihăilă et al. (1969, 1971), Istocescu, Ionescu (1970), Vasilescu et al. (1969), Vasilescu, Nichiti (1969), Crăciun (1980, non publ.), Marinescu, Popescu (1987) et Petrescu et al. (1987). Aussi, Patrulius et Drăgănescu (1970, non publ.), Patrulius et al. (1971, 1972, non publ.) nous ont offert des détails biostratigraphiques et structuraux concernant le secteur situé au nord de la ville d'Oradea.

On présente dans cette note l'étude biostratigraphique et partialement pétrographique des carottes de 40 forages emplacés le long de la frontière avec la Hongrie, entre Tăşnad-Pişcolţ au nord et Jimbolia-Timişoara au sud.

Stratigraphie

La sédimentation des dépôts néogènes a eu lieu dans des conditions très différentes de subsidence, fait qui a déterminé (Patrulius et Drăgănescu, 1970) une différentiation faciale entre les secteurs fortement enfouis, par rapport à ceux élevés. Pour exemplifier les successions les plus complètes nous avons choisi spécialement les forages percés dans des zones fortement abaissées ou la sédimentation a connu un processus presque continu. Ainsi, nous avons séparé 3 formations géologiques: formation de Sînnicolau de Munte (formation marine, Languien-Kossovien), formation saumître (Sarmatien-Volhinien+Bessarabien inférieur) et formation de Sînnicolau Mare (Pannonien sens restr. -Malvensien, Pontien-Dacien-Romanian).

Formation de Sînnicolau de Munte (formation marine-Languien+Kossovien)

Les dépôts de la base de cette formation reposent sur un soubassement comportant du cristallin et des formations paléozoïques, mésozoïques et paléogènes. Toute la succession est argilo-gréseuse surtout, ayant des intercalations de tufs. Dans quelques forages de la partie septentrionale de la dépression (Biharea, Borş, Tămăşeu, Tăuteni, Ciocaia, Pişcolţ et Carei), les dépôts de la base de la formation ont un caractère fortement détritique (Istocescu, Ionescu, 1970), étant représentés par conglomérats, marnes gréseuses couleur café-brune et grès. Ces dépôts ont été attribués au Miocène inférieur, étant donné les similitudes lithologiques avec les dépôts de la dépression de Şimleu considérés ce temps-là ayant le même âge.

A des différents niveaux de cette série détritique qui comporte des cinérites aussi, on a rencontré des marnes tufacées à orbitolinidés. Patrulius et Drăgănescu

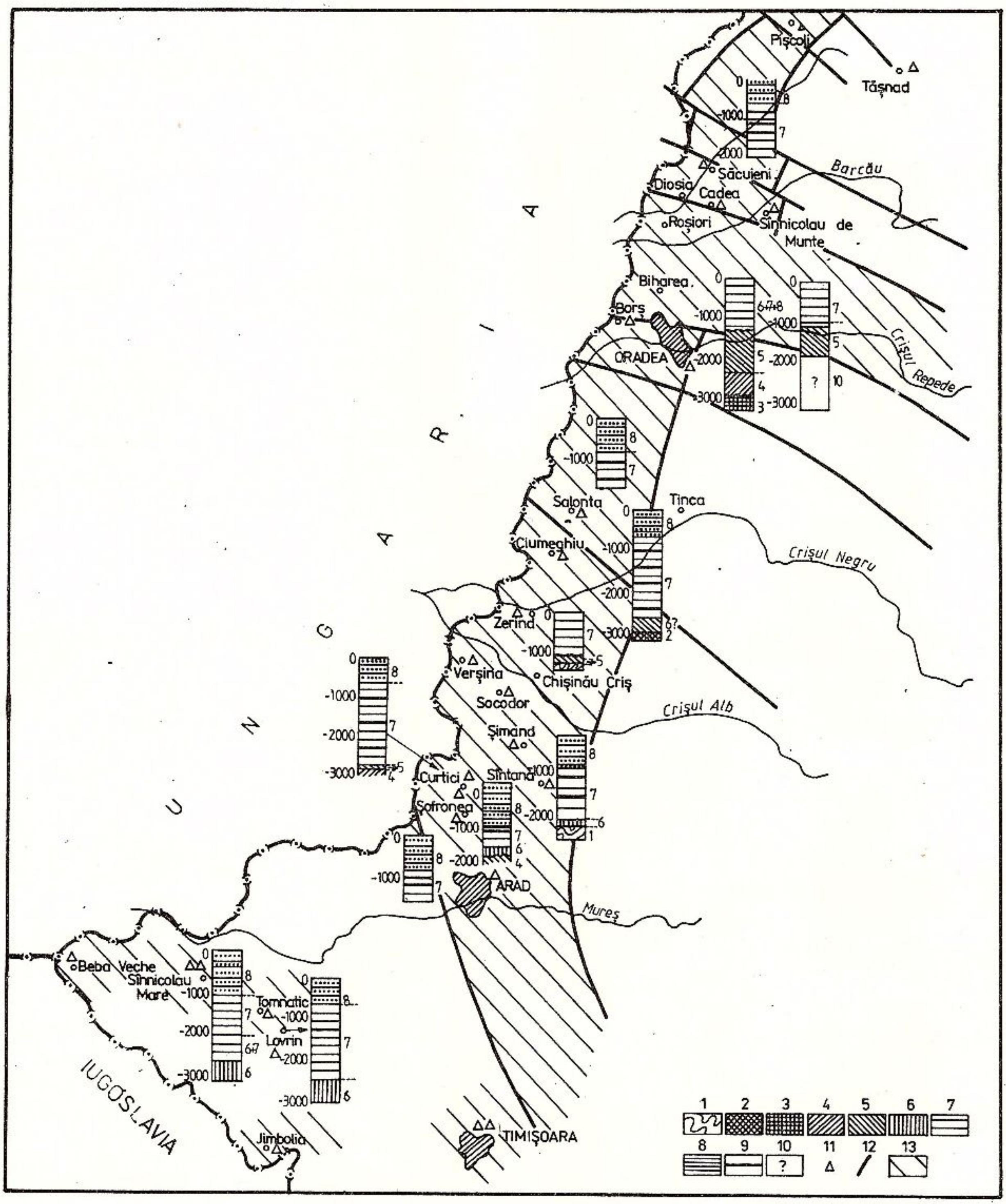

Schéma des principaux grabens post-tectogénétiques du soubassement de la Dépression Pannonique (selon Săndulescu (1984), avec des nouveaux apports) avec l'emplacement des forages étudiés et de quelques colonnes stratigraphiques de forages. 1, Cristallin; 2, Mésozoïque; 3, Paléogène; 4, Badénien; 5, Sarmatien; 6, Pannonien s. rstr.; 7, Pontien; 8, Dacien, Romanien; 9, couches de lignite; 10, non défini; 11, forages; 12, ligne de faille; 13, grabens.

(1970, rapport nonpubl.) signalent des orbitolinidés dans la partie basale de la formation à cinérites de Cadeu, sur une épaisseur de 80 m, ainsi que des "globigérines, *Nonion*, rares unionidés et foraminifères planctoniques" dans les intercalations pélitiques du forage de Borş.

La microfaune présente dans les forages de Borş, dans l'intervalle de 2850 à 2950 m et de Sîn Mihaiul German (NO de Timişoara), de 2140 à 2150 m, est assez riche: *Candorbulina suturalis* (BRONN), *Globigerinoides triloba* (REUSS), *G. immatura* LE ROY, *G. sacculifera* CITA, PREMOLI-SILVA, ROSI, *Globigerina bulloides* D'ORB., *G. regina* CRESCENTI, *Globorotalia (Turborotalia) bykovae* AIS., *Uvigerina semiornata* PAPP. TURN., *Heterelopa dutemplei* (D'ORB.). On signale aussi la présence des mollusques dans le forage de Sin Mihaiul German, dans l'intervalle de 2148 à 2152 m: *Nucula nucleus* (1), *Lucina* sp., *Corbula* sp.

Dans le graben de Borş-Sînnicolau de Munte la succession continue par des grès argileux et grès quartzeux ou calcaires, marnes siltiques, marnes gréseuses, grès pyroclastiques et pyroclastites. Les intercalations pyroclastiques sont quand même beaucoup plus rares.

Patrulius et al. (1971, données non publ.) cite de cet horizon: "*Globigerina triloba, Cibicides lobatusus, C. pseudoungerianus, Spirialis andrussowi* et *Gyroidina soldanii*. Les mêmes auteurs mentionnent la présence dans les grès calcaires et les calcaires sableux des "globigérines grandes, miliolidés de grande taille, *Borelis mello, Elphidium, Cibicides* et *Articulina*.

Aussi dans le secteur septentrional investigué par nous, à Păţal, entre 1295 et 1310 m, et à Curtici, entre 2660–2700 m, dans des grès calcaires grisâtres semblables on a rencontré: *Elphidium crispum* (LINNÉ), *Asterigerina planorbis* D'ORB., *Baggatella lappa* (CUSH.-PARKER), *B. elongata* D'ORB., *Heterolepa dutemplei* (D'ORB.), *Velapertina luczkowska* POPESCU, *Florilus communis* (D'ORB.), *Melonis pomillioideus* (FICHTELL et MOLL.), *Spiratella andrussowi* (KITTL.).

Selon Patrulius et al. (1971, données nonpubl.), dans quelques forages situés au nord d'Oradea, les couches terminales badéniennes sont marno-argileuses surtout sur une épaisseur de 30 à 100 m et comportent des niveaux ayant des "faciès faiblement suamatres" à *Elphidium, Nonion, Ammonia beccari*, aussi bien que des espèces d'*Articulina*.

L'épaisseur des dépôts badéniens (présents dans les forages de Borş et Sînnicolau de Munte dans l'intervalle de 2280 à 2950 m) varie entre 10 et 600 m.

La faune y présente indique que dans la Dépression Pannonique, notamment dans le secteur du nord d'Oradea, le Badénien a été entièrement représenté. Nous avons reconnu ainsi tant le Languien basal (zone à Candorbulina universa) *Globorotalia bykovae*, aussi bien que le Kossovien – zone à Velapertina, ainsi que des associations à *Spirialis* div. sp. et *Borelis mello* caractéristiques pour cette zone aussi.

Formation saumâtre
(Sarmatien-Volhinien+Bessarabien inférieur)

Il paraît exister un passage graduel, au moins dans le secteur roumain de la partie septentrionale de la Dépression Pannonique, entre la formation de Sînnicolau de Munte et la formation saumâtre inférieure. La suite de dépôts gréso-calcaires devient selon ce qu'on a déjà mentionné, marno-argileuse surtout, sur un intervalle de 30 à 200 m, et puis exclusivement détritique, comportant des grès à intercalations de conglomérats et microconglomérats (2 à 300 m). La formation saumâtre rencontrée dans le secteur situé entre Zădăreni-Chişinău Criş et Arad, présente un caractère transgressif, reposant directement sur les schistes cristallins à des différentes profondeurs (Zădăreni 792 m, Chişinău Criş 1263 m, Socodor 2663 m et Arad 1106 m, selon Liteanu et al., 1965).

Les dépôts trouvés à la base de la formation saumâtre portent des foraminifères y predominant *Anomalinoides dividens*. Ainsi, dans les forages de Păţal (1245 à 1295 m), Curtici (2610 à 2660 m), Gerăuşa-Crişana (1160 à 1170 m) nous avons reconnu des associations à *Anomalinoides dividens* LUCZK., *Elphidium macellum* (FICHTELL et MOLL.), *Nonion* sp., *Quinqueloculina reussi* BOGD., *Mutillus* sp., fragments de bryozoaires. A Tăşnad (département de Bihor), dans l'intervalle de 1040 à 1042 m on a rencontré, dans les grès microconglomératiques gris-noirâtres, des lumachelles à *Obsoletiforma (Obsoletiforma) obsoleta* (EICHW.) et *Musculus* sp.

La suite de dépôts gréso-calcaires, superposée aux horizons de la base de la formation, comporte des foraminifères, stratolites de mysidés et ostracodes. A Păţal (entre 1228 et 1245 m) et Episcopia Bihorului (de 1060 à 1810 m), font leur apparition, dans les marnes grisâtres à intercalations de calcaires gréseux grisatres, la suivante association faunique: *Articulina problema* BOGD., *Miliolina akneriana* D'ORB., *Elphidium rugosum* D'ORB., *Elphidium sculpturatum* CUSH., *E. macellum* (FICHTELL ET MOLL.), *Nonion martkobi* BOGD., *Ammonia beccari* (L.), *Porosononion granosum* D'ORB., ostracodes.

Dans les forages effectués au nord de la ville d'Arad, à Sîntana (960 à 1000 m), Sofronea (1790 à 1800 m), Cebza (1907 à 1910 m), dans les grès calcaires apparaissent: *Paramysis michai* VOICU, *Protelphidium granosum* (D'ORB.), *Ammonia beccari* (L.), *Elphidium macellum* (FICHT. et MOLL.), *E. hauerinum* (D'ORB.), *E. rugosum* (D'ORB.), *Quinqueloculina akne-*

riana D'ORB., *Q. consobrina* D'ORB., *Florilus bogdanowiczi* (VOL.), *Xestoleberis elongata* STANCEVA.

La formation saumâtre, ayant une épaisseur de 30 à 800 m est suivie d'une manière transgressive ou discordante parfois, de dépots argilo-gréseux, grisâtre-noirâtre à des intercalations de grès noiratres à *Congeria banatica* et *Silicoplacentina*, ou d'argiles à des intercalations de sables portant une faune pontienne.

La formation saumâtre est sarmatienne en sens pannonique. La base du Sarmatien, zone à Anomalinoides dividens, a été intercepté par plusieurs forages; ainsi, dans le forage de Gerăuşa (Crişana), a été percée, sur une épaisseur de 40 m, une suite d'argiles grisâtres, comportant uniquement *Anomalinoides dividens*; on y rencontre puis *Elphidium macellum, Quinqueloculina reussi*, des foraminifères qui indiquent le Volhinien. Dans la séquence surjacente apparaissent *Protelphidium granosum* (D'ORB.), *Elphidium* div. sp., *Quinqueloculina* div. sp., *Xestoleberis* div. sp., qui indiquent la présence du Volhinien supérieur et du Bessarabien inférieur.

Formation de Sînnicolau Mare (Pannonien s. restr. terminal, Pontien-Dacien-Romanien)

Cette formation repose transgressivement sur des entités stratigraphiques plus anciennes: cristallin, Mésozoïque, Paléogène, Badénien. Elle est suivie par des formations quaternaires sableuses surtout.

Les horizons de la base de la formation, à une éppaisseur de 2 à 300 m, identifiés par nous dans une zone septentrionale à Săcuieni-Arad, tout comme dans une autre méridionale à Sînnicolau Mare-Lovrin, sont représentés par des grès, argilites, argiles et argiles marneuses à restes de mollusques et ostracodes. Dans les argilites noirâtres traversées par les forages de Simand (au nord d'Arad) (2066 à 2068 m), Lovrin (2244 à 2900 m), Sînnicolau Mare (3197 à 3200 m), Sîn Mihaiul German (1294 à 1297 m) nous avons identifié la faune suivante: *Congeria banatica* (R. HÖRN.), *Paradacna abichiformis* (GORJ. KR.), *Pontalmyra (Pontalmyra) deserta* (STOL.), *P. (P.) promultistriata* JEK., *Silicoplacentina hungarica* MEHES, *Amplocypris abscisa* (REUSS), *A. subacuta* ZAL., *Candona (Caspiola) elongata* SOKAĆ.

Dans le forage de Sofrona (au nord d'Arad), dans l'intervalle de 1520 à 1710 m, dans les argiles grisâtres, noirâtres on rencontre: *Candona (Pontoniella) unguiculus* (REUSS), *C. mutans* POK., *Candonopsis arcana* KRSTIĆ, *Loxoconcha hodonica* POK., *Hemicytherea foliculosa* (REUSS), *Amplocypris abscisa* (REUSS).

Les analyses micropaléontologiques réalisées dans les dépôts argileux percés par le forage de Gerăuşa (Crişana) (1060 à 1130 m) ont mis en évidence, près d'*Orygoceras*, une riche ostracofaune comportant: *Pontoleberis attilata* (STANCEVA), *Loxoconcha lacunosa* (REUSS), *Hemicytherea folliculosa* (REUSS), *Leptocythere nasa* (MEHES), *Xestoleberis* sp., *Amplocypris* sp., *Hungarocypris* sp.

Les horizons gréseux et les argilites de la base sont recouverts par une suite pélitique formée de marnes grisâtre-verdâtres, marnes sableuses à concrétions calcaires et des intercalations de charbons, ainsi que des sables fins grisâtres. Dans le forage de Sînnicolau Mare, dans l'intervalle de 1180 à 1550 m on a évidentié 17 intercalations de lignite à des épaisseurs réduites (inférieures à 1,5 m). Dans le même forage ont été rencontré aussi, jusqu'à la profondeur de 2830 m des restes charbonneux et des intercalations minces de charbons. Des accumulations de charbons ont été découverts dans la zone de Roşiori-Oradea où apparaissent 12 couches de charbon épaisses de 0,2 à 3 m (Nicorici, Nicorici in Petrescu et al., 1987).

La faune caractéristique de la série pélitique à charbon abonde en paradacnes en association avec les ostracodes. Dans les forages emplacés au nord d'Oradea (Borş et Păţal, dans l'intervalle de 200 à 1185 m) nous avons rencontré: *Paradacna abichi* (HOERN.), *Cyprideis macrostigma* KOLLM., *C. seminulum* (REUSS), *C. pannonica* MEHES, *Candona (Caspiolla) balcanica* (ZAL.), *C. lactea* BAIRD. Dans un paquet inférieur, entre 1185 et 1210 m nous avons prélevé aussi: *Paradacna abichi* (HOERN.), *Congeria* ex. gr., *Congeria digitifera* ANDR., *Leptocythere lacunosa* (REUSS), *Loxoconcha hodonica* POK., *L. granifera* (REUSS), *Bakunella dorsoarcuata* (ZAL.), *Silicoplacentina hungarica* MEHES.

Pour les forages de Săcuieni (au nord d'Oradea) nous mentionnons dans les argiles charbonneuses noirâtres situées entre 625 et 1645 m: *Cyprideis heterostigma* POK., *C. pannonica* MEHES, *C. multipora* POK., *Leptocythere lacunosa* (REUSS), *Loxoconcha rombovalis* POK., *Bakunella dorsoarcuata* (ZAL.): du même intervalle ont été prélevés aussi des mollusques: *Paradacna abichi* (HOERN.), *P. asperocostata* (GORJ. KR.), *Congeria digitifera* ANDRUS., *Planorbis tenuistriatum* GORJ. KR.

Les forages de Salonta ont intercepté, entre 823 et 1010 m, des argiles noirâtres à: *Paradacna abichi* (HOERN.), *P. abichiformis* (GORJ. KRAMB.), *Pontalmyra otiophora* (BRUS.); entre 1565 et 1566 m les mêmes forages ont intercepté des argiles grisâtres-blanchâtres à *Congeria spathulata* PARTSCH et fragments de limnocardiidés. L'association microfaunique présente dans ces forages entre 100 et 1620 m comporte: *Silicoplacentina hungarica* MEHES, *Bakunella dorsoarcuata* (ZAL.), *Leptocythere lacunosa* (REUSS), *Cyprideis pannonica* MEHES, *C. triangulata* KRSTIĆ, *Candona* sp. et *Silicoplacentina hungarica* MEHES.

Dans les forages de Ciumeghiu (au nord de Crişul Negru, près de la frontière avec la Hongrie) les dépôts crétacés sont recouverts par des formations pontiennes qui renferment, entre 2662 et 3025 m, une riche association pallinologique identifiée par Filofteia Sîrbu: *Diconodinium inequicornutium* BALTEŞ, *Leiosphaeridia pannonica* BALTEŞ, *Leptodinium* aff. *churcilli* HORLAND, *Hystricosphaera* aff. *bentori* (REUSS), *Thalassiophora balcanica* BALTEŞ, *Palaeostomacystis operculatum* GERMAN. On a y rencontré aussi un microplancton marin remanié, d'âge crétacé. Dans les memes forages, dans la suite de dépôts surjacente, sur l'intervalle de 720 à 2600 m on a mis en évidence une riche association d'ostracodes pontiens: *Candona (Caspiolla) balcanica* (ZAL.), *C. (Caspiocypris) alta* (ZAL.), *C. (C.) lobata* (ZAL.), *Lepthocythere lacunosa* (REUSS), *L. multituberculata* (LIV.), *Cyprideis pannonica* MEHES, *Hemicytheria pajinovicensis* (ZAL.), *Bakunella dorsoarcuata* (ZAL.).

Dans les forages de Simand (au nord d'Arad), dans l'intervalle de 1530 à 1531 m et dans celui de Sîntana (1123 à 1912 m) on a rencontré: *Paradacna abichi* (HOERN.), *Dreissena simplex auricularis* (FUCHS), *D. serbica orevacensis* STEV., *Pseudocatillus* sp., *Candona (Pontoniella) acuminata striata* MAND., *C. (Caspiolla) balcanica* (ZAL.), *C. (Pontoniella) sagittosa* KRSTIĆ, *Cyprideis triangulata* KRSTIĆ, *Candona (Caspiocypris) alta* (ZAL.).

Dans les forages d'Arad, Curtici, Zerind, Verşind et Chişinău Criş le paquet de dépôts pontiens a été traversé sur une épaisseur variable de 10 à 2770 m. A Arad, par exemple, (profondeur de 823 à 1219 m) on a rencontré des silts marneux grisâtre-jaunâtres, argilites grisâtres et des grès fins argileux marneux à: *Paradacna abichi* (HOERN.), *Paradacna abichiformis* (GORJ. KR.), *Pseudocatillus simplex* (FUCHS), *Pontalmyra otiophora* (BRUS.), *Parvidacna chartacea rădmăneşti* (GILLET), *Silicoplacentina hungarica* MEHES, *Leptocythere lacunosa* (REUSS), *L. multituberculata* LIV., *Bakunella dorsoarcuata* (ZAL.), *Candona (Lyneocypris) trapezoidea* (ZAL.), *C. (Caspiocypris) alta* (ZAL.), *C. (Caspiolla) balcanica* (ZAL.), *Pontoleberis pontica* STANCEVA, *Cyprideis seminulum* REUSS, *C. pannonica* MEHES.

Dans un forage de Beba Veche nous signalons, dans les argilites noiratres, jusqu'à 3200 m: *Leptocythere* div. sp., *Bakunella dorsoarcuata* (ZALL.), *Candona (Pontoniella) acuminata striata* MAND.

A l'est de Beba Veche, dans les forages de Sînnicolau Mare (1796 à 1986,5 m) et de Sînnicolau German (jusqu'à la profondeur de 2000 m) les dépôts pontiens comportent: *Paradacna abichi* (HOERN.), *Congeria digitifera* ANDRUS., *Dreissenomya* sp., *Valenciennius* sp., *Gyraulus* sp.

Dans le forage de Tomnatec (près de Lovrin), dans l'intervalle de 1708 à 1709 m à argilites charbonneuses, micacées, nous avons rencontré des lumachelles à *Viviparus sadleri* PARTSCH, *Limnocardium* sp., *Unio* sp. et *Bakunella dorsoarcuata* (ZAL.), *Candona (Caspiolla) balcanica* (ZAL.) etc.

On a prélevé aussi des associations macrofauniques du forage de Jimbolia (1300 à 1757 m); *Paradacna abichi* (HOERN.), *Plagiodacna subcarinata* (DESH.), *Limnocardium* sp. et *Dreissenomya* sp.

Le Pontien est présent aussi dans les forages de Timişoara par les paquets argilo-sableux (1000 à 1200 m) comportant: *Silicoplacentina hungarica* MEHES, *Valvata simplex* FUCHS, *Leptocythere lacunosa* REUSS, *Pontoniella* sp., *Hemicytherea* sp., *Paradacna* sp.

La partie supérieure de la formation de Sînnicolau Mare-Tomnatec est sableuse surtout. Dans un forage de Sînnicolau Mare on observe la suivante succession: entre 0 et 580 m il y a des sables en alternance avec des graviers et des niveaux marneux et argileux; de 580 à 1050 m les marnes et les marnes sableuses se trouvent en alternance avec les sables; c'est à la profondeur de 1050 à 1740 m que la suite de dépôts devient pélitique surtout, comportant des intercalations de charbons.

Patrulius et al. (1971, données non publ.) cite, de la base du paquet sableux du forage de Săcuieni, une macrofaune déterminée par Florian Marinescu: *Limnocardium (Limnocardium) decorum vicinum* FUCHS, *L. (L.) decorum decorum* FUCHS et *L. (L.) secans* FUCHS. Istocescu, Ionescu (1970) citent d'un horizon sableux les suivantes espèces déterminées par Iosefina Stancu: *Lithoglyphus* sp., *Caladacna* sp., *Viviparus* sp., *Bulimus* sp., *Valvata* sp. et *Prosodacnomya vutskitsi* (BRUS.).

Dans le forage de Cebza (Lovrin) on signale, dans les couches supérieures de la formation de Sînnicolau Mare, des associations de mollusques formées exclusivement de gastéropodes. Ainsi, dans les argiles grisâtres sableuses, dans l'intervalle de 845 à 846 m font leur apparition: *Bulimus* sp., *Valvata piscinalis subcarinata* Brus, *Dreissena* sp. Entre 850 et 860 m les argiles grisâtre-blanchâtres, fragiles à l'état sec, englobent: *Melanopsis (Melanopsis) decollata* STOL., *M. (M.) pterochilla* BRUSINA, *Valvata eugeniae gibulaeformis* BRUS., *Valvata* cf. *Valvata eugeniae* NEUM. et *Pseudamnicola* sp. Entre 969 et 970,50 m sont trésentes les argiles grisâtre-blanchâtres à cassure concoïdale, fossilifères à: *Viviparus* sp., *Limnocardium* sp. et *Paradacna* sp. Toute cette suite de dépôts repose au-dessus des argiles dures, litées, noirîtres, qui comportent, sur l'intervalle de 1890 à 1891 m, *Congeria banatica* R. HÖRN.

Dans les couches terminales de la formation de Sînnicolau Mare prédominent les gastéropodes et les ostracodes. Ainsi, dans le forage de Lovrin, sur l'intervalle de 20 à 100 m, le paquet sableux à intercalations argileuses renferme des mollusques et des ostracodes, tels: *Melanopsis* sp., *Viviparus* sp., *Gyraulus* sp., *Valvata* sp., *Candona lactea* BAIRD, *Candoniella albicans* (BRADY), *Cypridopsis ridua* MÜLLER, *Iliocypris brady* M. SARS, *Cyprinotus* sp. Une association faunique similaire est citée dans le forage de Sofronea, où à 990 m de profondeur on a rencontré: *Candona lactea* BAIRD, *Candoniella albicans* (BRADY), *Cypridopsis ridua* MÜLLER, *Iliocypris brady* M. SARS, *Cyprinotus karasi* et *Darwinulla* sp.

La formation de Sînnicolau Mare a plus de 3000 m d'épaisseur. Son contenu faunique témoigne la présence dans la Dépression Pannonique du: Pannonien s. restr.=Malvensien terminal, Pontien et éventuelement Dacien et Romanien.

Ainsi, les dépôts sarmatiens ou plus anciens sont recouverts transgressivement par un paquet d'argiles et grès à des épaisseurs variables, comportant *Congeria banatica* surtout, en association avec d'autres espèces de *Pontalmyra*, *Silicoplacentina hungarica* et ostracodes: *Amplocypris subacuta* (REUSS), *A. abscisa* (REUSS), *Candona (Caspiolla) elongata* SOKAĆ, *Hungarocypris hierogliphica* (MEHES), *Loxoconcha granifera* (REUSS), *L. hodonica* POK., *Pontoleberis hilata* (STANCEVA) et *Hemycytheria folliculosa* (REUSS).

Dans quelques forages (Mailat, Gerăuşa), à la limite Sarmatien/Pontien s. restr.=Malvensien, est présente une microfaune sarmatienne remaniée en des dépôts pannoniens à de nombreux ostracodes caractéristiques. Il ne s'agit pas d'une "faune de transition" étant donné les milieux de provenance des deux associations microfauniques, nettement différents (foraminifères: *Quinqueloculina* div. sp., *Elphidium* div. sp. et *Amplocypris* div. sp., *Candona* etc.). Ces faunes attestent l'érosion qui a eu lieu avant le Pannonien, tout comme le caractère transgressif du Pannonien s. restr.

Le Pontien basal est présent dans dés dépots percés par les forages de Sintana, Arad, Lovrin où ont rencontré de riches ostracofaunes du même âge: *Cyprideis triangulata* KRSTIĆ, *Candona (Pontalniella) unguiculum* (REUSS), *C. (Typhlocypris) ornata* OLTEANU, *Hungarocypris hierogliphica* (MEHES), *Amplocypris abscisa* (REUSS), *Loxoconcha granifera* REUSS), *Candonopsis arcana* KRSTIĆ, *Hemicytheria dubokensis* KRSTIĆ.

Le faciès charbonneux appartenant au Pontien de la Dépression Pannonique se caractérise par la prédominance des espèces de *Paradacna abichi*, *Congeria digitifera*, *Viviparus radleri* et ostracodes dont les plus fréquents sont: *Cyprideis pannonica*, *C. heterostigma*, *C. triangulata* en association avec *Candona* div. sp., *Bacunella dorsoarcuata* etc.

D'age pontien, probablement portaferrien sont aussi les associations de mollusques citées par Istocescu, Ionescu (1970) et Patrulius et al. (1972), données nonpubl., dont on remarque: *Limnocardium (Limnocardium) decorum decorum* FUCHS, *L. (L.) secans*, *Caladacna* sp., *Prosodacnomya vutskitsi* etc.

La partie supérieure de la formation de Sînnicolau Mare aurait un âge compréhensif, Pontien supérieur-Dacien+Romanien. L'association de mollusques rencontrée dans le forage de Cebza comporte *Melanopsis (M.) pterochilla* BRUS., *Valvata eugeniae gibbulaeformis* BRUS., *Valvata eugeniae* NEUM. et *Pseudammicola* sp. et indiquent le Dacien et le Romanien. Dans le secteur de Roşiori-Oradea le faciès charbonneux se mentient aussi à ce niveau. Ainsi, dans les forages de Lovrin et Sofronea on a identifié des associations d'ostracodes qui indiqueraient un milieu très adouci, étant présentes aussi dans le bassin Dacique, au niveau du Dacien et du Romanien: *Candona lactea* BAIRD, *Candoniella albicans* (BRADY), *Cyprinotus karasi* et *Darwinula* sp. Des associations fauniques similaires ont été rencontrées dans les forages de Sinersig, Visag et Darova (Banat).

Les investigations récentes de Simionescu et al. (1989) ont relevé que les formations quaternaires qui achèvent la sédimentation dans la Dépression Pannonique, au moins dans la zone de Lovrin, sont représentées par des dépots pléistocènes qui englobent une riche faune de mollusques (surtout gastéropodes, plus rare ostracodes et fragments de foraminifères.

Donc, nous avons argumenté, du point de vue biostratigraphique, l'existence, dans la Dépression Pannonique (le secteur roumain situé entre Pişcolţ-Tăşnad N et Jimbolia-Timişoara au sud), de quelques formations géologiques appartenant au Badénien, Sarmatien pannonique, Pannonien, Pontien, Dacien et Romanien.

Le travail met aussi en évidence la similitude lithologique et biostratigraphique des formations néogènes percées par des forages situés dans la Dépression Pannonique avec celles qui affleurent dans les ramifications entre les montagnes de celle-ci.

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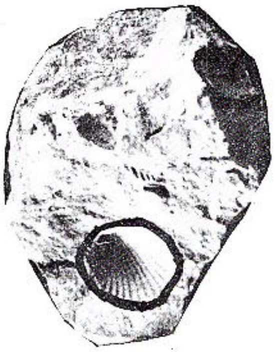

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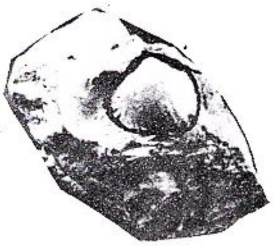

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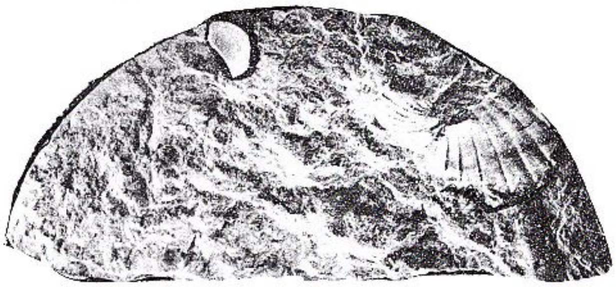

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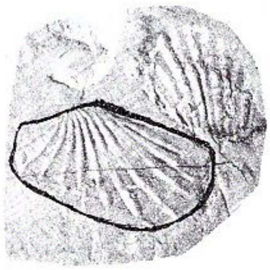

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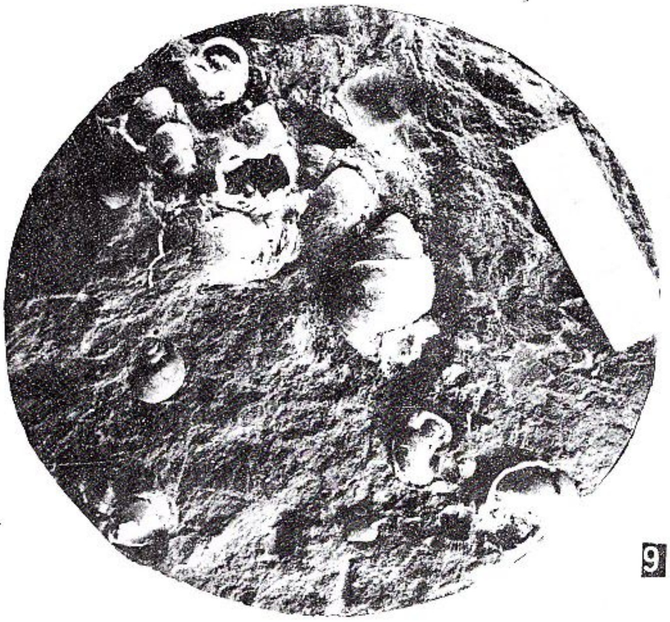

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Plate

Figs. 1, 2 - *Obsoletiforma (Obsoletiforma) Obsoleta* XI, 3, forage de Tăşnad, profondeur de 1040 à 1042 m, Sarmatien (Volhinien).

Fig. 3 - ? *Pseudocatillus* sp., (x 1, 7), forage d'Arad, 2439 m de profondeur, Pannonien s. rstr.=Malvensien.

Fig. 4 - *Limnocardium* sp. (x 1,5), forage d'Arad, 1110 à 1112 m de profondeur, Pontien.

Fig. 5 - *Paradacna abichi* (HOERN.), (x 1,5), forage d'Arad, 823 à 825 m de profondeur, Pontien.

Figs. 6, 7 - *Paradacna abichi* (HOERN.), (x 1), forage de Salonta, 1565 à 1566 m de profondeur, Pontien.

Fig. 8 - *Congeria digitifera* ANDRUSOV, (x 1,7), forage d'Arad, 1294 à 1297 m de profondeur, Pontien.

Fig. 9 - Lumachelle à *Viviparus sadleri* PARTSCH, (x 1), forage de Tomnatec, 1708 m de profondeur, Pontien.

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BADENIANUL DE LA MALU (BANATUL DE EST)

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S. C. "Prospecţiuni" S. A. Str. Caransebeş 1, 78344 Bucureşti 32.

Key words: Badenian. Mollusks. Echinoderms. Bryozoans. Foraminifers. Ostracods. South Carpathians – Intermontane depressions – Mehadia-Caransebeş.

Abstract: *Badenian Rocks at Malu.* The authors present a new Badenian outcrop in the Bistra Graben, at Malu (south-east of Bistra). The Malu fauna bears certain affinities with those at Balta Sărată (Caransebeş Depression), Var – area of passage between the Caransebeş Depression and the Bistra Graben, Bistra Valley, Bistra and Sarmisegetusa-Haţeg Basin. The micropaleontological assemblage is typical of the marine Lower Badenian, the Uppermost Langhian of the Paratethys-Candorbulina universa/Turborotalia bykovae Zone – upper Lagenidae – bearing Zone.

În apropiere de localitatea Malu, la aproximativ 6 km sud de Bistra (Banat), apare o nouă ivire de Badenian, care face obiectul notei de faţă.

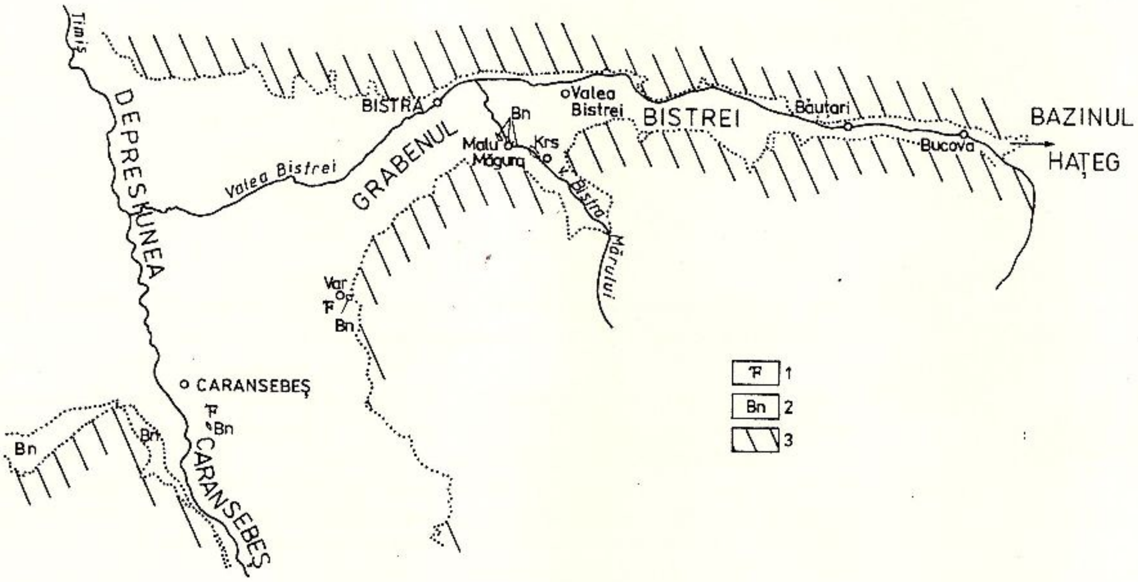
În grabenul Bistrei, Dincă şi Radu (1968) sînt primii care semnalează în vecinătatea comunei Valea Bistra (est de Bistra) un afloriment de gresii şi marne foarte bogat fosilifere, de vîrstă "tortonian superioară". Suita de strate se dispune după autori, peste o alternanţă de gresii tufogene, verzui, cenuşii şi albicioase ce aparţin Danian-Paleocenului.

Într-o zonă de legătură dintre grabenul Bistrei şi depresiunea Caransebeş, Lubenescu, Pavnotescu (1967, rap. nepubl., 1970) semnalează pe valea Var, un afloriment de marne albăstrui fosilifere, cu intercalaţii de marno-calcare albicioase, fosilifere, de vîrstă "Tortoniană superioară-Bugloviană".

Formaţiunile geologice badeniene descoperite de noi se dispun peste şisturi cristalofiliene, care sînt deschise în versanţii şi talvegul văii Bistra Mărului, la ieşirea din localitatea Măgura. La cca 5–600 m în aval de ivirea de şisturi cristalofiliene, în versantul stîng al văii, apar nisipuri masive, argiloase, cenuşii-albăstrui, micafere, nefosilifere, pe cca 4-5 m grosime. Aceleaşi depozite sînt deschise în versantul stîng al văii Bistra Mărului în dreptul bisericii din Malu. Ele conţin resturi indeterminabile de moluşte.

O ultimă ivire am întîlnit în aval, la cca 80 m, în malul stîng al unui fost braţ al Bistrei Marului. Aici aflorează marne cenuşii-vineţii compacte, nestratificate, fosilifere, deschise pe cca 1–1,5 m grosime. Asociaţia faunistică recoltată de noi este formată predominant din moluşte (mai ales gasteropode), fragmente de echinoderme, briozoare, corali şi foraminifere. Cităm: *Corbula (Varicorbula) gibba* (OLIVI), *Nucula (Nucula) nucleus* (LINNÉ), *N. (N.) sulcata* BRONN., *Glycymeris (Glycymeris) glycymeris pilosus* (DESH.), *Columbella* (ANACHIS) *zitelli* HOERN.u.AUING., *C. (A.) corrugata* BELL., *Lunatia catena helicina* BROCCHI, *Ancilla (Baryspira) glandiformis* (LAM.), *Turritella (Archimediella) erronea erronea* COSSM., *Clavatula jouaneti* DESMOULIN, *Mitrella (Macrurella) nassoides* (GRATELOUP.), *Conus (Chelyconus) fuscocingulatus* BRONN., *Nassa (Hinia) restitutiana* FONT., *Murex* sp.

Ansamblul microfaunistic este alcătuit din foraminifere planctonice şi bentonice, precum şi din ostracode, briozoare, spini de echinoderme reprezentate prin speciile: *Spiroplectinella carinata* (D'ORB.), *Spiroplectammina mariae* (D'ORB.), *Martinotiella communis* (D'ORB.), *Amphicoryna armata* (NEUG.), *Dentalina paronai* (DERV.), *Lenticulina cultrata* (MONTF.), *Lenticulina serpens* (SEG.), *Stilostomella elegans* (D'ORB.), *Plectofrondicularia medeligensis* (KARRER), *Alomorphina trigona* REUSS, *Baggatella subulata* (CUSHM. şi PARKER), *Bulimina inflata* SEG., *Uvigerina proboscidea* SCHWAGER, *Uvigerina longistriata* PERCONING, *Uvigerina semiornata* D'ORB., *Uvigerina venusta* FRANZENAU, *Valvulineria complanata* D'ORB., *Globorotalia (Turborotalia) bykovae* AIS., *Globigerina leroy* BLOW, *Zeoglobigerina druryi* (AKERS.), *Globigerinoides trilobus* (REUSS), *Globigerinoides sicanus* DI STEFANI, *Globoquadrina praealtispira* POPESCU, *Neoponides schreibersii* (D'ORB.), *Gyroidinoides soldanii* (D'ORB.), *Cibicidoides ungerianus*



Schiţa geologică a sectorului Carnsebeş-Bistra, după Harta geologică a R.S.R., sc. 1:200000, cu amplasarea unor formaţiuni badeniene. 1, formaţiuni de ramă; 2, formaţiuni badeniene; 3, puncte fosilifere.

(D'ORB.), *Heterolepa dutemplei* (D'ORB.), *Cytherella vulgata* RUGGIERI.

Asociaţia microfaunistică este tipică pentru Badenianul inferior marin, partea terminală a Langhianului din aria Paratethysului – zona micropaleontologică cu Candorbulina universa/Turborotalia bykovae – subzona superioară cu Lagenidae.

Fauna de la Malu prezintă o serie de afinităţi (în parte) cu fauna de la Balta Sărată (Pavnotescu et al., 1973, rap. nepubl.; Marinescu, Popescu, 1987), cu fauna de la Var (Lubenescu, Pavnotescu, 1967, date nepubl.,1970; Breban et al., 1988) şi cu fauna de la Valea Bistrei-Bistra (Dincă, Radu, 1968).

Alături de ivirea de Badenian de la Var, cît şi cea de la Valea Bistrei, noile iviri de la Malu, demonstrează pe baze paleontologice extinderea Badenianului inferior (Langhian) atît în Depresiunea Caransebeş, cît şi în grabenul Bistrei (fig.). Prezenţa Kossovianului a fost pusă în evidenţă (în grabenul Bistrei) pînă în prezent numai în forajul 1004 situat în apropiere de localitatea Var (Breban et al., 1988; date nepubl.).

Menţionăm că încă din 1901 Schafarzik paralelizează, pe considerente litologice, depozitele nisipoase-marnoase de la Bucova (est de Băuţari - grabenul Bistrei), cu cele din bazinul Haţeg de la Sarmisegetusa, bogat fosilifere şi atribuite Mediteranului II.

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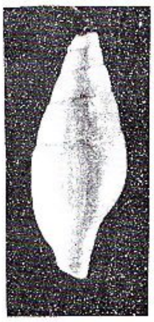

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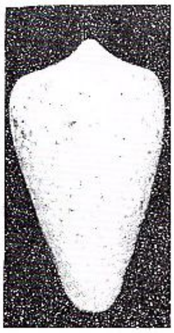

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Plate

Fig. 1 – *Glycimeris (Glycimeris) glicimeris pilosus* (DESH.). (x 1), Malu, Badenian.
Fig. 2 – *Turritella (Archimediella) erronea erronea* COSSM. (x 1,5), Malu, Badenian.
Fig. 3 – *Mitrella (Macrurella) nassoides* (Grateloup.) (x 1,6), Malu, Badenian.
Fig. 4 – *Conus (Chelyconus) fuscocingulatus* BRONN., (x 1), Malu, Badenian.
Figs. 5, 6 – *Clavatula jouanneti* DESMOULIN, (x 1,6), Malu, Baden.
Figs. 7, 8 – *Turritella (Archimediella) erronea erronea* COSSM., (x 11), Malu, Badenian.
Fig. 9 – *Lunatia catena helicina* BROCCHI, (x 1,6), Malu, Badenian.
Figs. 10, 11 – *Columbella (Anachis) corrugata* BELL., (x 1,6), Malu, Badenian.
Fig. 12 – *Nassa (Niotia) restitutiana* FONTANNES, (x 1,6), Malu, Badenian.

Rom. J. Stratigraphy, 1993, 75, p. 121-125

CONSIDERAŢII STRATIGRAFICE ASUPRA FORMAŢIUNILOR NEOGENE DIN BAZINUL CARANSEBEŞ (ZONA CARANSEBEŞ-TEREGOVA)

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S. C. "Prospecţiuni" S. A. Str. Caransebeş 1, 78 344 Bucureşti 32.

Key words: Neogene. Sedimentary basins. Stratigraphic units. Mollusks. Foraminifers. Biostratigraphy. Faults. South Carpathians – Intermontane depressions – Mehadia-Caransebeş.

Abstract: *Stratigraphic Data on the Neogene Formation of the Caransebeş Basin (Caransebeş-Teregova Area).* The present paper represents the synthesis of a rich geological material obtained by geological, geophysical prospections and explorations by mining and drilling works for coal. The processing of this material has made it possible for us to obtain a new stratigraphic image of the Neogene formations in the Caransebeş-Teregova area.

Lucrarea de faţă se bazează pe acumularea unui bogat material geologic obţinut prin prospecţiuni geologice, geofizice şi de explorare. Prelucrarea acestui material ne-a permis în final obţinerea unei imagini stratigrafice noi asupra formaţiunilor neogene din sectorul Caransebeş-Teregova.

Multiplele cercetări geologice efectuate în bazinul Caransebeş, pe mai bine de o sută de ani, au contribuit la elucidarea unor probleme stratigrafice privind formaţiunile de ramă şi fundament, cît şi a celor de bazin-neogene.

Într-o primă etapă cuprinsă între 1850-1949 se cunosc cercetări geologice sumare, cît şi lucrări miniere de explorare şi exploatare. Cităm: Hauer (1856), Halaváts (1880).

Etapa a doua se caracterizează prin lucrări de prospecţiuni geologice, geofizice, explorări prin lucrări miniere şi foraje pentru diverse substanţe minerale utile şi, în special, pentru cărbuni. Intre acestea amintim studiile şi documentaţiile întocmite de: Pop (1959), Iliescu (1956, 1974, rap. nepubl.), Iliescu et al. (1968), Lubenescu, Pavnotescu (1970), Radu, Pavnotescu (1968, 1969, rap. nepubl.), Voicu (1982), Marinescu, Popescu (1987), Petrescu et al. (1987), Radu (1988, rap. nepubl.).

Litostratigrafie

În sectorul Caransebeş-Teregova se recunosc în parte formaţiunile separate de Marinescu, Popescu (1987): formaţiunea de Rugi, formaţiunea de Delineşti, formaţiunea salmastră, formaţiunea de Valea Timiş şi formaţiunea de Turnu Ruieni.

După datele noastre se mai poate pune în evidenţă în cadrul formaţiunii de Delineşti o subformaţiune de Petroşniţa-Bela Reca (Kossovian), iar formaţiunea salmastră poate fi denumită formaţiunea de Sadova-Armeniş (Sarmaţian inferior şi mediu).

Formaţiunea de Rugi (Langhian)

Apare la zi atît pe rama vestică, pe cea estică, în zona centrală, la est de falia Balta Sărată-Armeniş între valea Armeniş şi valea Feneş, precum şi în lucrările geologice de cercetare cu foraje executate în zonă. Pe rama vestică, între Caransebeş-Petroşniţa, aceste depozite sînt reprezentate prin nisipuri şi pietrişuri, care la sud de Petroşniţa se împănează treptat cu conglomerate poligene, puternic cimentate pe care le substituie apoi, aproape în întregime, pe valea Ruianului şi pe valea Goleţ.

Pe rama estică, formaţiunea de Rugi a fost întîlnită între Borlova-Vîrciorova şi între Ilova- Feneş, unde se disting pietrişuri slab cimentate şi conglomerate poligene.

Prin lucrările de cercetare cu lucrări miniere şi foraje din sectoarele Zerveşti-Balta Sărată- Buchin, Vîrciorova, Goleţ, Feneş-Armeniş-Satu Bătrîn s-a constatat că depozitele formaţiunii de Rugi prezintă în general aceeaşi constituţie litologică. Spre centrul bazinului devin predominante gresiile, nisipurile cu intercalaţii subţiri de marne, argile şi tufuri dacitice groase de pînă la 0,50 m.

Grosimea formaţiunii de Rugi este cuprinsă între 50-180 m.

Atît în intercalaţiile pelitice ale nivelelor conglomeratice, cît şi în nisipurile grosiere sau microconglomerate s-a întîlnit o bogată faună de moluşte în apropierea localităţii Vîrciorova, pe Valea Pietroasă şi valea Poiana. Cităm: *Neopycnodonte navicularis* (BROCCHI), *Ostrea* sp., *Turritella (Archimediella) turris* SACCO., *Cerithium* sp. etc. (Radu, Pavnotescu, 1968; Lubenescu, Pavnotescu, 1970, rap. nepubl.).

Formaţiunea de Delineşti (Langhian-Kossovian)

Se dispune în continuitate de sedimentare peste gresiile şi conglomeratele bazale şi este reprezentată litologic prin pelite, strate de cărbuni, tufuri şi pelite cu o faună îndulcită (G[illegible], Vîrciorova, Ilova, Satu Bătrîn, Armeniş, Zerveşti, Balta Sărată, Buchin).

Astfel în sectorul Zerveşti-Balta Sărată-Buchin apar marne cenuşii cu rare intercalaţii subţiri de nisipuri gălbui-albicioase şi argile nisipoase, negricioase, cu mici fragmente de cuarţ rulat, strate de cărbuni şi tufuri bentonitice cu grosimi de 0,45-2,60 m.

În sectorul Armeniş-Feneş, formaţiunea de Delineşti este constituită din alternanţe de marne cenuşii, foioase, uşor grezoase, nisipuri gălbui, friabile, micacee, strate de cărbuni şi nivele de tufuri bentonitice cu grosimi cuprinse între 0,10–0,60 m.

În împrejurimile localităţii Goleţ remarcăm şi prezenţa unor nivele de calcare compacte cu grosimi de 0,10-0,15 m intercalate în complexul cu cărbuni şi tufuri bentonitice.

Asociaţia faunistică de moluşte cantonate în intercalaţiile pelitice ale formaţiunii de Delineşti este extrem de bogată şi variată. Menţionăm punctele fosilifere de la Balta Sărată, pîrîul Sadoviţa, Iloviţa etc., de unde s-a determinat: *Cardita (Cardita) partschi* GOLD., *Corbula (Varicorbula) gibba* OLIV., *Glycimeris pilosus deshayesi* (MAY), *Turritella* div. sp., *Strombus (Strombus) coronatus* DEFR., *Ostrea* div. sp., *Terebralia bidentata bidentata* DEFR., (Lubenescu, Pavnotescu, 1970; Pavnotescu et al., 1973; Radu et al., 1987, rap. nepubl.).

Faciesul marin a fost interceptat şi în foraje unde este reprezentat prin alternanţe de marne cenuşii, foioase, uneori compacte, marne slab consolidate, gălbui, gresii albăstrui-cenuşii, curbicorticale, tufuri.

În probele prelevate din forajele executate în zona Zerveşti-Balta Sărată-Buchin, Goleţ, Armeniş-Feneş-Satu Bătrîn s-a identificat zona cu lagenide, respectiv, zona Candorbulina universa/ Globorotalia bykovae. Asociaţia de foraminifere determinată cuprinde: *Uvigerina asperula*, *U. macrocarinata* PAPP şi TURN., *Elphidium crispum* (LINNÉ), *Orbulina suturalis* BRONN., *Candorbulina universa* D'ORB. (Radu et al., 1987, rap. nepubl.).

Din partea terminală a formaţiunii de Delineşti Constanţa Balteş identifică în foraje o asociaţie caracteristică zonei cu Velapertina, şi anume: *Bolivina dilatata* REUSS., *Valvulineria complanata* (D'ORB.), *Spirialis andrussovi* KITTL., *Spirialis hospes* ROLLES, *Gyroidinoides soldani* (D'ORB.), *Globocassidulina oblonga* (REUSS.), *Uvigerina venusta* FRANZENAU, *U. helicostata* SZ., *Allomorphina macrostoma* KARRER, *Gyroidina girardana* REUSS., *Bulimina inflata* SEG., *Baggatella subulata* CUSH. şi PARK., *Velapertina luczkowska* POPESCU.

Subformaţiunea de Petroşniţa-Bela Reca. La partea superioară a formaţiunii de Delineşti (în parte) sau direct peste şisturile cristaline (în sud) şi suportînd depozite sarmaţiene, am separat o suită de strate cu o mare varietate litologică, alcătuită preponderent din calcare organogene şi pietrişuri (Pietrişurile de Bela Reca, Iliescu et al., 1967, rap. nepubl.).

Astfel pe rama vestică, între Buchin, Petroşniţa şi Bucoşniţa, apar calcare albicioase, detritice, fosilifere, în bancuri de 0,20–1,5 m, cu intercalaţii de marne, argile şi gresii.

Pe valea Jurov aceste calcare stau direct pe şisturi cristaline, avînd în bază nisipuri gălbui- cafenii, grosiere şi pietrişuri poligene slab cimentate.

Către partea superioară a succesiunii apar intercalaţii de marne cenuşii şi calcare oolitice (sectorul Teregova-Armeniş-Sadova Veche). Suita calcaroasă trece lateral către centrul bazinului la est de falia Balta Sărată-Armeniş şi spre rama sa estică, la o succesiune detritică de nisipuri şi pietrişuri asemănătoare cu cea separată în Mehadia de Iliescu et al., 1967, rap. nepubl.) ca "Pietrişurile de Bela Reca". Pe rama estică s-au întîlnit pietrişuri, gresii cu intercalaţii fine de marne şi gresii conglomeratice.

În subformaţiunea de Petroşniţa-Bela Reca a fost întîlnită o bogată asociaţie de moluşte ce atestă prezenţa Badenianului: *Ostrea* div. sp., *Ancilla (Garyspira) glandiformis* (LAM.), *Echinocyathus stellatus* (CAPEDER), *Turritella bicarinata* D'ORB. etc. (Lubenescu, Pavnotescu, 1970).

Grosimea formaţiunii descrise atinge 80–150 m.

Formaţiunea de Sadova-Armeniş (Sarmaţian)

Depozitele bazale ale acestei formaţiuni se dezvoltă pe rama vestică între Goleţ şi Sadova Nouă. Între Armeniş şi Feneş se disting marne cenuşii nisipoase, friabile, micacee, fin stratificate, cu intercalaţii subţiri de nisipuri gălbui-albicioase, cenuşii şi gresii slab cimentate, micacee, fosilifere. Deasupra urmează un pachet predominant marnos-argilos cu intercalaţii subţiri de gresii cenuşiu-albăstrui şi lentile de pietrişuri.

Către zonele centrale şi pe rama de est a bazinului, la izvoarele văii Jurov, în depozitele bazale sînt intercalate argile şi marne cenuşiu-negricioase cu nisipuri gălbui-roşcate. Aceste argile, trec lateral la gresii cenuşiu-verzui-albăstrui. Întregul pachet de strate are un aspect rubanat.

Fauna de moluşte determinată de noi din depozitele bazale ale formaţiunii de Sadova Veche-Armeniş este destul de săracă şi reprezentată prin: *Abra reflexa* (EICHW.), *Ervilia trigonula* SOK., *Ervilia dissita* (EICHW.), *Acteocina lajonkaireana* (BAST.).

Peste marnele cenuşiu-verzui, în valea Armeniş, se dispune o serie detritică reprezentată prin pietrişuri polimictice, nisipuri gălbui-cenuşii, micacee, intercalaţii centimetrice de gresii cenuşii în bancuri decimetrice şi lentile cărbunoase. În continuare se dezvoltă marne şi argile fosilifere cu: *Obsolletiforma (Obsolletiforma) obsoleta vindobonense* (PARTSCH.), *Chartocardium ni rum* (ZHIZH.), *Ervilia dissita podolica* (EICHW.), *Paphia gregaria* (D'ORB.) (Lubenescu, Pavnotescu, 1967, rap. nepubl., 1970; Iliescu et al., 1971, rap. nepubl.).

Probele micropaleontologice din foraje conţin: *Anomalinoides dividens* LUCZKOWSKA, *Elphidium antoninum* (D'ORB.), *Miliolina seminulum* LINNÉ, *Miliolina akneriana* KARRER, *M. consobrina* D'ORB., *Ammonia beccari* (LINNÉ).

Grosimea suitei de strate descrise este de circa 200 m.

Menţionăm că în forajele executate în partea de nord a zonei cercetate a fost depistat şi un scurt episod cu depozite de precipitaţie chimică, reprezentat prin gipsuri breccioase cu intercalaţii de argile fosilifere cu: *Ervilia dissita* (EICHW.), *Abra reflexa* (EICHW.) şi foraminifere, avînd o grosime de 1–12 m (Petrescu et al., 1987; Marinescu, Popescu, 1987).

Partea superioară a formaţiunii de Sadova Veche-Armeniş debutează cu conglomerate polimictice, friabile sau slab cimentate, pietrişuri poligene şi nisipuri cenuşii sau cenuşiu-verzui, micacee cu intercalaţii subţiri de argile nisipoase cenuşii, fosilifere. În continuare se dispun marne şi argile cenuşii-vineţii, gălbui sau cu tente brun roşcate, bogat fosilifere: *Ervilia dissita podolica* (EICHW.), *Plicatiforma plicatofittoni* SINZOW., *Obsoletiforma obsoleta vindobonense* (PARTSCH.), *Mactra (Sarmatimactra) vitaliana* (D'ORB.), *Paphia ponderosa* (d'Orb.), *Pirenella* div. sp.

Stratele terminale ale formaţiunii de Sadova Veche-Armeniş sînt reprezentate prin argile şi marne argiloase stratificate în bancuri de 0,50–0,60 m, fosilifere, cu intercalaţii subţiri de nisipuri cenuşiu-gălbui, cu feţe limonitice şi pungi de pietrişuri cuarţoase. Subordonat, apar nivele de gresii cenuşiu-albăstrui cu feţe curbicorticale, dure, cenuşii sau cu intercalaţii de marnocalcare tari, aşchioase de 0,03–0,30 m (valea Satu Bătrîn).

Transgresiv se dispun depozite malvensiene.

În probele micropaleontologice prelevate din foraje s-a identificat o asociaţie de foraminifere cu: *Porosononion subgranosum* EGGER., *Ammonia beccari* (LINNÉ), *Miliolina consobrina* D'ORB., *Miliolina circularis* BRONN., ostracode şi statolite de myside: *Paramysis vancouveringi* VOICU şi *P. michaii* VOICU.

Grosimea suitei superioare a formaţiunii sarmaţiene este de 100–150 m.

Formaţiunea de Valea Timişului (= formaţiunea salmastră superioară; Malvensian mediu)

Separată în regiune de Marinescu, Popescu (1987), are un caracter transgresiv şi este reprezentată prin depozite detritice şi pelitice. Astfel, depozitele detritice sînt dezvoltate în două arii de sedimentare, sudică şi nordică. Aria sudică se extinde la sud de pîrîul Feneş, unde apar pietrişuri şi bolovănişuri poligene. În cealaltă arie de dezvoltare, care începe la nord de Satu Bătrîn, apare o suită de pietrişuri, conglomerate, nisipuri gălbui-reginii, cu intercalaţii de marne şi argile cenuşii şi verzui, compacte sau stratificate fosilifere cu *Congeria ramphophora* BRUS., *Melanopsis impressa* STRAUS şi *M. fossilis* (MARTINI-GMELIN).

Din succesiunea argilo-nisipoasă de pe valea Copaciului, Slatina Timiş şi Valea Ilovei s-a determinat o asociaţie de moluşte caracteristică Pannonianului s. rstr. mediu=Malvensian mediu, între care menţionăm: *Congeria banatica* R. HÖERN., *C. ramphophora* BRUS., *Melanopsis* div. sp., *Velutinopsis velutina* DESH. etc. (Lubenescu, Pavnotescu, 1967, rap. nepubl., 1970; Radu, Pavnotescu, 1968, rap. nepubl.).

Formaţiunea de Turnu Ruieni (= formaţiunea continental-fluviatilă; Pannonian s. rstr. superior-Ponţian inferi)

Este constituită predominant din pietrişuri şi nisipuri cu stratificaţie oblică de origine fluviatilă. Pietrişurile poligene sînt constituite din elemente de şisturi cristaline (gnaise, micaşisturi, cuarţite) cu diametrul de 5–10 cm, cu forme rulate şi subrulate şi bordonat angulare.

Nisipurile sînt de culoare gălbui-cenuşie, grosiere, foarte rar cu intercalaţii de marne şi argile de 0,10–0,30 m grosime. Din argilele nisipoase care aflorează în apropiere de localitatea Zerveşti au fost colectate: *Congeria* sp., *Paradacna lenzi* R. HÖERN. şi ostracode. În probele micropaleontologice s-au determinat speciile: *Bacunella dorsoarcuata* ZALL., *Caspyiocypris labiata* ZAL., *C. allata* ZALL., *Leptocythere multituberculata* (LIV.) etc.

Biostratigrafie

După cum arată Marinescu şi Popescu (1987), în formaţiunea de Rugi nu s-au întîlnit fragmente de organisme. În sectorul Caransebeş-Teregova într-un singur afloriment, la Vîrciorova, se disting asociaţii de moluşte badeniene. În formaţiunea de Delineşti au fost puse în evidenţă asociaţii de foraminifere care atestă prezenţa Langhianului inferior şi superior. Nu a fost evidenţiată baza Langhianului, şi anume zona cu Praeorbulina. După analizele micropaleontologice în forajele executate în sectorul Balta Sărată, în depozitele bazale ale Badenianului (peste seria detritică) s-a întîlnit nivelul inferior cu Cyprinide (facies îndulcit), faune marine (de moluşte), din nou un nivel cu Cyprinide şi foraminifere, suportat de un alt facies marin. În continuitate de sedimentare depozitele suprajacente conţin foraminifere care atestă prezenţa Kossovianului (zona cu Velapertina).

În stratele terminale ale formaţiunii de Delineşti, subformaţiunea de Petroşniţa-Bela Reca, fauna de moluşte întîlnită atestă, de asemenea, prezenţa Badenianului.

Formaţiunea de Sadova Veche-Armeniş conţine o faună salmastră caracteristică Volhinianului. A fost întîlnit nivelul cu *Abra reflexa, Ervilia trigonula, Anomalinoides*, miliolide etc.

Stratele terminale sarmaţiene ar putea aparţine Bessarabianului inferior. Astfel, pe lîngă moluşte ca *Mactra (Sarmati mactra) vitaliana* D'ORB., *Plicatiforma plicatofittoni, P. subfittoni*, întîlnite frecvent în Bessarabian apar şi asociaţii cu *Porosononion subgranosum, Paramysis* div. sp. şi ostracode.

Menţionăm că depozitele bessarabiene au o dezvoltare limitată, fiind probabil îndepărtate în mare măsură prin eroziune. Acest fapt a fost surprins şi în forajele din sectorul Balta Sărată-Zerveşti (Petrescu et al., 1987).

Formaţiunile pannoniene (Malvensian+Ponţian) din sectorul Caransebeş-Teregova conţin moluşte şi ostracode întîlnite frecvent în Pannonianul s. rstr. mediu + superior (Zona cu *Congeria banatica, Melanopsis* div. sp.). În partea terminală a formaţiunii de Turnu Ruieni asociaţiile cu ostracode determinate sînt, însă, ponţiene.

Consideraţii tectonice

Perimetrul Caransebeş-Teregova face parte din bazinul Caransebeş-Mehadia, cu orientare generală nord-sud, are o lăţime maximă în zona Caransebeş-Turnu Ruieni şi una minimă în zona Feneş, după care se lărgeşte din nou către sud.

Lucrările de prospecţiuni geologice, geofizice şi de explorare au condus la formarea unei imagini de ansamblu asupra bazinului, acesta conturîndu-se ca o structură depresionară asimetrică cu axul deplasat către est şi cu fundamentul împărţit în blocuri care se afundă treptat de la vest la est pe un sistem de falii nord-sud. În lungul bazinului se poate observa o ridicare a fundamentului în zona Turnu Ruieni-Caransebeş şi Armeniş-Satu Bătrîn şi coborîri ale acestuia ca urmare a unei subsidenţe accentuate, în zonele Vîrciorova şi sud Feneş-Teregova.

Din analiza întregului material geologic şi geofizic existent, s-au evidenţiat şase falii longitudinale, denumite de la vest la est astfel:

- falia Caransebeşul Nou-Jurov;
- falia Balta Sărată-Armeniş;
- falia vest Zerveşti-Bolvaşniţa;
- falia Zlagna-Bolvaşniţa;
- falia Borlova-Ilova şi
- falia Plugova.

Acest sistem de falii longitudinale este completat cu falii orientate est-vest, falii determinate prin lucrări geofizice sau prin lucrările de cercetare geologică însoţite de foraje şi lucrări miniere.

La nord de Caransebeş-Zerveşti, prin lucrările de prospecţiuni geofizice s-a identificat o falie orientată vest-est, falie care coboară fundamentul cristalin la 600 m.

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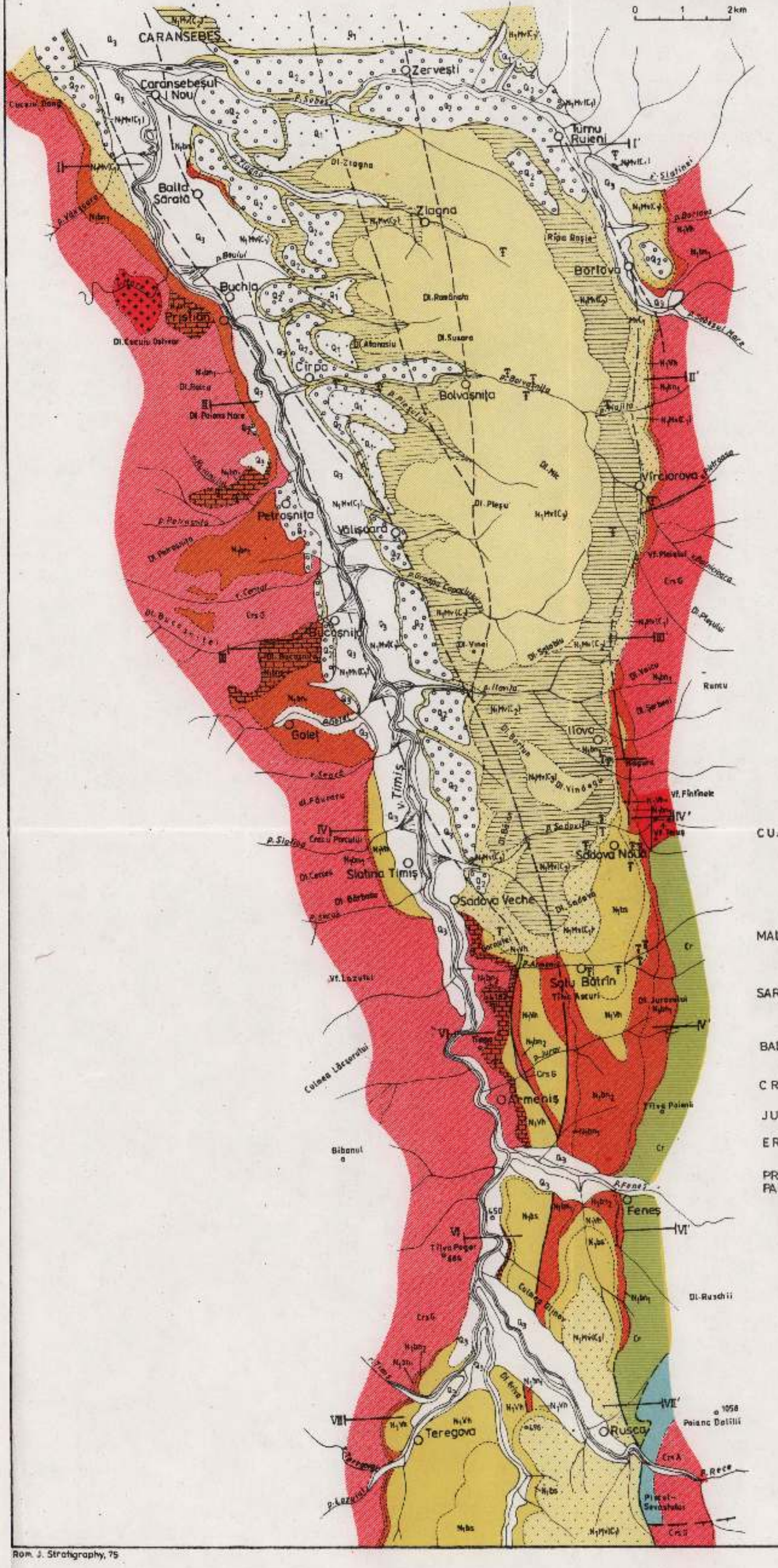
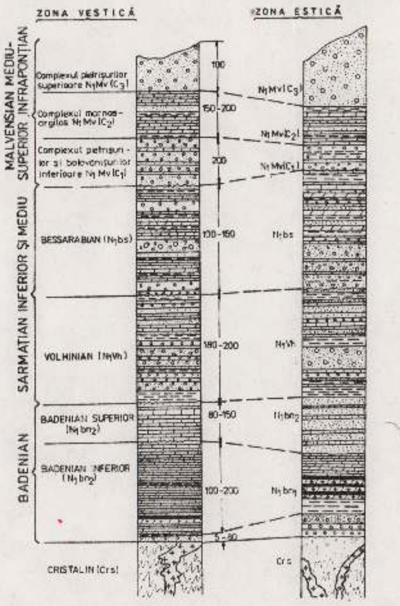
HARTA GEOLOGICĂ A FORMAŢIUNILOR NEOGENE DIN DEPRESIUNEA CARANSEBEŞ (SECTORUL CARANSEBEŞ-TEREGOVA)



COLOANE LITO STRATIGRAFICE



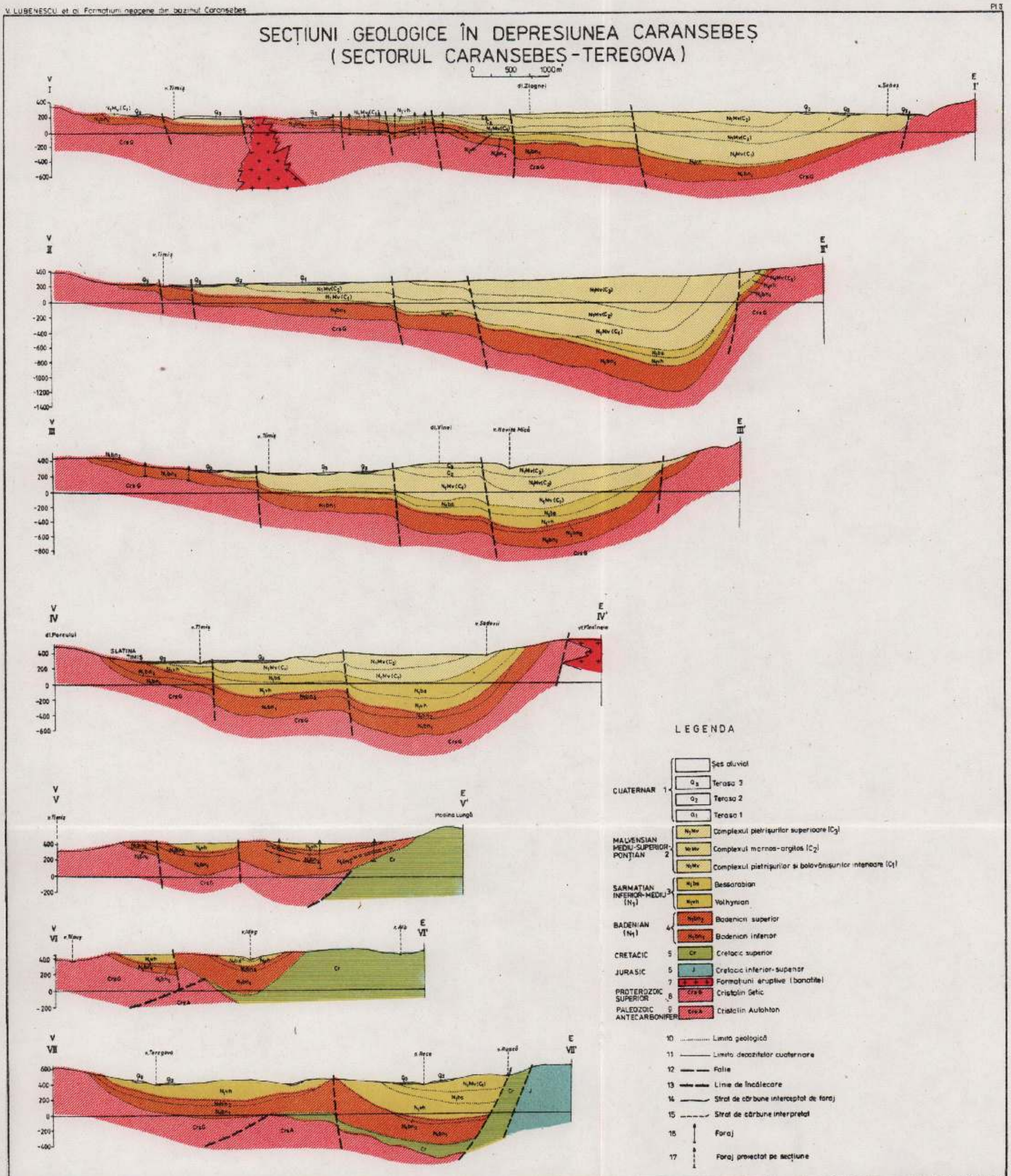
LEGENDA

- Calcar
- Marnă calcaroasă
- Marnă tufacee
- Marnă
- Argilă
- Nisip
- Gresie
- Pietriş
- Eruptiv
- Cristalin
- Strat de cărbune
- Limită de discordanţă

LEGENDA

CUATERNAR	1		Con de dejecţie
	2		Şes aluvial
	3	Q_3	Terasa 3
	4	Q_2	Terasa 2
	5	Q_1	Terasa 1
MALVENSIAN MEDIU-SUPERIOR-PONŢIAN	6	$N_1Mv(C_3)$	Formaţiunea de Turnu Ruieni
	7	$N_1Mv(C_2)$, $N_1Mv(C_1)$	Formaţiunea de Valea Timişului
SARMAŢIAN: BESSARABIAN, VOLHYNIAN	8	N_1bs, N_1Vh	Formaţiunea de Sadova Veche-Armeniş
BADENIAN: SUPERIOR	9	N_1bn_2	Subformaţiunea de Petroşniţa-Bela Reca
BADENIAN: INFERIOR	10	N_1bn_1	Formaţiunea de Rugi şi formaţiunea de Delineşti
CRETACIC	11	Cr	Cretacic superior
JURASIC	12	J	Jurasic inferior-superior
ERUPTIV	13		Formaţiuni eruptive (banatite)
PROTEROZOIC SUPERIOR-PALEOZOIC-ANTECARBONIFER	14	Crs G	Cristalin Getic
	15	Crs A	Cristalin Autohton
	16	T	Punct fosilifer
	17		Linie de încălecare
	18		Linie de falie
	19		Limită geologică
	20		Limita depozitelor cuaternare
	21	I—I'	Poziţia secţiunii geologice

Imprim. Atel. Inst. Geol. Geofiz.



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Rom. J. Stratigraphy, 1993, 75, p. 127–128

CONTRIBUŢII LA STUDIUL MOLUŞTELOR SARMAŢIENE DIN FORAJELE SĂPATE ÎN "GOLFUL" FĂGET (BANAT)

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S. C. "Prospecţiuni" S. A. Str. Caransebeş 1, 78344 Bucureşti 32.

Key words: Sarmatian. Mollusks. Biostratigraphy. Cores. Pannonian Depression. Adjacent depressions – Lipova Depression.

Abstract: *Contributions to the Study of Sarmatian Mollusks from Boreholes Drilled in the Făget "Gulf" (Banat).* Rich assemblages of Sarmatian mollusks – among which *Cardiacea* and *Mactra* lamellibranch prevail – have been found in the cores from certain boreholes drilled in the Făget "Gulf". Their study has made it possible for us to obtain biostratigraphic conclusions. The mollusks assemblages attest the presence of the Upper Volhynian and of the Lower Bessarabian. In this area, the Upper Bessarabian seems to be represented by *Mactra (Sarmatimactra) vitaliana* and *Plicatiforma latisulca* bearing beds.

În carotele prelevate din cîteva foraje, executate de I.P.E.G.-Banat în "golful" Făget, a fost întîlnită o bogată faună de moluşte, al cărei studiu face obiectul acestei note.

Date mai amănunţite asupra Sarmaţianului din regiune apar în nota întocmită de Mihăilescu et al., 1987. Cu această ocazie se menţionează şi asociaţii de moluşte sarmaţiene recoltate din foraje şi aflorimente. În nota de faţă, prezentăm moluşte sarmaţiene din foraje situate într-un sector vestic celui studiat anterior.

Forajul 32. Primele depozite sarmaţiene pelitice au fost întîlnite la 375 m. În partea bazală a acestei serii pelitice s-au întîlnit: *Ervilia dissita dissita* (EICHW.), *Paphia gregaria* (D'ORB.), *Plicatiforma plicata* EICHW., *Chartocardium nigrum* (ZHIZCH.), *Acmaea socenii* JEK., *Cerithium rubiginosum* (EICHW.), *Pirenella picta picta* (DEFR.), *Duplicata* sp. şi numeroase foraminifere. Între 325–370 m apare o bogată faună în care sînt frecvent: *Mactra (Sarmatimactra) vitaliana* D'ORB., *Plicatiforma plicata* (EICHWALD), *Plicatiforma latisulca* (MÜNSTER), *Obsoletiforma (Obsoletiforma) obsoleta obsoleta vindobonense* (PARTSCH), *Obsoletiforma (Obsoletiforma) obsoleta obsoleta* (EICHWALD), *Obsoletiforma (Obsoletiforma) obsoleta ghergutai* (JEKELIUS), *Obsoletiforma (Obsoletiforma) lithopodolica* (DUBOIS), *Paphia* cf. *Paphia ponderosa* (D'ORB.), *Hydrobia (Hydrobia) andrussowi uiratamense* KOL., *Hydrobia (Hydrobia) frauenfeldi suturata* FUCHS, *Valvata (Turrivalvata) sarmatica* PAPP, *Valvata (Turrivalvata) soceni wiesenensis* PAPP, *Acteocina (Acteocina) lajonkaireana* (BAST.), *Gibbula* sp.

În *forajul 33*, numai la 131 m, în argile cenuşii, s-au întîlnit gasteropode sarmaţiene: *Pirenella picta picta* (DEFRANCE), *Cerithium (Thericium) rubiginošum* (EICHWALD), *Theodoxus* div. sp., *Acteocina* sp., foraminifere div. sp.

Forajul 34. Între 250–260 m a traversat o succesiune de argile cenuşii şi verzui, fosilifere, cu nivele de nisipuri cenuşii. Moluştele identificate de noi sînt rare şi slab conservate: *Plicatiforma* sp., *Pirenella* div. sp., *Hydrobia* div. sp., *Ervilia dissita podolica* (EICHWALD) etc.

În continuare, pînă la 240 m a fost străbătută o alternanţă de argile cenuşii-verzui şi argile albicioase cu nivele de nisipuri fosilifere. Dintre moluştele prezente în asocioaţie faunistică, mai frecvente sînt: *Mactra (Sarmatimactra) vitaliana* D'ORB., *Plicatiforma latisulca* (MÜNSTER), *Plicatiforma plicata* (EICHWALD), *Paphia* ex gr. *Paphia ponderosa* (D'ORB.), *Pirenella picta picta* (DEFR.), *Cerithium (Thericium) rubiginosum* EICHWALD, *Acteocina (Acteocina) lajonkaireana* (BAST.) şi numeroase ostracode.

Forajul 35. A interceptat Sarmaţianul în întregime, de la limita sa cu Badenianul, pînă la limita sa cu Pannonianul s. rstr.=Malvensian. Astfel, la 390–392 m se disting argile cenuşii- albicioase fosilifere de vîrstă badeniană. Ele conţin fragmente de Ostrea, diverse alte moluşte indeterminabile, scaphopode etc. În continuare se dispune o secvenţă detritică de pietrişuri, nisipuri cenuşii şi nivele de argile nefosilifere, care suportă argile cu intercalaţii de nisipuri cenuşii micacee. Ele conţin moluşte ce atestă prezenţa sarmaţianului: *Pirenella* div. sp., *Plicatiforma* sp., *Paphia* sp. O bogată faună a fost remarcată între 249–255,40 m, în argile cenuşii: *Plicatiforma plicata* (EICHWALD), *Plicatiforma latisulca* (MÜNST.), *Chartocardium nigrum* (ZHIZHCHENKO), *Obsoletiforma (Obsoletiforma) lithopodolica* (DUBOIS), *Paphia vitaliana* (D'ORB.), *Mytilaster* sp.

Între 225–242 m apar: *Modiola navicula* DUB., *Paphia gregaria* D'ORB., *Plicatiforma latisulca* MÜNST., *Mactra (Sarmatimactra) vitaliana* D'ORB., *Musculus sarmaticus* GAT., *Modiola incrassata* D'ORB.

Deasupra, în intervalul cuprins între 198,50–225,00 m este întîlnită o succesiune de argile cenuşii fosilifere cu intercalaţii fine de nisipuri. Din fauna de moluşte foarte bogată s-au determinat: *Mactra (Sarmatimactra) vitaliana* D'ORB., *Plicatiforma plicatofittoni* (SINZOV), *P. latisulca* (MÜNST.), *Paphia ponderosa* (D'ORB.), *Mytilaster incrassatus* D'ORB., *Hydrobia* div. sp., *Gibbula (Rolandiana) picta* (EICHW.), precum şi numeroase ostracode.

Analizînd asociaţia de moluşte de mai sus, constatăm că în partea superioară a suitei de depozite sarmaţiene străbătute de foraje apar frecvent: *Mactra (Sarmatimactra) vitaliana, Paphia ponderosa, Plicatiforma latisulca* şi *P. plicatofittoni*, specii cunoscute frecvent la nivelul Bessarabianului inferior. Partea bazală a formaţiunii ar putea fi atribuită Volhinianului inferior.

Un preţios ajutor în scopul obţinerii unor precizări suplimentare va fi oferit şi de studiul microfaunei, care este în curs.

După cum remarcă Kojumdgieva (1987, p. 4), grupul de moluşte cel mai variat din fauna endemică salmastră şi care se întîlneşte în Volhynian şi Bessarabian este cel al cardiidelor. Autoarea propune schema filogenetică şi a distribuţiei stratigrafice şi faciale a genurilor de cardiide sarmaţiene: *Plicatiforma, Planacardium, Obsoletiforma, Inaequicostata* şi *Chartocardium*. Această clasificare generică şi subgenerică a fost folosită şi de noi cu unele amendamente asupra dispoziţiei stratigrafice a unor moluşte.

Am remarcat că o frecvenţă deosebită o are *Chartocardium nigrum* (ZHIZCH.), specie descrisă sub mai multe nume de către diverşi autori (*Cardium niger* ZHIZCH., 1954; *Cardium barbotensis* ATANASIU, MACAROVICI, 1950; *Cardium gleichenbergense* PAPP, 1954; *Cardium transcarpaticum* GRISK. in Merklin şi Neveskaia, 1955).

În "golful" Făget, exemplarele întîlnite nu au o variabilitate intraspecifică deosebită. Numai cîţiva indivizi au un număr de coaste ceva mai mare, dar îşi menţin aspectul morfologic. Asupra poziţiei stratigrafice a acestei specii credem că ea poate depăşi limita Volhynian/Bessarabian cu toate că apariţia şi frecvenţa sa mare este în Volhynianul superior.

– *Plicatiforma latisulca*, specie cu o mare variabilitate intraspecifică (Kojumdgieva, 1987) apare bine individualizată la limita Volhynian/Bessarabian şi chiar în Bessarabianul inferior. Unele exemplare studiate ating peste 3 cm diametru anteroposterior şi au cochilii robuste şi bine conservate. Am întîlnit numeroase cochilii de indivizi tineri şi adulţi.

– *Plicatiforma plicatofittoni* este o specie rară în Sarmaţianul din regiune (la nivelul Bessarabianului).

– *Plicatiforma plicatum* are o mare frecvenţă numai în forajele 2 şi 5 unde se disting exemplare tinere şi adulte, slab conservate. Remarcăm marea variabilitate intraspecifică a speciei. Am întîlnit exemplare cu contur rotunjit şi cu contur oval.

– *Mactra (Sarmatimactra) vitaliana* (D'ORB.) are o frecvenţă şi o talie mare în stratele terminale ale suitei sarmaţiene. Au fost întîlnite cochilii cu contur triunghiular şi alungit (pl. I, fig. 9, 10). A fost atribuită subgenului "Sarmatimactra" conform schemei întocmită de Ionesi (1986).

În ansamblu, menţionăm că moluştele studiate caracterizează partea terminală a Sarmaţianului pannonic. Depozitele bessarabiene au fost în mare parte îndepărtate prin eroziune. În "golful" Făget, în forajele studiate, nu a fost surprins Volhynianul bazal.

Apariţia la diferite nivele în foraje a formelor de moluşte se datorează jocului pe verticală a unor blocuri tectonice sau apropierii de rama golfului Făget.

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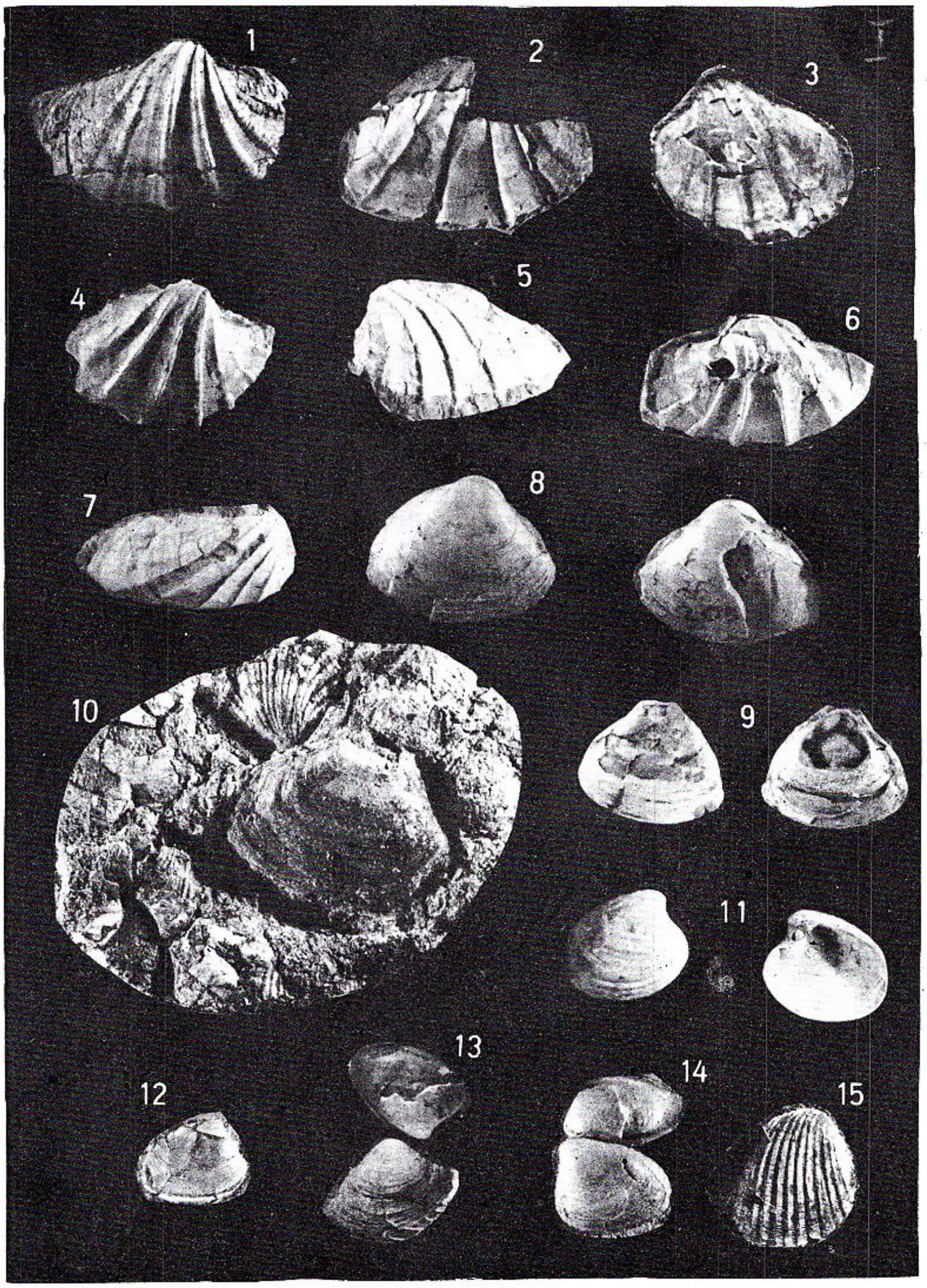
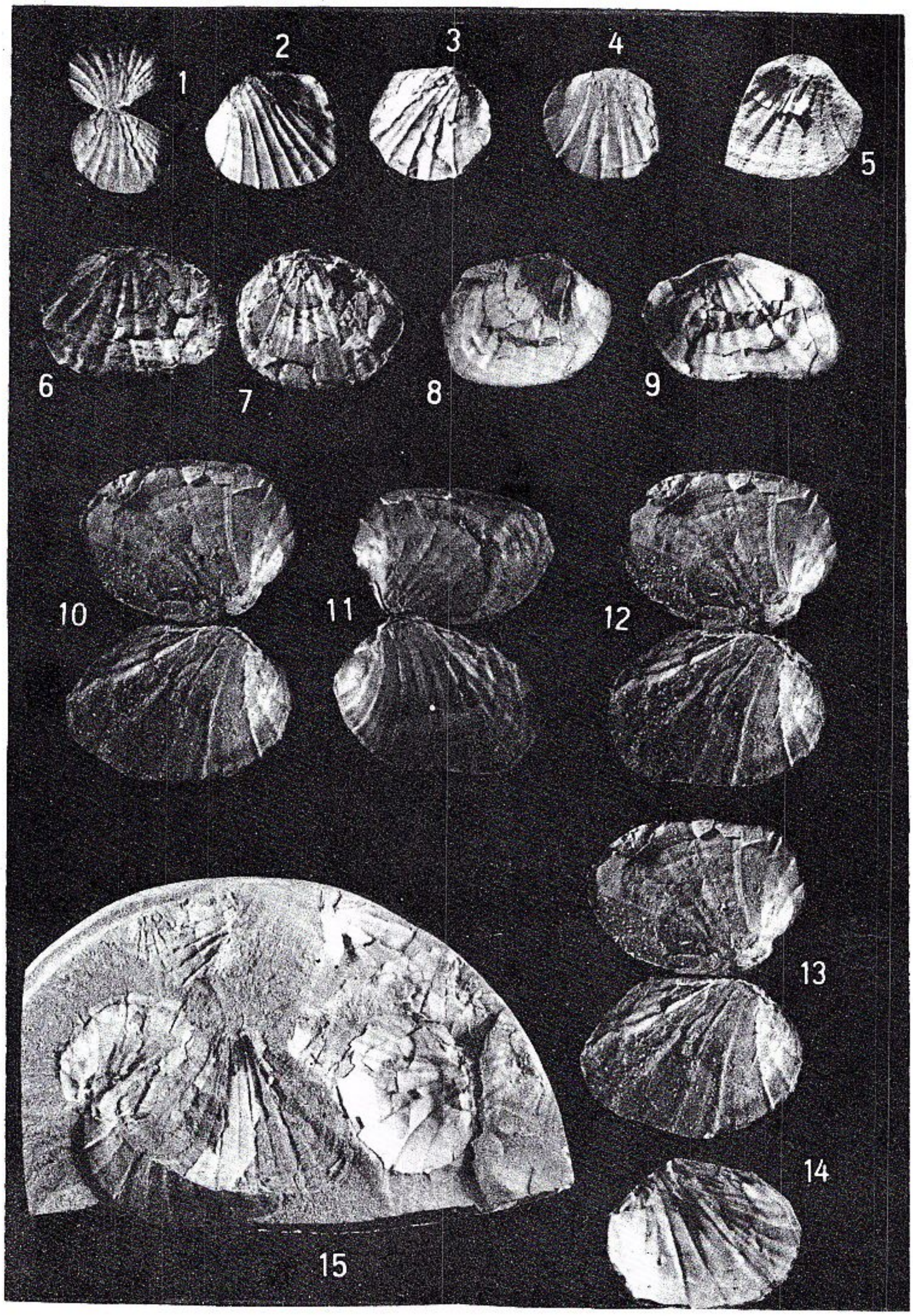


Planşa I

Figs. 1 – 7 – *Plicatiforma latisulca* (MÜNST.), (x 1), Bessarabian inferior, fig. 1–4, forajul 5, adîncime 198,50–205 m.
Figs. 8 – 10 – *Mactra (Sarmatimactra) vitaliana* (D'ORB.), (x 1), Bessarabian inferior, forajul 5, adîncime 198,50–205 m.
Figs. 11 – 14 – *Paphia* ex. gr. *Paphia gregaria* (PARTSCH), (x 1), forajul 3, adîncime 173 m. Volhynian superior-Bessarabian inferior.

Fig. 15 – *Obsoletiforma (Obsoleta) obsoleta ghergutai* JEK., (x 1), forajul 3, adîncime 249 m. Volhynian superior.



Planşa II

Figs. 1 – 4 – *Plicatiforma* sp. (x 1), Volhynian superior, forajul 2, adîncime 372 m.
Figs. 6 – 8 – *Plicatiforma* ex. gr. *Plicatiforma plicata* MÜNST.), (x 1), Volhynian superior, forajul 2, adîncime 372 m.
Figs. 10 – 15 – *Chartocardium nigrum* (Zhizhchenco), (x 1), Volhynian superior, forajul 2, adîncime 372 m.

Rom. J. Stratigraphy, 1993, 75, p. 129-130

PREZENŢA UNOR DEPOZITE CU MATERIAL VULCANIC ÎN CUATERNARUL DIN SUDUL PODIŞULUI MOLDOVENESC

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Key words: Quaternary. Sand. Volcanic ash. Cinerite. Moldavian Plateau – Zone between the Siret and Prut Rivers – Southern sector.

Abstract: *Presence of Volcanic Material in the Quaternary Deposits from the Southern Part of the Moldavian Plateau.* The note presents the results of the mineralogical analysis carried on heavy minerals in the Pleistocene alluvial deposits in the southern part of the Moldavian Plateau. The great frequency of slightly rounded volcanic minerals is considered to be due to the erosion, the transport and the redeposition achieved by the Siret and the Bîrlad rivers.

Intercalaţii de roci piroclastice în formaţiunile sedimentare din Podişul Moldovenesc au fost evidenţiate pentru prima dată în 1911 de Athanasiu, care descrie "cenuşe şi gresii andezitice" în Sarmaţianul superior, pe malul drept al Siretului, la sud de Bacău.

Enculescu (1911) descrie depozite similare pe care le atribuie aceluiaşi interval stratigrafic, pe malul sting al Siretului, la Parincea şi Bibireşti.

Ulterior, Sevastos (1920) menţionează extinderea tufurilor andezitice între rîurile Siret şi Tutova, atribuindu-le Meoţianului.

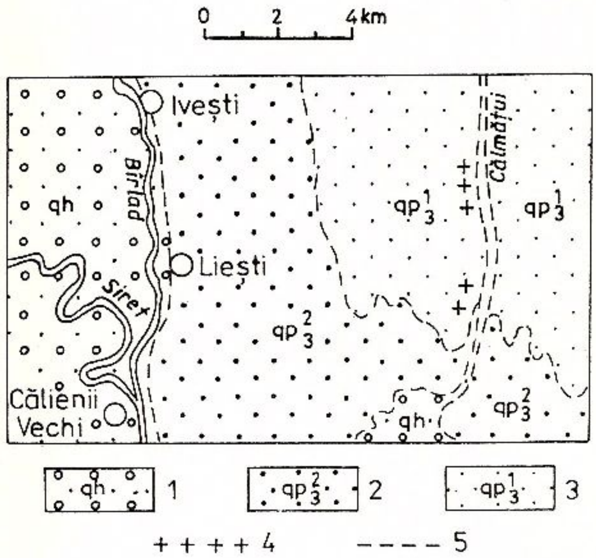
Macarovici (1956) aduce completări asupra grosimii, originii şi vîrstei cineritelor din aflorimentele semnalate de autorii precedenţi.

Jeanrenaud (1966) consideră cineritele meoţiene de Nuţasca-Ruseni ca orizont reper pentru trasarea limitei Sarmaţian-Meoţian. Orizontul cineritic ocupă suprafeţe mari, ajungind în nord pînă aproape de Iaşi, iar în sud pînă la o linie ce ar trece pe la nord de Bîrlad şi de localitatea Podu Turcului.

Prezenţa nisipurilor piroclastice în depozite exclusiv de vîrstă meoţiană este contestată de către Ghenea (1965, 1968), care menţionează material vulcanic în depozite fosilifere atribuite Sarmaţianului superior. Autorul extinde aria de răspîndire a nisipurilor piroclastice pe interfluviul Bîrlad-Prut pînă la paralela oraşului Bîrlad.

Din acest scurt istoric se desprinde clar faptul că în Podişul Moldovenesc, în intervalul Sarmaţian superior-Meoţian, s-au acumulat, ca urmare a activităţii vulcanice din lanţul muntos Călimani-Harghita, depozite cineritice cu grosimi importante, care reprezintă o particularitate geologică a acestei unităţi.

In cele ce urmează considerăm interesantă apariţia unor depozite cu material vulcanic în extremitatea sudică a Podişului Moldovenesc, în formaţiuni de vîrstă cuaternară. Astfel, pe Valea Călmăţuiului, la nord-est de confluenţa Bîrladului cu Siretul şi est de localitatea Lieşti, apar aflorimente în formaţiuni pleistocene din alcătuirea sistemului aluvionar Siret-Bîrlad (fig.).



Aflorimente cu material vulcanic. 1, Holocen (qh), nisipuri şi pietrişuri din alcătuirea luncilor; 2, Pleistocen superior (qp_3^2), aluviuni grosiere din alcătuirea terasei de 10-15 m; 3, Pleistocen superior (qp_3^1), aluviuni din alcătuirea terasei de 30-40 m; 4, nisipuri cu material vulcanic; 5, limită geologică-morfologică.

In zona amintită au fost identificate două terase generate de acţiunea comună a Siretului şi Bîrladului, un nivel inferior cu altitudine relativă de 10–15 m, acoperit de nisipuri de dune, şi un nivel superior, cu altitudine relativă de 30–40 m, ambele cu extindere mare spre nord.

Deschiderile menţionate au fost urmărite pe malul drept al Călmăţuiului, în zona lacului de acumulare situat la est de Lieşti. In acest sector, torenţii laterali au deschis în versant formaţiunile aluvionare din alcătuirea terasei superioare Siret-Bîrlad. Apar, astfel, nisipuri grosiere pînă la mediogranulare, de culoare gălbuie şi cu stratificaţie paralelă evidenţiată de existenţa unor benzi brun-roşcate, mai cimentate.

La circa 2 km în aval de lacul de acumulare, într-o deschidere situată la punctul denumit "Movila Chifului" aflorează pe circa 4 m grosime aceste nisipuri care, macroscopic, atrag atenţia prin frecvenţa mare a elementelor melanocrate.

Analizele mineralogice executate în laboratorul de minerale grele din Institutul de Geologie şi Geofizică (mulţumim pe această cale dr. A. Popescu) au pus în evidenţă o cantitate mare de minerale de origine vulcanică, dintre care hornblenda verde ajunge la procente ridicate (50 %). Subordonat se recunoaşte augitul.

Mineralele vulcanice îşi mai păstrează contururile prismatice iniţiale, remarcîndu-se o slabă rotunjire a acestora, dovadă a reluării şi transportului pe o distanţă relativ redusă. In rest, asociaţia de minerale grele se caracterizează prin prezenţa granatului, epidot-zoizitului, rutilului, staurolitului, de provenienţă cristalofiliană.

Existenţa nisipurilor pleistocene cu material vulcanic în extremitatea sudică a Podişului Moldovenesc necesită sublinierea următoarelor concluzii:

– provenienţa produselor vulcanice este legată de existenţa formaţiunilor cineritice, bine dezvoltate în intervalul Sarmaţian superior-Meoţian;

– transportul mineralelor de origine vulcanică s-a făcut de către Siret şi Bîrlad, care în această perioadă aveau, probabil, un curs comun în zona de confluenţă, aşa cum dovedesc terasele comune pleistocene din sudul Podişului Moldovenesc.

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REMARKS ON THE CORBICULA FLUMINALIS BEDS IN THE QUATERNARY FORMATIONS OF THE ROMANIAN PLAIN

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Key words: Quaternary. Deposition. Biostratigraphy. Mindel-Riss Interglacial. Romanian Plain.

Abstract: On the occasion of finding out certain beds containing *Corbicula fluminalis*, an analysis is made on the stratigraphic position of this form in the Quaternary Formations of the Romanian Plain. The conclusion is drawn that the deposits concerned should be assigned to the last but one interglaciation.

In the Quaternary formation of the Romanian Plain several sites have been mentioned where *Corbicula fluminalis* remains have been found. As there exist certain confusions concerning the correlation of the deposits containing this form, we have decided to make certain specifications on the stratigraphic position of the species *Corbicula fluminalis*. The conclusion has also been occasioned by a recent outcrop found during certain geological researches in the central part of the Romanian Plain, in the zone of the Prahova Valley-Ialomiţa Valley confluence.

In this area, the alluvial Prahova Plain is clearly delimited to the north-west, a vast area of meandering in which the river, because of an intense activity of erosion and accumulation that characterized at least the Upper Pleistocene and the Holocene, led to the formation of a low plain, made up only of recent deposits. Their structure can be followed in the bank of the Prahova River, facing the village of Patru Fraţi where, on about 2–2.5 m there occur new alluvia of the Prahova River, in at last three horizons of finely grained sands, separated by alluvial soils (organic clays, blackish in colour) with frequent vegetal remains.

The age of the Prahova alluvia in the flood plain of the Prahova-Teleajen Rivers is also proved by the 3860±90 years BP age obtained on a wood fragment collected from alluvial sands and assigned by C^{14} by professor M. Stuiver from the Washington University in Seattle.

But the erosion of the Prahova southwards has led to an erosion activity also in the higher plain situated on the right side of the Prahova River. In the zone of confluence of the Prahova River with the Ialomiţa one, the older geological sequence of the structure of this plain is evidenced by a series of outcrops at whose upper part there appear loesses as non-structured yellowish clayey silts, that are quite compact (Fig.).

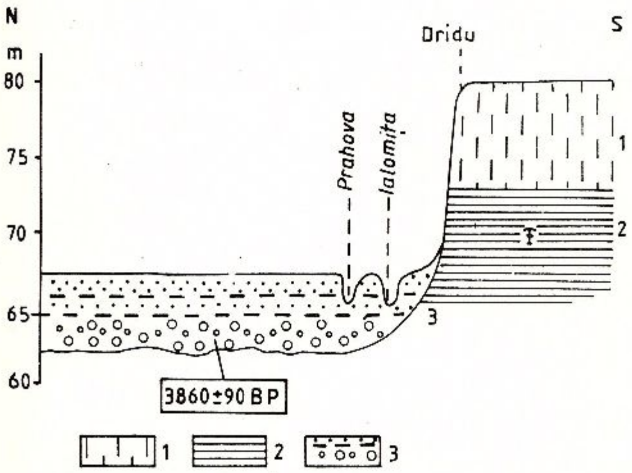

Fig. Geological sequence of the structure of the Romanian Plain.
1, loess; 2, clays; 3, alluvia.

At the lower part of the loess, where the erosion of the Ialomiţa and the Prahova Rivers was greater, there is an outcrop in horizon of grey, sometimes blackish clays. In an outcrop at the western border of the Dridu village, this clayey horizon contains *Corbicula fluminalis* MÜLL. (very frequent), *Unio* sp., *Valvata piscinalis* MÜLL., *Valvata naticina* MEW, *Planorbis corneus* L., *Tropidiscus planorbis* L., *Lymnaea palustris* MÜLL etc.

For appreciating the stratigraphic position of this horizon, certain data concerning the presence of the form *Corbicula fluminalis* in the Quaternary deposits of the Romanian Plain are mentioned. It was mentioned by Sabba Ştefănescu (1896) in the Mărculeşti borehole, between depths 29 and 72 m in sands and gravels assigned to the Quaternary. Researches carried out by Liteanu (1952) point out important elements concerning the stratigraphic level at which *Corbicula fluminalis* is situated. Their presence in the Frăteşti Beds in the underground of the Bucureşti zone seems to make them be assigned to the Lower Pleistocene. The mentioned author also cited them at Balta-Potcoava, in deposits described as the "Marly Complex", a term that has subsequently been largely accepted and would include the interval of the two glaciations, Mindel and Riss, and the Mindel-Riss Interglaciation (1953). Liteanu cites *Corbicula fluminalis* too in the Brăila and Zagna-Vădeni Zone, in deposits that would represent the transition between the Frăteşti Beds and the deposits at Barboşi, assigned to the Middle Pleistocene. In the paper treating about the presence of the species *Corbicula fluminalis* in the Holocene of the Danube Basin, Liteanu et al. (1961) points out this form in several sites in Recent alluvia of the Danube Delta, considering it has no sure stratigraphic value.

At the end of the presentation of the sites where it has been cited, the I. C. Frimu-Obileşti outcrop is mentioned, where Coteţ and Prisnea (1957) describe the "Mostiştea Complex" containing *Corbicula fluminalis*, which it considered of Mindel-Riss age.

Considering what has been presented, *Corbicula fluminalis* seems to be placed in the deposits referred to the Middle Pleistocene, but equally in Holocene formations. Referring to the Holocene age of this species, it is mentioned to live at present in Iran, Afganistan, Syria, India, Transcaucazia as well as the Danube Delta and the marshes around the Black Sea (Ložek, 1954; Liteanu, 1961). The specimens found at Dridu have an obviously lower stratigraphic position (under the loess). Besides, their shells have been analyzed by C^{14} in the Seattle Laboratory of absolute ages, the analyses indicating an age older than 40,000 years B.P.

In this situation, it is the stratigraphic position of the species from the Pleistocene formations in Europe that is of interest.

The thermopile form *Corbicula fluminalis* characterizes the warm periods of the Middle Pleistocene in Europe. It is cited in the Holsteinian Interglaciation, whose revised stratotype indicates at the Lower boundary a climate optium corresponding to the passage from sub-arctic conditions to boreal ones, just like in the deposits of the terraces in Czechoslovakia, situated at the upper part of the Holsteinian Interglaciation (Ložek, 1964).

Soviet literature contains more exact data concerning the stratigraphic position of this species. Although the migration of the genus *Corbicula* is mentioned since the beginning of the Middle Pleistocene, the largest development of the fauna containing *Corbicula fluminalis* is mentioned in the Lihvinian Interglaciation (Mindel-Riss Interglaciation), when the area of distribution of this subtropical form extends up to 55–60^0 north latitude (Cepaliga, 1982).

The synthesis of the data obtained points out that the clayey deposits in the Romanian Plain, in the Dridu Zone, containing *Corbicula fluminalis*, are assigned to the Mindel-Riss Interglaciation. This specification raises the problem of the future revision of the stratigraphic position of the whole complex of the fluvial-lacustrine beds that filled the area of the Romanian Plain all along the Quaternary.

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NEW DATA ON THE AGE OF THE LOESS IN DOBROGEA

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Key words: Quaternary. Loess. Paleosols. Absolute age. Paleomagnetism. Magnetic declination. Magnetic inclination. Magnetization. Thermoluminiscence. Dobrogea.

Abstract: For paleomagnetically dating the loess of Dobrogea (a unit where it extends on a large surface) two areas have been selected, characterized by complete sections. The results have shown normal magnetization associated with the Brunhes epoch, from the surface up to the loess bottom. The paleomagnetic data have been partially correlated equally with absolute ages obtained by thermoluminiscence, which have shown the age of 138,000 ± 20,000 years for paleosol III and the age of 650,000± 90,000 years for paleosol VI.

Loess extends on large areas in Dobrogea, covering almost completely the southern zone, in the central part developing on quite large areas at the upper part of the green schists, and in the northern region of Dobrogea being present especially in the Danube Valley zone. Because of its great extension, loess has been of interest for specialists for a long time. More systematic researches were carried out by Brătescu (1934, 1935) who, after examining certain cross-sections between Constanţa and Eforie, has assigned the loess horizons to the glacial periods and the fossil soils to the interglacial ones. Subsequently, pedogenetic criteria become determinant for the relative dating of paleosols. Pedochronologic analyses were used by Hasse and Richter (1951), Popovăţ et al. (1964), Conea (1970) for certain sections in South Dobrogea. Interpreting the types of paleosols, a general picture of the climate in Dobrogea has been obtained, in which brown-red clayey paleosols with very well developed profile correspond with the interglacial periods and the chernozem-like fossil soils with the Würm interstages. Starting from the identification of seven fossil soils that interrupt the loess series, the stratigraphic diagram obtained has placed the last paleosol (Ps VII) in the Günz-Mindel interglacial period; the upper loess horizons (L I, L II, L III) have been referred to the Würm, the intermediary loess to the Riss, and two basal horizons of altered loess to the Mindel glacial period (Conea, 1970).

The problem of the age of the loesses in Dobrogea is discussed, taking into account either the Paleolithic inventory mentioned in certain sections in South Dobrogea (Valoch, 1968; Păunescu et al., 1972; Ghenea, Codarcea, 1974; Circiumaru, Păunescu, 1976) or the fossil mammal faunas studied in the Adam Cave (Samson, Rădulescu, 1961). The archeological and the paleontological data at our disposal generally refer to a more recent period of loess deposition in Dobrogea.

Recently, the Quaternary formations have been dated using more and more the new techniques of absolute age datings which have replaced the classical methods used so far. In this paper we intend to put forth the results of paleomagnetic investigations made in a few type sections in South Dobrogea as well as certain data of absolute ages obtained by thermoluminiscence, the first of this kind in Romania.

Geological Data

The region with the most representative loess sections corresponds with the Black Sea zone and the Danube banks, where, because of erosion, loess horizons and paleosols are well exposed.

The plateau zone, characterized by a slightly morphological incline is generally not exposed and it is only on big valleys (Caraşu) or in the slopes of the Danube-Black Sea canal that the loess peculiarities can be observed. Seemingly, in a uniform facies, loess varies as far as its grain size composition, the thickness of the loess and paleosol horizons, the carbonate contents are concerned.

As a rule, the loess in South Dobrogea shows features specific to the regions considered classical for this type of sediment: lack of bedding, high porosity (45–50 %), the contents in carbonates ranging between 15 and 10 %, yellowish colour (10YR). Most of the definitions of eolian origin, made up of quartz grains, frequently with a percent of clay (ϕ<0.002 mm) smaller than 20 % and of sand smaller than 15 %. In Dobrogea, taking into account the grain size composition, one could recognize: typical loess in which the silty fraction generally represents more than 70 %, sandy loess (on the bank of the Danube) in which the sandy fraction (0.2 0.002 mm) sometimes reaches 50 % and clayey loess (clay contents more than 30 %), marking the passage to the so called "loess-like deposits". The last ones are characterized by the occurrence of so called beddings or laminations, the presence of coarser fragments in the silty sediment mass, by the absence, as a rule, of carbonates. In recent literature, the attention is called to the confusion created around the term of "loess-like deposit". Certain authors consider it to be exclusively formed by the redeposition of eolian material. In other cases it has been described as solifluction or colluvial deposit, formed, as a rule, out of sediments, wherefrom it has been reworked, and many researchers consider it polygenetic.

In what follows we shall only refer to typical loess, identified in the south of Dobrogea. Its thickness is related to the morphology of the Ante-Quaternary relief. On a relatively small area, zones of loess accumulation, 11–20 m thick, have been found, corresponding to depressions in the Post-Sarmatian relief, while in other, more uplifted regions, its thickness is of only 1–2 m.

The outcrops on the Black Sea coast, as well as some ones situated on the plateau between the Black Sea and the Danube, have allowed the identification of several loess horizons separated by paleosols. At the upper part three loess horizons are identified, separated by two fossil soils, generally of chernozem type. They overlie a series characterized by an alternance of loess and brown-reddish clay paleosols. Towards the lower part of the loess and paleosols sequence, loess is weathered, diagenetically transformed and it is difficult to separate it from the brown-reddish paleosols in the cover and the bottom.

It is interesting to mention that, on the Carasu Valley (the present Poarta Albă-Năvodari canal), east of Nazarcea, in the base of the loess and the paleosols, there is a horizon of grey and red clays, up to 7–8 m thick, with a lot of gypsum crystals, a formation also identified in other areas in Dobrogea, at the loess base.

For paleomagnetically dating the loess, two sections have been chosen, representative because of their great number of well developed loess and fossil soils horizons, peculiarities on the basis of which they can be considered to be among the oldest deposits of this type in Dobrogea.

The first section analyzed is situated in the Black Sea zone, just south of Costinești. From top to bottom, the section looks like (Fig. 1):

– Recent soil, 1 m thick;

– upper loess horizons (L I, L II, L III), represented by yellowish silt, 9 m thick.

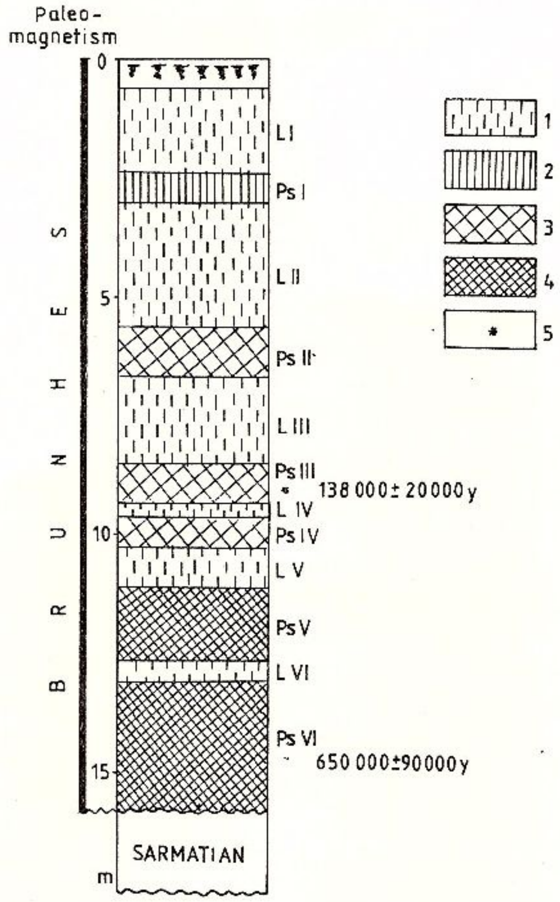

Fig. 1 – Costinești Section.
1, loess (L I .... L VI); 2, paleosol (chernozem type) Ps. I; 3, brown-reddish paleosol (Ps. II .... Ps. IV); 4, reddish paleosol rich in clay (Ps. V, Ps. VI); 5, ages established by thermoluminiscence.

The carbonate contents range between 15 and 20 %. Between the depths 2.80–3.50 m and 6.20–7.30 m, two slightly developed paleosols occur, the chernozem type being more obvious in the second horizon.

Between the depths of 9.40 m and 19 m, where Sarmatian limestones lie, there is a sequence characterized by the great development of brown-reddish clay soils. The first two horizons (Ps. III and Ps. IV) are clay soils, with prismatic structure, separated by a thin loess horizon. The next paleosols (Ps. V and Ps. VI) are represented by red and brown clays, with manganese traces and frequent calcareous concretions. In the base, the last paleosol, well developed, is represented by red clays. The loess at the lower part of the section is generally weathered and the features typical of a loess sediment have almost gone away. In the base of paleosol VI, the loess can no longer be identified.

Paleomagnetic researches have been carried out on oriented samples collected both from loess and from paleosols. Paleomagnetic datings on loess, as well as on other types of formations, are based on polarity inversions, excursions and secular variations of the magnetic field recorded in sediments, and rendered equivalent to corresponding absolute ages. For the Costineşti loess, the features of primary remanent magnetization indicate a normal polarity, corresponding to the Brunhes epoch, for the whole loess section. It results that the Costineşti loess is younger than 730 000 years, i.e. the lower limit of the normal polarity Brunhes epoch.

Paleomagnetic data have been correlated with the two absolute ages obtained by thermoluminiscence, carried out in the Lublin Laboratory, by dr. E. Krol's kindness (Rădan et al., 1987).

The first result that refers to a sample collected from paleosol III has indicated the age of 138 000 ± 20 000 years, that places the period of this soil genesis in the Riss-Würm interglacial period. The second dating has been made out on a sample from paleosol VI and has shown the age of 650 000 ± 90 000 years, that would correspond with the Günz-Mindel interglacial period.

Another cross-section analyzed is situated on the line of the present Poarta Albă-Năvodari canal, between Nazarcea and Ovidiu (Fig. 2). Between depth 0.40 m and 9 m there are three loess horizons, the silty fraction of which reaches 70–80 percents. The loess shows the typical features of eolian loess and within it *Helicopsis dejecta* and *Chondrula tridens* specimens have been identified. Upper paleosols (I and II) are represented by soils made up of two horizons – a brown, clayey one and a chernozem type one. The next paleosols (Ps. III and Ps. IV) are clayey, red-brown in colour, separated by a compact yellowish loess (L IV). Paleosol V is the thickest clay horizon of this section. It is red in colour, has ferromagnetic patches and prism structure. Paleosol VI, partially eroded in the base, is a yellowish-reddish clay, with rare calcareous concretions. The loess at its lower part (L IV and L V) is weathered, more clayey, difficult to separate from the paleosol in the bottom and the top.

In the base of this section, on about 6–7 m thickness, there occurs a continental formation, equally identified in other areas of Dobrogea. In the section described, it is made up of brown-reddish clays with manganese oxide patches, grey clays, glazy in aspect, that make up concoidal aggregates, red clays with grey patches, with quite many big gypsum crystals.

The Nazarcea section can be compared, taking into account the types of fossil soils and the features of the loess out of which these soils were formed, with the Costineşti section. Just like at Costineşti, the paleomagnetic investigation of the whole section, from the surface to the base (Ps. VI) shows normal polarity (Brunhes) for loess and paleosols.

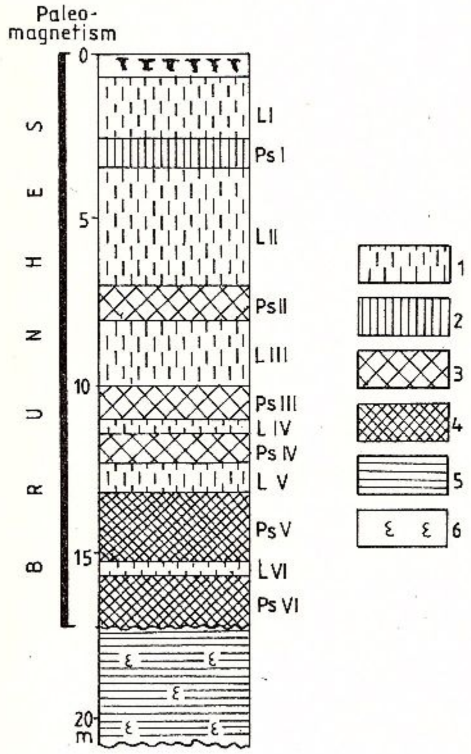

Fig. 2 – E Nazarcea Section.
1, loess (L I .... L VI); 2, paleosol (chernozem type) Ps. II; 3, brown-reddish paleosol Ps. III-Ps. IV; 4, reddish paleosol rich in clay (Ps. V, Ps. VI); 5, clayey complex: red and grey clays; 6, gypsum.

Chronological significances

In view of obtaining reliable data concerning the age of the loesses in Romania, South Dobrogea has been considered to represent the region with good premises for containing the oldest formations of this type, developed on the territory of Romania. Considering the types of paleosols developed on loess as well as their morphological features, a few of the most edifying sections have been selected, appearing on the territory of South Dobrogea. In the paper the results are given, obtained south of Costineşti and at Nazarcea-Ovidiu. The paleomagnetic data show, for the two sections, a normal magnetization, associated with the Brunhes epoch. According to the paleomagnetic time scale for the Quaternary (Mankinen and Dalrymple, 1979), it results that the Dobrogea loess is younger than 730 000 years.

The data obtained seem to correspond, in broad lines, with the results obtained concerning the absolute dating of the Eurasian loess. In Hungary, genuine loess is a little bit younger than 1 million years (Pecsi, 1984). In Czechoslovakia, in the Cerveny Kopec section, under paleosol IX the oldest loess in Central Europe would lie, corresponding with the Matuyama (Kukla, 1970) epoch of polarity. For the loess formation in Austria, Fink (1979) mentions loess older than Matuyama-Gauss only in the Stranzendorf section.

In Normandy, loess generally indicates normal polarity and only the basal paleosol in the Mesnil-Esnard section is of reversed magnetization (Lautridou, 1979). In the European part of the Soviet Union, loess is younger than 1 million years (Veklich, 1979). In Central Asia, many paleomagnetic analyses show an age younger than 730 000 years B.P. (Shermatov, Toiciev, 1982). In China, where in the last few years numerous studies have been carried out on the loess with thicknesses that locally reach 200 m, the oldest formations of this type are older than 1.1 million years (Liu Tung Sheng, 1982).

We should also point to something important related to the presence of the red clay horizon in the base of the Nazarcea loess. In the papers concerning the Dobrogea Quaternary, this deposit has already been pointed out. Examined especially of late, on the occasion of the Danube-Black Sea canal works, it has been assigned to the Lower Pleistocene, based on its position in the base of the Dobrogea loess.

The attempts of paleomagnetic dating, carried out on samples from the Nazarcea section, have not yielded conclusive results. Variations of the declination and inclination of primary remanent magnetization are noticed nevertheless compared with those registered in the loess at the upper part, but insufficient for considering them to represent the Matuyama reversed magnetization epoch.

In the last period of time, foreign literature mentions similar deposits at the base of the loess formations. In Hungary, the Dunaföldvar Formation, represented by red or brown-reddish clays, sometimes by glazy clays, paleosols and weathered clayey silts, that lie at the base of the loess, has been considered, based on paleomagnetic data, to have a stratigraphic position that corresponds with the Upper Pliocene and the Lower Pleistocene (Pecsi, 1984). In China, the Wucheng Formation, made up of loess deposits with red clays in the base, has been dated as representing the Matuyama epoch. Concluding then, we should point to the presence in this facies of deposits, associated in all cases, to loess formations. They seem to have a similar stratigraphic position in the regions where they have been cited.

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Prima pagină a textului va cuprinde: a) titlul lucrării (concis, dar informativ), cu un spaţiu de 8 cm deasupra; b) numele întreg al autorului (autorilor); c) instituţia (instituţiile) şi adresa (adresele) pentru fiecare autor sau grup de autori; d) colontitlu de maximum 60 semne. Notele de subsol se vor numerota consecutiv.

Citările din text trebuie să includă numele autorului şi anul publicării. Exemplu: Ionescu (1970) sau (Ionescu, 1970). Pentru doi autori: Ionescu, Popescu (1969) sau (Ionescu, Popescu, 1969). Pentru mai mult de doi autori: Ionescu et al. (1980) sau (Ionescu et al., 1980). Pentru lucrările care se află sub tipar, anul publicării va fi înlocuit cu "in press". Lucrările nepublicate şi rapoartele vor fi citate în text ca şi cele publicate.

Abstractul, maximum 20 rînduri, trebuie să fie în limba engleză şi să prezinte pe scurt principalele rezultate şi concluzii (nu o simplă listă cu subiecte abordate).

Cuvintele cheie (maximum 10) trebuie să fie în limba engleză sau franceză, corespunzător limbii în care este lucrarea (sau abstractul, dacă textul este în română), prezentate în succesiune de la general la specific şi dactilografiate pe pagina cu abstractul.

Bibliografia se va dactilografia la două rînduri, în ordine alfabetică şi cronologică pentru autorii cu mai mult de o lucrare. Abrevierile titlului jurnalului sau ale editurii trebuie să fie conforme cu recomandările respectivelor publicaţii sau cu standardele internaţionale.

Exemple:

a) jurnale:

Giuşcă, D. (1952) Contributions à l'étude cristallochimique des niobates. *An. Com. Geol.*, XXIII, p. 259–268, Bucureşti.

—, Pavelescu, L. (1954) Contribuţii la studiul mineralogic al zăcămintului de la Muşca. *Comun. Acad. Rom.*, IV, 11–12, p. 658–691, Bucureşti.

b) publicaţii speciale:

Strand, T. (1972) The Norwegian Caledonides. p. 1–20. In: Kulling, O., Strand, T. (eds.) Scandinavian Caledonides, 560 p., Interscience Publishers.

c) cărţi:

Bălan, M. (1976) Zăcămintele manganifere de la Iacobeni. Ed. Acad. Rom., 132 p., Bucureşti.

d) hărţi:

Ionescu, I., Popescu, P., Georgescu, G. (1990) Geological Map of Romania, scale 1:50,000, sheet Cîmpulung. Inst. Geol. Geofiz., Bucureşti.

e) lucrări nepublicate sau rapoarte:

Dumitrescu, D., Ionescu, I., Moldoveanu, M. (1987) Report. Arch. Inst. Geol. Geofiz., Bucureşti.

Lucrările sau cărţile publicate în rusă, bulgară, sîrbă etc. trebuie menţionate în bibliografie transliterînd numele şi titlurile. Exemplu:

Krasheninnikov, V. A., Basov, I. A. (1968) Stratigrafiya kainozoia. Trudy GIN, 410, 208 p., Nauka, Moskow.

Ilustraţiile (figuri şi planşe) trebuie numerotate şi prezentate în original, pe coli separate (hîrtie de calc), bune pentru reprodus. Dimensiunea liniilor, a literelor şi simbolurilor pe figuri trebuie să fie suficient de mare pentru a putea fi citite cu uşurinţă după ce au fost reduse. Dimensiunea originalului nu trebuie să depăşească suprafaţa tipografică a paginii: lăţimea coloanei 8 cm, lăţimea paginii 16,5 cm, lungimea paginii 23 cm, pentru figuri, iar pentru planşele liniare nu trebuie să depăşească dimensiunile unei pagini simple (16,5/23 cm) sau duble (23/33 cm) şi trebuie să fie autoexplicativă (să includă titlul, autori, explicaţie etc.). Scară grafică obligatorie.

Ilustraţiile fotografice (numai alb-negru) trebuie să fie clare, cu contrast bun şi grupate pe planşe de 16/23 cm. În cadrul fiecărei planşe numărătoarea fotografiilor se repetă (de. ex. Pl. I, fig. 1, Pl. II, fig. 1).

Tabelele vor fi numerotate şi vor avea un titlu. Dimensiunea originală a tabelelor trebuie să corespundă dimensiunilor tipografice menţionate mai sus (8/16,5 sau 16,5/23).

Autorii vor primi un singur set de corectură, pe care trebuie să-l înapoieze, cu corecturile corespunzătoare, după 10 zile de la primire. Numai greşelile de tipar trebuie corectate; nu sînt acceptate modificări.

Autorii vor primi gratuit 30 de extrase pentru fiecare lucrare.

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Author(s) should add a separate sheet with a summary indicating the hierarchy of headings from the text listed in decimal classification (1; 1.1; 1.1.1) but not exceeding four categories.

Text should be double-spaced typed (31 lines/page with 64 strokes each line) on one side of the paper only, holding an empty place of 3-4 cm on the left side of the page. The text cannot exceed 20 typewritten pages (including references and figures).

Front page (first page of the text) should comprise: a) title of the paper (concise but informative) with an empty space of 8 cm above it; b) full name(s) of the author(s); c) institution(s) and address(es) for each author or group of authors; d) short title (colontitle) of maximum 60 strokes. Footnotes should be numbered consequtively.

Citations in the text should include the name of the author and the publication year. Example: Ionescu (1970) or (Ionescu, 1970). For two authors: Ionescu, Popescu (1969) or (Ionescu, Popescu, 1969). For more than two authors: Ionescu et al. (1980) or (Ionescu et al., 1980). For papers which are in course of print the publication year will be replaced by "in press". Unpublished papers or reports will be cited in the text like the published ones.

Abstract, of maximum 20 lines, must be in English, summarizing the main results and conclusions (not a simple listing of topics).

Key words (max. 10 items), in English or French, following the language used in the text (or the Resumé if the text is in Romanian), given in succession from general to specific, should be typed on the abstract page.

References should be typed in double-line spacing, listed in alphabetical order and chronological order for authors with more than one reference. Abbreviations of journals or publishing houses should be in accordance with the recommendations of the respective publications or with the international practice.

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Giuşcă, D. (1952) Contributions à l'étude cristallochimique des niobates. *An. Com. Geol.*, XXIII, p. 259-268, Bucureşti.

— , Pavelescu, L. (1954) Contribuţii la studiul mineralogic al zăcămîntului de la Muşca. *Comm. Acad. Rom.*, IV, 11-12, p. 658-991, Bucureşti.

b) special issues:

Strand, T. (1972) The Norwegian Caledonides. p. 1-20. In: Kulling, O., Strand, T. (eds.) Scandinavian Caledonides, 560 p., Interscience Publishers.

c) books:

Bălan, M. (1976) Zăcămintele manganifere de la Iacobeni. Ed. Acad. Rom., 132 p., Bucureşti.

d) maps:

Ionescu, I., Popescu, P., Georgescu, G. (1990) Geological Map of Romania, scale 1:50,000, sheet Cîmpulung. Inst. Geol. Geofiz., Bucureşti.

e) unpublished papers or reports:

Dumitrescu, D., Ionescu, I., Moldoveanu, M. (1987) Report. Arch. Inst. Geol. Geofiz., Bucureşti.

Papers or books published in Russian, Bulgarian or Serbian etc. should be mentioned in the references transliterating the name and titles. Example:

Krasheninnikov, V. A., Basov, I. A. (1968) Stratigrafiya kainozoia. Trudy GIN, 410, 208 p., Nauka, Moskow.

Illustrations (figures and plates) must be numbered and submitted as originals on separate sheets (tracing papers), ready for reproduction. The thickness of the lines, lettering and symbols on figures should be large enough to be easily read after size-reduction. The original size should not extend beyond the print area of the page: column width 8 cm, page width 16.5 cm, page length 23 cm for figures; the width of line drawings should not extend over a single (16.5/23) or double (23/33 cm) page area and must be selfexplanatory (including title, authors, legend etc.). The graphic scale is obligatory.

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