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ПАЛЕОЗОЙ

МЕЗОЗОЙ

НЕОЗОЙ

NEW PALYNOLOGICAL PROOFS ON THE CAMBRIAN AGE OF THE TULGHEŞ GROUP (EAST CARPATHIANS)[1]

BY

VIOLETA ILIESCU[2], HANS-GEORG KRÄUTNER[2], FLORENTINA KRÄUTNER[2], HORST HANN[2]

With the increase of palynological and radiometric data the age of the Tulgheş Group has been highly disputed for the last 15 years.

First a Riphean age was presumed by Iliescu and Codarcea-Dessila (1965) and Codarcea-Dessila (1967) on the basis of a poor association of sphaeromorphs, ultramicrospores and *Laminarites*, reported from the Bistriţa Mts at Cîrlibaba and Isipoaia. At the same time Timofeev (1966) mentioned Upper Precambrian sphaeromorphs from similar rocks of the Rahov Mts.

The Cambrian age, at least for a part of the Tulgheş Group, was supposed first by Iliescu and Mureşan (1970) on the basis of *Lophotriletes rugosus* Naum. identified at Bălan. Later Iliescu and Mureşan (1972) provided new proofs for this statement by reporting Cambrian sphaeromorphs as *Granomarginata vulgaris* Naum., *Acanthosphaera cambriensis* Naum., *Psophosphaera obscura* Pich., *Microconcentrica atava* Naum. Considering also the available radiometric data and the known stratigraphic range of *Leioligotriletum bistrovi* Tim., the mentioned authors conclude on the Lower Cambrian age of the whole Tulgheş Group. In 1973 the possible presence of Cambrian rocks is the sequence of the Delovetsk Series (the equivalent of the Tulgheş Group in the Soviet literature), considered mainly as Vendian by Timofeev (1973 a), was also admitted by this author on the basis of *Lophodiacrodium* Tim. in association with sphaeromorphs frequently reported from the Vendian of the Russian Platform. Concluding on the existent data, Iliescu and Kräutner (1975) presumed that the upper part of the Tulgheş Group (formation Tg_3) represented the Cambrian, while the lower part (formations Tg_1 and Tg_2) could be a stratigraphic equivalent of the Vendian from the Russian Platform.

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1 Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, 1981 September 8 – 13, Bucharest, Romania.

2 Institute of Geology and Geophysics, str. Caransebeş 1, 78344 Bucharest, Romania.

New data confirming the Cambrian age of the Tulgheş Group were reported from its equivalents in the Rahov Massif (Delovetsk Series) and in the Civcin Mts (Boerovsk Series) where S e r g e e v a (1974) identified *Baltisphaeridium* sp. and *Baltisphaeridium* cf. *varium* V o l k . associated with *Laminarites*. Cambrian palynomorphs as *Achantosphaera cambriensis* N a u m ., *Granomarginata vulgaris* N a u m . and *Lophotriletes rugosus* N a u m . were mentioned later also in the higher parts of the Tulgheş Group (Tg_3 and Tg_4) from the Bistriţa Mts near Broşteni (F o m o v , in V o d ă et al., 1976, unpublished data).

The palynological data were correlated with the available radiometric data by B e r c i a et al. (1976) who concluded on the Vendian-Cambrian age of the Tulgheş Group.

Lithostratigraphy of the Tulgheş Group

The lithostratigraphic research of the Tulgheş Group started in 1962 by C o d a r c e a - D e s s i l a et al. (1964) at Broşteni-Crucea and was later extended over large areas by many geologists. A review was given by B e r c i a et al. (1976).

The new palynological data refer to the Broşteni-Zugreni zone of the Central Bistriţa Mts (Fig. 1). In this area the following lithostratigraphic sequence has been recognized in the Tulgheş Group of the Putna Unit from the Bucovinian Nappe (K r ä u t n e r et al., 1981, unpublished data) (Fig. 2).

The formation Tg_1 (*blastodetrital-quartzitic*, 800 m) represents the lowest known part of the sequence. In the lower part it is cut by the overthrust plane of the Putna Nappe. It consists of quartzite-sericite-chlorite schists with a member of quartzites and quartz-feldspar rocks (Toancele Member) in its middle part.

The formation Tg_2 (*graphitic with metalydites*, 450 — 600 m) consists of an alternation of sericite-graphite schists, sericite schists and black quartzites (metalydites) with intercalations of limestones and a marker horizon of sericite quartz-feldspar rocks with chlorite and biotite (Pîrîul Ursului Horizon).

The formation Tg_3 (*volcano-sedimentary rhyolitic*, 1200 — 2400 m) corresponds to the products of five main phases of rhyolitic extrusions, separated by metasedimentary rocks. In this volcano-sedimentary pile four lithostratigraphic units have been mapped at the rank of members, called Isipoaia, Leşul Ursului, Moroşan and Fundul Moldovei. The volcanic sequences consist of quartz-feldspar schists mainly derived from rhyolitic tuffs and epiclastites. In the metasedimentary rocks associated with the volcanic sequences stratiform pyrite and base metal ore deposits of Kuroko type do appear (Fig. 2).

The formation Tg_4 (*blastodetrital with phyllites*, 1500 — 2500 m) derived from the detrital material deposited after the main volcanic phases. It is represented by an alternation of various terrigenous rocks in which scarce intercalations of rhyolitic metatuffs and metaepiclastites appear within two horizons only, named Praşca and Dealul Fagi. The following three members have been recognized : the B a ş c a m e m b e r, a quartz-

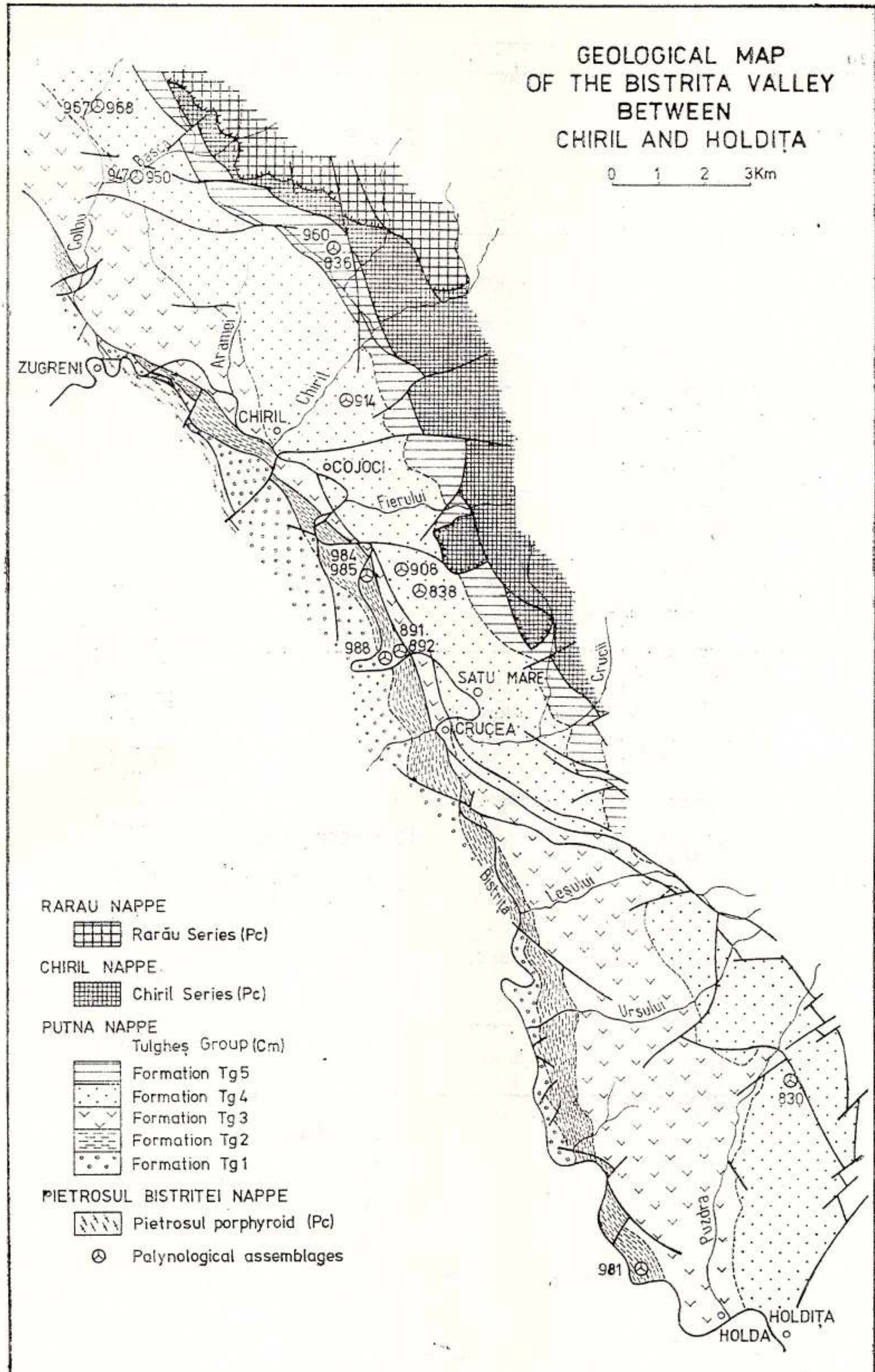
GEOLOGICAL MAP
OF THE BISTRITA VALLEY
BETWEEN
CHIRIL AND HOLDIȚA
0 1 2 3Km
967
968
947
950
960
836
914
984
985
906
838
891
892
988
830
981
Golbu
Arameiei
Chiril
Fierului
Crucii
Bistrița
Leșului
Ursului
Puzdra
ZUGRENI
CHIRIL
COJOCI
SATU MARE
CRUCEA
HOLDIȚA
HOLDA
RARAU NAPPE
Rarău Series (Pc)
CHIRIL NAPPE
Chiril Series (Pc)
PUTNA NAPPE
Tulgheș Group (Cm)
Formation Tg5
Formation Tg4
Formation Tg3
Formation Tg2
Formation Tg1
PIETROSUL BISTRITEI NAPPE
Pietrosul porphyroid (Pc)
Palynological assemblages

Fig. 1

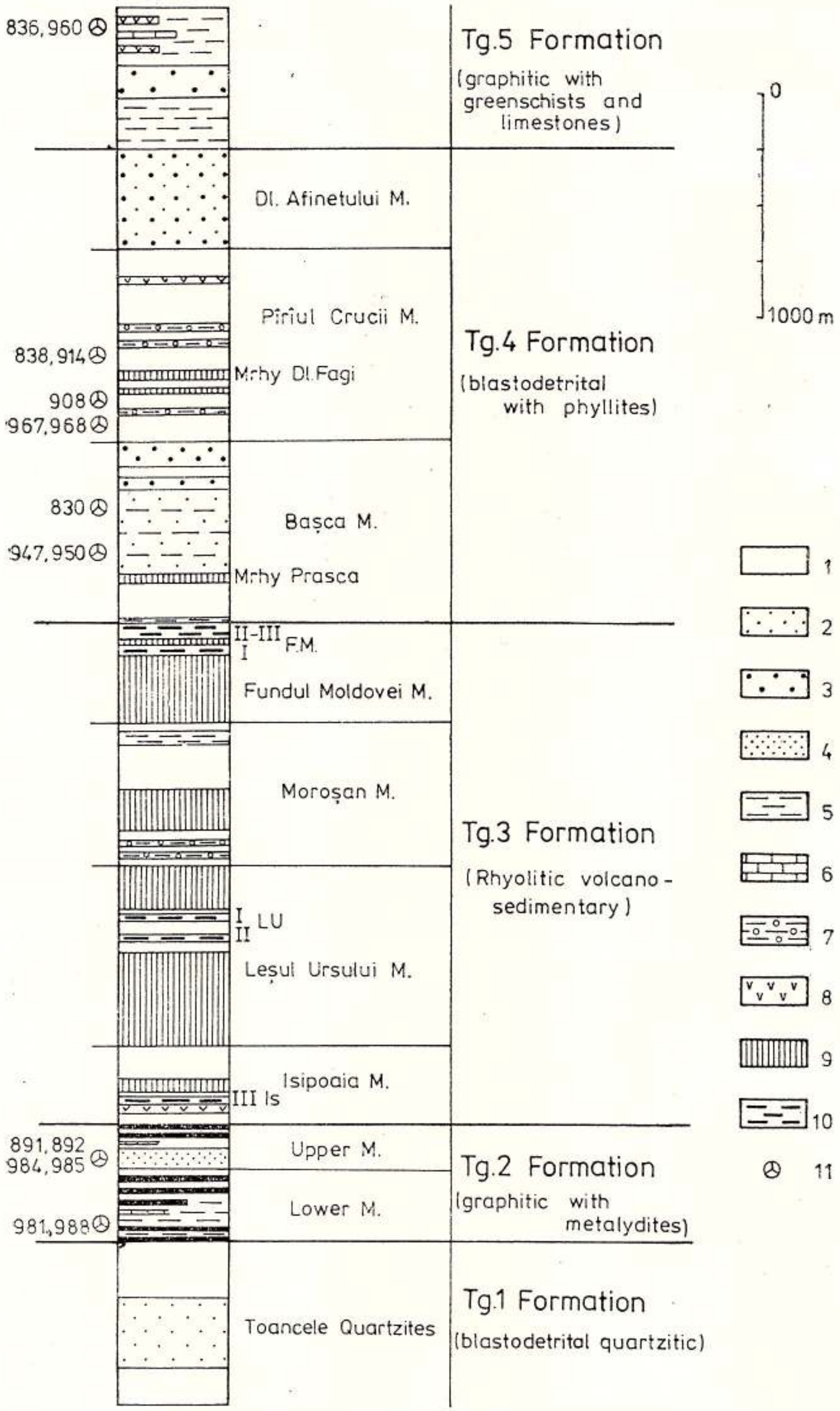
836,960
Tg.5 Formation
(graphitic with greenschists and limestones)
0
1000 m
Dl. Afinetului M.
Pîrîul Crucii M.
838,914
Mrhy Dl.Fagi
908
967,968
Tg.4 Formation
(blastodetrital with phyllites)
830
Bașca M.
947,950
Mrhy Prasca
II-III
I
F.M.
Fundul Moldovei M.
Moroșan M.
Tg.3 Formation
(Rhyolitic volcano-sedimentary)
I
II
LU
Leșul Ursului M.
Isipoaia M.
III Is
891,892
984,985
Upper M.
Lower M.
981,988
Tg.2 Formation
(graphitic with metalydites)
Toancele Quartzites
Tg.1 Formation
(blastodetrital quartzitic)
1
2
3
4
5
6
7
8
9
10
11

Fig. 2

ite-phyllitic association with quartz-feldspar rocks at the top (Arşiţa Rea Horizon); the Pîrîul Crucii member, a specific association of sericite-chlorite schists with albite porphyroblasts, phyllites, sericite-chlorite schists and greenschists; the Afinet member consisting of various types of quartzitic and quartz-feldspar rocks.

The formation Tg_5 (graphitic with greenschists and limestones, 600 m) is represented by an alternation of sericite—graphite schists with sericite schists, quartz-feldspar schists and intercalations of greenschists, limestones and metalydites.

Palynological Content

Samples from the whole sequence of the Tulgheş Series, excepting its lower part, the formation Tg, were collected from seven partial profiles between Zugreni and Holda (Putna Unit of the Bucovinian Nappe in the Bistriţa Mts, Fig. 1) as follows: 1, Upper Colbu Valley; 2, Valea Başca 3, Culmea Afinet; 4, Valea Fierului; 5, Dîmbul Colacului-vîrful Runc; 6, Valea Dorneanu; 7, Bistriţa Valley at Holda.

All the analysed rocks are in the greenschist faciest reaching locally the biotite isograde. Palynomorphs have been identified in approximately 20% of the collected samples. They are often deformed (compressional folds; kinks) or fragmentarily preserved and show different degrees of incipient carbonization (darkening). Some samples furnished palynomorphs in good state of conservation, without diagnosis difficulties. Such associations have been found especially in the metapsammitic rocks of the Pîrîul Ursului Horizon, in the phyllitic rocks of the formation Tg and in the calcschists and graphite schists of the formations Tg_2 and Tg_5.

The palynological content of these rocks consists mainly of sphaeromorphs without special biostratigraphic importance, associated with Cambrian acritarchs. Contaminations with Mesozoic spores were observed in some samples from the Bistriţa Valley at Dîmbul Colacului (no. 985) as well as from the Colbu Valley and Başca Valley (no. 951, 968).

Formation Tg_1. No samples were collected from this formation. Ruxandra Fomov (fide Vodă et al., 1976)[3] reported from the respective sequence, near Broşteni, an association of sphaeromorphs with *Protosphaeridium* sp., *P. tuberculiferum* Tim., *P. flexuosum*, *P. acis* Tim., *P. densum* Tim., *Trematosphaeridium* sp., *Stictosphaeridium* sp., *S. sinapticuliferum* Tim., *Orygmatosphaeridium* sp., *Trachysphaeridium* sp., *Favososphaeridium* sp., *Leiosphaeridia undulata* Tim., *Symplassosphaeridium* sp., *Glaeocapsomorpha*.

Fig. 2. — Lithostratigraphy of the Tulgheş Group and Acritarch content.
1, sericite-chlorite schists; 2, quartzitic rocks; 3, feldspar quartzites (Arşiţa Rea type); 4, quartz-feldspar rocks with chlorite and biotite (Pîrîul Ursului Horizon); 5, phyllites and sericite—graphite schists; 6, limestones; 7, chlorite-sericite schists with albite porphyroblasts; 8, greenschists (basic metatuffs); 9, rhyolitic metavolcanics; 10, stratiform pyrite and base metal sulphides; 11, palynological assemblages.

Formation Tg_2. In the lower part of the sequence, the following associations of acritarchs were identified in the graphite schists on the road near Holda (sample no. 981)[4] :

Baltisphaeridium sp.
B. dubium Volk.
Cymatiosphaera cf. *membranacea* Kir.
Trachyopsophosphaera exilis Lop.
Podoliella irregulare Tim.
Protosphaeridium sp.
P. tuberculiferum Tim.
Leiosphaeridia sp.
Trachysphaeridium sp.
Orygmatosphaeridium sp.
Trematosphaeridium sp.
Stictosphaeridium sp.
Pterospermopsimorpha sp.

and at Dîmbul Colacului (sample no. 988) :
Trachysphaeridium atenuatum Tim.
Protosphaeridium sp.
Leiosphaeridia sp.
Orygmatosphaeridium sp.
Stictosphaeridium sp.

At the upper part of the sequence a palynological assemblage was found in the metapsammitic rocks of the Pîrîul Ursului Horizon. The samples (no. 984, 985, 891, 892) were collected from the road on the Bistrița Valley, north and south of Dîmbul Colacului.

Cymatiosphaera sp.
Baltisphaeridium sp.
B. dubium Volk.
B. ciliosum Volk.
B. cerinum Volk.
B. cf. *compressum* Volk.
Granomarginata squamacea Volk.
Spumosata sp.
S. prima Naum.
Protosphaeridium sp.
P. laccatum Tim.
Synsphaeridium conglutinatum Tim.
Leiosphaeridia sp.
L. undulata Tim.
Trachysphaeridium incrassatum Sin-Liu
T. atenuatum Tim.
Orygmatosphaeridium sp.
Stictosphaeridium sp.
Symplassosphaeridium sp.
Trematosphaeridium sp.
Kildinella sp.

Formation Tg_3. No palynomorphs have been found in the samples collected from the rhyolitic volcano-sedimentary formation on the mentioned profiles. But Iliescu and Mureşan (1972) reported from a probable lithostratigraphic equivalent of this formation, namely the Băilor Horizon at Bălan, an association (samples no. 4005, 4043) with *Achantosphaera cambriensis* Naum., *Microconcentrica atava* Naum., *Leiodiscina* sp., *Granomarginata* sp. Near Broşteni, Vodă et al. (1976)[3] mentioned a similar association in the formation Tg_3 : *Achantosphaera cambriensis* Naum., *Granomarginata vulgaris* Naum., *Protosphaeridium densum* Tim., *Favososphaeridium favosum* Tim., *Synsphaeridium* sp., *Orygmatosphaeridium* sp., *Kildinella* sp.

Formation Tg_4. Palynomorphs from the lower part of this lithostratigraphic unit, equivalent to the Başca member, were reported by Vodă et al.[3] in 1976: *Archaeopsophosphaera asperata* Naum., *Lophotreiletes rugosus* Naum., *Synsphaeridium conglutinatum* Tim., *Favososphaeridium* sp., *Leiosphaeridia* sp., *Protosphaeridium* sp. It seems that the samples 2737, 2738, 2740, 2710, 4006 collected by Iliescu and Mureşan (1972) from the Sîndominic complex and the upper part of the Sadocut complex of the Bălan region are also in equivalent lithostratigraphic position with the formation Tg_4. The mentioned authors report from these samples *Granomarginata vulgaris* Naum., *Spumosata* sp., *Archaeosacculina* sp., *Asperatopsophosphaera* sp., *Psophosphaera* sp. and *Trematosphaeridium* sp.

On our profiles associations with Cambrian acritarchs were found in the Başca member and within two horizons in the Pîrîul Crucii member.

In the Başca member there were identified in phyllitic sericite-graphite schists of the Dorneanu Valley (sample 830)

Baltisphaeridium cf. *compressum* Volk.	*Granomarginata* sp.
	Leiosphaeridia sp.

and in the same rocks on the Başca Valley (samples 947, 950)

Veryhachium sp.	*Protosphaeridium* sp.
Pterospermella vitalis Jank.	*Leiosphaeridia* sp.
	Trachysphaeridium sp.

In the Pîrîul Crucii member similar acritarchs were identified in sericite-graphite schists of the upper Colbu Valley, below the horizon of the Dealul Fagi rhyolitic metatuffs (samples 967, 968):

Veryhachium sp.	*Archaeofavosina* sp.
Spumosata prima Naum.	*Asperatopsophosphaera* sp.
	Pseudotasmanites sp.
	Protosphaeridium sp.
	P. flexuosum Tim.
	Synsphaeridium conglutinatum Tim.
	Trachysphaeridium sp.

as well as in quartz-sericite-graphite schists of Dealul Runc in higher lithostratigraphic position in respect to the Dealul Fagi rhyolitic metatuffs (sample 908):

Baltisphaeridium sp.	*Trachysphaeridium* sp.
Orygmatosphaeridium distributum Tim.	*Protosphaeridium* sp.
	P. acis Tim.
	Leiosphaeridia sp.

Formation Tg_5. In this highest known part of the Tulgheş Group, palynomorphs were found only in sericite-graphite schists and in calcschists associated with the limestones near Schitu Rarău (samples 836, 960):

Veryhachium sp.	*Granomarginata* sp.
Cymatiosphaera sp.	*Synsphaeridium* sp.
	Leiosphaeridia sp.

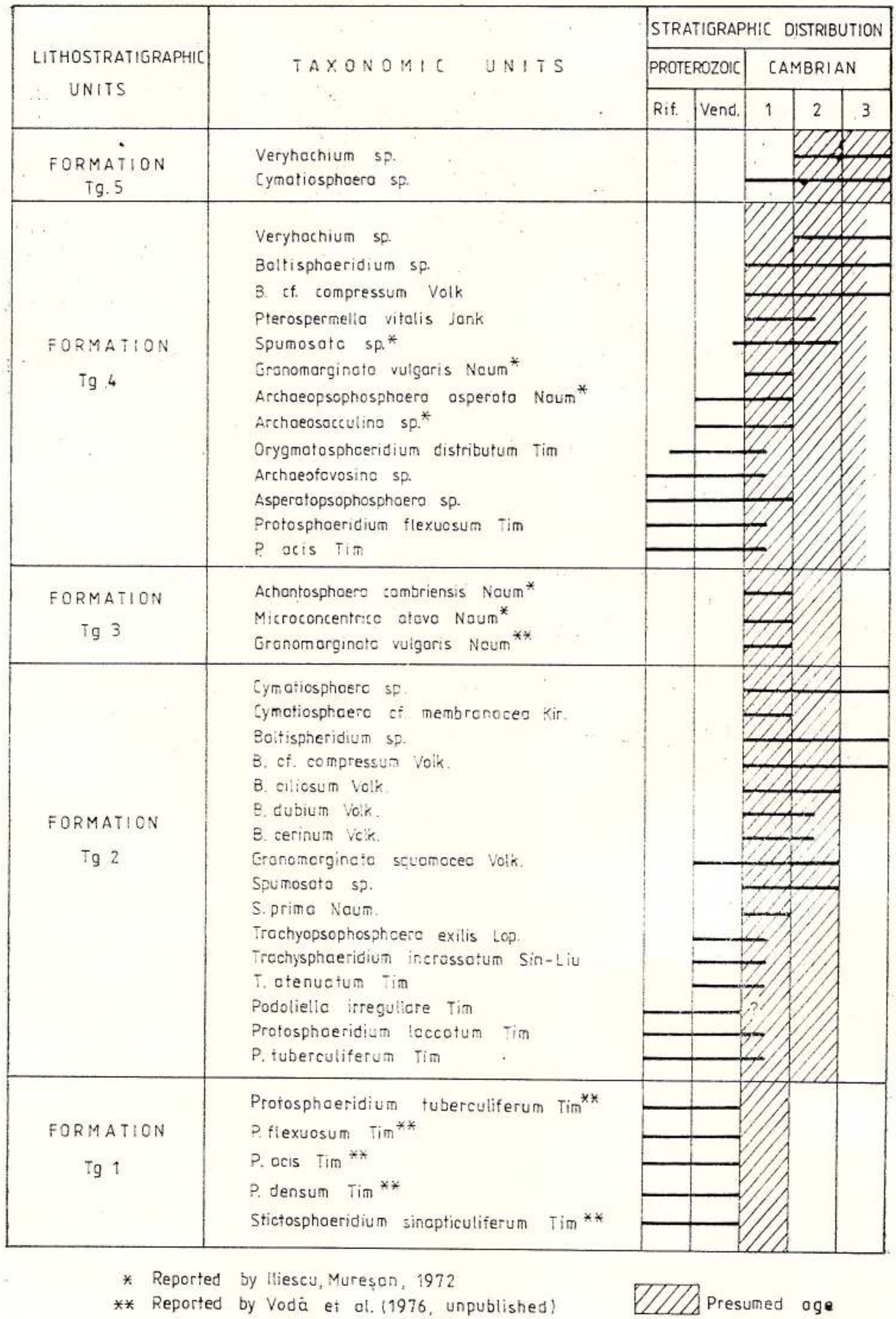

Fig. 3. – Lithostratigraphic and stratigraphic distribution of the Acritarchs in the Tulgheș Group.

Conclusions on the Age

Considering the stratigraphic distribution of the palynomorphs identified in the Tulgheș Group (Fig. 3) four types of assemblages may be recognized: 1, acritarchs of the types *Baltisphaeridium*, *Cymatiosphaera* and *Veryhachium* considered to appear first in the Cambrian (Volkova et al., 1979); 2, acritarchs mentioned by Naumova (1968) only in the Lower Cambrian, as for example *Granomarginata vulgaris* Naum., *Achantosphaera cambriensis* Naum., *Microconcentrica atava* Naum., *Spumosata prima* Naum.; 3, sphaeromorphs as *Archaeopsophosphaera asperata* Naum., *Archaeosacculina* sp., *Granomarginata* cf. *squamacea* Volk., *Trachyopsophosphaera exilis* Lop., *Trachysphaeridium incrassatum* Sin-Liu, *T. atenuatum* Tim., mentioned by Volkova (1969), Sin Yu Sheng and Liu Kui-Zhih (1973), Timofeev (1973), Timofeev et al. (1976) in the Vendian and Lower Cambrian; 4, sphaeromorphs with decline in the Lower Cambrian, covering most of the Middle and Upper Proterozoic as for example *Protosphaeridium flexuosum* Tim., *P. acis* Tim., *P. tuberculiferum* Tim., *P. laccatum* Tim., *Orygmatosphaeridium distributum* Tim., *Asperatopsophosphaera* sp., *Archaeofavosina* sp., *Podoliella irregulare* (Timofeev, 1973a, b).

If contaminations or resedimentation of the older palynomorphs are excluded, the coexistence of the mentioned acritarch assemblages (1 and 3, 4) may be considered as a proof of the Lower Paleozoic age (Volkova, 1969; Volkova et al., 1979). It seems therefore that the palynological data of Figure 3 indicate the Cambrian age of the Tulgheș Group.

For the lower part of the Tulgheș Group (formation Tg_1) rests the possibility to include the Upper Vendian. As it was supposed by Kräutner (1980) the formation Tg_1 may be considered a lithostratigraphic equivalent of the lower part of the Muncel Series (Apuseni Mts) in which Visarion (unpublished data) mentioned *Cymatiosphaera*. Thus, it is highly probable that the formation Tg_1 represents also the Lower Cambrian.

For the upper part of the Tulgheș Group (formations Tg_4, Tg_5) with *Veryhachium*, at least a Middle Cambrian age may be supposed by the available data, but it is possible that the Upper Cambrian and even the Lower Ordovician may be included also, especially in the Formation Tg_5

We may conclude that the whole Tulgheș Group may be assigned to the Cambrian, with possible and probable time span for Tg_1 in the Lower Cambrian, for Tg_2, Tg_3 in the Lower-Middle Cambrian and for Tg_4, Tg_5 in the Middle-Upper Cambrian, may be also lowest Ordovician.

Correlation with Radiometric Ages

The available radiometric data refer to the age of the volcanic activity of the formation Tg_3 and the associated metallogenesis, as well as to the age of the regional metamorphism of the Tulgheș Group.

U-Pb zircon ages of 560 – 640 m.y. were reported by Boiko et al. (1975) for the rhyolitic metavolcanics of the Delovetsk Series (equivalent of the Tulgheș Group in the Rahov Mts). The upper part of the mentioned time span is in concordance with the Lower Cambrian age indicated by acritarchs (Fig. 3) for the rhyolitic volcano-sedimentary formation Tg_3.

Pb-Pb ages of 540 – 600 m.y. were obtained by Vîjdea, Anastase (1975 and unpublished data) and Popescu (unpublished data) on syngenetic stratiform ores associated with the rhyolitic volcanism of the formation Tg_3. These data are in concordance with the mentioned zircon ages and support the Lower Cambrian age of the volcano-sedimentary formation Tg_3.

K/Ar whole rock and sericite ages were reported from rocks of the Tulgheş Group by Semenenco et al. (1969), Pop et al. (1974), Mînzatu et al. (1975), Boiko et al. (1970) and Kräutner et al. (1976). The ages were frequently regenerated by Ar loss during Variscan or Alpine events. Maximal values of K/Ar modal ages are of 472 m.y. ($\lambda k\ 10^{-10}$ ans^{-1}= 0.584) and the K/Ar isochrone ages suggest 505 ± 5 m.y. Therefore the regional metamorphism of the Tulgheş Group may be accepted around the Cambrian-Ordovician boundary or in the Lower Ordovician. This proves that the rocks of the Tulgheş Group with acritarchs such as *Veryhachium*, *Baltisphaeridium* and *Cymatiosphaera* cannot be younger than the lower Ordovician.

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[3] A. Vodă, D. Vodă, I. Popescu, C. Velio. Unpublished data.

[4] Diagnosis palynomorphs and sphaeromorphs without special stratigraphic importance are listed in different columns.

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ON THE PRE-NEOGENE FORMATIONS IN THE NORTH-DOBROGEAN PROMONTORY [1]

BY

DUMITRU PARASCHIV [2], CORNELIA PARASCHIV [2], CORNELIA ANDREI [2], MIRCEA POPESCU [2], NURHAN DĂNEŢ [2]

The North-Dobrogean Promontory — an extension of North Dobrogea W—NW of the Danube under the Neogene cover — has been lately the object of intensive and systematic geological and geophysical investigations. Over 300 wells have been drilled up to now between the Danube, the Prut and the Siret rivers and the Bîrlad parallel, out of which more than half have also reached pre-Neogene deposits.

As already known, the geological formations making up the area corresponding to the North-Dobrogean Promontory may be assigned to two structural stages : a lower one, with folded, more or less metamorphosed, deposits of Precambrian (?), Paleozoic and Triassic age ; and an upper one, present in a quasihorizontal Neogene structure. The lower stage corresponds to the North-Dobrogean Orogene, consolidated during the uppermost Triassic, while the upper stage terms make up the epicratonic flank of the Carpathian Foredeep.

Despite the high degree of rock transformation and almost complete lack of determinable fossil remains, the stratigraphic sequences crossed by wells under the Neogene cover are similar, sometimes even identical to those which outcrop in Dobrogea. This has allowed registering a similar succession in the buried promontory, shown by the geological mappings S—SE of the Danube. The latter includes *the Orliga-Megina Catamesometamorphites*, *the Boclugea Formation*, *the Ţepu Formation*, *the Măxineni Formation*, *the Carapelit Formation*, as well as *postcarapelitic sedimentary sequences* and numberless *magmatic bodies*.

1. *The Orliga-Megina Catamesometamorphites* described in the relevant literature (Giuşcă, 1934; Ianovici et al., 1961; Mirăuţă, Mirăuţă, 1962; Mirăuţă 1966, etc.) make up a whole

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[2] Ministry of Petroleum, Bucharest, Romania.

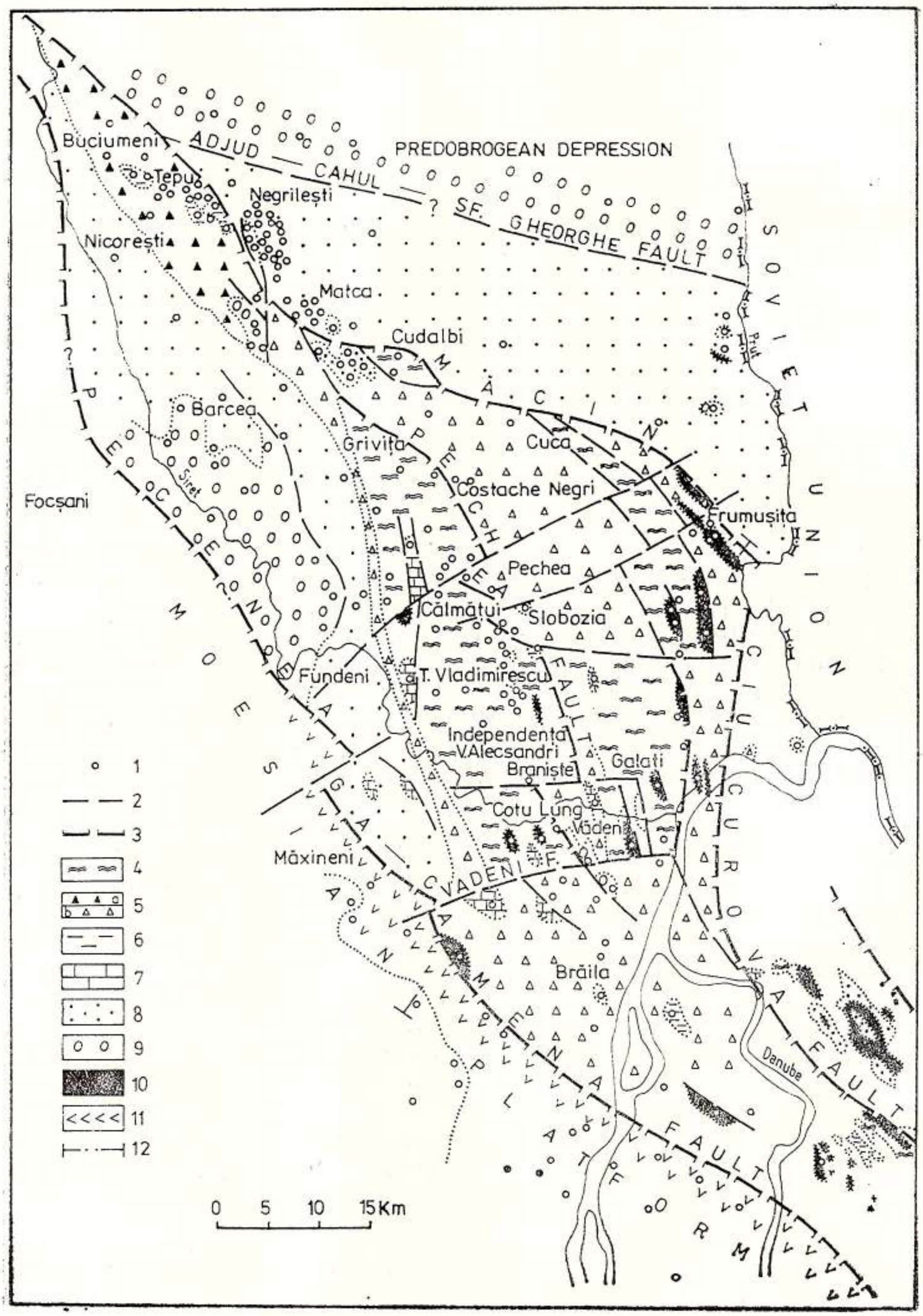
PREDOBROGEAN DEPRESSION
ADJUD — CAHUL — ? SF. GHEORGHE FAULT
SOVIET UNION
Buciumeni
Tepu
Negrilești
Nicorești
Matca
Cudalbi
Barcea
Grivița
Cuca
Costache Negri
Frumușița
Pechea
Călmățui
Slobozia
Focșani
Siret
Prut
Fundeni
T. Vladimirescu
Independența
V.Alecsandri
Braniște
Galați
Cotu Lung
Vădeni
Măxineni
VADENI F.
Brăila
Danube
MĂCIN
PECHEA FAULT
PECENEAGA CAMENA FAULT
MOESIAN PLATFORM
CIUCUROVA FAULT
1
2
3
4
5
6
7
8
9
10
11
12
0 5 10 15Km

Fig. 1

complex of crystalline schists consisting of amphibolites, amphibolic schists, shaley white quartzites and micaschists with muscovite and biotite in the Măcin Mts. Similar rocks have been crossed within the North-Dobrogean Promontory, mainly in the area of well concentration and greater depth of pre-Neogene investigations. Catametamorphic rocks are represented by gneisses, amphibolites and amphibolic schists, quartzites, micaschists with biotite and muscovite, partially slightly cornified.

Within the North-Dobrogean Promontory, the Orliga-Megina Catamesometamorphites make up two buried ridges, oriented SE–NW (Fig. 1). One is the axial promontory zone which seems to extend continuously between Cotu Lung and Grivița; the other, revealed by the wells drilled between Vădeni and Cuca, follows the line of the Măcin-Ciucurova Fault, still lying west of the accident in question.

According to the researches carried out by Mirăuță (1966) in the southern part of the Măcin Mts, the age of the catamesometamorphic complex must be Precambrian, possibly including part of the Cambrian as well. A very recent paper (Seghedi, 1980) also foresees a Precambrian age for these rocks.

2. *The Boclugea Formation* includes a sedimentogene, more or less epimetamorphosed, association, where phyllites and quartzites are prevailing. In North Dobrogea, where it outcrops, the sedimentogene association has been defined (over a thickness of about 1500 m) as "the Phyllito-Quartzitic Series" or "the Boclugea Series" (Ianovici et al., 1961; Mirăuță, Mirăuță, 1962). Phyllites are dominant in the lower part of the profile and appear as metric intercalations. Generally weak, their metamorphism amplifies with sericitous and chloritous phyllites. White, grey or red quartzites are either massive (banks of 2–4 m) or bedded and prevail in the upper part. "The Phyllito-Quartzitic Series" has been estimated as belonging to the Cambro-Ordovician, conformably overlying the Orliga-Megina Catamesometamorphites.

Within the North-Dobrogean Promontory, an equivalent of the "Phyllito-Quartzitic Series" is considered to be the "Boclugea Formation", a complex of epimetamorphic rocks or, locally, a sequence of orthoquartzites, grauwacke, subgrauwacke and argillites, almost unaffected by metamorphism and known as "*the Buciumeni Orthoquartzites*". In the authors' opinion, the Buciumeni Orthoquartzites, comprising a Cambro-Ordovician palyno-protistologic assemblage, could be the stratigraphic equivalent of the Boclugea Formation. The discrepancy between the two lithostratigraphic units lies in the rock transformation degree, a variation equally registered in North Dobrogea. Lying in the north-

Fig. 1. – Map of the pre-Neogene formations location within the North-Dobrogean Promontory: 1, well reaching pre-Neogene deposits; 2, fault; 3, overthrust line; 4, the Orliga-Megina Catamesometamorphites; 5, the Boclugea Formation: a, Buciumeni Quartzites; b, epimetamorphites; 6, the Țepu Formation; 7, the Măxineni Formation; 8, the Carapelit Formation; 9, post-carapelitic deposits; 10, magmatites; 11, greenschists; 12, section direction.

western end of the promontory (Fig. 1), at a certain distance from the Paleozoic magmatic hearth, the Țepu-Buciumeni area has been estimated as less affected, dynamically and thermally, during its evolution.

The epimetamorphites (the Boclugea Formation) consist of : chloritic and chlorito-sericitic phyllites; chloritic, sericitic and chlorito-sericitic schists; chlorite-quartzitic schists; quartzitic schists with epidote; at Frumușița, these rocks are strongly mineralized with hematite.

The Buciumeni Orthoquartzites have been crossed near the locality bearing the same name; still, they have been opened by drillings but for a section of 433 m at the utmost.

In the Buciumeni Orthoquartzite sequence, the well 117 Țepu has encountered fossil remains at the depths 2372. 5 – 2374.5 m, 2379 – – 2382 m and 2382 – 2385.5 m, consisting of *Microhystridium shintonensis* Dow., *M. stellatum* Defl., *Archaeohystrichosphaeridium* sp., *A. arenigum* Tim., *Acanthodiacrodium* sp., *Cymatiogalea* sp., *Baltisphaeridium hirsutoides* (Eis.), *B. brevifurcatum* (Eis.), *B. trifurcatum paucifurcatum* Eis., *Gonyosphaeridium polygonale* (Eis.) Eis., *Veryhachium europaeum* Stock., Will., *V. aculeatum* Deunff., *Lophominuscula* sp., *Leiominuscula* sp., *Protosphaeridium asaphum* Tim., *P. densum* Tim., *Leiosphaeridia fragile* Dow., *Leiofusa* sp., undeterminable Acritarchs, undeterminable Leiosphaere, algal remains.

The identified palyno-protistological assemblage seems to indicate uppermost Cambrian-lowermost Ordovician.

3. *The Țepu Formation* is composed of a succession of argillites, limestones and sandstones, abundant in organic matter and pyrite. More or less affected by diagenetic processes, these deposits are fragmentarily distributed (Fig. 1). The lithostratigraphic term in question looks like better developed in the north-western promontory end, in the Țepu-Buciumeni area, where it appears as the conformable extension of the Cambro-Ordovician sequence. These deposits have neither been completely crossed; the maximal section, of 124 m, has been opened by the well 118 Țepu.

Over the investigated thickness, the Țepu Formation comprises the following petrographic types : slightly microgritty, pyrito–carbonaceous argillites, locally looking shaley or strongly breccified; gritty grey limestones with vegetal remains, lithic limy–detrital sandstones, slightly pigmented by organic matter and pyrite; sandstones with siliceous-chlorito-sericitic binder, sometimes affected by metamorphic processes.

The deposits belonging to the Țepu Formation lack determinable fossil remains. Taking into account that the described sequence has accumulated within an euxinic, reducing environment, that it has been crossed both on the flank and in the Țepu uplift pericline (the axis of the latter equally including the Boclugea Formation) and that carbonate episodes also occur in the deposit succession, the Țepu Formation is finally estimated as part of the Silurian.

Similar deposits have been encountered in the southern Măcin Mts (Mirăuță, Mirăuță, 1962), where the "Grey Phyllite and Limestone Series" follows over a thickness of about 1000 m. The latter comprises crinoid and coral remains (ex. *Cyathophyllum*) and supports the fossiliferous Devonian, which decides for its assignment to the Silurian.

4. *The Măxineni Formation* appears in patches on the western promontory flank (Fig. 1), at Măxineni, Muchia, Brăila, Piscu and Pechea. Being crossed only over 166 m (at Măxineni), the deposits in question consist of crystalline limestones and dolomitic crystalline grey limestones, impurified by clayey matter and pyrite; they also include grey pseudoskarns with vesuvian (Măxineni).

Considering certain similarities with the profile opened in the Măcin Mts (Simionescu, 1924; Paeckelman, 1935; Mirăuţă, Mirăuţă, 1962), it is temporarily considered that the carbonate deposits affected by contact-metamorphic processes belong to the Devonian.

5. *The Carapelit Formation*, defined in North Dobrogea (Mrazec, Pascu, 1896), consists of two horizons: a lower one — predominantly conglomeratic — and an upper one, made of grauwacke and phyllites (Ianovici et al., 1961). The outcropping "Carapelit Beds" reach thicknesses of about 1500 m. Their age has been estimated as Dinantian.

Within the North-Dobrogean Promontory, Mureşan et al. (1971) identified and described, first north of Frumuşiţa, the Carapelit Formation. Subsequently, while extending the research, the Carapelit Formation has been crossed by other 40—50 wells, most of them concentrated in the Matca area (Fig. 1). These deposits may also be present in the western promontory flank, between Nicoreşti and Fundeni. According to the present knowledge, it seems that the Carapelit Formation, preserved in some big synclines, is part of the North-Dobrogean Promontory flank and pericline.

None of the wells drilled up to now has entirely opened the lithostratigraphic term in question. The thickest sections have been crossed by the wells 147 Matca (550 m) and 4200 Nicoreşti (685 m).

The rocks in this formation show certain anchymetamorphic aspects, expressed by the effects of micaceous mineral orientation, as well as quartz and feldspar grain crush and extension. The clayey and tuffaceous material included in the binder and in the rock components globally change into sericite and partially into chlorite and silicas.

As in North Dobrogea, the Carapelit Formation is represented within the lowermost North-Dobrogean Promontory by predominantly conglomeratic deposits, overlain by a gritty-clayey sequence.

The Carapelit Formation deposits lack in determinable fossil remains. Only in the reworked limestone fragments, there have been identified numberless foraminifera and conodont species. Among conodonts, the following species are to be noticed: *Belodella devonica* (Staufer), *Ligonodina* aff. *salopia* (Rhodes), *Ozarkodina media* Walliser, *O. typica denckmanni* Ziegler, *Pelekysgnatus furnishi* Klapper, *Polygnatus foveolatus* Philip., Yakson, *P. linguiformis linguiformis* Hinde, *Spathognathodus remscheidensis* Ziegler, S. *optimus* Moskalenko, *S. steinhornensis optimus* Moskalenko (Lower

Devonian); *Polygnatus docorosus* Stauffer, *P. pseudofoliatus* Wittekindt (Middle Devonian); *Angulodus waleathi* (Hibbard), *Palmatodella delicatula* Ulrich, Bassler, *Palmatolepsis delicatula delicatula* Branson, Mehl., *P. gigas* Miller, Youngquist, *P. triangularis* Sannemann, *Polygnathus communis* Branson, Mehl, *P. normalis* Miller, Younquist, *Spathognathodus bidentatus* Bischoff, Ziegler (Upper Devonian).

Besides the elements already discussed for North Dobrogea outcrops, namely: relationships to main magmatic phases, the facies, correlable to that of Culm, the presence of an *Asterocalamites* fragment and some coral remains in the limestone lenses, the present paper adds another two:

— the Carapelit "Beds" rework magmatic and sedimentary rocks up to the Famennian (included);

— the formation under discussion includes a carbonaceous facies with siderite, the latter being probably of Carboniferous, maybe post-Tournaissian age.

6. *Post-carapelitic sedimentary deposits*. The sedimentary realm of the North-Dobrogean Promontory includes other deposits besides the above-described sequences, whose age cannot be precised for the moment. This is the case of the conglomerates and of the grey-greenish sandstones in the Drăgănești area, of the sandstones, clay and more seldom of the garnet-red conglomerates in the Barcea-Fundeni and Vădeni-Zagna areas, as well as of the conglomerates and brown — garnet-red sandstones in the Matca area. These rocks have been provisorily assigned to the Werfenian (through analogy with the Bogza and Mihai Bravu sequences), but part of them might equally be Permian or Carboniferous, as they sometimes display slight metamorphic changes or antecarapelitic igneous rock reworkings (aspects also present in the Carapelit Formation). Moreover, the respective conglomerates rework post-carapelitic elements specific to North Dobrogea, such as: the granites and the basic sequences of the Greci type, the quartzo-feldspathic porphyries of the Camena type and the red, hematitic porphyries of the Consul Hill eruption type located in the Upper Carnian-Lower Norian.

7. *Magmatic rocks*. The presence of intrusive and effusive magmatic rocks in the investigated area has been testified by numerous wells. Other magmatic bodies— the strongest as it seems — have not been crossed by wells, but are suggested by the existence of the contact metamorphic rocks and by magnetometric surveys.

The igneous rocks show variations from acid to basic, being predominantly represented by intrusive and very seldom by effusive terms (Fig. 1). Granites, granodiorites, pyroxenites, metagabbros and vein rocks (porphyries and lamprophyres) are the most widespread.

It is difficult to precise the moment of the magmatic event. This should be estimated within the geological framework of North Dobrogea (Mirăuță, Mirăuță, 1962; Mirăuță, 1966; Mureșan et al., 1971).

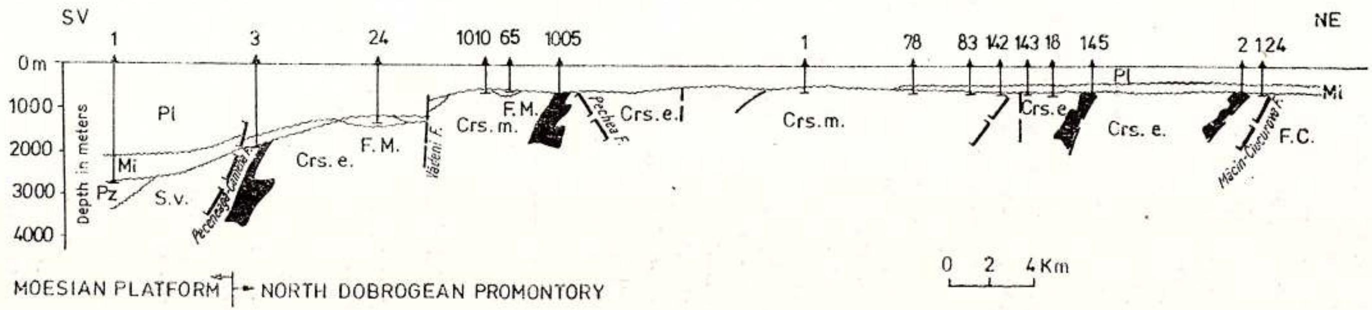

Fig. 2. — Geological cross-section in the North-Dobrogean Promontory; Pl = Pliocene; Mi = Miocene; Pz = Paleozoic; S.v. = Green-schists; Crs.e. = epizone crystalline schists; Crs.m. = meso- and catazone crystalline schists; F.M. = the Măxineni Formation; F.C. = the Carapelit Formation.

8. *Tectonical considerations*. The sedimentogene-volcanogene complex in the North-Dobrogean Promontory is part of the folded, more or less metamorphosed, formations category (6° — 85° dip). The metamorphism degree generally varies with rock age.

The drilling data and the similarity between the formations in the regional pre-Tertiary basement and those identified in North Dobrogea have led to the conclusion that the tectonic style of this buried basement resembles that in the outcrop area south-east of the Danube: structural-facial zones, tectonic overthrusted compartments, folds, wedges, lots of disjunctive accidents, eastward vergence (Fig. 2).

The North-Dobrogean Promontory, a buried paleohorst, is delimited N—E and S—W by the main tectonic dislocations Sf. Gheorghe — Cahul -Adjud and Peceneaga-Camena, respectively. The former is subsequent to the main regional dislocations, as it tectonically contacts, in a successive manner, the Pre-Dobrogean Depression to the Tulcea Unit (eastwards) and then, to the Măcin Unit (westwards). In the vicinity of Adjud, the line in question might encounter the Peceneaga-Camena Fault, thus forming the north-western boundary of the North-Dobrogean area.

Besides the most debated Peceneaga-Camena Fault, mention must be made of other longitudinal dislocations, such as the Măcin-Ciucurova and Pechea faults (Figs. 1,2), equally oriented eastwards. All along the Măcin-N Ciucurova Fault, there is a tectonic contact between the Carapelit Formation and the laminated granites (Frumușița) or the epimetamorphic deposits in the Boclugea Formation. Other longitudinal or transversal accidents result from the distribution of the various lithostratigraphic terms, of the metamorphite and from the geophysical (seismic) data collected at the Neogene/Basement contact.

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THE PRESENT STAGE OF PRE-JURASSIC DEPOSITS KNOWLEDGE IN SOUTH DOBROGEA [1]

BY

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South of the Capidava-Ovidiu Fault — a tectonic accident which delimits South Dobrogea from Central Dobrogea — there are Jurassic, Cretaceous, Paleogene and Neogene outcropping deposits. Their study has started even since the previous century. The knowledge of formations older than the Jurassic is, however, much more recent, owing to wells drilled during the last three decades. With the continuous progress of research, there have been identified, in a first stage, Silurian deposits at Costinești-Tuzla (Grigoraș, 1956), then the Mangalia Devonian (Răileanu et al., 1965, 1966, 1967) as well as possible Ordovician within the same area have been referred to. Further on, there have been signalled the Comana (Năstăseanu, Paraschiv, 1973) and the Negru Vodă Carboniferous (Paraschiv et al., 1973), while lately (Paraschiv, Dăneț, 1981) the presence of the Triassic at Independența and possibly at Castelu has been paleontologically substantiated. Out of isolated papers, published during the last quarter of a century, the conclusion is reached that the South Dobrogea pre-Jurassic is represented by the Ordovician (?), Silurian, Devonian, Carboniferous and Triassic. A general view on the succession and distribution of these deposits is not available yet; that is the reason why the present paper attempts such a synthesis.

The South Dobrogea is a part of the Moesian Platform, a unit located between the Carpathian and the Balkanian realms. Therefore, the pre-Jurassic south of the Capidava-Ovidiu Fault should be located within the geological framework of the platform in question. This holds true all the more that most lithostratigraphic terms with their stratotype north of the Danube also extend throughout Dobrogea, south of the great river.

The pre-Jurassic deposit basement in South Dobrogea seems to consist of the catamesometamorphites making up the gneissic (archaic)

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[2] Ministry of Petroleum, Bucharest, Romania.

series and the Palazu Mare Series (500 — 1100 m thick), the latter considered as Lower Proterozoic, as well as of the anchymetamorphosed formations of the Greenschists Series, of an Infracambrian-Lower Cambrian age. In the above-mentioned constitution, the basement has been crossed only at Palazu Mare. This basement is uncertainly accepted to extend throughout Dobrogea.

The various published and unpublished studies on the Moesian Platform make room for the conclusion that the pre-Jurassic in South Dobrogea develops unevenly, its thickness varying between a few hundreds and 3500 m. Within this lithologically heterogeneous sequence, several lithostratigraphic units of a formation rank may be separated. These terms are briefly discussed further on.

1. *The Mangalia Orthoquartzites* are the first term of the cover and consist of quartzwacke, quartzitic sandstones, gray or black orthoquartzites and sericitic argillite intercalations. Out of the above-mentioned petrographic varieties, quartzites are prevailing.

The wells which encountered this formation, Mangalia 5083, Ciobănița 5054 and Cobadin 5067, have not entirely crossed the discussed arenitic sequence. The longest (523 m) profile opened has been signalled at Mangalia.

According to data recorded north of the Danube (Țăndărei), the orthoquartzitic formation seems to unconformably overlie the Greenschists Series and diachronically supports various terms of the upper lithostratigraphic unit — the Țăndărei Argillites.

The Mangalia Orthoquartzites have been crossed by wells on the eastern, raised margins (Fig. 1) of some major blocks, but their presence might be continuous in South Dobrogea.

The core samples from the Cobadin 5067 well, together with other samples which belong to the borehole Țăndărei 1052, contain a Tremadocian-Arenigian palyno-protistological association, out of which mention should be made of the species *Archaeohystricosphaeridium arenigum* Tim., *A. minor* Tim., *Cymatiosphera boulardi* Deunff., *Cymatiogalea polygonomorpha* Gorka, *C. bellicosa* Deunff., *Leiofusa* sp. etc. Taking into account the much thicker Mangalia Quartzites and the incomplete penetration of the series, the Mangalia Orthoquartzites are likely to include, besides the Ordovician, part of the Cambrian, too.

2. *The Țăndărei Argillites* comprise a sequence of predominantly pelitic, 11—495 m thick deposits, overlying the Mangalia Quartzites (in other parts of the Moesian Platform argillites directly overlie the metamorphic basement) and conformably, but diachronically support the Smirna Quartz-Sandstones Formation.

In South Dobrogea, the Țăndărei Argillites have been signalled by the boreholes Costinești 53, 62, 64, 67, Tuzla 5055, Biruința 5082 and Mangalia 5083. According to them and to other boreholes carried out in the region, the argillitic formation occurs in the sunken sectors, better sheltered against denudation. On the raised margins of the main tectonic blocks, the respective formation lacks because of erosion (Fig. 1).

The lithostratigraphic term under discussion is mostly composed of argillites (sericito-chloritic, sericitic, siltic, gritty, sideritic, occasionally bituminous, calcareous and slightly dolomitic, tuffaceous argillites), clays, graywacke and sparite intercalations.

The paleontological content of the Ţăndărei Argillites, consisting of macrofauna (graptolites, orthocone nautiloids, bivalve, gastropoda) and microfauna indicates a Ludlowian-Eodevonian, even Lower Eifelian age for the deposits in question. North of the Danube (Ţăndărei, Bordei Verde), the argillitic facies gets down to Lower Ordovician. The graptolites regis-

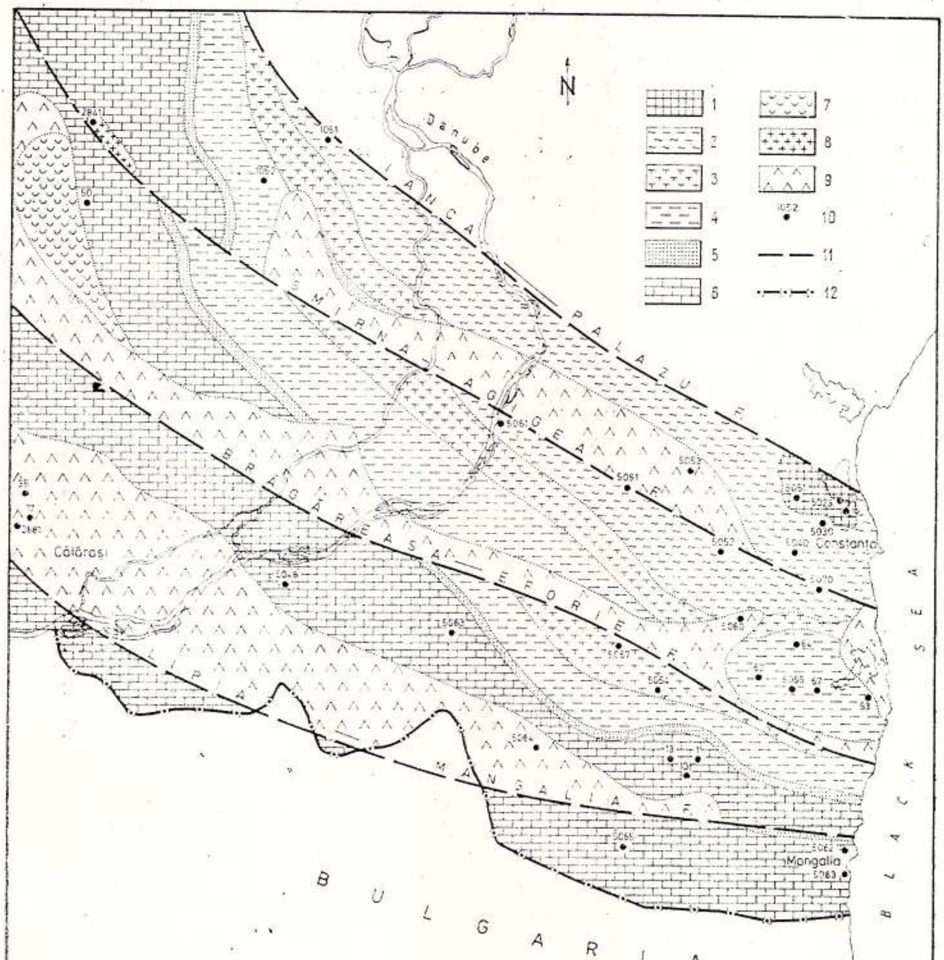

Fig. 1. — Pre-Jurassic formations distribution map in South Dobrogea and in the south-eastern part of the Romanian Plain. 1, crystalline basement; 2, anchymetamorphic basement (the Greenschists Series); 3, Mangalia Quartzites; 4, Ţăndărei Argillites; 5, Smirna Quartz- Sandstones; 6, the Călăraşi Formation; 7, the Vlaşin Formation; 8, the Ciureşti volcanogene-sedimentary Complex; 9, the Segarcea Formation; 10, well crossing pre-Jurassic formations; 11, main fault; 12, Romanian-Bulgarian boundary.

tered at Mangalia and Tuzla belong to the Neodiversograptus nilssoni and Lombograptus scanicus zones (I o r d a n, 1977) pointing to the Lower Ludlowian. The logs of the Mangalia 5083 well let us suppose the Upper Ludlowian, and even the Pridolian, terms found north of the Danube, but paleontologically unidentified in Dobrogea because of poor investigations. In exchange, the Eodevonian is proved by a rich macrofauna, determined (R ă i l e a n u et al., 1966 ; I o r d a n, 1977) in the Mangalia 5082 well. Out of the mentioned species, *Tentaculites gyracanthus* (E a t o n), *T. ornatus* S o w., *T. acuaria* R i c h., *Chonetes omaliana* (D e K o n i n k), *Schuchertella euzona* (F u c h s), *Delthyris dumontianus* (D e K o n i n k), *D. infans* D a h m e r – characterize the Gedinian; *Tentaculites straeleni* M a i l l., *Prolationus praelongus* L j a s, *Leptostrophia* cf. *index* H a v l., *Fimbrispirifer trigeri* (V e r n.), *Volinites velaini* (N u m.- C h a l m.), *Dignomia hunsurckiana* F u c h s, *Limoptera orbicularis* (O e h l.) – point to the Siegenian, while the trilobites *Pilletina asiatica* (V e r n.), *P. hammerschmidti* (R o e m e r), *P. pectinata* (R o e m e r), *Pseudocryphaeus prostellans* (R i c h t e r) and the bivalves *Cardyum inflatum* D i e n s t, *Goniophora nassoviensis* K a y s er – testify to the Emsian.

The Lower Devonian presence is supported by an abundent microfauna, identified in the Mangalia 5082 well. The Pontocypris delicata Wilson, Icriodus Woschmidti Ziegler and Dizygopleura trisinuata Van Peet biozones point to the Eodevonian and the transition to the Eifelian. This means that the Țăndărei Argillites also extend in the lowermost Eifelian, a fact actually proved by the macrofaunal forms : *Schellwienella umbraculum* S c h l., *Chonetes coronatus* C o n r a d, *Spirifer* (*Mucrospirifer*) *thedfordensis* S c h l., *Orthonata triplicata* F u c h s, etc. (R ă i l e a n u et al., 1965).

3. *The Smirna Quartz-Sandstones* are characterized by a predominantly arenitic sequence, overlying the Țăndărei Argillites, and probably conformably supporting the Călărași Formation.

Actually crossed by the Mangalia 5082 and 5083 wells, this formation might be present in other borehole profiles, too. The quartz-sandstones thickness varies between 120 m (the well 5082) and 648 m (the well 5083). The considerable difference in thickness between the two wells, over only 4 km distance, might have been caused by lithofacial variations, i.e. by arenitic facies invasion, especially on account of the lower argillitic term. In fact, the limits of all three geological formations discussed up to now have a diachronic character. Thu, the Țăndărei Argillitic Facies starts north-east of the Danube in the Lower Ordovician and transgressively progrades south-westwards in Dobrogea, up to the Ludlowian. The Smirna Quartz-Sandstone Facies starts in South-East Dobrogea in the Pridolian and regressively progrades north-eastwards up to the Eifelian.

The formation under discussion is made of quartzitic sandstone, gritty lithic sandstone with bioclasts, subfeldspathic or siltic grauwacke with siltic clay and shelly limestone interbeddings. The arenitic sequence has a variegated aspect, which the violet, garnet reddish, black and white shades lend to it. This aspect, together with the fish-fauna (Smirna 2841)

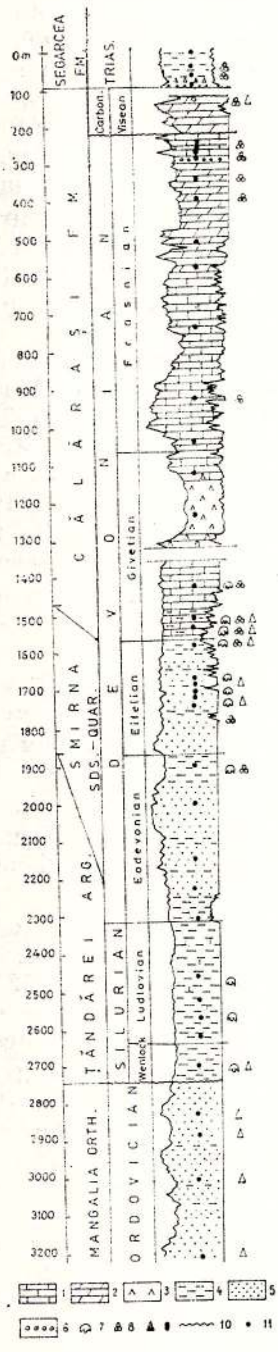

Fig. 2. – South Dobrogea pre-Jurassic synthetic profile. 1, limestones ; 2, dolomites ; 3, anhydrites, gypsum ; 4, marls, clays ; 5, sandstones, sands ; 6, conglomerates ; 7, macrofauna ; 8, microfauna ; 9, spore pollen ; 10, unconformity ; 11, core sample.

and the vegetal remains (Smirna 2341, Mangalia 5082) remind the Old Red Sandstone facies.

According to paleontological and log studies, the Smirna Arenitic Facies is specific to the Eifelian. Locally (Mangalia 5083, Călărași 2881), this facies begins, as already said, from the Eodevonian and it is not impossible for the quartz-sandstone sequence in the Mangalia 5083 well to equally comprise part of the Pridolian. At the same time, in the Mangalia 5082 well, the Eifelian terrigenous facies also extends into the Givetian base, in a clay and marly-limestone complex.

The Emsian, in the Smirna Facies, has been identified at Călărași (well 2881), where I o r d a n (1977) points ont numerous fossil forms, out of which *Pilletina asiatica* (V e r n.). The Mangalia 5082 well has registered the following species of brachiopoda and bivalvia, Eifelian in look: *Fimbrispirifer* sp., *Isortis* sp., *Ctenodonta* sp., *Pholadomiacus* sp., as well as the vegetal remains: *Pseudosporachnus* K r e j c i, P o t., B e r m., *Aneurophyton germanicum* K r ä u s, W e i l., *Calamophyton primaevum* K r ä u s, W e i l., *Hyenia* sp. The overlying horizon, made of black clays and marly-limestones, transiting to the carbonate term (the Călărași Formation), comprises macrofossil forms which might be assigned to the Givetian (R ă i l e a n u et al., 1966), such as: *Nowakia maureri* Z a g o r a, *Homoctenus* cf. *banusi* (B o u c e k, P r a n t l), *Fimbrispirifer* sp., *Isortis* sp., *Uncinulus* sp., *Ctenodonta* sp., *Bellerophon* sp.

All the above-described lithostratigraphic units make up the Ialomița Group.

4. *The Călărași Formation* consists of a thick, predominantly carbonate sequence normally overlying the Smirna Quartz-sandstones and, in most cases, unconformably underlies the Vlașin Formation and other younger stratigraphic terms.

In South Dobrogea, the Călărași Formation has been crossed by the wells Oltina 5048 (?), Negru-Vodă 5065, Comana 1, 10, 13, Mangalia 5082 and 5083. The Adamclisi 5063 well might be added too, if data supplied by it were not so imprecise.

The above-mentioned carbonate sequence has been completely crossed only in Mangalia, i.e. where subject to denudation effects. At Comana, the well 13 drilled 1227 m in these deposits without wholly crossing them. North of the Danube, in the Călărași 2881 well, the formation bearing the same name reaches 2520 m.

The paleontological content furnished by South Dobrogea wells demonstrates that the Călărași Formation includes the Givetian, the Frasnian and the Visean. The presence of the Famennian and Tournaisian has not been proved. Still, north of the Danube (Călărași), the Călărași Formation seems to represent a complete succession from Givetian to Visean included. In the Mangalia 5082 well, overlying the already-mentioned Givetian fauna clay complex, a marly-limestone horizon is first encountered, then a calcareous one. The following forms have been registered in the former: *Mucrospirifer mucronatus* (C o n r a d), *Punctatrypa nalivkini* H a v l., *Atrypa reticularis kuzbassica* R z n o s, *Chonetes* aff. *scitulus* H a l l., while *Athyris* aff. *nuculoidea* C o o p e r, *Chonetes rowei*

Schl., *Mucrospirifer mucronatus* (Conrad), *Spirifer audaculus* Conrad, *Spinocyrtia martinofi* (Stuck.), *Eleuterokoma leducensis* Krik. etc. occur in the latter. According to Răileanu et al. (1966), the first complex is likely to belong to the Givetian, while the second — to the Frasnian, possibly to the Upper Givetian.

The microfauna proves the presence of the Givetian in the Parathurammina and Icriodus curvatus (wells Comana 10, 13) and Pontocypris? bythocyproidea (Mangalia 5082 well) biozones. The Upper Givetian-Frasnian interval is attested by the Icriodus and Polygnatus Biozone, equally registered at Comana (1,13). Frasnian microfossils, grouped within the Palmatolepsis distorta Zone, have been reported from Comana (well 10) and Mangalia (well 5082). The Upper Visean has been identified at Comana (well 13) and Negru-Vodă (well 5065) by the Millerella and Valvulinella youngi (Brady) Zone. Palynologically, only the Givetian and the Frasnian have been identified — the former by the Dybolisporites, Calyptosporites, Ancyrospora and Hystrichosporites Zone, the latter by the Geminospora and Samarisporites triangulatus Zone. As to the Famennian and Tournaisian, apparently lacking in South Dobrogea, they are to be added to the sunken Călărași Zone with complete sedimentation (depocenter) and, possibly, to the north-eastern part of Bulgaria.

With respect to the Călărași Formation it is worth emphasizing that in the south-eastern extremity of South Dobrogea, the Dinantian might pass on into an argillitic facies, i.e. the Vlașin Formation facies, identified north of the Danube. Once again, this might be a matter of diachronic boundary. This hypothesis relies on the fact that in the Varna Depression, the Dinantian is terrigenous, sometimes paralic. This situation is attested by Ograjden 120 well, near the Romanian-Bulgarian boundary, south of Negru-Vodă.

5. *The Segarcea Formation*. In South Dobrogea, the Middle and Upper Carboniferous, Permian, Lower and Middle Triassic deposits have not been signalled yet. They insert in the stratigraphic succession north of the Danube and in north-eastern Bulgaria. Consequently, the Călărași Formation or the terms older than the Givetian are directly overlain by the Segarcea Formation, which is in fact an almost exclusively terrigenous sequence of marine, lagunary and continental origin.

The Segarcea Formation has been encountered by the Independența 5064 well and probably by the boreholes Castelu 5053 and Techirghiol 5069. None of the mentioned wells crossed it completely, the thickest section (103 m) being opened at Independența. The equivalent of the respective deposits is known both north of the Danube and over the Bulgarian territory where, concomitantly with their considerable thickening (in the well R1 Dulovo they reach about 800 m), they mostly change into limestones and dolomites.

As already shown, the Segarcea Formation transgressively and unconformably covers various older formations and also transgressively and unconformably underlies the Jurassic, locally the Upper Cretaceous, too.

Petrographic analyses from Independența 5064 well indicate, in the crossed profile, ferruginous feldspathic sandstone, grauwacke, ferruginous — occasionally mottled clay, calcareous ferruginous clay, marl — frequently gritty and ferruginous, calcareous — gritty, ferruginous argillite, gritty limestone.

The microfauna identified in the Segarcea Formation has been grouped into three paleocenoses: the Upper Ladinian-Lower Carnian Zone with *Epigondolella hungarica*, the Zone with *Stellatochara* and *Darwinulla* — typical of the Carnian — and the Zone with *Polytaxis seelandensis* and Ostracoda 800—which might correspond to the Rheto-Liassic. In other words, in South Dobrogea, the Segarcea Formation is of a Carnian age, with transitions both to the Upper Ladinian and the Retho-Liassic. It is not impossible for the sequence opened by the Techirghiol 5069 well to belong to the Permian—Lower Triassic i.e. to the Roşiori Formation.

The pre-Jurassic deposits are affected by a fault network, only partly known. Starting from the pattern conceived north of the Danube, it would be possible to prove, on the South Dobrogea territory, the presence of four regional NW—SE trending dislocations, namely: Ianca-Palazu (more or less an equivalent of the Capidava-Ovidiu Fault, at the pre-Jurassic level), Smirna-Agigea, Brăgăreasa-Eforie and Lipia-Mangalia. The fault throw might exceed 500 m. Taking into account the situation present at Palazu (V i s a r i o n et al., 1979) and Ianca, the main blocks are supposed to successively overthrust from south-west to north-east, along the separating faults. Within the large blocks (megablocks), there might be numerous disjunctive, still unidentified, accidents.

The presence of slight folds at the outcropping Lower Cretaceous and Jurassic level (C h i r i a c, 1979), associated to the low and moderate (5^0 — 40^0) values of strata dipping, determined by wells, also admit in South Dobrogea a plicative tectonics at the pre-Jurassic formations level. Both with disjunctive accidents and with plicative elements, the main geodynamic element seems to be the pushing north-east coming process, i.e. the East-European Platform displacement.

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QUESTIONS

N. Balteş. 1. What is, in your opinion, the extension of the pre-Jurassic formations of South Dobrogea in the continental platform of the Black Sea?

2. What do you think about the importance of the pre-Jurassic formations you have presented concerning oil potential source-rocks and also as reservoir rocks?

Answers : 1. The boreholes drilled in the field indicate differentiations in the distribution of the Paleozoic formations on an east-west trend, particularly nearby the sea shore. Consequently, as there is no evidence in the acquatorial sector, the spreading of the Paleozoic beyond the shore line is questionable.

2. The Paleozoic of the Moesian Platform (Ţăndărei Argillites, Călăraşi Formation and Vlaşin Formation) could provide industrial quantities of hydrocarbon as pointed out by the oil and gas occurrences in the mentioned formations, the Lower Triassic deposits as well as the results of geochemical analyses.

As regards the accumulation capacity of the Paleozoic, it has been proved by tests and geophysical and laboratory analysis for the Mangalia quartzites, Smirna arenaceous quartzites, and Călăraşi Formation.

DISCUSSIONS

M. Mureşan. I think that there are enough arguments to state that the Ianca-Palazu line represents an overthrust, as mentioned previously in several papers concerning the Dobrogea (Visarion, Săndulescu, Maier). First of all, the fact that the Cocoşu Series and the Greenschist Series are heterogeneous isochronous deposits (Mirăuţă) indicates that they have been brought into contact by an important tangential movement (overthrust). Secondly, it has been proved at Palazu that the Carelian gneissic series overlies the greenschist series.

REVIEW OF THE PALEOZOIC-MESOZOIC OF NORTH HUNGARY[1]

BY

KÁLMÁN BALOGH[2]

The North Hungarian Palaeozoic-Mesozoic is outcropped in the Aggtelek Mts, i.e. the southern continuation of the South Slovakian Karst, in the Rudabánya Mts, further in the Szendrö-Uppony range and in the Bükk Mts. Tectonically, the first two mentioned mountains belong to the Silicicum, and the two latter ones, however, to the Bükkium (Fig. 1).

This paper gives a rapid glance over the stratigraphical changes resulted in by newer micropalaeontological and faciological studies in the named areas.

1. The unmetamorphosed Permian and Mesozoic of the Aggtelek Karst and its Slovakian continuation begins with the Perkupa Evaporite Formation of Upper Permian age (Pl. I). It is overlain by predominantly terrigenous "Seis Beds", then by "Campil Beds", being far richer in both carbonates and fossils. The base of the Middle and Upper Triassic carbonate platforms is formed by the dolomites and limestones of the Gutenstein Formation, but their bulk consists of Steinalm, Wetterstein, Tisovec, as well as Furmanec limestone.

The deep-water zones among the reef—and reef—lagoon facies are indicated in both Slovakian and Hungarian territory by Schreyeralm Limestone in the Illyrian and by Hallstatt Limestone from the Ladinian to the Norian. The grey and (at least partially) cherty Reifling Limestone was still known among the basinal facies.

But the borehole Szölösardó-1 crossed such a basinal sequence that strongly differs from all the mentioned ones. Namely, the dasycladacean-bearing Steinalm limestones transit upwards gradually to a grey-coloured and cherty dolomarl complex. The latter enclosing *Gondolella regalis* Mosher, a leading fossil of the Bithynian, presses down the age of its footwall far deeper as it was supposed till now. The deepest part of this complex contains a thin acid tuff layer, too.

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[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association 1981 September 8 – 13, Bucharest, Romania.

[2] Budapest, XIII. Visegràdi-u. 17. III. 6. Hungary.

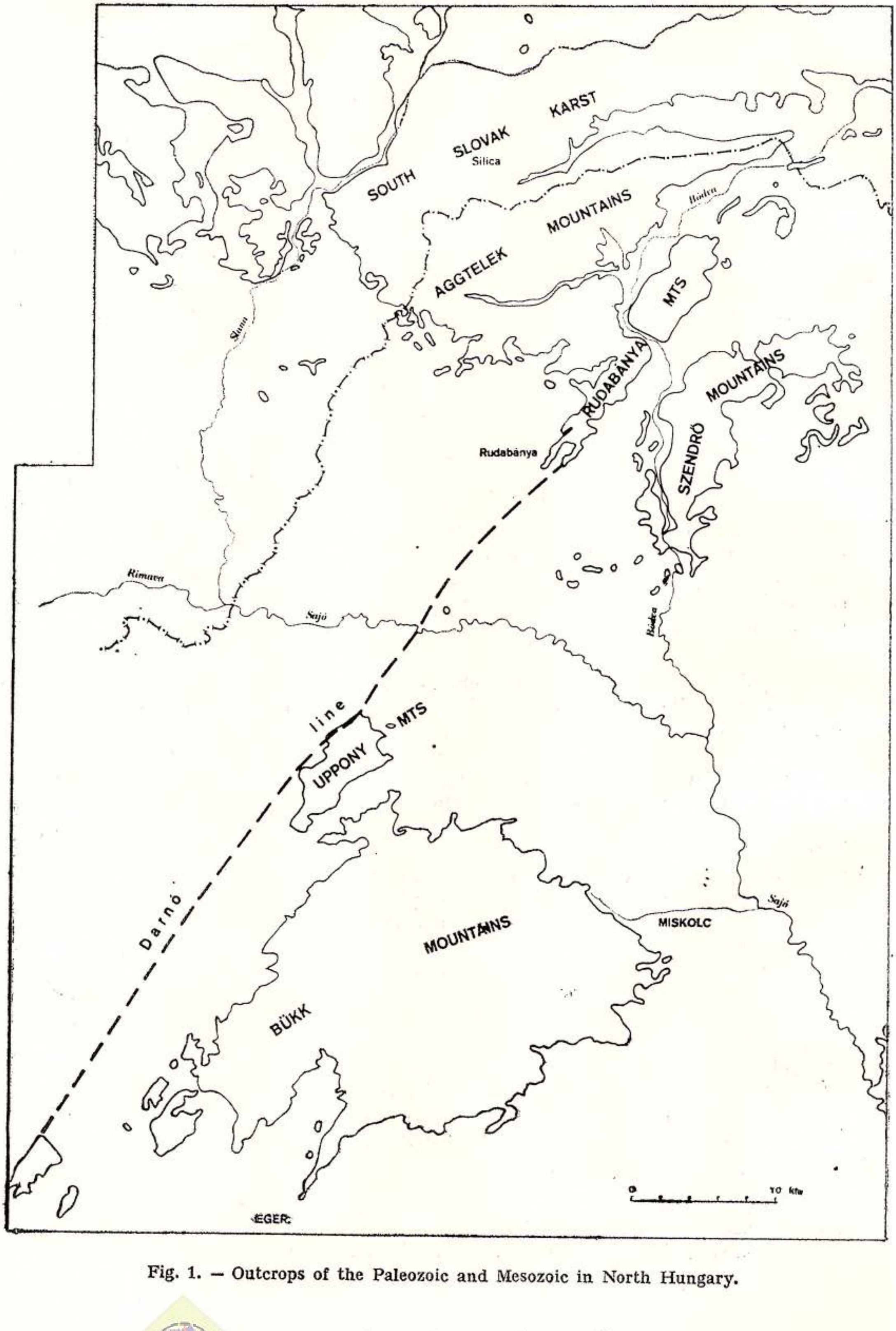

Fig. 1. — Outcrops of the Paleozoic and Mesozoic in North Hungary.

The dolomarl complex is separated from the overlying, varicoloured Nádaska Limestone – representing a submarine slope-sediment – by a fault zone. Therefore the latter begins here only with its upper Illyrian part. But in other places of the karst region the Pelsonian-Lower Illyrian horizons of the Nádaska Limestone were also evidenced by conodonts. Higher up, the grey-coloured Cordevolian-Lower Tuvalian Szölösardó Marl Formation, totally unknown hitherto in the Slovakian/Hungarian karst region, then the Tuvalian part of the Pötschen Limestone were crossed.

		Local assemblage biozones		Substages
PÖT-SCHEN LIM.	16	M. communisti–M. nodosus	/≈ Kl. macrolobatus/	TUVALIAN
	15	M. angustus	/≈ Tr. subbullatus/	
	14	G. tadpole	/≈ Tr. dilleri/	
SZÖ-LÖS-ARDÓ	13	Gl. malayensis	/≈ Sirenites + Tr. austriacum/	JULIAN
	12	G. polygnathiformis	/≈ Tr. aonoides/	CORDEVOLIAN
NÁDASKA LIMESTONE	11	G. foliata n. ssp.	/≈ Tr. aon/	
	10	M. mostleri–M. mungoensis	/≈ Fr. sutherlandi/	
	9	G. foliata n. ssp.–G. excelsa		LANGOBARDIAN
	8	M. hungaricus	G. n.sp. D	
	7	M. truempyi		FASSANIAN
	6	G. trammeri–G. transita		
	5	G. trammeri–G. excelsa		
	4	G. excelsa		
	3	G. excelsa–G. constricta		
	2	G. cornuta		ILLYRIAN
	1	G. constricta		

Fig. 2. – Local assemblage zones above the lower boundary of the Nádaska Limestone in the borehole Sözlösardó – 1 after Kovács and Balogh (1981).

In these basinal sediments 16 local assemblage biozones are distinguished (Fig. 2) by means of conodonts as well as macrofossils (*Daonella cassiana* Mojs., *Halobia rugosa* Mojs., *Austrotrachyceras* sp., *Sirenites* sp.). The naturally outcroped part of the Pötschen Limestone enclosing *Halobia styriaca* Mojs. is already of Lower Norian age; its top, however, may reach up to the Upper Norian.

On the Hungarian territory, the Rhaetian sediments (i.e. the Kössen Beds of Drnava) and the Zlambach Marls of Silicka Brezova are not – or only in a very small extension – outcropped.

The fact that the Hungarian part of the karst region was also transgressed by the Jurassic sea after a gap at the beginning of the Liassic, was already supposed on the basis of remnants of the Hierlatz, Adneth and spotted marl facies as well as the varied Bathonian-Callovian radiolarites found in some places of the South Slovakian Karst. However, only the studies of M. Sýkora and M. Mišík (1980) on the pebbles of Gosau and Miocene conglomerates have made evidence that the Jurassic

PALAEOGENE			? Gap ?
CRETACEOUS		SENONIAN	Fresh-water limestone
		MIDDLE AND LOWER CRETACEOUS	?
JURASSIC	MALM	TITHONIAN	*Clypeina* -bearing limestone
		KIMMERIDGIAN	Light-coloured oncolithic limestone
		OXFORDIAN	Cherty limestone
	DOGGER	CALLOVIAN	Varied radiolarites
		BATHONIAN	
		BAJOCIAN	Spotted limestone and spotted marl
		AALENIAN	
	LIAS	TOARCIAN	
		PLIENSBACHIAN	Adnet limestone; Hierlatz limestone
		SINEMURIAN	Gresten beds
		HETTANGIAN	Gap

Fig. 3. — Stratigraphy of the Jurassic and Cretaceous sediments of the Aggtelek Karst and its continuation in South Slovakia completed with new data of Prof. Mišik.

sequence embraces also the Malm (taken as a whole), and it ends with a shallow-marine Tithonian, and the Upper Cretaceous is represented by fresh-water limestones (Fig. 3).

2. Up to the Pelsonian, the nonmetamorphosed Mesozoic of the Rudabánya Mts is substantially identical with that of the Aggtelek Karst. The Steinalm Limestone, however, is also overlain here by several basinal facies classified as Ladinian or Lower Carnian in 1949. After newer data, however, this basinal sequence reaches as high as the Liassic, where it is accompanied by a quartz-porphyry volcanism.

It is possible that some members of the slightly metamorphosed Meliata sequence underlying the Silicicum are present also in the Rudabánya Mts and in the tectonic zone of the upper Bódva Valley.

The age and appurtenance of the diabase serpentinite and natron-gabbro masses in the upper Bódva and the Zósva Valley to a distinct tectonical unit is also questionable. In most instances they appear as enclosed into an evaporitic complex considered as old as Late (?) Permian. But they form only smaller bodies cut out from their original connections. Their K/Ar age runs between 197 and 212 m.y. A magmatism of similar character is hardly imaginable in the Silicicum.

Summarizing, the following theses seem to be evident.

a) The nonmetamorphosed Permian and Triassic of the Silicicum is very similar to those of the Hallstatt nappe of the Northern Limestone Alps.

b) With its Jurassic featuring a North Alpine character, the Silicicum forms a tectonic unit which is independent of the ranges of the Central Transdanubian Mountains.

c) In contradiction to this, the Mesozoic of the Silicicum may be interlinked with the slightly metamorphosed Meliata sequence through the nonmetamorphosed Rudabánya-type series.

3. In consequence of the conodont-findings of Kozur and Mock (1979) and mainly of Kovács, Kozur (1980) the age of the Szendrö and Uppony sequences (Fig. 4) must have been significantly modified. They are slightly metamorphosed. The degree of the metamorphism varies by zones, but it is never stronger than the quartz-albite-muscovite-chlorite subfacies of the greenschist facies.

The oldest member is the light-grey Nekézseny Limestone with traces of contemporaneous submarine diabase volcanism. Previously it was considered of Middle Triassic age; it belongs to the Lower Devonian (Kovács, 1980).

The Middle Devonian-lowermost Frasnian is represented by the dark-grey Szendrölád Limestone Formation bearing *Heliolites*, *Favosites* *Pachyfavosites*, *Thamnopora*, *Gracilopora*, *Alveolites*, *Syringopora* and conodonts. The upper part of this formation seems to be interfingered with the dark-grey Irota Phyllites, the uppermost beds of which reaches as far as Lower Famennian. The light-grey Abod Limestone (and perhaps also the light-coloured parts of the Borsod Limestone in the Szendrö Mts as well as the light-grey Uppony Limestone) can be placed to the Frasnian /Famennian boundary, but — for the time being — only the age of the Abod Limestone is evidenced by Kovács, Kozur (1980).

Individual parts of the "Lázbérc Formation" in the Uppony Mts consisting of schists, siltstones, grey limestones, diabase masses and tuffs bear Upper Devonian conodonts (Kozur, Mock, 1977 a). The bulk of the formation contains no fossils, its other parts belong to the Lower Viséan, its uppermost beds, however, seem to be as old as Bashkirian. The age of the overlying Tapocsány Formation, forming a thick, dark-grey schist and siltstone complex with manganese layers and radiolarites, but without limestones interbedded, is not definitely known. Its deeper-

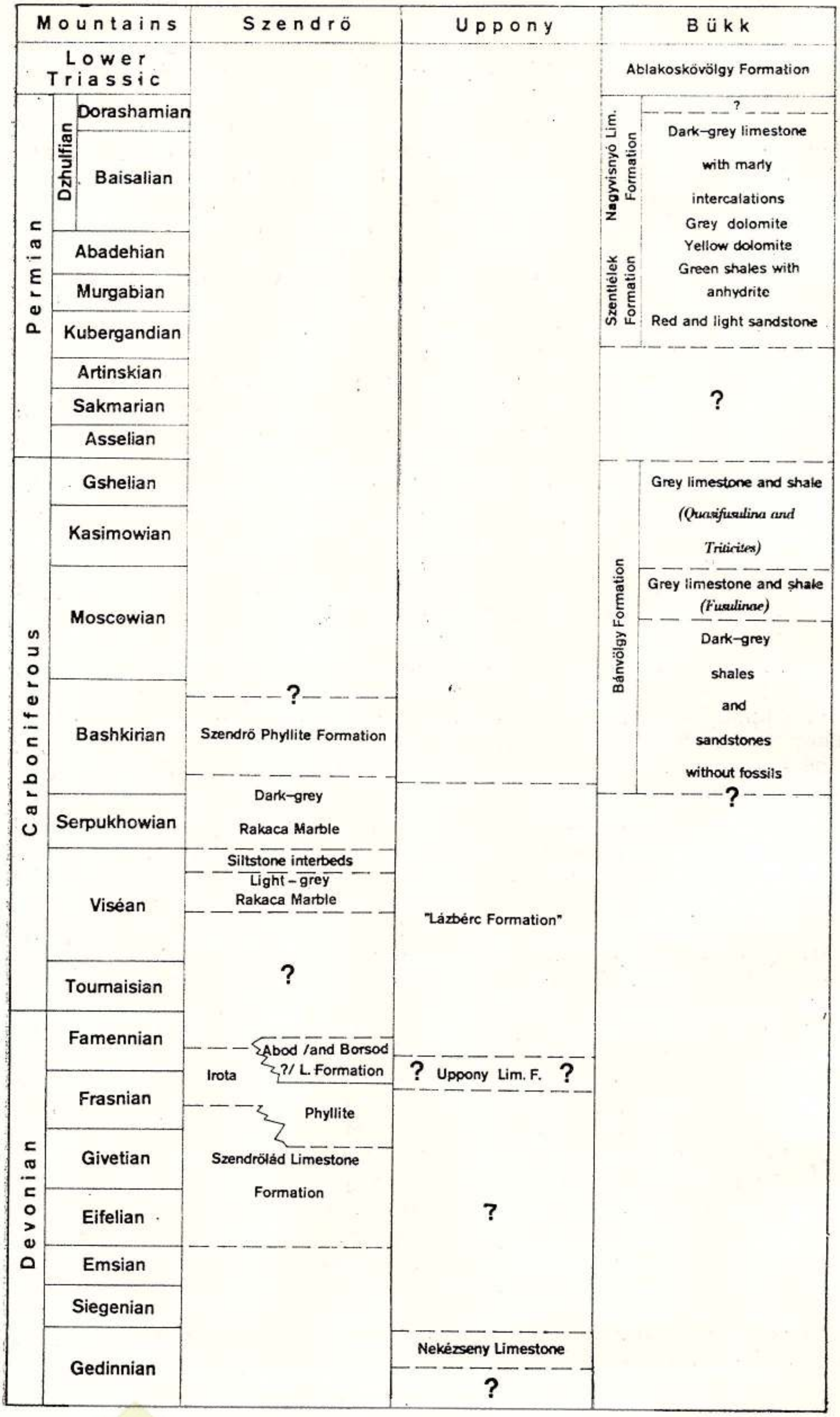
Mountains
Szendrö
Uppony
Bükk
Lower Triassic
Ablakoskövölgy Formation
Permian
Dzhulfian
Dorashamian
Baisalian
Abadehian
Murgabian
Kubergandian
Artinskian
Sakmarian
Asselian
Nagyvisnyó Lim. Formation
Dark-grey limestone with marly intercalations
Szentlélek Formation
Grey dolomite
Yellow dolomite
Green shales with anhydrite
Red and light sandstone
?
Carboniferous
Gshelian
Kasimowian
Moscowian
Bashkirian
Serpukhowian
Viséan
Tournaisian
Bánvölgy Formation
Grey limestone and shale (Quasifusulina and Triticites)
Grey limestone and shale (Fusulinae)
Dark-grey shales and sandstones without fossils
Szendrö Phyllite Formation
Dark-grey Rakaca Marble
Siltstone interbeds
Light-grey Rakaca Marble
"Lázbérc Formation"
Devonian
Famennian
Frasnian
Givetian
Eifelian
Emsian
Siegenian
Gedinnian
Abod /and Borsod ?/ L. Formation
Irota
Phyllite
Szendröládi Limestone Formation
Uppony Lim. F.
Nekézseny Limestone

Fig. 4

marine facies differs from the other Carboniferous sequences of the Bükkium.

Undoubted Upper Viséan, Serpukhowian and lowermost Bashkirian were found by Kovács and Kozur (1980) in the Rakaca Marble Formation of the Szendrö Mts, being divisible by means of a shaly intercalation into a lower, light-grey part and upper, dark-grey one. Its cover, the dark-grey Szendrö Phyllite Formation consisting of a rhythmic alternation of schists and coarse-grained metasiltstones intercaled by grey limestone beds and olistolites, belongs in all probability to the Bashkirian.

The present data are not sufficient to reach clear conclusions as regards the age of the gaps and the metamorphosis of the mentioned sequences. The manifestations of the Hercynian metamorphism are undistinguishable from those of the Alpine metamorphosis.

4. The profile of the Bükk Mts begins with a thick alternation of grey sericitic shales, silt- and sandstones (Fig. 4). Unfortunately, their immediate connection with the Uppony Mts could not be stated because of an important thrust plane and several faults; whereupon the connecting area is covered by Miocene sediments. The lowermost shales being free from fossils can be classified as Upper Serpukhowian-Bashkirian. They are similar to the South Alpine Hoochwipfel flysch, but their ages not known precisely. This lower part is separated by coarser-grained silt- and sandstones from the higher shales containing crinoids, brachiopods, gastropods, trilobites, plant-remnants and *Hemifusulina moelleri* Rauser that make probable the deepest horizons of the Moscowian stage.

The Upper Moscowian, Kasimowian and Gshelian shales and limestones include a rather rich fauna (Fusulinids, *Rugosa*, Chaeteids, Gastropods, Bivalves). This part of the section contains local lenses of lydite-bearing conglomerates and quartzose sandstone beds reminding of the Auernig Facies of the Carnic Alps and the Dinarides.

It did not succeed to prove the Lower Permian. Likewise, the time span of the evaporitic series overlying different horizons of the Middle and Upper Carboniferous and characterized by red and light-grey sandstones, green shales, anhydrite-layers and (on the top) yellow or grey, early-diagenetic dolomites was left open. However, it seems to be comparable with the Košna or the Gröden beds. The upper boundary of this Szentlélek Formation is much clearer because the overlying Nagyvisnyó Limestone includes a rather rich Baisalian flora and fauna, moreover some Dorashamian microfossils. This fact consolidates the old supposition concerning the uninterrupted transition between the carbonates of the Permian and Scythian.

After the Lower Triassic and Anisian the formation of the Middle Triassic carbonate platform is interrupted in the Illyrian (Pl. II). This is proved by the intercalated dolomite-breccias being comparable with the Uggowitz breccia and by a marine porphyrite volcanism. After this acci-

Fig. 4. – Stratigraphy of the Paleozoic complexes of the Szendrö, Uppony and Bükk Mts completed with the new results of Kovács.

dent the forming of the carbonate platform continued up to the end of the Ladinian. The Carnian sericitic shales with diabase and cherty limestone intercalations refer to deepening of the sea. But from the start of the Norian probably to the end of the Triassic, the grey, deep-water Felsötárkány Limestone, including also products of a diabase and quartz-porphyry volcanism, is interfingered with different light-coloured (partly coral-bearing) platform facies. The succession of the Jurassic rocks incorporated previously with the Carnian sericitic shales is not cleared as yet. Appearing mainly in the SW Bükk, they contain, beside a flysch-like alternation of dark-grey shales, silt- and sandstones also siliceous and manganiferous shales or oolithic limestones. In places they include Middle Liassic foraminifers and Dogger radiolarians. In the Szarvaskö-Darnóhegy zone they include also masses of diabases, spilites and pillow-lavas, as well as gabbros and peridotites which seem to belong to a probably Jurassic magmatic activity.

The slight Alpine metamorphism of the Paleozoic and the Mesozoic of the Bükk Mts varies by zones. But the Gosau Beds at Nekézseny were not touched by this metamorphosis. Subsequent to the post-Gosau movements the Upper Eocene transgresses over the evolved structural elements.

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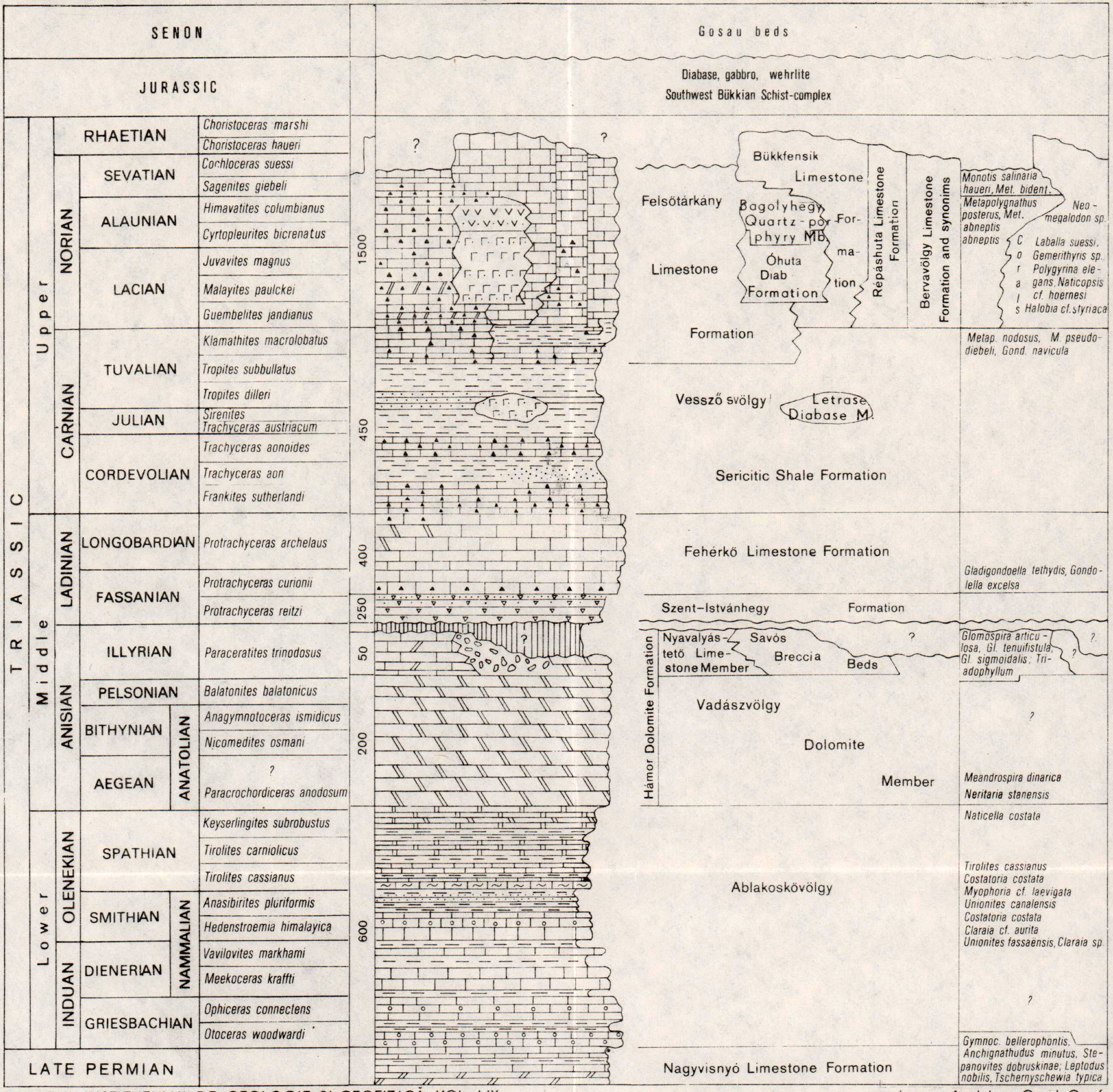
LITHO - AND BIOSTRATIGRAPHIC SKETCH OF THE TRIASSIC IN THE AGGTELEK MOUNTAINS (North Hungary) AFTER BALOGH (1980)

Imprim. Atel. Inst. Geol. Geof.

STRATIGRAPHY OF THE MESOZOIC OF THE BÜKK MOUNTAINS MODIFIED BY MEANS OF THE NEW DATA OF KOZUR



Imprim. Atel. Inst. Geol. Geof.

FORMATIONS SALIFÈRES DE LA PLATE-FORME MOESIENNE (ROUMANIE)[1]

PAR

DUMITRU PARASCHIV [2]

Les travaux de forage effectués dans la Plaine Roumaine ont signalé des dépôts salifères dans trois secteurs de la Plate-forme moesienne : dans la partie nord-ouest, à Bibești, dans la zone centrale, à Cartojani et dans la partie sud, délimitée à peu près par les rivières d'Argeș et d'Olt (fig. 1). La séquence sédimentaire à sel de Bibești, représentant l'extension vers le sud de la zone évaporitique badénienne qui se développe dans le flanc interne, plissé, de l'Avant-fosse carpathique, n'a pas constitué une surprise. Ces dépôts, connus depuis longtemps à la surface aussi bien que par les forages, ont fait l'objet de nombreux études et rapports. Par conséquence, le travail ci-présent sera concentré sur les formations salifères des secteurs sud et central de la Plate-forme moesienne.

Les premiers indices sur les halites du fondement de la Plaine Roumaine proviennent de son secteur sud. Elles datent de 1956, lorsque le puits 3 Putinei, arrivé à 2746 m de profondeur, a identifié, à l'aide d'une carotte mécanique, de l'anhydrite massive et du sel blanc grossier.

A cause de certaines difficultés techniques, on n'a plus répété le carottage mécanique et on a arrêté le forage à 2839 m. Selon la diagraphie électrique, les bancs à sel et à anhydrites commencement à environ 2720 — 2740 m (planche) et se succèdent jusqu'à la profondeur finale. Dans les conditions ci-dessus, le puits 3 Putinei n'aurait pas pu traverser la séquence évaporitique en entier, cette dernière allant au-dessus de 2839 m.

On aurait bientôt confirmé le sel de Putinei grâce à trois autres forages, achevés en 1962. Il s'agit des puits 3 Lița, 60 Chiriacu et 100 Vlașin, tous situés dans le même secteur sud de la Plate-forme moesienne.

Dans le puits 3 Studina, on a rapporté du sel blanc grossier aux éléments d'anhydrite, dans la carotte récupérée de l'intervalle 2340 — — 2341. On n'a continué le forage que le long d'un seul mètre ; ainsi, il est très probable que le puits se soit arrêté dans le sel. Selon les mêmes

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[1] Note présentée au 12 ème Congrès de l'Association géologique Carpatho-Balkanique, 8 — 13 septembre 1981, Bucarest, Roumanie.

[2] Ministère du Pétrole, Bucarest.

diagraphies électriques, les bancs de sel à anhydrites pourraient se développer à partir d'environ 2300 m vers le fond. Donc, ce puits n'a pas entièrement traversé la séquence évaporitique.

Dans le puits 60 Chiriacu (fig. 1 planche), on a extrait une première carotte à 2661–2662 m, au sel blanc, partiellement rouge et aux impurités marno-calcaires gris-verdâtres, ayant l'aspect d'une brèche. On a

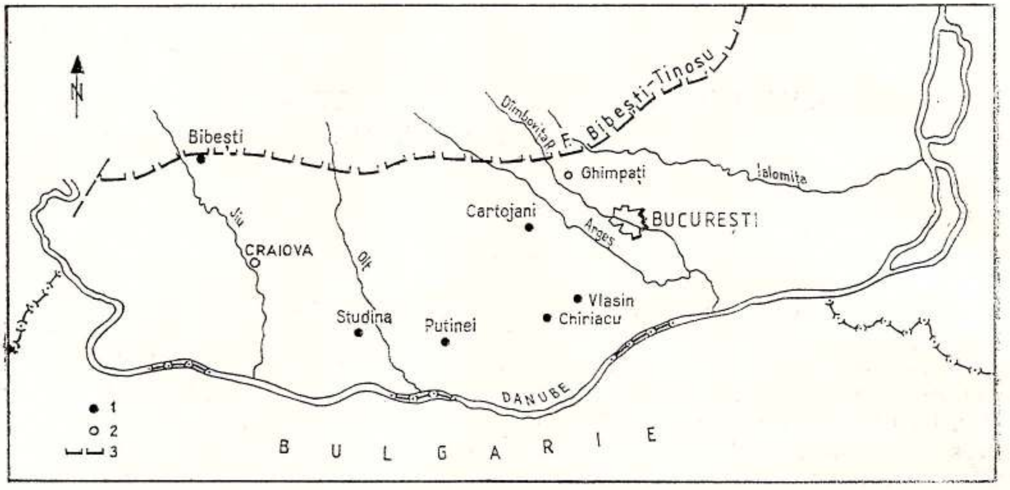

Fig. 1. – La plate-forme moesienne et la position géographique des puits qui ont rencontré des formations salifères : 1, forages ; 2, localités ; 3, ligne de Bibești-Tinosu.

récupéré du sel blanc grossier, aux intercalations dolomitiques et anhydritiques à la profondeur de 2915 – 2916 m. Entre ces deux prises de sel, on a également extrait deux carottes des intervalles de 2733 – 2734 m et de 2848 – 2849 m qui ont représenté des marnes, respectivement des calcaires aux diaclases remplies, dans les deux cas, par de cristaux de sel. Le forage a continué sur les 278 m sous la dernière carotte à sel, sans signaler pourtant de nouvelles halites. Selon le diagramme électrique, corrélé aux carottes mécaniques, on voit que la formation salifère se développe sur une épaisseur d'environ 360 m (2640 – 3000 m). Le long de cet intervalle, on pourrait séparer trois bancs masifs de sel associé aux anydrites. La formation à sel est bordée de calcaires et de dolomies. On y trouve également des dépôts carbonatés et marno-calcaires entre les bancs de sel, ce qui suggère que la formation salifère constitue un épisode dans la séquence carbonatée triasique.

Le puits 100 Vlașin a rencontré, à 2545 – 2547 m, du sel blanc aux tâches rougeâtres, associé aux débris calcaires et marno-gréseux, brun-rougeâtres. A 2802 – 2804 m on a signalé de l'argile aux intercalations anhydritiques, les diaclases des roches étant remplies de sel rougeâtre. Les diagrammes géophysiques du puits suggèrent que la formation à sel commence à environ 2520 m et s'étend jusqu'à peu près 2830 m. De là jusqu'à 4215 m (la profondeur totale du forage) on a traversé successivement les roches carbonatées du Trias, les grès et les argiles quasicontinentaux permiens (possible triasiques inférieures aussi) et les pélites charboneuses du Namurien.

Tous les quatre forages ci-discutés appartiennent à une zone dépressionnaire de la Plate-forme moesienne, connue sous le nom de „la Dépression Roșiori-Alexandria''. Formée à l'époque du diastrophisme sudétique — sinon plus tôt, cette dépression de type plate-forme a fonctionné jusqu'à la fin du Trias. Le terme lithofacial le plus ancien connu de la dépression respective est constitué par la Formation de Călărași, une séquence presque exclusivement carbonatée, épaisse de 2000 — 2500 m, qui débute au Givétien et s'étend jusqu'au Viséen y compris. Localement, la formation en question peut être partiellement érodée.

La Formation de Vlașin y suit d'une manière discordante. Elle se compose d'argillites, d'argiles calcaires et de microgrès, plus ou moins charbonneux, grauwackes, sousgrauwackes, microconglomérats, dont l'épaisseur pourrait totaliser 1000 — 1200 m. L'âge de la Formation de Vlașin est namurien et westphalien inférieur.

Une séquence de dépôts quasi-continentaux, localement lagunaires, généralement brun-rougeâtres, suit dans l'échelle stratigraphique. Elle est représentée par des argiles, des silts, des grès, des microconglomérats, aux intercalations ou aux inclusions gypsifères et anhydritiques. Il paraît que la séquence en question, épaisse de 300 — 2500 m est d'âge permien et triasique inférieur.

Les dépôts marins-lagunaires (0 — 1200 m) généralement carbonatés, du Trias suivent transgressivement et d'une manière discordante. Formée de calcaires, dolomies, argiles calcaires, et plus rarement de grès calcaires et d'argilles, cette succession forme aussi, localement, des évaporites comme des anhydrites, des grès et du sel. Ces évaporites ont été signalées par les puits 3 Studina, 3 Putinei, 60 Chiriacu et 100 Vlașin. Selon un rapport récemment publié (P a r a s c h i v et al., 1978), les dépôts marins-lagunaires triasiques sont hétérochrones, c'est-à-dire leur dépôt a débuté au Trias inférieur (synchrone au Werfenien) dans le sud de la Plate-forme, tandis que dans le secteur nord de la région, l'épisode carbonaté débute plus tard, pendant, l'Anisien. Il ne faut pas ignorer le fait que vers les périphéries de la Plate-forme moesienne, la formation marine-lagunaire a été érodée, parfois entièrement. Dans le secteur Cartojani par exemple, l'érosion est arrivé jusqu'à la moitié inférieure de l'Anisien (planche).

Le Trias s'achève par une séquence de dépôts quasi-continentaux, presque exclusivement terrigène, épaisse de 0 — 900 m. Composée d'argiles, argiles calcaires, rarement calcaires, microgrès, grès, microconglomérats et localement, de magmatites, associés aux intercalations et aux nids anhydritiques et gypsifères ; la séquence respective, que l'on appelle la Formation de Segarcea (P a r a s c h i v , 1981), se caractérise par une couleur rouge dominante. Elle se trouve en rapports de discontinuïté avec le terme lithostratigraphique sous-jacent aussi bien qu'avec le Jurassique qui la recouvre.

Les quelques éléments ci-mentionnés tâchent de définir la position de la séquence salifère dans le contexte stratigraphique de la Plate-forme moesienne et surtout dans celui de la Dépression Roșiori-Alexandria.

A ce sujet-là, une importance particulière gagnent les puits 60 Chiriacu et 100 Vlașin qui ont traversé des séquences assez complètes, permettant ainsi d'établir l'âge des dépôts à sel et de les localiser dans la succession triasique. En outre, on a préservé tout le matériel documentaire fourni par le puits 100 Vlașin, y compris les carottes mécaniques récemment soumises aux réexaminations.

Dans l'un des rapports récemment parus (Paraschiv et al., 1978) on mentionne que l'échantillon de terrain (la carotte) récupéré du puits 100 Vlașin sur l'intervalle 2802 — 2804 m (la séquence salifère) contient une association palynoprotistologique [3], c'est-à-dire une palynozone à *Angustisulcites* et *Triadispora*, spécifique pour l'Anisien. Cette association comprend les formes suivantes : *Duplicisporites* sp., *Punctatisporites* sp., *Verrucosisporites* sp., *V.* cf. *thuringiaccus* Mäd., *Alisporites* sp., *A. grauvogeli* Kl., *A. minutisaccus* Clar., *Angustisulcites* sp., *A. grandis* (Freund), *A.* cf. *Klaussi* (Freund), *Chordasporites* sp., *Falcisporites* sp., *F. snopkovae* Viss., *Illinites* sp., *Klaussipolenites* sp., *K. schaubergeri* (Paut., Kl.,), *Limitisporites* sp., *Microcahrydites doubingeri* Kl., *M. sittleri* Kl., *Platysaccus* sp., *P.* cf. *triassicus* Mäd., *Pityosporites* sp., *Potonieisporites* sp., *Sahnisporites* cf. *reticulatus* Mäd., *S. thomassii* Mäd., *Striatites* sp., *S. richteri* (Kl.), *Striatoabietites* cf. *richteri* (Kl.), *S.* aff. *aytugii* Vis., *Sulcatisporites reticulatus* Mäd., *Triadispora* sp., *T. crassa* Viss., *T. epigona* Kl., *Vitreisporites pallidus* (Reiss.) Paut., *Voltziaceaesporites heteromorpha* Kl.

Environ 90m plus bas, la carotte provenant de 2890 — 2890,5 m bien qu'également d'âge anisien (selon la microfaune), ne contient plus du sel.

Au dessus des deux échantillons déjà mentionnés, la carotte située à 2545 — 2547 m se fait remarquer par un contenu microfloral ladinien (palynozone à *Ovalipollis* et *Taeniaesporites sulcatus*) qui comprend les formes : *Apiculatisporites* sp., *Calamospora* aff. *mesozoica* Coup., *Punctatisporites* sp., *Verrucosisporites* sp., *Alisporites* sp., *A.* aff. *grauvogeli* Kl., *Chordasporites* sp., *C. magnus* Kl., *C. singulichorda* Kl., *Classopollis* sp., *Cuneatisporites* sp., *C. radialis* Lesch., *Falcisporites* sp., *Gardenasporites* sp., *Gigantosporites* sp., *G.* cf. *hallstatensis* Kl., *Gingkocycadopites* sp., *Klausipollenites* sp., *Limitisporites parvus* Kl., *Microcahryidites doubingeri* Kl., *Ovalipolis* sp., *O. ovalis* Kl., *O. lunenzis* Kl., *O. grebae* Kl., *Parcisporites* sp., *P. cirratus* Lesch., *Sulcatisporites* sp., *Taeniaesporites* sp., *T.* cf. *sulcatus* (Paut.), *T.* cf. *alutus* Kl., *Vitreisporites pallidus* (Reiss.) Paut., *Voltziaceaesporites heteromorpha* Kl.

Au même niveau on a rencontré une microfaunelad inienne (Paraschiv et al., 1978) composée de nombreuses espèces de foraminifères et d'ostracodes [4] : *Sorosphaera scabra* Trif., *Ammodiscus* sp., *Glomospira* sp., *G. irregularis* Lip., *Glomospirella* sp., *Tolypammina rotula* Guts., Treck., *Lituotuba indistincta* Trif., *Ammobaculites* sp., *Placopsilina lacera* Trif., *Placopsilius florae* Trif., *Gaudryina racenna* Triff., *G. triassica* Triff., *Gaudrynella kotlensis* Trif., *Pliammina densa* Pantič, *Agathammina austroalpina* Kris, *Meandrospira deformata* Sălaj, *M. insolita* (Ho), *M. pusilla* (Ho), *Nodosaria libera* Trif., *N. ordinata* Trif., *Dentalina hoi* Trif., *Pseudoglandulina rosenbergi* Ober., *P. simpsonensis* Tappan, *Lingulina Klebelsbergi* Ober., *L.* aff. *L. major* (B o r n.), *Trocholina acuta* O b e r., *Globigerina ladinica* O b e r., *G. mesotriassica* O b e r. (foraminifères); *Bairdia anisia* K o z u r, *Darwinula fragilis* S c h n., *Bythocypris triassica* K o z u r, *Triasselina bicuspidata* K o z u r, *Monoceratina minuta* K o z u r, *Pojanites striatus* K o z u r, *Lutkevichinella gruenndeli* K o z u r, *L. lata* K o z u r *L. (Cytherissinella) rectagona* K o z u r, *L.*(c) *schneiderae* K o z u r, *L. simplex* K o z u r, *Limnocythere triassica* K o z u r, *Telocythere tollmanni* K o z u r, *Speluncella sulcata* K o z u r, *Healalia (Hungarella) reniformis* (M e h e s).

On a également signalé des associations palynoprotistologiques et microfauniques dans la carotte de 2473—2475 m, située immédiatement au-dessus de la séquence salifère.

Les bulletins d'analyse plus anciens indiquent des éléments microfloraux similaires dans les puits 60 Chiriacu et 3 Studina. A Chiriacu, les carottes à sel provenues de 2661 m et 2915 m contiennent entre autres: *Ovalipolis* sp., *O. ovalis* K l., *O. grebeae* K l., *O. brevis* K r., *O. lunzensis* K l., *Taeniaesporites* sp., *T. sulcatus* P a u t., *Cuneatisporites* sp., *C. radicalis* L e s c h., *Parcisporites* sp., *Sulcatisporites* sp. Dans le puits 3 Studina, les carottes extraites des intervales de 2300—2301 m (à travers le sel) et de 2215—2216 m (au-dessus du sel) contiennent entre autres: *Apiculatisporites* sp., *Cuneatisporites radialis* L e s c h., *Ovalipolis* sp., *O. lunenzis* K l., *O. grebeae* K l., *Parcisporites* sp., *Sulcatisporites* sp., *Taeniaesporites* sp. etc.

Le contenu paléontologique des carottes examinées, aussi bien que la corrélation des profils lithologiques traversés par le puits (planche) suggèrent que le dépôt de la formation à sel a commencé pendant l'Anisien et a continué jusqu'à la fin du Landinien. Pendant tout cet intervalle, des dépôts carbonatés marins se sont accumulés à travers la Plate-forme moesienne, excepté la Dépression Roșiori-Alexandria où des lagunes ont pris naissance, localement et temporellement. Il s'ensuit que la séquence sedimentaire à sel représente une variation locale de faciès dans l'épisode carbonaté triasique de la Plate-forme. Cette conclusion vient de s'ajouter aux éléments déjà connus (dispersion, épaisseur, contenu paléontologique, etc.) présentées dans un rapport antérieur (P a r a s c h i v et al., 1978) pour soutenir l'idée que les dépôts marins-langunaires triasiques dans la Plate-forme moesienne soient assemblés dans une unité lithofaciale distinctive, c'est-à-dire „La Formation d'Alexandria". On pourrait nommer la séquence à sel „L'horizon de Chiriacu".

Les dépôts salifères de la zone centrale de la Plate-forme moesienne ont été mis en évidence par le puits 4517 Cartojani, récemment foré. Du Jurassique terrigène (la microflore du Bathonien supérieur — Callovien inférieur), ce dernier a pénétré l'Anisien (microfaune spécifique dans le puits voisin 10 Cartojani) et l'a traversé entre 2222 m et 2267 m. Dans la région de Cartojani, les calcaires et les dolomies correspondant au Ladinien et à l'Anisien supérieur ont été érodés de telle maniére que la Formation d'Alexandria reste uniquement représentée par l'Anisien basal. Au-dessous, on a rencontré un horizon de grès aux intercalations d'argiles et de silts rouges, épais d'environ 300 m (2267—2562 m). Dans plusieurs puits de la plate-forme, cet horizon contient une association palynoprotistologique triasique inférieure (association à *Lumbaldispora* et *Alisporites* *cymbatus*). A partir de 2562 m jusqu'au fond (4500 m), c'est-à-dire sur une épaisseur de 1938 m, le puits a pénétré une succession de dépôts continentaux-lagunaires, sans la traverser complètement pourtant. Ces der-

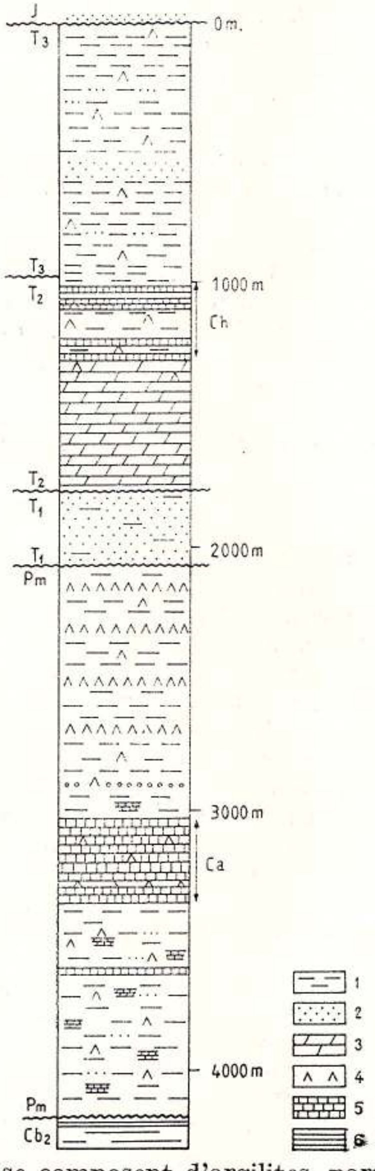

Fig. 2. – Profil synthétique du Permo-Trias salifère de la Plate-forme moesienne (selon les puits 100 Vlașin et 4517 Cartojani) : 1, argiles, marnes ; 2, grès, sables ; 3, dolomies, calcaires ; 4, gypses, anhydrites ; 5, sel ; 6, charbons ; Ch, l'horizon de Chiriacu ; Ca, l'horizon de Cartojani.

niers se composent d'argilites, partiellement calcaires, ferrugineuses, marnes ferrugineuses, microgrès et grès ferrugineux aux intercalations ou nodules de gypse et d'anhydrite. La pente des couches comporte un pendage de 6° – 10°. La carotte prise entre 3606–3610 m a indiqué du sel gemme blanc et rose aux impurités formées d'argile dolomitique gréseuse-ferrugineuse, de grès et de nodules d'anhydrites. Environ 20 m plus bas (3628—3633 m) on a de nouveau récupéré du sel blanc et rose aux impurités marneuses brunes. Les échantillons de terrain provenus des profondeurs au-dessous de 3800 m n'ont plus indiqué le sel, mais seulement les nodules gypsifères et anhydritiques. Selon la diagraphie géophysique du puits, l'horizon à sel correspondrait à l'intervalle de 3545 — 3830 m, ce qui signifie une épaisseur de presque 300 m. Si l'on tient compte des occurrences de gypse et d'anhydrites au-dessus de 3830 m, des intercalations minces et des nodules de sel pourraient également paraître jusqu'au fond du puits.

La carotte à sel de 3628—3633 m contient quelques formes microflorales et microfauniques qui indiquent l'âge ladinien. Etant donné l'inconsistance des éléments paléontologiques et la position stratigraphique contrôlée de la séquence à sel de Cartojani, on considère ces évaporites comme appartenant au Permien. On connaît également du sel permien au sud du Danube, dans le territoire bulgaire, à Mirovo.

A partir de l'âge, du contexte stratigraphique et de la nécessité de la faire distinguer du sel triasique moyen, on propose de nommer la séquence permienne aux halites „L'horizon de Cartojani".

On pourrait expliquer l'occurrence du sel permien dans la Plate-forme moesienne au nord du Danube par la formation d'un sillon orienté est-ouest, au contact de deux grands segments de l'écorce. Presque continental en exclusivité dans le reste de la plate-forme, le Permien a évolué à l'intérieur du trench vers un faciès lagunaire puissant, même lacustre, à calcaires (dans la zone de Ciurești). Par conséquent, il est possible que l'Horizon de Cartojani s'étend par intermittence le long du fosse mentionné, environ entre les méridiens Ghimpați (à l'est) et Craiova (à l'ouest).

En conclusion, les puits forés sur la Plate-forme moesienne ont rencontré trois niveaux stratigraphiques à sel. Le premier est localisé dans le Badénien et appartient en effet à l'Avant-fosse carpathique. Le deuxième niveau, l'Horizon de Chiriacu, correspond à la partie supérieure de l'Anissien et au Ladinien presqu'en entier. Le troisième, l'Horizon de Cartojani, appartient au Permien. En effet, il n'y a que ces deux derniers horizons qui soient propres à la plate-forme. La figure 2 exprime d'une manière synthètique le contexte stratigraphique des horizons salifères permo-triasiques.

[3] Déterminations faites par D. Beju.

[4] Déterminations effectuées par N. Dăneț.

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PROFILE SYNTHÉTIQUE DU PERMO-TRIASSIQUE SALIFERE DE LA PLATE-FORME MOÉSIQUE
(selon les puits 100 Vlaşin et 4517 Cartojani)

D. PARASCHIV. Formations salifères de la Plate-forme Moesienne (Roumanie).

Imprim. Atel. Inst. Geol. Geof.

THE GEOLOGICAL CONSTITUTION OF THE DANUBE DELTA[1]

BY

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Introduction

From the geological point of view, the Danube Delta represents the south-eastern part of the Pre-Dobrogean Depression, a unit of foredeep characters, comprised between the North Dobrogean Hercynian and Kimmerian Orogene and the East-European Platform. The young Delta sedimentary has been investigated by shallow wells (Liteanu, Pricăjan, 1963) and further geological surveys have been carried out, followed by drilling of several deep stratigraphic and structural wells. The results of these works have been synthetized in the oil industry by Pătruț et al. (1976) and this synthesis lies at the basis of the present paper.

1. Stratigraphic Considerations

Deep drillings in the Danube Delta have not reached the basement and its constitution is still a problem.

The sedimentary cover comprises Paleozoic to Quaternary deposits and may be separated into sedimentary cycles, delimited by stratigraphic gaps, out of which the most significant and general are those corresponding to the Upper Devonian-Permian, Lower Jurassic and Upper Cretaceous-Lower Miocene intervals.

A fault line, the Sf. Gheorghe Fault, roughly outlining the Sf. Gheorghe branch of the Danube, separates the Delta sedimentary from the North Dobrogean one.

1.1. *The Paleozoic sedimentary cycle.* In the north-eastern part of the Delta, the Rosetti well has been completed in a sequence with dolomites and limestones, after opening it over 300 m (Pl.—A). No fossil remains have been identified, but not far eastwards, in the Snake Island (USSR), fossiliferous Devonian limestones, of about 100 m in thickness,

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have been identified in a shallow well (B o g a e t et al., 1976) and it was estimated that these limestones plunge westwards into the Danube Delta area, being even better developed. Upper Devonian-Carboniferous carbonate deposits are equally mentioned north of the Danube, in the USSR (K a p t a n et al., 1963).

In the eastern part of the Delta, the North Lacul Roșu well has opened, on about 180 m in thickness, a grey-blackish sequence of feldspathic sandstones and silicified argillites (Pl.—A). No organic remains have been identified, but a vitrocrystallolitoclastic tuff (ignimbrite) identified in the lower part of the sequence suggests a Silurian-Devonian age. In the Snake Island (USSR), a thin layer of effusive rocks has been noted in the Ludlovian (B o g a e t et al., 1976).

No clear remarks can be made on these deposits real thickness and spreading. An area of limestones and dolomites rather amply developed seems to be located, however, between the Sulina and Chilia branches of the Danube, with a northward extension in the USSR.

The carbonate Paleozoic is lacking at Lacul Roșu, as well as in Dobrogea, suggesting the uplifted position of the areas at this stratigraphic level.

1.2. *The Lower Triassic sedimentary cycle* overlies, slightly unconformably, the Paleozoic and consists of thick red continental deposits (400—over 2500 m), fine-coarse sandstones with argillites and marly clays. Diaclases, nodules, and thin anhydrite layers have been noted throughout the sequence.

Thick acid effusive rock interbeddings (up to 200 m in thickness) were encountered in the eastern part of the Delta at Letea (feldspathic porphyries) and Lacul Roșu North (quartziferous porphyries), as well as basic ones in its middle-western part, at Stipoc and Obretin (diabases and melaphyres) (Pl. — A, B). The age of the volcanics is still a problem.

Only some continental vegetal remains of gymnosperms and spores have been identified out of which a constant frequency seems to present the species *Taenisporites noviaulensis* (L e s c h i k) and *Falcisporites zapfei* (P o t. & K l a u s) L e s c h i k, as well as some undeterminable specific types of the genera *Striatoabietites*, *Klausipollenites*, *Hemiapollenites* and *Alisporites*. All these are derived from Paleozoic types, but their association rather points to a Lower Triassic age.

The thickness variations of the red sedimentary, as well as its lack in the Snake Island (USSR), suggest an area of maximal deposit accumulation in the western half of the Delta, in front of the Sf. Gheorghe Fault.

In North Dobrogea, the red continental deposits, dated as Lower Triassic (Seissian) were only noticed in the Tulcea — Mahmudia area. Their thickness is comparatively unsignificant and obviously indicates the uplifted position of the respective area at this stratigraphic level too.

1.3. *The Middle-Upper Triassic sedimentary cycle* transgressively overlies the red sedimentary and consists of thick marine deposits, carbonate in the lower part and arenaceous at the top.

— The carbonate deposits are about 1000 m thick and consist of limestones at the base (350—450 m thick) and dolomites at the upper part (500—600 m thick) (Pl. —A, B).

The limestones are grey coloured, micro- and mesocrystalline, sometimes argillaceous. No determinable macrofossils have been registered and a poor foraminifera fauna has been solely identified at the base of the sequence (Lacul Roșu). It includes different forms of *Glomospirella, Nodosinella, Spirillina, Agathamina, Meandrospira*, suggesting an Anisian age. This age is equally substantiated by the continental palyno forms, present in all wells. Out of these, mention must be made of : *Paravesicaspora planderoväe* (Vischer), *Protodiploxypinus doubingeri* (Klaus) Warrington, *Thiradispora crassa* Klaus, *Alisporites graivogeli* Klaus.

The dolomites are micro- and macrocrystalline, whitish or yellowish grey-coloured and, occasionally, red-spotted. No macrofossils have been identified and the poor foraminifera content (forms of *Nodosaria, Dentalina, Placopsilius, Nodosinella)*, as well as the palyno one (lesser than at the limestone level) are only suggesting a Middle Triassic age.

— The arenaceous deposits have been encountered only in the well Caraorman, where they are overlying the dolomites and consist of a thick sequence (450 m) of fine-coarse grey calcareous sandstones. A very rich palynological content of continental forms has been identified, out of which most frequent and characteristic are : *Limbosporites lundbaldi* Nellson, *Riciisporites tuberculatus* Lundbald, *Ovalipollis ovalis* Krutzsch, *Chordosporites platysaccus* Mädler, etc. The flora indicates a late Upper Triassic age with transition to the Rhaetian.

The Middle-Upper Triassic deposits in the Delta are strikingly different from those in North Dobrogea, both lithologically and paleontologically, and therefore their correlation is not possible yet.

1.4. *The Jurassic sedimentary cycle* overlies transgressively the above-mentioned various Triassic complexes and consists of thick marine deposits, detrital at the base (Middle Jurassic) and carbonate at the upper part (Upper Jurassic). The sedimentation does not start, however, by the same deposits everywhere in the Delta.

In the Maliuc and Lacul Roșu wells, the dolomites and the Middle Triassic limestones, respectively, are overlain by grey-yellowish calcarenites (100—130 m thick) and microcrystalline limestones (about 70 m thick), including a continental palynoflora with Triassic and Jurassic forms. On this account and since they are overlain by the Bajocian, the respective deposits have been considered as late Liassic.

The Liassic limestones or various Triassic levels are covered by dark grey argillaceous limestones with interbedded thin, grey-blackish argillaceous shales. The thickness of the complex is of 20—25 m at Stipoc, Obretin, and Caraorman, of 150 m at Lacul Roșu and of 400 m in the Maliuc well. Here the limestones are predominantly lacustrine at the base (with numerous vegetal remains) and contain a palynoflora of Lower Bajocian age. Peculiar mention should be made of the species : *Chasmatosporites apertus* (Rogalska) Nillson, *Baculatisporites commannensis* (Cookson) Potonié, from the continental forms, and *Nannoceratopsis gracilis* Alberti and *Nannocodinium semitabulatum* Morgenroth from the dinoflagellates.

The Bajocian and the Lower Triassic limestones (the Letea and Rosetti wells) are overlain by a thick sequence of dark grey-blackish argillaceous shales with pyrite concretions. Decimetric intercalations of grey fine sandstones and, occasionally, of limestones, marly limestones and sandy limestones are to be noted mainly at the upper part, where these are making up even thicker beds.

Bositra buchi (Roemer), *Nuculana* sp., *Leda* sp. have been identified, together with a rich microfauna and microflora, testifying to a comprehensive Bajocian-Bathonian age with extension into the Callovian-Lower Oxfordian.

The microfauna comprises numerous foraminifera of the Rhizaminidae, Reophacidae, Textulariidae, Lituolidae, Spirillinidae families and especially of the Ophtalmidiidae which sometimes occur exclusively (the Opthalmidium microfacies). Towards the upper part of the sequence, *Epistomina mosquensis* (Uhlig) has been identified (an index-fossil for the Callovian), while at its more calcareous top, frequent microfilaments have been noticed in the Caraorman well (in the Moesian Platform they characterize the Upper Callovian-Lower Oxfordian interval).

The palynoflora in the shales particularly consists of dinoflagellates (but also continental forms) out of which are mentioned : *Nannoceratopsis spiculata* Stover and *Chytroeisphaeridia variabilis* Pocock (usually present in the Middle-Lower Bajocian), *Meiourogonyaulax valensii* Sarjeant and *Pareodinia ceratofora* Deflandre (specific to the Bathonian), *Ctenidodinium tenullum* Deflandre, *Scriniodinium dementii* Pocock, *Nannoceratopsis pellucida* Deflandre, *Meiourogonyaulax rioulti* Sarjeant (defining the Upper Callovian-Lower Oxfordian interval).

The total thickness of the Middle Jurassic deposits widely varies between 500 m in the north (Stipoc) and over 1700 m in the south-west (Maliuc), the maximal thickness areas being located in front of the Sf. Gheorghe Fault (Plate —C).

— The upper part of the Jurassic consists of mesocrystalline grey or grey-yellowish limestones, alternating with marly limestones and marls. Remains of lamellibranchs, gastropoda and brachiopoda were identified, together with a relatively rich microfauna which proves a Malm age.

Out of foraminifera, peculiar attention should be paid to : *Textularia jurassica* (Gümbel), *Bigenerina minima* (Iovceva & Trifonova), *Spirillina orbicula* (Terquem & Berthelin), *S. elongata* (Bielecka & Pozariska), *Paalzowella feifeli* (Paalzow), *Turispirillina polygyrata* (Gümbel), *T. amoena* (Dain), etc., to which the holothurid species, *Theelia heptalampra* (Bartenstein) and *Hemisphaeranis seiboldi* (Schwager), should also be added. In the Lacul Roșu well, there have been identified microoncolites and *Mercierella dacica* (Dragastan), characteristic, in the Moesian Platform and the Carpathians, of the Upper Kimmeridgian-Lower Tithonian interval.

The palynoflora especially consists of dinoflagellates : *Scriniodinium dictyotum* (Cookson & Eisenack), *Ctenidodinium panneum* (Norris)Lentin & Williams, *Cyptarcheodinium calcaratum* (Deflandre) Gitmez.

The mentioned paleontological content has been found only in the lower part of the sequence, predominantly made of limestones ; its upper part is invaded by marls and marly limestones, and, according to certain palyno data, the latter seem to stand for an extension of the Tithonian into the Portlandian facies.

The Upper Jurassic sedimentary develops over an area somewhat smaller than that of Middle Jurassic, but the areas of maximal deposits thickness (of about 1000 m or even thicker) also lie in front of the Sf. Gheorghe Fault (Plate — C).

In North Dobrogea, the Jurassic is only represented by thin Liassic deposits and it is obvious that in this time it roughly acted as an uplifted area. Some data, however, suggest a southern invasion of the Jurassic over the Sf. Gheorghe Fault, along the border of the Black Sea.

1.5. *The Lower Cretaceous sedimentary cycle* overlies the Upper or Middle Jurassic and consists of red continental deposits, clays, marls, sands and sandstones with thin anhydrite interbeddings, especially in the Stipoc area.

The thickness of the deposits is highly variable (Plate — D), being greater at Stipoc-Rosetti and Lacul Roşu (over 500 m). In this last area a palyno association of spores-pollen and subordinately of dinoflagellates has been identified, pointing to a Lower Cretaceous age (previously, these deposits have been considered as Upper Jurassic — Liteanu E. & Pricăjan A., 1963, etc.). Out of these forms mention must be made of: *Circulina parva* (Brenner), *Classopolis classoides* (Pfulg), *Broomea jaegeri* (Alberti), *Dingodinium cerviculum* (Cookson & Eisenack).

No such red deposits are present in North Dobrogea, but from the Lacul Roşu area they seem to expand southwards over the Sf. Gheorghe Fault, covering the eastward plunging of the Mahmudia Uplift.

1.6. *The Sarmatian-Pliocene sedimentary cycle* transgressively overlies the various Mesozoic complexes and roughly consists of alternating marls and sands. Detailed data on this sequence, of about 200 — 350 m thick, have been furnished by Liteanu and Pricăjan (1963).

2. Evolutionary and Structural Considerations

The Pre-Dobrogean Depresssion, of which the Danube Delta is a part, has been regarded as a Jurassic depression, which came into being after the North Dobrogean old Kimmerian foldings. It, however, settled down on an area whose depressionary character was inherited at least from the first stage of the Hercynian foldings (old Bretonic), a long time, between the Devonian and Upper Triassic, the area acted irrespective of North Dobrogea and in this respect, the doubts expressed by certain researchers as to its foredeep character are partly justified.

In the south, the depression is bordered by the Sf. Gheorghe Fault, a deep fracture which acted as far as the Paleozoic, if not before. Along the fault, the North Dobrogean sedimentary overthrusts the Delta one, but the thrust amplitude has not been defined yet.

According to certain researchers, the Sf. Gheorghe Fault continues westwards along the Galaţi—Tecuci line (Mutihac and Ionesi

1973), but it is much more probable that it crosses the Danube, merging into the Ismail-Cahul line (USSR), equally regarded as the south-eastern margin of the Pre-Dobrogean Depression (D r u m e a et al., 1959). Eastwards, the fault continues into the marine realm.

The northern depression margin lies in the USSR, on the Kangaz-Glubkoe line (D r u m e a et al., 1959) which also seems to correspond to a deep fracture. This line extends into the Black Sea, west of the Snake Island.

During the Paleozoic and Lower Triassic, the area of maximal sedimentation seems to have been located north of the Sulina branch of the Danube, with extension into the USSR, but, in the Middle Triassic, it moved into the southern part of the Delta, between the Sulina and the Sf. Gheorghe branches of the Danube (Pl.—A, B).

During the Jurassic, the Depression extended progressively northwards, overpassing the Middle Triassic border, but its depocenter still remains in front of the Sf. Gheorghe Fault (Pl.—A, C).

The thickness variations of the Middle and Upper Jurassic sedimentary (Pl. — C, D) show not only a striking asymmetry of the depression at these stratigraphic levels (a very large northern flank and an extremely narrow southern one), but also a deep geological condition in the Stipoc-Caraorman area, which has strangled the depression, separating two depocenters; one in the west, at Maliuc, where the Middle Jurassic is widely developed with continuity in the USSR (at Bolgrad), and another one in the east, at Sulina, with continuity into the Black Sea realm.

The deep geological conditions having generated the ridge separating these depocenters have not been defined yet, but it lies in the southern sunken extremity of a promontory in the Letea-Stipoc area and could correspond to a deep fault system along which basic effusive events have taken place (Obretin and Caraorman, Pl.—A, B).

At the end of the Jurassic, the Danube Delta was uplifted, without being folded and attached to the North-Dobrogean dryland. The old movement tendencies of sinking were only resumed for a short time in the Lower Cretaceous and then in the Upper Miocene and Pliocene.

The Danube Delta succession of deposits is very thick and, generally, lithologically different from the North Dobrogean one; many sequences in the Delta are lacking in Dobrogea and the other way round, all these remarkable differences took place over a very short distance and are rather difficult to explain. An ample thrusting of the North Dobrogean sedimentary over the Delta one along the Sf. Gheorghe Fault could offer a more convenient explanation, but there are no positive data to prove such a thrusting.

Inferences

The Danube Delta is a part of the Pre-Dobrogean Depression and comprises a thick sequence of deposits, carbonate and detrital, from Paleozoic to Quaternary, disposed in major sedimentary cycles, separated by stratigraphical gaps.

The red continental Lower Triassic deposits, as well as the detrital marine Jurassic ones, are amply developed throughout the Delta, but the carbonate deposits of Middle Triassic and Upper Jurassic are only developed in front of the Sf. Gheorghe Fault, which separates the Delta from the North-Dobrogean Orogene.

In the red continental Lower Triassic, there are to be noted interbeddings of volcanics — acid (porphyries) and basic (diabases, melaphyres) — but their age is still a problem.

The Delta sedimentary is very thick and to a great extent lithologically different from the North Dobrogean one.

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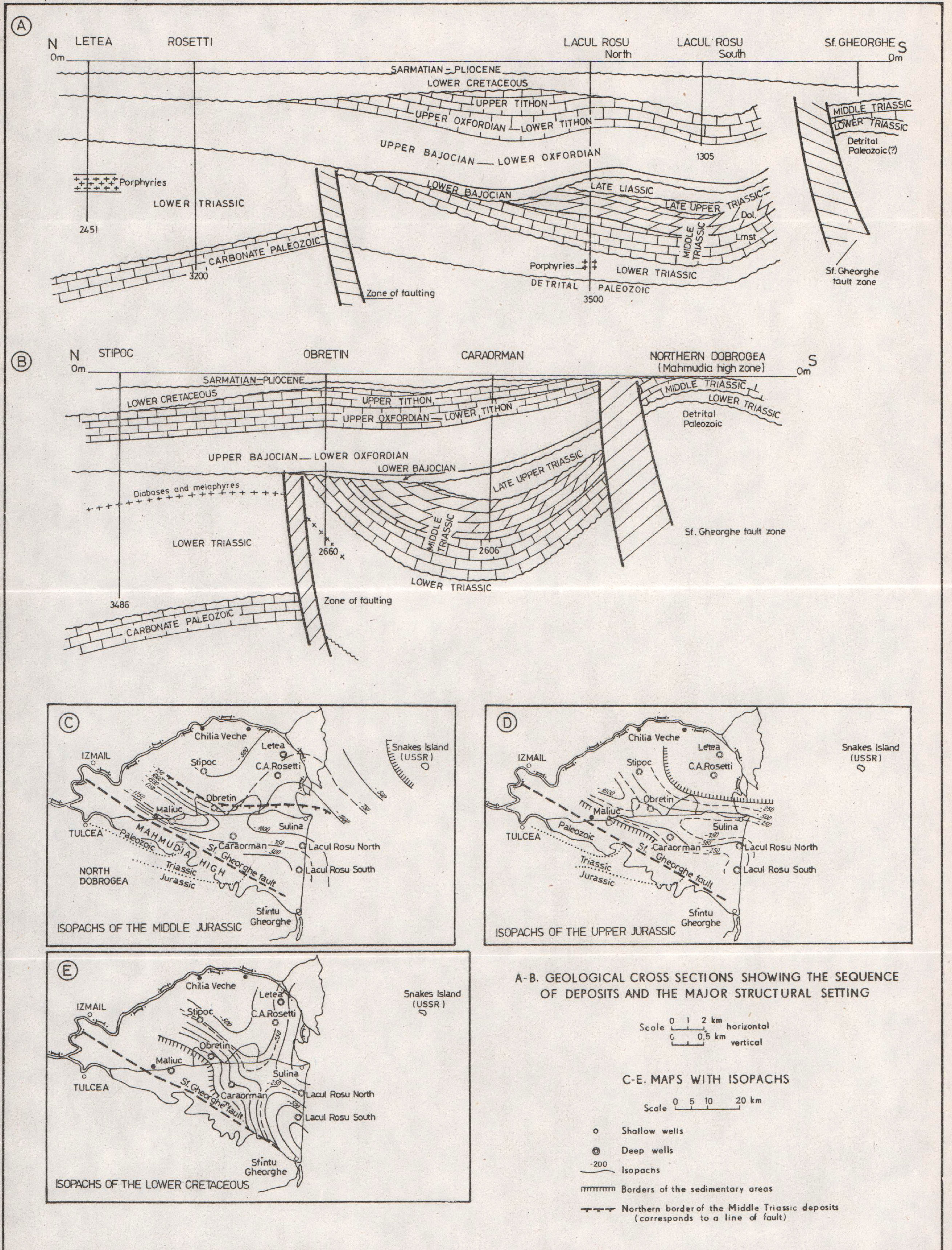
DISCUSSIONS

R. Cădere : This very interesting paper on the stratigraphy and tectonics of the Danube Delta is of a great importance for the hydrological interpretation of the lack of drinking water in this zone.

Likewise, the stratigraphy of the Delta is completed by a Quaternary rich in alluvia with significant contents in titaniferous ores resulting from the deposition of heavy alluvia, drifted by the Danube and sedimented here and in the zones of overflow into the sea. It will constitute an important ore deposit for the titanium mining in the future.



A-B. GEOLOGICAL CROSS SECTIONS SHOWING THE SEQUENCE OF DEPOSITS AND THE MAJOR STRUCTURAL SETTING

C-E. MAPS WITH ISOPACHS

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METAMORPHOSED PALEOZOIC AND MESOZOIC FORMATIONS OF THE MEHEDINȚI-RETEZAT UNIT [1]

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Within the Danubian area ("Danubian Autochthon") several Alpine (Laramian) tectonic units are distinguished: Arjana Nappe (Codarcea, 1940), Presacina Nappe (Stănoiu, 1973 a), etc. The shear plane of the Presacina Nappe divides the Danubian area into two important subdivisions: the internal part (internal Danubian area) and the external part (external Danubian area). The external part of the Danubian area has been called by Stănoiu (1973 a) the Mehedinți-Retezat Unit, by Pop (1973) the Mehedinți Threshold, by Kräutner et al. (1978) the Parîng Retezat Unit, and by Kräutner et al. (1981) the Lower Danubian Unit. The Arjana Nappe in the Țarcu Mts is also known under the name of the Căleanu Duplicature (Morariu, Morariu, 1981) or the Feneș Unit (Kräutner et al., 1981). The Presacina Nappe has also been called the Poiana Mărului Unit (Kräutner et al., 1978; Kräutner et al., 1981).

Manolescu (1932, 1937 a, 1937 b, 1940) and Paliuc (1937), who founded the lithostratigraphy of the metamorphosed Paleozoic and Mesozoic formations, referred them mainly to the Mesozoic. After the year 1953, all the rocks of the above-mentioned formations have been assigned to the Paleozoic, being included into the Tulișa Series (Pavelescu, 1953). Mention should be made of Pavelescu and Răileanu's paper (1963) in which the unconformity between the limestones and basal conglomerates of the Tulișa Series is pointed out. Recently, important progresses have been achieved due to the papers of Stănoiu (1971, 1972, 1973 b, 1976, 1980 a, 1980 b), Gherasi et al., (1968, 1973, 1975), Năstăseanu (1973, 1974, 1975), Pop (in Pavelescu et al., 1974), Solomon, Visarion (1974), Solomon et al. (1976), Schuster (1980, in Năstăseanu et al., 1974).

As the Paleozoic formations of the external part of the Danubian area have already been presented by one of the authors of this paper in

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the Guidebook to Excursion A_1 of the 12th Congress of the CBGA, we shall not deal with them any more.

The metamorphosed Paleozoic and Mesozoic formations of the Mehedinţi-Retezat Unit may be referred to the Jieţ Supergroup. The Paleozoic formations constitute the Tulişa Group and the Mesozoic formations make up the Schela Group.

Cambrian

A lithostratigraphic entity with limestones (carbonate complex) has been delimited by Bercia, Bercia (1963) in the Mehedinţi Mts, at the upper part of the Lainici-Păiuş Group. This lithostratigraphic entity has also been found in the Vîlcan, Retezat and Parîng mountains, almost always immediately under the Paleozoic formations. Lithologic and palynologic evidence (*Acanthodiacrodium* sp., *Cymatiosphaera* sp., *Baltisphaeridium* sp.) seems to indicate that at least some of the rocks of this entity belong to the Cambrian.

The Tulişa Group

Ordovician-Silurian(?)

The Valea Izvorului Formation (Stănoiu, 1972), unconformable, appears in the Mehedinţi Mts. It is represented by a quartzitic lower member (cca 10 – 60 m thick), consisting of white quartzites, often with interbeddings of chlorite-sericite schists or sericito-chlorite schists, and a schistous upper member (300 m thick), constituted of chlorite-sericite or sericito-chlorite ± graphite greenschists. The rocks of the Valea Izvorului Formations in the Vîlcan, Retezat and Parîng mountains have been temporarily separated by Stănoiu (1976) under the name of the Coarnele Formation.

From the Valea Izvorului Formation Stănoiu (1971, 1972, 1973 b) and Iliescu (in Stănoiu, Iliescu, 1976) reported corals, bryozoa, brachiopods and trilobites (*Caleidocrinus artifex*, *Flexicolymene* sp., *Encrinurus* or *Cromus* sp.) beside acritarchs, chitinozoa, scolecodontes and microspores.

Acritarchs (*Baltisphaeridium* sp., *B. brevispinosum* Eis., *Priscogalea simplex* Dunff., *Acanthodiacrodium* cf. *achrasi* Martin, *Veryhachium lairdi* (Defl.) Deff., *Multiplicisphaeridium* sp. cf., *M. varians*, St. Will)., chitinozoa (*Lagenochtina*, *Conochitina* sp.), and scolecodontes have been reported from the Coarnele Formation.

Devonian-Lower Carboniferous

The Tusu Formation (Stănoiu, 1976) – unconformable, up to 800 m thick – is apt to be found in the Vîlcan, Retezat and Mehedinţi mountains. It is represented by graphitic schists, chlorito-epidotic schists, chloritoid schists, lenses of quartzites, metapsamites, green metapsamites, intraformational metabreccias, mostly acid-intermediary metavolcanic conodont bodies.

The paragenesis of the Tusu Formation is represented by quartz, calcite, albite, microcline, sericite, chlorite, chloritoid, epidote, actinote, biotite, muscovite, stilpnomelane, graphite.

From the Tusu Formation (Upper Devonian or more recently) Adina Visarion (in Solomon et al., 1976) and Violeta Iliescu (in Stănoiu, Iliescu, 1976) reported a palynologic assemblage and Stănoiu (1980 b) pointed out the presence of macroflora remains (*Palaeophyllales* group, *Neuropteris* sp., etc.). The palynological assemblage is represented by *Auroraspora* sp., *Samarisporites* sp., *Hymenozonotriles* sp., *Archaeozonotriletes vasjamicus* Cibrik, *Endoculeospora* sp., *Brochotriletes triangularis* Cibrik. The precise age will be obtained only after having finished the paleontological studies.

The Sgura Formation (Stănoiu, in Kräutner et al., 1981) — unconformable, cca 600 — 700 m thick — is represented by metapsamites, metasiltites and metapelites with a lot of plant remains. It is admitted (Stănoiu, 1976) that there is a facial lateral transition between the rocks of the Tusu Formation and those of the Sgura Formation.

The mineralogical paragenesis of the Tusu and Sgura formations is represented by quartz, calcite, albite, microcline, sericite, chloritoid, epidote, actinote, biotite, muscovite, stilpnomelane, graphite.

The Vidra Formation (Codarcea, Gherasi, 1959) — cca 700 m thick — appears in the Petreanu, Țarcu, and Retezat mountains. Here, the overthrust plane of the Presacina Nappe overlies a homoclinal succession of metapsephites, black (coarse) metapsamites and metapelites and crystalline limestones with a mineralogical paragenesis consisting of quartz, albite, sericite, chloritoid, epidote, albite, graphite, biotite (Gherasi et al., 1968).

From the Vidra Formation (Devonian) there have been reported remains of crinoids (Gherasi et al., 1968), *Michelinoceras* sp. (Magdalena Iordan — in Gherasi et al., 1975) and a palynologic assemblage (Adina Visarion — in Gherasi et al., 1975) similar to that found in the Tusu Formation. The lithologic aspect points to the Mesozoic age.

The Latorița Formation (Schuster, in Năstăseanu et al., 1974) is apt to be found only in the Parîng Mts. West of the Coasta lui Rus it consists of a calcareous lower member (cca 100 m thick) and a schistous upper member (300 — 500 m thick), represented by chlorito-epidote greenschists in association with calcareous shales and serpentinites. East of Coasta lui Rus, the Latorița Formation becomes more heterogeneous (including metapsephites, metapsamites, metaarkoses, metagrauwackes, albite-sericite-quartz-epidote schists, sericito-graphite schists, sericite schists, crystalline limestones, metabasites, serpentinites, metaquartz-keratophyres, and rodingites); it shows numerous lateral facies variations, becomes thicker (up to 1000 m), and the basal limestones branch out at several levels.

The age of the Latorița Formation is confined to the Ordovician-Lower Carboniferous interval.

The Oslea Formation s.s. — cca 400 m thick, unconformable — — occurs in the Vîlcan and Retezat mountains. The upper part of the

Oslea Formation *s.l.* has been attributed to the Mesozoic (Năstăseanu, 1973). The Oslea Formation *s.s.* has been divided by Stănoiu (1980 a) into a metapsamitic lower member, a carbonatic middle member, and a metapelitic (schistous-graphitous) upper member. The paragenesis of the Oslea Formation is represented by quartz, calcite, dolomite, micrite, chlorite, biotite, stilpnomelane, epidote, actinote.

A palynologic assemblage, represented by *Convolutispora* sp., *Apiculatisporites* sp., *Cristatisporites* sp., *Waltzispora* sp., *Triquitrites* sp., *Murospora kosankei* Som., *Knoxisporites literatus* (Waltz) Playford, *Lophozonotriletes cristifer* (Luber) Kedo, which would indicate the Carboniferous (Viséan-Namurian?), has been pointed out at the lower part of the metapelitic member.

The Hercynian Molasse

Middle Carboniferous-Permian

Semaka (1963) reported "remains of *Annularia stellata* Schloth., *Calamites (Calamitina) undulatus* Sternb., *Precopteris feminaeformis* Schloth. and *Stigmaria?* sp. from the schists of the Schela Formation of Vai de Ei and remains of *Sphaenophyllum longifolium* Germar, pointing to the Upper Carboniferous, in the Rafailă chloritoid schists.

In the Retezat Mts, slightly metamorphosed, red conglomerates, assigned to the Permian, are overlain by the Mesozoic rocks and overlie the crystalline schists. The Permian might also include the Piatra Cloșani metapsephites (Stănoiu, 1973), found in the Cerna Mts and on the southern slope of the Vîlcan Mts.

The rocks of the Tulișa Group, belonging to the Mehedinți-Retezat Unit, have undergone a Hercynian, regional metamorphism in the greenschist facies, quartz-albite-chlorite subfacies, between the contact area with the prehnite-pumpellyte facies (in the outer parts) and the contact area with the quartz-albite-biotite subfacies (in the inner parts). The intensity of the metamorphism processes increased from the exterior to the interior (Stănoiu, 1973; 1980 a); an obvious difference may be observed as regards the intensity of the deformations, blastesis, and the orientation of the minerals in the Paleozoic formations of the Mehedinți Plateau as compared to the synchronous formations in the Retezat Mts and the northern part of the Vîlcan and Parîng mountains. Pyrophyllite occurs in the Tusu Formation (Poiana Mică Formation) in the Mehedinți Mts.

The Schela Group

The Mesozoic rocks of the Mehedinți-Retezat Unit may be grouped into four main formations: Baia de Aramă Formation (Semaka, 1963) — Liassic; Poiana Mare Formation (calcareous formation) — Middle Jurassic-Lower Cretaceous; Nadanova Formation — Cenomanian-Lower Turonian; Mehedinți Formation (of olistostrome type) — Middle Turonian? — Senonian.

The Schela "Formation" (Mrazek, 1898) in the southern slope of the Vîlcan Mts is mostly constituted of rocks (metapsamites, metapelites and metapsephites) with Liassic plants within which chloritoid, pyrophyllite (80%), chlorite, sericite, prehnite, graphite (as powder and anthracite strips) and illite (Paliuc, 1972; Anton, 1974) are to be found. These rocks point out (Gurău, Şerbănescu, 1972) blastesis, deformations (boudinages, microfolds, S_1 foliation) and mineral trending. All this points to low-grade metamorphism (at the level of the entrance domain in the greenschists). The lithology, rendered evident particularly by drillings, indicates that the upper part of the Schela "Formation" might be represented by the Mehedinţi Formation.

On the northern side of the Vîlcan and Parîng mountains, Stănoiu (1980 a) has separated provisionally the Răstovanu Formation — unconformable, cca 800 m thick — represented by metapsamites, metapelites and metapsamites with *Pterophyllum* sp., *Cladophlebis aldanensis* Vochraneev, assigned to the Liassic, possibly to the Upper Carboniferous, too. In this formation, beside Liassic spores, Upper Carboniferous spores are found in places. The Răstovanu Formation is unconformably overlain by the Nadanova Formation (sericitous and calcareous grey metapelites) with lamellibranchs, gastropods, belemnites and ammonites (Stănoiu, 1980 a) east of the Jiu Passage (Pop in Pavelescu et al., 1975), on the Baleia Valley and south of the Cîmpu lui Neag locality. This formation is unconformably overlapped by the rocks of the Mehedinţi Formation (Codarcea, 1941; Solomon, 1978; Pop, in Pavelescu et al., 1975; Năstăseanu and Stănoiu in Stan et al., 1979; Năstăseanu, 1980; Stănoiu, 1980 a), represented by the lower member (cca 300 m thick) with limestone olistoliths and the upper member (cca 200 m thick) with dolerite-basalt olistoliths, serpentinites and rocks of the Sinaia and Azuga beds type. Solomon et al. (1976) have separated a part of the rocks of the Mehedinţi Formation in the north-eastern part of the Vîlcan Mts under the name of the Petroşani Formation. From the "lenses" (=olistoliths) of limestones associated to the Paroşeni Formation, Magdalena Iordan reported coral remains belonging to the group *Thamnasterida*, *Phacelostilophylum* sp., *Cyatocoenia alpina* (Gimbel) (Ronoeevicz), crinoids, bivalves, brachiopods. In the same formation, Adina Visarion has identified pollen of Pinaceae and *Ginkgocycadophytus* sp., beside spores (*Cyathidites* sp., *Gleicheniidites* or *Toroisporites* sp. and *Dictyophyllidites* sp.). On the basis of the mentioned fossils the Paroşeni Formation has been assigned to the Rheto-Liassic. *Cyathidites* sp., *Classopollis* sp., *Piceapollenites* sp., *Deltoidospora* sp., and *Toroisporites* sp. have also been reported from the Răstovanu and Mehedinţi formations.

The intense deformations (a marked S_1 foliation, an obvious lineation mostly due to the crenulation microfolds and the intersection of S_0 planes — bedding — with S_1 planes, intense boudinages, b_1 folds beside crenulation microfolds and b_2 incipient folds — almost perpendicular to b_1 — which affects also S_1 planes), obvious blastesis, well-marked trending of the minerals with tabular habitus and the mineralogical paragenesis (quartz, calcite, albite, sericite, chlorite, chloritoid, pyrophyllite, prehnite, pumpellyte, paragonite and graphite—anthracite with over 90%

graphite) prove that the rocks of the Răstovanu and Mehedinți [3] formations in the northern flank of the Vîlcan and Parîng mountains were metamorphosed in the uppermost part of the greenschist facies.

On the northern slope of the Retezat Mts, a Mesozoic lithostratigraphic entity (Mehedinți Formation and possibly the Baia de Aramă Formation) appears immediately under the overthrust plane of the Presacina Nappe. This entity is represented by quartz metapsamites, metasiltites and black (graphitous) metapelites with lenses (olistoliths) of limestones and serpentinites, partly included into the Rîușoru Formation.

North of the Godeanu Mts, in the Lăpușnic Valley basin, there have been described Mesozoic (Schafarzik, 1898; Streckeisen, 1934; Gherasi et al., 1973; Pop, 1964) or Paleozoic rocks (Morariu, 1972), represented by metaarkoses (Liassic) at the lower part, followed by calcareous metapsamites (Dogger), crystalline limestones (Upper Jurassic-Lower Cretaceous), schists with "lenses" (olistoliths) of limestones (lower member of the Mehedinți Formation) and schists with huge blocks — gravity sliding "lambeaux" consisting of crystalline schists associated with Permian (?) red conglomerates (upper member of the Mehedinți Formation). On the Bran and Lăpușnic valleys, from this succession Adina Visarion reported (in Gherasi et al., 1973) a Mesozoic (Jurassic) spore-pollen assemblage represented by *Cyathidites* sp., *Deltoidospora juncta* (Kr. — Mz) Singh, *Ovalipollis* sp., *Ginkgocycadophitus nitidus* (Balme) Dettman, *Bennettitinaepolenites* sp. The mineralogic paragenesis (sericite, chlorite), the intense deformations (highly microfolded rocks, marked foliation and lineation, boudinages), the trending of minerals with tabular habitus, intense recrystallizations (calcite inclusions in quartz) point to changes at the level of the greenschist facies.

N—E, E and S—E of the crystalline of the Sebeș-Lotru Group in the Godeanu Mts, the recrystallized calcareous formation (Gherasi, 1937; Pop, 1964) and the Nadanova Formation are unconformably overlain by the lower member of the Mehedinți Formation, consisting of a metamorphosed clayey-gritty olistostrome. The upper member of the Mehedinți Formation is represented, at the lower part, by metaconglomerates, metabreccias and sericite $\pm$ chlorite schists with elements and blocks of crystalline schists, rocks of the Nadanova Formation and Permian (?) red conglomerates, followed by sericite $\pm$ chlorite schists and breccias with elements and blocks of crystalline schists. In the middle part of this succession and especially at the terminal part one may observe blocks, often huge, of crystalline schists associated with Permian (?) red conglomerates. The intense deformations (microfolds, lineations, boudinages, foliation), the blastesis, the trending of the minerals with tabular habitus, the mineralogical paragenesis (chlorite, sericite) point out a metamorphism at the greenschist level.

Urcan (1977) reported pyrophyllite, sericite, chlorite and ? chloritoid (small crystals) from the rocks of the Baia de Aramă Formation in the Mehedinți Mts, between the Vîrful lui Stan Summit and the Motru Sec Valley. Deformations (boudinages, microfolds, lineations, S_1 foliation), blastesis, mineral trending, mineralogical paragenesis and the alteration stage of the coal matter (the stage of transition from anthracite to graphite) indicate metamorphism at the level of the transition domain from the prehnite-pumpellyte facies to the greenschist facies, less intensive than the metamorphism of the Mesozoic rocks described above.

Almost all limestones of the calcareous formation in the Mehedinți-Retezat Unit are intensely recrystallized.

The textural and structural characters (foliation, satinated aspect, incipient blastesis — the frequency of the so-called pressure shadows, beginning of trending of the minerals with a tabular habitus and the mineralogical paragenesis (chlorite + micaceous minerals: illite + sericite) indicate that the rocks of the Mehedinți Formation in the Mehedinți Plateau and the southern side of the Vîlcan and Parîng mountains have also undergone intense alteration processes of low-grade metamorphism (probably the prehnite-pumpellyte facies).

It is only in the rocks occurring in the outermost parts (SE of the Baia de Aramă tectonic alignment) that the sedimentary characters have not been altered.

One may infer (Stănoiu, 1973 a, 1980 a) that all the Mesozoic rocks — in almost the whole Mehedinți-Retezat Unit — have undergone a low-grade Alpine (Laramian) regional metamorphism and the intensity of these processes has increased from the inner towards the outer parts. The Alpine metamorphism has been felt in the formations of the Variscan and pre-Variscan crystalline formations (more intensely metamorphosed) by the effects of retrograde metamorphism. Many of the plicative (microfolds, large regional folds, etc.) and ruptural deformations in the pre-Mesozoic crystalline schists are the result of the Alpine orogenesis.

[3] The rocks of this formation have also been included into the Tulișa Series (Pavelescu, 1953), the Schela Formation (Manolescu, 1937; Paliuc, 1937) and the Vidruța Formation (Schuster, 1980).

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CORRÉLATION PALYNOLOGIQUE DU CRÉTACÉ TERMINAL DU SUD-EST DES MONTS METALIFERI ET DES DÉPRESSIONS DE HAŢEG ET DE RUSCA MONTANĂ [1]

PAR

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A) Introduction

Des formations continentales comprenant des dépôts détritiques, parfois rouge-violacés, avec de restes de dinosauriens, à intercalations de roches volcanoclastiques et des charbons, existent au sud-ouest de la Roumanie. Ces formations ont été attribuées au Maestrichtien, au Danien, à l'Eocène supérieur-Oligocène ou à l'Aquitanien. Nous avons entrepris des recherches afin de préciser l'âge et de corréler ces dépôts au moyen de la microflore et des associations des gastropodes.

B) Cadre géologique

a) Sud-est des Monts Metaliferi

Au sud-est des Monts Metaliferi, dans la rive droite de la rivière Mureş, dans la région de la vallée Pîclişa, aux environs de la ville d'Alba-Iulia, une formation détritique, rouge-violacée repose en continuité de sédimentation sur la formation molassique marine d'âge campanien supérieur-maestrichtien inférieur. Celle-ci a été attribuée soit au Maestrichtien (Dimian, Popa-Dimian, 1964; Tomescu et al., 1969), soit à l'Oligocène (Bleahu, Dimian, 1967). Antonescu (1973) identifie une association palynologique à *Pseudopapilopollis praesubhercynicus* (Góczán, 1967, d'âge maestrichtien moyen et supérieur, commune aux dépôts détritiques rouges de Pîclişa et aux dépôts à charbons de Rusca Montană, corrélable avec celle du Maestrichtien supérieur d'Hongrie.

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b) **Dépression de Haţeg**

Dans la dépression de Haţeg une formation détritique, continentale, puissante d'environ 1500 mètres (Dincă, Tocorjescu, Stilla, 1972) repose sur un faciès de flysch dont la partie supérieure est d'âge campanien supérieur-maestrichtien inférieur (Pop, Neagu, Szász, 1972). Dans la partie centrale-ouest du bassin, la série continentale commence par un faciès pyroclastique suivi par un faciès fluvio-lacustre qui contient des restes de dinosauriens à la partie inférieure (Nopcsa, 1902; selon Dincă et al., 1972). L'âge des dépôts compris entre le faciès de flysch et les premiers dépôts à faune marine du Miocène de la partie centrale-ouest de la dépression a été attribuée soit au Danien (Nopcsa, 1905; Laufer, 1925; Mamulea, 1953; Drăghindă, Mihalache Paula, 1963; selon Pop, Neagu, Szász, 1972) soit au Maestrichtien (Pop, 1971; selon Pop et al., 1972; Pop et al., 1972; Dincă et al., 1972). Rădulescu, Iliescu, Iliescu (1976) considèrent la partie inférieure du faciès fluvio-lacustre à dinosauriens d'âge éocène supérieur-oligocène inférieur (les dinosauriens étant remaniés), et la partie supérieure de la même formation à dépôts détritiques rouges des environs de Ciula d'âge oligocène étant comparée avec les dépôts sous faciès similaire oligocène de l'est de la dépression. Nous considérons que dans la partie centrale-ouest de la depression de Haţeg il y a deux formations; 1) la formation de Densuş à roches volcanoclastiques, inférieure, d'âge maestrichtien selon la flore (Baikovskaia, 1956; selon Mărgărit, Mărgărit, 1967), qui repose sur le flysch sénonien et, 2) la formation de Ciula, supérieure, comprenant, les dépôts détritiques rouge violacé à dinosauriens et les dépôts sous le même faciès, qui au sud de Ciula Mică comportent une intercalation charbonneuse. Au-dessus de la formation de Ciula suit avec discordance le Miocène marin, l'âge de la formation de Ciula sera discuté plus-bas.

c) **Dépression de Rusca Montană**

Dans cette région, au-dessus du faciès de flysch d'âge santonien-campanien (Dincă, 1977), suit le faciès molassique à roches volcanoclastiques et intercalations charbonneuses. Dincă (1977) décèle dans cette molasse un horizon détritique, inférieur, et le second, supérieur, piroclastique, qui dans l'est de la dépression comporte des charbons à la partie supérieure. Dincă et al. (1972), et Dincă (1977) attribuent au Maestrichtien les horizons détritiques et piroclastiques de Rusca Montană.

C) **Corrélation au moyen de la microflore et des gastropodes du Maestrichtien continental du sud-ouest de la Roumanie**

a) **Gastropodes**

Les dépôts détritiques continentaux à dinosauriens de la formation de Ciula ont fourni aussi des associations de gastropodes dulcicoles ou terrestres. Les spécimens appartenant aux familles Pomatiasidae, Anadromidae, Viviparidae, Planorbidae et Streptaxidae ont été prélevés des localités Sînpetru et Vălioara. L'association des espèces *Anastomopsis rotellaris* (Math.), *Bauxia bulimoides* (Math.), *Bauxia* sp., *Strophomella reussi* (Stol.), *Helix* sp., *Vidadiella darderi* (Vidal), *Gastrobulimus munieri* (Hantken), *Bauxia* aff. *bulimoides* (Math.), *Gastrobulimus* sp., *Lychnis* sp., ? *Paludinopsis* sp., *Planorbis* sp., indiquent le Maestrichtien supérieur ; des associations similaires ont été décrites dans le Garoumien français aussi bien que dans le Crétacé terminal d'Espagne, Iles Baléares, Portugal (Wenz, 1939, 1959, 1960).

b) **Microflore**

1. *Provenance de la microflore.* Au sud-est des Monts Metaliferi, dans la vallée de Pîclişa nous avons identifié un second niveau palynologique dans la formation détritique rouge (la premier ayant décrit par Antonescu, 1973) dans la partie moyenne de la formation. La microflore est semblable à celle décrite par Antonescu (1973) ; de plus il y a des rares exemplaires de pollen attribuables à *Proteacidites* cf. *subscabratus* Couper, 1960 et au genre *Convexipollis* Krutzsch, 1967.

Dans la dépression de Haţeg, (formation de Ciula) deux niveaux contenant de la microflore ont été découverts, dans l'affleurement de la vallée Sibişel à Sînpetru (rive gauche) et dans la rive droite de la vallée de Ciula, au sud du village, dans un affluent qui met à jour une intercalation d'argiles charbonneuses dans les dépôts détritiques rouges.

Dans la partie est de la dépression de Rusca Montană, les intercalations d'argiles charbonneuses de l'horizon piroclastique du Maestrichtien ont fourni une riche microflore dans la coupe de la vallée Slătioara (sept niveaux à microflore) et dans les affluents de la vallée Loznicioara.

2. *Composition de la microflore.* Les espèces déterminées à Sînpetru sont : *Plicapollis* cf. *conserta* Pflug, 1953, *Oculopollis* cf. *baculotrudens* (Pflug) Zaklinskaia, 1963, cf. *Oculopollis sibiricus* Zaklinskaia, 1963, *Papilopollis* ? spp., *Pseudopapilopollis praesubhercynicus*, *Convexipollis* cf. *convexigerminalis* Krutzsch, 1967.

Pour les autres régions — Pîclişa, Ciula, Loznicioara, Slătioara — nous allons présenter ensemble les listes, les espèces étant communes, avec spécification dans le cas d'une espèce trouvée seulement dans une région. Ce sont : *Leiotriletes* cf. *paramaximus* Krutzsch, 1959 (Pîclişa) ; *Leiotriletes* spp. ; *Dictyophyllidites* spp. ; *Trilites* sp. ; Spore type B ; *Echinatisporis longechinus* Krutzsch, 1959 ; *Corrugatisporites toratus* Weyland, Greifeld, 1953 ; *Cicatricosisoprites* spp. (fort rares) ; *Retitriletes* sp. (Slătioara) ; *Faveolatisporites* sp. (Pîclişa) ; *Polypodoaceoisporites* spp. ; cf. *Lusatisporis* sp. (Slătioara) ; *Laevigatosporites* spp. ; *Polypodiisporites* spp. ; *Zlivisporites blanensis* Pacltova, 1961 ; *Ephedripites* spp. (Pîclişa) ; *Triporoletes* cf. *tornatilis* Srivastava, 1972 ; *Triporoletes* spp. ; *Inaperturopollenites* sp. ; *Pistilipollenites* sp. ; *Vacuopolis* sp. ; *Interporopollenites proporus* Weyland, Greifeld, 1953 ; *Interporopollenites* cf. *gracilis* Krutzsch, 1960 ; *Interporopollenites* sp. (Slătioara) ; *Plicapollis* cf. *pseudoexcelsus* (Krutzsch) Krutzsch, 1961 ; *Plicapollis* cf. *conserta* Pflug, 1953 ; cf. *Minorpollis* sp. ; *Pseudoplicapollis peneserta* (Pflug) Krutzsch, 1967 ;

Suemeghipollis triangularis G ó c z á n, 1964; *Oculopollis* cf. *orbicularis* G ó c z á n, 1964; *Oculopollis* cf. *baculotrudens*; *Pseudoculopollis* sp.; *Semioculopollis praedicatus* (W e y l a n d, K r i e g e r) K r u t z s c h, 1967; *Papilopollis*? spp., *Pseudopapilopollis praesubhercynicus*, *Pompeckjoidaepollenites subhercynicus* (K r u t z s c h) K r u t z s c h, 1967 (Slătioara); *Trudopollis*? spp., *Trudopollis* ex gr. *imperfectus* (P f l u g) P f l u g, 1953; *Hungaropollis* sp. (Pîclișa, remanié?); *Pseudotrudopollis pseudalnoides* (K r u t z s c h) K r u t z s c h, 1967 (Slătioara); Normapolles non-identifiés, *Convexipollis* cf. *rotundatus* P o r t n i a g h i n a, 1971 (Rusca Montană), *Convexipollis* sp. (Pîclișa), cf. "*Comptonia*" sp.; "*Tricolporites*"? sp., *Subriporopollenites* aff. *constans* P f l u g, 1953 subfsp. 1; *S.* aff. *constans* P f l u g 1953 subfsp. 2; *Triatripollenites* cf. *plicatus* (P f l u g) T h o m s o n, P f l u g, 1953 (Ciula); cf. *Engelhardtioidites* sp.; *Tricolpites* sp. 1 (exine lisse); *Tricolpites* sp. 2 (exine fovéolé); *Tricolporopollenites* spp., *Liliacidites* sp.; *Polyvestibulopollenites* sp. (Pîclișa); Pollen monoporé (Pîclișa); cf. *Oculopollis giganteus* Z a z l i n s k a i a, 1963 (Pîclișa); *Trileites* sp. (Slătioara); *Azolla* sp. (Ciula); *Spermatites* spp., *Microcarpolithes hexagonalis* W a n g e r o v, 1954 (Rusca Montană).

La microflore de toutes les régions, excepté le niveau de Sînpetru plus pauvre, est unitaire, très riche en genres, espèces et nombre d'exemplaires. Le palynofaciès est constitué par des microspores, pollens, tissus végétaux, et plus rarement, à Ciula et Rusca Montană par des mégaspores et restes *incertae sedis* de type *Spermatites* et *Microcarpolithes*. C'est un palynofaciès continental typique dans lequel les dinoflagllés manquent; un seul exemplaire *d'Ophiobolus* sp. a été observé à Loznicioara. Les microspores trilètes lisses de type *Leiotriletes* et *Dictyophyllidites* sont les plus fréquents. Suivent ensuite les spores à cingulum de type *Polypodiaceoisporites*, ensuite les spores ou pollens du groupe Zlivisporites — Triporoletes. Le pollen disaccate est presque absent, le pollen de type *Inaperturopollenites* fréquent.

Les pollens du groupe Normapolles Pflug, 1953 et Postnormapolles Pflug, 1953, constituent le fond de l'association. *Pseudopapilopollis praesubhercynicus* est l'espèce index de l'association étant restreinte au Campanien-Maestrichtien (G ó c z á n, G r o o t, K r u t z s c h, P a c l t o v a, 1967; A n t o n e s c u, 1973) et commune dans toutes les régions. Les espèces du genre *Convexipollis* sont aussi restreintes seulement au Maestrichtien (G ó c z á n et al., 1967; P o r t n i a g h i n a, 1973). *Suemeghipollis triangularis*, *Semioculopollis praedicatus* et *Pseudoculopollis* sp. sont communs au Sénonien. La plupart des espèces présentes dans l'association ont une distribution stratigraphique allant du Crétacé supérieur au Paléogène; nous signallons *Interporopollenites proporus*, *Oculopollis* cf. *baculotrudens*, *Pompeckjoidaepollenites subhercynicus*, *Trudopollis imperfectus*, *Plicapollis conserta*, *Plicapollis peneserta*, *Tricolpites* spp. + *Tricolporopollenites* + *Liliacidites* (du groupe Postnormapolles), *Pseudotrudopollis pseudalnoides* (Maestrichtien-Paléocène). Les espèces qui font leur début au Paléogène ou qui sont cantonnées exclusivement au Paléocène manquent; nous nous rapportons aux espèces des genres *Stephanoporopollenites* T h o m s o n, P f l u g, 1953; *Thomsonipollis* K r u t z s c h, 1960; *Tetrapollis* P f l u g, 1953; *Intratriporopollenites* T h o m s o n, P f l u g, 1953; *Duplopollis* K r u t z s c h, 1959; etc. Aussi nous considérons que l'association à *P. praesubhercynicus* appartient au Maestrichtien supérieur. Enfin, nous remarquons qu'il y a des différences entre nos espèces et celles signalées par Balteș (1966 ; 1980, dans Dragastan, Petrescu, Olaru) à Rusca Montană.

3. *Corrélations palynologiques.* La corrélation palynologique de l'association à *P. praesubhercynicus* des formations susmentionnées se fait avec la microflore du Maestrichtien supérieur des Monts Bakony — Hongrie (Goczan, 1973) et le Maestrichtien supérieur de l'Allemagne de l'Est (Oebisfelder Bild, Góczán et al., 1967). Des traits communs existent avec les microflores de la partie inférieure des couches de Stryi de l'Unité de Skolle (Portniaghina, 1973) et du Maestrichtien de Provence-France et du nord-est de l'Espagne (Médus, 1972). La corrélation avec les dépôts du Maestrichtien des autres unités de Roumanie (les Monts Metaliferi par exemple) s'avère difficile à cause du palynofaciès continental des dépôts examinés.

4. *Problème de l'âge de la formation de Ciula.* Il s'ensuit que l'âge des dépôts à dinosauriens et des dépôts rouges détritiques à intercalations charbonneuses de la région de Ciula appartiennent au Maestrichtien supérieur. Il importe d'élucider le problème de l'âge du reste de la formation de Ciula, donc des dépôts détritiques rouges situés au-dessus des niveaux à microflore maestrichtienne de la partie centale-ouest de la dépression de Hațeg, jusqu'au premiers dépôts miocènes à faune marine. Nous avons découvert dans la vallée de Peștenița, dans une intercalation argileuse grise de la formation de Ciula, une association palynologique dont la composition est semblable à celle de l'association à *P. praesubhercynicus*, mais dans laquelle il y a de nombreux exemplaires de *Proteacidites* sp. Cette association pourrait représenter un niveau plus élevé du Maestrichtien ?. Il est possible pourtant, que cette association soit plus récente (*Proteacidites* persistant au Paléogène) et que la formation de Ciula se continue au Paléogène ? . La question reste ouverte, tenant compte de l'étrange allure de cette association à *Proteacidites* (genre rare en Europe, commun au Maestrichtien et au Paléocène des provinces siberienne et américaine) et du fait que nous n'avons identifié qu'un seul niveau ayant cette association. Aussi la partie inférieure de la formation de Ciula comprenant les dépôts à dinosauriens, ceux des environs de Ciula et, peut-être, ceux de Peștenița, appartiennent donc au Maestrichtien et ne doit pas être comparée avec le faciès semblable mais à faune de molusques et mamifères oligocènes de l'est de la dépression. L'âge de la partie supérieure de la formation de Ciula (maestrichtien ou paléocène) reste à être précisé.

D) Conclusions

Nos recherches ont mises en évidence une microflore à traits communs — une association palynologique à *Pseudopapilopollis praesubhercynicus* d'âge maestrichtien supérieur dans quelques formations au sud-est de la Roumanie. Cette association permet de corréler les formations suivantes : 1) la formation détritique rouge de la rive gauche du Mureș entre Vințul de Jos et Alba-Iulia à Pîclișa ; 2) la formation de Densuș et la partie inférieure de la formation de Ciula comprenant les dépôts à dinosauriens et les dépôts détritiques rouges à intercalations charbonneuses de Ciula de la partie centrale-ouest de la dépression de Haţeg; 3) la partie supérieure contenant des charbons de l'horizon pyroclastique de la partie est de la dépression Rusca Montană. Ces formations appartiennent au Maestrichtien, qui dans cette région de la Roumanie présentent un faciès continental avec des dépôts détritiques rouges, des roches volcanoclastiques (plus fréquentes vers l'ouest) et des charbons. Les gastropodes découverts confirment l'âge maestrichtien supérieur de la formation de Ciula.

Le contenu palynologique du Maestrichtien continental du sud-ouest de la Roumanie peut être corrélé avec celui du Maestrichtien des Monts Bakony de la Hongrie et avec celui du Maestrichtien supérieur de l'Allemagne de l'Est. Les traits communs palynologiques existent aussi avec la partie inférieure des couches de Stryi supérieures de l'unité de Skolle des Carpates Orientales et avec le Maestrichtien de la Provence—France et du nord-est de l'Espagne.

Les réconstitutions paléogéographies qui peuvent être faites par l'intermédiaire de ces formations continentales — résultat de l'activité des paléofleuves qui devaient charrier du matériel détritique et végétal qui s'accumulait dans les tourbières, les manifestations volcaniques, les lacs et les dinosauriens qui vivaient aux voisinage — donnent du paysage à l'époque une image particulière.

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THE BLACK SHALES FORMATION OF THE EAST CARPATHIANS, LITHO-BIOSTRATIGRAPHY AND OIL POTENTIAL [1]

BY

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The Lower Cretaceous deposits of the External Flysch Zone of the East Carpathians (Audia, Tarcău and Marginal Folds Units) developed in a peculiar lithofacies, generally known as the Black Shales Formation (BSF) with different names for the various units (Audia Beds, Sărata Beds, Streiu Beds) where this formation has several distinct lithofacial features, previously mentioned by Filipescu et al. (1952), Dumitrescu (1952), Băncilă (1955), Mirăuță, Mirăuță, (1964), etc.

The BSF represents the oldest deposits in all three structural units of the External Flysch Zone. Their initial basement is unknown because of shearing during overthrusting. At the upper part, this formation is conformably covered by the Variegated Shales Formation (Vraconian-Coniacian).

The age of the BSF was established by several assemblages or isolated specimens of ammonites, belemnites, inocerams, foraminifera, etc. (Table 1), which proved its stratigraphical extension from Hauterivian to Albian. Mention should be made of the Vrancea Half-Window (Marginal Folds Unit) where the micropaleontological evidences of J. Săndulescu (in Dumitrescu et al., 1971– unpublished) established the presence of the Turonian in the upper part of the Streiu Beds.

Recently, very detailed palynological studies (Balteș, 1977 and unpublished; Antonescu et al., 1978; Antonescu et al., in press) carried out in the whole BSF area enriched its paleontological content and established several characteristic associations for each lithostratigraphic subdivision (Table 2).

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TABLE 1

Paleontological content (ammonites, bivalves, etc.) of the Black Shales Formation

BLACK SHALES (= AUDIA BEDS)		AUDIA UNIT	TARCĂU UNIT	SĂRATA BEDS	MARGINAL FOLDS UNIT (Bistrita Half - window)
Up. m. glauc. sandstones	ALBIAN - Lower Vrac.	Neohibolites minimus (List.) - Filipescu et al. (1952); Marinescu (1962) Neohibolites minor Stoll. - Filipescu et al. (1952)	Macrofossils unfound until now	Upper member (Silicified limestones)	Inoceramus sp. - Micu (unpublished)
Middle member with cherts and calcareous concretions	APTIAN - MIDDLE ALBIAN	Neohibolites aff. strombecki G. Müll. emend. Stoll.; Neohibolites aptiensis var. strombeckiformis Stoll.; Neohibolites strombecki G. Müll. - Filipescu et al. (1952) Inoceramus sp. (II) ex gr. neocomiensis d'Orb. - Alexandrescu (1971) Hamites sp. - Alexandrescu (1971) Acanthohoplites sp. - Alexandrescu and J. Ion (1982) Acanthohoplites aff. laticostatus Sinz. - Nicolaescu et al. (1966) Ancyloceras cf. matheronianum d'Orb. - Aghiorghesei et al. (1965) Cheloniceras (Procheloniceras) sp. - Băncilă and Aghiorghesei (1962) Cheloniceras seminodosus (Sinz.) - Ionesi (1971) Anahoplites mantelli Spath - Marinescu (1962)		Middle member with cherts and calcareous concretions	Leymeriella sp. - Micu (unpublished) Douvilleiceras sp.; Hypacanthoplites aff. milletianus d'Orb.; Cleoniceras sp.; Leymeriella sp.; Leymeriella regularis (Bruguière); Leymeriella rencurelensis Jacob; Leymeriella tardefurcata d'Orb. - Joja et al. (1970) Acanthohoplites sp. ex gr. bergeroni - bigoureti (Turculeț, Patrulius, in Mirăuță and E. Mirăuță, 1964)
Lower member with sideritic rocks	UPPER HAUTERIVIAN - BARR.	Cardium sp., Metacerithium sp., Rhynchonella sp. Inoceramus sp. (I) ex gr. neocomiensis d'Orb.; Silesites cf. vulpes (Coq.); Costidiscus ap. ex gr. recticostatus (d'Orb.) - Alexandrescu and J. Ion (1982) Costidiscus recticostatus (d'Orb.) - Filipescu et al. (1952); Pseudothurmannia angulicostata (d'Orb.); Pseudothurmannia sp. - Filipescu and Grigorescu (1966) Neocomites neocomiensis d'Orb.; Breistrofferella (=Leopoldia) castellanensis d'Orb. - Herbich (1878) Phyllopachyceras infundibulum d'Orb. (Müntz, Patrulius, in Bucur, 1980)		Lower member with sideritic rocks	—

1, Sideritic rocks; 2, cherts (lydites); 3, argillaceous shales; 4, calcareous concretions; 5, sandstones; 6, glauconitic sandstones; 7, silicified limestones.

TABLE 2

Palyno-stratigraphical Zonation of the Black Shales Formation

<table>
<tr><th colspan="3" rowspan="3">Chrono- and Phytostratigraphy</th><th colspan="4">Palyno-stratigraphic subdivisions</th></tr>
<tr><th rowspan="2">Audia Unit</th><th rowspan="2">Tarcău Unit</th><th colspan="2">Marginal Folds Unit</th></tr>
<tr><th>Bistrița Half-window</th><th>Vrancea Half-window</th></tr>
<tr><td rowspan="7">Upper Valanginian – Lower Vraconian (Lowermost part)</td><td rowspan="7">Black Shales (= Audia Beds)</td><td rowspan="2">Upper member with glauconitic sandstones</td><td colspan="3">VII Ovoidinium verrucosum and Pseudoceratium dettmannae assemblage</td><td></td></tr>
<tr><td colspan="4">VI Litosphaeridium siphoniphorum, Carpodinium obliquicostatum, Hexagonifera chlamidata and Stephodinium coronatum assemblage</td></tr>
<tr><td rowspan="2">Middle member with lydienes and calcareous concretions</td><td colspan="4">V Chlamidophorella nyei, Cribroperidinium orthoceras, Astrocysta cretacea and Ovoidinium scabrosum assemblage</td></tr>
<tr><td colspan="4">IV Prilixosphaeridium deirense, Cribroperidinium sepimentum and Aptea polymorpha assemblage</td></tr>
<tr><td rowspan="3">Lower member with sideritic rocks</td><td>Dingodinium albertii Druggidium deflandrei and Meiourogonyaulax stoveri assemblage</td><td>III</td><td></td><td></td></tr>
<tr><td>Broomea exigua and Pseudoceratium pelliferum assemblage</td><td>II</td><td></td><td></td></tr>
<tr><td>Muderongla tetracantha and Kleithriasphaeridium fasciatum assemblage</td><td>I</td><td colspan="2">I – VII Number of the assemblages</td></tr>
</table>

Lithostratigraphy. Except for the Streiu Beds present only in the Vrancea Half-Window (Marginal Folds Unit), the BSF is lithologically subdivided into three main members (Table 1).

In the Audia Unit the lower member of the BSF consists of black shales, siltstones interbedded with calcareous sandstones. Sideritic limestones are also present, as lenses or, more scarcely, as continuous layers. The next member is generally made of argillaceous black shales and siltstones with lydites (cherts) intercalations. Sandstones are rare or absent. The upper member of the BSF is mostly represented by well-bedded glauconitic sandstones of the orthoquartzitic type with black shales and siltstones intercalations. BSF sandstones sometimes contain greenschists fragments derived from the foreland of a Dobrogean type area which proves not only a Carpathian, but also a foreland origin for this formation.

The middle and upper members contain levels of sedimentary breccias rich in granitoids, granodiorite fragments with red feldspar, more frequent in the Variegated Shales which overlie the BSF. These fragments originated in a cordillera (The Cuman Cordillera, Murgeanu, 1937), most probably developed west of the BSF sedimentary area. The uneven distribution of granodiorite-bearing breccias along the Audia and Tarcău units suggests that this cordillera was rather a chain of islands discontinuously supplying the adjacent small areas with materials.

The BSF of the Tarcău Unit generally shows the same lithofacies as the Audia Unit, the main differences chiefly consisting in the thinning tendency, especially southwards, of the glauconitic sandstones member (Săndulescu & Jana Săndulescu, 1964), locally replaced by quartzitic sandstones with calcareous cement (Băncilă, 1955).

The Sărata Beds, outcropping in the Bistriţa Half-Window (Marginal Folds Unit), have also been subdivided into three members (Mirăuţă & Elena Mirăuţă, 1964), only the upper member lithologically differs from those separated in the innermost units. This sandy limestones member, regarded as an outer equivalent of the glauconitic sandstones, consists of sandy, organogenous limestones, interbedded with black, bituminous shales. The laminae or lenses of spongolitic rocks are present in the middle part of each sandy limestone layer.

The Streiu Beds, the equivalent of the BSF in the Vrancea Half-Window (Marginal Folds Unit), are generally similar to the BSF. They consist of a comparatively monotonous, rhythmical alternation of black, sometimes bituminous shales with graded calcareous sandstones. Rather rrequent, beds or lenses of sideritic marly-limestones and thin conglomefates with greenschists elements of a Dobrogean type are also present.

Palynological assemblages. The BSF has supplied rich palynoassemblages. The lower member with sideritic rocks is dominated by continental microfloral elements (microspores and gymnospermic, coniferal and benttital pollen). Towards the middle member and the upper member, the marine microflora (dinoflagellates) becomes quantitatively predominant. Up to the present, in the whole BSF there have been identified seven specific palyno-assemblages (Table 2, numbered I — VII) corresponding to the Upper Valanginian (?) — lowermost Vraconian stratigraphic interval.

In the lower member with sideritic rocks, three palyno- assemblages have been identified (I, II, III) and in only one place — the Marginal Folds Unit in the Bistriţa Half-Window — at the top of the deposits in this member, the 4th assemblage has also been located, which is otherwise fairly represented in the basal middle member with cherts and calcareous concretions.

I. *The Muderongia tetrachanta and Kleithriasphaeridium fasciatum Association* (*? Upper Valanginian*). It has been solely identified in the lower part of the complex, only in the Audia Unit (the Moldova Valley, the Ostra Valley, Petriceni, near Tg. Secuiesc, Chiruş and Zagonul Mic streams, south of Covasna). The insertion of this association in the Upper Valanginian interval is temporary, since dinoflagellates are scarce, and the existing spore and pollen species have a wider stratigraphic distribution.

II. *The Broomea exigua and Pseudoceratium peliferum Association (Middle-Upper Hauterivian)*. It has been revealed in the Audia Unit (the Moldova Valley, at Moldova Sulița, then the Corlăţeni-Cîmpulung Moldovenesc Stream, the Suha Valley, Petriceni, Chiruş and Zagonul Mic streams, south of Covasna). The dinoflagellates content is richer as compared to Association I, which constituted the separation criterion for the latter.

III. *The Dingodinium albertii, Meiourogonyaulax stoveri and Druggidium deflandrei Association (Upper Barremian)*. It has been located in the Audia Unit (the Demăcuşa-Moldovița Valley and the Corlăţeni-Cîmpulung Moldovenesc stream), and in the Marginal Folds Unit (the Bistrița Half-Window) in the Horăicioara Valley. The association is characterized by acmezone genera and species occurrence in the Upper Barremian. This association, prevailing in phytoplankton, extends over quite a large areal. In the middle member with cherts, the Associations IV, V and VI have been identified, the last passing on into the upper member as well (the upper member with glauconitic sandstones).

IV. *The Prolixophaeridium deirense, Cribroperidinium sepimentum and Aptea polymorpha Association (Aptian)*. It has been dsclosed in all three structural units. In the Audia Unit it occurs in the northern part (the Demăcuşa-Moldovița Valley, the Paltin Stream, the Sadova Stream, the Suha Mică and Suha Mare streams and the Ostra Valley) and in the southern part (the Păpăuț and Delțeg valleys near Zagon, the Bota and Crasna-Buzău valleys). Within the Tarcău Unit, it has been also discovered at Stulpicani-Suha Bucovineană and Plotonița. It has equally been crossed by certain drillings (Aţa-Brateş, Ojdula, Comandău, Bîsca-Cernat and Băile Siriu). In the Putna-Vrancea Half-Window, the association has been encountered on the Strei, Cireşu streams, a.s.o.

It is necessary to stress that this association has also been identified in the Bistrița Half-Window, at the top of the lower member with sideritic rocks, namely on the Horăicioara Valley. The association is characterized by species either with a first occurrence during the Aptian or showing part of the acmezones at this level.

V. *The Chlamidophorella nyei, Cribroperidinium orthoceras, Astrocysta cretacea and Ovoidinium scabrosum Association (Lower-Middle Albian)*. It has been encountered in the same places as the previous (IV). Moreover, it also occurs in the Audia Valley, the Bicaz Valley, the Zăbrătău – Buzău Valley and in the Bîsca-Cernat well (the Tarcău Unit), as well as in the Cuejdiu and Sărata valleys around Piatra Neamț (the Marginal Folds Unit). Taking into account the larger stratigraphic distribution of the index species of the association, it is possible for its lower part to include, partly, the Upper Aptian.

VI. *The Litosphaeridium siphoniphorum, Carpodinium obliquicostatum, Hexagonifera chlamidata and Stephodinium coronatum Association (Upper Albian)*. It has been encountered at the uppermost middle member (with cherts and calcareous concretions) in all structural units. It has been equally identified in the upper member, in the following places : the Corlăţeni — Cîmpulung Moldovenesc Stream in the Audia Unit, the Măguricea-Găineşti Valley (the Tarcău Unit)) and the Cuejdiu Valley (the Marginal Folds Unit), as well as in certain boreholes, such as Bîsca cu Cale, etc. (the Tarcău Unit). The association is characterized by the presence of some genera and species first occurring in the Upper Aptian (the "inflatum zone"), thus being one of the richest and best individualized associations. The deposits of the upper subdivision (the upper member with glauconitic sandstones) include in their lower part the Association VI and in their upper part the Association VII, which also extends in the formation basis over the Black Shales, i.e. in the Variegated Shales Formation (Vraconian-Coniacian).

VII. *The Ovoidinium verrucosum and Pseudoceratium dettmannae Association (Vraconian)*. It has been found in the same places as the previous one, only in the central and northern parts of the East Carpathians (the Măguricea-Găineşti Valley — the Tarcău Unit and the Corlăţeni Stream—the Audia Unit). The association is characterized by index species occurrence, which at a European level are solely stocked in the Vraconian, the rest of their componence resembling Association VI.

The (?) Valanginian and Hauterivian Associations (I and II) correlate to the palyno-associations in the Sinaia Beds (Neocomian) especially to those in the pelitic sequence from the Prahova Valley, to the Hauterivian from Dealul Sasului (the Dîmbovicioara Tunnel) and to the lower part of the Braşov marls. The Associations III and IV correlate to those in the upper part of the Comarnic Beds in the Ceahlău Unit, the Barremian limestones in the Dîmbovicioara Tunnel, the upper part of the Braşov Marls the lower part of the Sviniţa Marls (Barremian, in Banat), as well as to the palyno-associations of the Lower Aptian in the stratotype. The Association V (and VI?) correlates to the associations identified in the Palanca Beds (the Bistriţa, Bicaz and Trotuş valleys), especially at the Lower-Middle Aptian level.

The Associations VI and VII allow larger interregional correlations and parallel may be drawn to the Upper Albian and Vraconian microfloras in the Carpathian Foreland, as well as to those in the stratotypes, in southern France and south-eastern England.

Sedimentology and basinal evolution. The East Carpathians BSF represents a Flysch Formation deposited in a sedimentary basin in which reducing (euxinic and subeuxinic) conditions prevail. As a whole, this formation has a pararhythmic character, with a large variability (from 1 : 1 to 1 : 20) between the thicknesses of the epiclastic and lutitic terms. The lithological, stratonomic and sedimentological features of the BSF indicate, at least in case of the Audia Unit, two informal subdivisions, outlining the major basinal evolutionary stages : a lower, prediastrophic subformation, corresponding to the first two members of this formation ; and an upper, syndiastrophic subformation, represented by the upper member, of glauconitic silicified sandstones.

The lower subformation is synchronous with the sedimentary basin separation , when subsidence is still uncompensated by sedimentation.

During the deposition of this subformation, two sedimentogenetic stages can be distingushed : the "vacuity stage", generally characterizing the lower member, and the "shaly flysch", corresponding to the median member with cherts of the BSF. The "vacuity stage" is characterized by typical turbiditic sequences, represented by subgraywackes and bioclastic sandstones alternating with pelagic episodes. The sandstones are gradually bedded with scour marks at their lower part. Microbiosparitic limestones, representing the pelagic episodes, are also gradually bedded, planktonic foraminifera accumulating in the lower part of each bed. Geochemically, the euxinic episodes favourable to organic matter preservation and to syngenetic pyrite accumulation, and the subeuxinic episodes (represented by sideritic rocks) alternate with neutral or weakly oxidant ones, while authigenic glauconite formed.

The presence of bioclasts, recorded at several levels, is connected to the massive supplies of organogeneous remains (briozoans, molluscs, sponges, red and green algae), thus deriving from a shelf zone and being mixed autochthonous bioclasts (crinoids, benthic forams, etc.). The latter demonstrate that aerobic conditions were, at least for a while, reinstalled. This situation determined Grigorescu (1971) to separate a "bioclastic horizon" in the Covasna area, subsequently generalized in this lower part of the BSF in the Audia and the Tarcău units (Grigorescu & Alexandrescu, 1977).

The "shaly flysch stage" is characterized by a greater stratonomic and lithostratigraphic uniformity. Euxinic conditions prevail here, subeuxinic episodes being known only in the lower part of the middle member with cherts. Silicolites are mostly represented by cherts, more rarely by spongolites. Towards the upper part of this "stage", as a consequence of the Mesocretaceous tectogenic phase, some intrageosynclinal cordilleras were formed, furnishing "exotic" rocks such as granodiorites with biotite, porphyric diorites, microdiorites, etc. (Filipescu & Alexandrescu, 1962; Grigorescu & Anastasiu, 1976).

Paleocurrent measurements (Dumitriu & Dumitriu, 1965; Joja & Dumitriu, 1972) indicate transverse or longitudinal transport directions. The presence of iron in the sideritic rocks requires a large emerged area in the basin vicinity, where lateritic soils widely developed. During the deposition of this subformation, the bilateral supply was occasionally accompanied by an "internal" one, due to the presence of the cordilleras.

The upper subformation is represented by a sandy-shaly flysch in which fine- to medium-grained glauconitic sandstones are interbedded with siltstones and black or gray shales. The petrographic study of the sandstones in the Covasna Valley outcrop (Grigorescu, 1970) reveals some important features as follows : the sandstones (0.1–1.5 m thick) gradually bedded, with numerous sole marks, are oligomictic, of a quartzarenite type (more than 95 % of epiclasts are represented by quartz and metaquartzite grains) ; fine fraction (less than 0.2 mm in diameter) is 70 % of the granulometric fractions of these sandstones ; the sorting index varies between 1.2 – 1.6.

In comparison with the lower subformation the bioclasts are more scarce here, mostly represented by sponge spicules. Geochemically, the environment during the sedimentation of the upper subformation was generally neutral to weakly oxidant, allowing for the glauconitic authigenesis. Subeuxinic conditions diminished as compared to those for the lower subformation. During the sedimentation of the upper subformation in the Audia and Tarcău Units, in the Marginal Folds Unit sandy-limestones with cherts accumulated in the Bistriţa Half-Window area, while

TABLE 3

Oil potential of the Black Shales Formation

Chrono- and lithostratigraphy			Microvegetal substance analysis						Oil potential	Hydrocarbons possibly generated	
			Kerogen type	OMI	Ro	Fl(UV)	T(%)	OD		Types	Quantities (Kg/t)*
					average values						
Black shales (= Audia Beds)	Upper Hauterivian – Lower Vraconian (Lower most part)	Upper member with glauconitic sandstones	F.X.	0.8–2.1	0.55	v.m. (480 nm)	75	weakly	null	–	–
		Middle member with cherts and calcareoos concretions	F.X.	3.8–3.95	1,57	g.o.m. (580 nm)	63	strongly	moderate	Condence and weakly oil	2.01–5
			X.F.		1.30	1.30 g.m (530 nm)			weakly	Condence, heavy oil and dry gases	0.5–2.0
			A.F. (s)	3.92–4.15			60				
		Lower member with sideritic rocks	X.F. (A)	2.4–2.8	2.02	v.s. (500 nm)	70	moderate	very weakly	oil and semi-dry methan	0.01–0.5

Legend

v = green; g = yellow; p = orange; m = matly; s = brightly; nm = nannomicrons; OMI = organic metamorphism index; Ro = vitrinite reflectance; Fl(UV) = Fluorescence (ultraviolet) T = Transglucency; OD = organic dyagenesis; F = Phytoplanktogenous; X = Xylogenous; A = amorphogenous; S = Sporogenous; * = after Espitalie et al. (1977)

in the Vrancea Half-Window a monotonous sequence of black shales and sandstones or silt sandstones continuously deposited from Upper Hauterivian to Turonian (?).

Oil potential. The following conclusions were reached through an original investigation methodology of the oil potential (Balteș, 1973): The BSF metamorphism generally displays different values between 2.5 — 4.1 for the lower member with sideritic rocks and lesser values for the other two members (Table 3). The microvegetal material predominantly belongs to the amorphogenous and subordinate sporo-and phytoplanktogenous groups, in the lower part, while in the middle and upper parts it is characterized by the phytoplanktogenous and subordinate xylogenous materials. The lower member with sideritic rocks displays all the oil source-rock features, while the hydrocarbons possibly generated by it could belong to heavy oil or thermic methane. The uppermost lower member with sideritic rocks, together with the shaly member with lydites (cherts), are affected by a strong metamorphism, including values between 3.8 — 4.1. Their petroligenous potential is limited to heavy oil and to restricted thermic methane amounts. The uppermost BSF, i.e. the upper member with glauconitic sandstone, has a lower metamorphism, its values varying between 0.8 — 2.1, unable to generate hydrocarbons.

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RÉFLEXIONS SUR LES CALPIONELLES REMANIÉES DANS LE CRETACÉ SUPÉRIEUR DE L'UNITÉ DES PLIS MARGINAUX (CARPATHES ORIENTALES, ROUMANIE)[1]

PAR

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La surprenante présence de Calpionelles, essentiellement empruntées au Tithonique terminal-Berriasien inférieur dans des sédiments du Crétacé supérieur appartenant à la zone la plus externe du flysch des Carpathes roumaines ("unité des plis marginaux") pose divers problèmes, d'ordre sédimentologique, structural et paléogéographique Nous allons tenter d'y répondre.

I. Situation géologique

Dans les demi-fenêtres de Vrancea et de Bistriţa, les couches de Lepşa représentent le faciès le plus externe du Sénonien des Carpathes orientales. Placées dans le Turonien par Dumitrescu (1958) ou dans le Turonien-Sénonien inférieur par Mirăuţă et Mirăuţă (1964), leur âge sénonien, voire paléocène inférieur — au moins dans la région de Vrancea — a été prouvé par une riche microfaune (Săndulescu in Dumitrescu et al., 1970, 1971) associée à de nombreux restes d'Inocérames, grands Foraminifères, Bryozoaires, Algues calcaires etc. Ultérieurement Micu (1973) a signalé la présence d'un microfaciès à *Pithonella ovalis* à la partie inférieure de ces Couches de Lepşa, en concluant que celles-ci comportent également du Turonien supérieur, dans la demi-fenêtre de Bistriţa.

Situées au-dessus des Couches de Tisaru supérieures dans la demi-fenêtre de Vrancea et reposant sur les Argiles bariolées (Vraconien-Turonien) dans la demi-fenêtre de Bistriţa, les Couches de Lepşa sont surmontées respectivement par les Couches de Caşin (Paléocène) dans la région de Vrancea et par les Couches de Runcu dans la région de Bistriţa. Récem-

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ment M i c u (1977), en s'appuyant sur des arguments micropaléontologiques, a démontré l'âge paléocène des Conglomérats de Horăicioara, considérés auparavant (M i r ă u ț ă, M i r ă u ț ă, 1964) comme appartenant au Sénonien. Ces conglomérats constituent en réalité une particularité locale de la partie inférieure des Couches de Runcu ; ils représentent ainsi le parfait équivalent des Conglomérats de Piatra Streiului qui remplacent localement, sur le flanc est de l'anticlinal de Streiu dans la région de Vrancea, les Couches de Cașin inférieures (D u m i t r e s c u, 1952 ; 1963).

La lithologie des Couches de Lepșa, assez monotone, est représentée par une alternance plus ou moins rythmique de grès calcaires granoclassés, souvent microconglomératiques à leur partie basale, de calcaires micritiques en couches épaisses de 15 — 30 cm, contenant fréquemment des chailles, et de marnes grises en couches ne dépassant pas quelques centimètres. Parfois la couleur des marnes et des argiles est rouge ou gris-verdâtre. Vers le haut, les Couches de Lepșa présentent des passées plus marneuses, dans lesquels sont intercalés plusieurs niveaux de brèches organogènes à éléments verts de type dobrogéen.

La puissance des Couches de Lepșa varie de 200 à 300 m. Dans le cas de la demi-fenêtre de Bistrița leur épaisseur est plus difficile à établir à cause de nombreux accidents tectoniques, produits pendant le diapirisme des Couches de Sărata (Crétacé inférieur) qui affleurent dans l'axe de l'anticlinal de Horaița-Doamna.

II. Les Calpionelles des couches de Lepșa

Parmi les échantillons recueillis à la partie inférieure des Couches de Lepșa de la région de Vrancea quelques-uns ont révélé des Calpionelles. Celles-ci se trouvent dans des micrites à fréquents petits quartz détritiques, riches en spicules siliceux de Spongiaires, en Pithonelles, avec également parfois des Radiolaires sphériques et des Gümbélines. Dans quelques cas, les sections de Calpionelles sont relativement fréquentes. La douzaine d'exemplaires déterminables peuvent provenir d'un même niveau primitif, du Tithonique terminal-Berriasien inférieur, caractérisé par : *Calpionella alpina* plus ou moins typique (espèce dominante), *Calpionella* ex gr. *alpina* (grande forme, relativement haute, ébauchant un passage à *Calpionella elliptica)*, *Crassicollaria parvula*, plus ou moins typique, et *Crassicollaria* sp., forme haute. En outre, quelques *Tintinnopsella* ex gr. *carpathica* proviennent des horizons post-tithoniques. D'autres échantillons ont montré la présence de quelques *Calpionella alpina* typiques et *Calpionella* cf. *alpina* à petit collier, d'un âge imprécis, du Tithonique supérieur au Berriasien franc. Un calcaire gréseux riche en spicules de Spongiaires et en Radiolaires a montré une section de ? *Lorenziella* cf. *hungarica*, espèce connue dans le Berriasien supérieur.

Les échantillons provenant des Couches de Lepșa de la demi-fenêtre de Bistrița ont également fourni des exemplaires typiques de *Calpionella alpina* et *Crassicollaria* sp., toujours associés à de nombreux spécimens de *Pithonella ovalis* et *Pithonella sphaerica*, spicules de Spongiaires, Gumbélines etc. Il faut noter le fait que dans les deux régions quelques échantillons contiennent, à coté des Calpionelles, diverses espèces de *Globotruncana*, *Rotalipora*, *Heterohelix* etc.

III. Le problème du remaniement des Calpionelles

Le remaniement des Calpionelles est un phénomène relativement commun si les assises qui les englobent, encore insufisamment indurées et déposées sur des pentes instables, sont affectées de glissements sous-marins entraînant une resédimentation. Cette situation est banale dans le Néocomien marneux de la fosse vocontienne dans les Alpes françaises externes (B e a u d o i n, 1977), dans le préflysch néocomien de la nappe du Jebel Tisirene (Rif marocain septentrional, inédit) ou dans la zone de Svinița dans les Carpathes méridionales (R u s u , 1970 ; A v r a m, 1976). Toutefois, dans ces divers cas le remaniement suit de très près la sédimentation initiale, effectuée dans une région très voisine, et les microorganismes sont englobés dans de petits, galets, plus ou moins "dissous" au sein du sédiment final.

Dans le cas des Couches de Lepșa, il s'agit au contraire d'une véritable érosion de couches déposées 50— 55 M.A. plus tôt, et théoriquement séparées des Couches de Lepșa par une importante colonne sédimentaire. On peut donc soupçonner que ce remaniement traduit une phase tectonique ayant amené— latéralement — des couches relativement anciennes à l'affleurement. Des exemples similaires sont connus en divers points de l'Apennin et des Alpes méridionales : ainsi dans les Couches à Rotalipores des unités toscanes de la région de Livourne (C o n t i et A n d r i, 1966) ou dans le Cénomanien-Turonien à faciès flysch de la région du lac de Varese (R e g g i o r i , 1958). Mais, alors que dans ces exemples la matrice calcaire se différenciait en général assez bien de la micrite plus sombre, remplissant les loricas des Calpionelles — chacune d'elles, protégée par son remplissage, correspondant à un véritable microgalet — dans le cas de Couches de Lepșa, sans exception, la micrite de l'intérieur des loricas ne peut pas être différenciée, sous le microscope optique, du ciment micritique sénonien : les assises remaniées devaient donc être ici particulièrement tendres, évidemment marneuses et de milieu probablement bathyal. Ces Calpionelles ont en effet des loricas de taille normale et à test fin, telles qu'on en connaît dans des faciès de haute mer et d'eaux relativement froides.

Il faut noter que les Calpionelles n'ont pas pu être remaniées "individuellement". Elles ont été primitivement englobées dans de petits galets de micrites ou de marnes, puis progressivement dégagées de ceux-ci Sans exception, on ne retrouve pas les contours de galets primitifs.

IV. Origine paléogéographique des niveaux à Calpionelles

Quant à la source des Calpionelles, nous pensons qu'une origine à partir de l'avant-pays carpathique est la plus plausible en nous basant sur les raisons suivantes :

a) Le matériel ayant cette origine externe, et au sein duquel les schistes verts de type dobrogéen représentent de loin la grande majorité, se trouve à tous les niveaux stratigraphiques dans le flysch externe des Carpathes orientales. Assez souvent, dans les grès et même dans les micrites des Couches de Lepșa, on trouve des fragments anguleux de schistes verts ayant des dimensions allant jusqu'à 4 — 5 cm ou plus. Donc une source externe active de détritus existait sans aucune doute durant le dépôt des Couches de Lepşa. En outre la présence, dans ces niveaux ou dans leurs équivalents plus internes, de spécimens remaniés de Dinoflagellés ainsi que de spores et pollens paléozoïques, jurassiques et du Crétacé inférieur (Antonescu in Săndulescu et al., 1979) s'explique facilement si ces restes proviennent de l'extérieur des Carpathes, compte tenu du fait que des depôts de ces âges sont connus dans la couverture de la plate-forme.

b) Une source interne (ouest) pour les Calpionelles remaniées nous semble plus douteuse, d'une part parce que le "passage" de tels débris n'a pas été jusqu'à présent signalé dans les assises sénoniennes des nappes plus internes, et d'autre part parce que durant le Crétacé supérieur la "morphologie" en creux du sillon carpathique a dû logiquement empêcher un transport de matériel provenant de ce côté.

c) La présence de galets de micrite contenant des Calpionelles a été signalée aussi dans les Conglomérats de Horăicioara d'âge paléocène, situés au-dessus des Couches de Lepşa (Micu, 1977, pl. IV) et aussi dans les dépôts détritiques grossiers paléocènes de la nappe de Tarcău (Alexandrescu, 1971). Sans exception, dans tous ces dépôts les schistes verts dobrogéens prédominent, montrant sans aucun doute l'origine externe du matériel détritique. La présence des galets à Calpionelles s'explique d'une manière satisfaisante car, dans le cas de ces conglomérats, les galets n'ont pas eu la possibilité d'être "débités" en menus fragments à cause de la grande vitesse de sédimentation.

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QUESTIONS

S. Gasiorowski. Remaniement possible des Calpionelles contenues dans les divers calcaires dits de Stramberg.

Réponse : Dans les galets de calcaires de type Stamberg englobés dans divers dépôts conglomératiques plus jeunes que les Couches de Lepşa nous n'avons pas trouvé des Calpionelles.

BIOSTRATIGRAPHISCHE UND FAZIELLE MERKMALE DER „GOSAUFORMATION" IM APUSENI GEBIRGE [1]

VON

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Einleitung

Die, in den letzten Jahren durchgeführten Revidierungen an den Alpinen Gosauschichten (Wiedmann, 1978; Herm et al., 1979, u.a.) haben zu neuen Angaben zur Gliederung und Stratigraphie dieser Schichtfolgen gedeutet, die zu Unterschiede im Vergleich zu der bis jetzt bekannten Untergliederung geführt haben.

Innerhalb der rumänischen Karpaten stellt das Apuseni Gebirge durch seine, hauptsächlich im nördlichen Teil, markannte vorgosauischen Deckenbautektonik, ein Gebiet dar, wo die Gosauablagerungen typisch entwickelt sind.

Die gegenwärtige Arbeit nimmt sich vor, infolge einer kurzen Analyse die biostratigraphischen Merkmale in Augenscheinlichkeit hervorzuheben und auf dieser Basis die Ähnlichkeiten und Unterschiede mit der Ostalpinen-Gosau zu erörtern.

Der Gosau-Begriff und die Biostratigraphie der „Gosau"-Ablagerungen im Apuseni Gebirge

Der Gosau Begriff so wie er noch heute angenommen wird umfasst zwei Elemente: der erste betrifft den transgresiven Charakter Über einen Untergrund dessen Deckenbau in der Vorgosau ischen Phase stadtfand und der zweite betrifft den Schelf öfters litoralen Charakter der Ablagerungen, die durch Rudistenkalke, Inoceramen und Ammonitenmergel, manchmal kohlenführende Ablagerungen, gekennzeichnet sind.

In dem nördlichen Apuseni Gebirge sind beide die oben erwähnten Merkmale vorhanden, in den südlichen Apuseniden gibt es aber keine Deutungen über eine borgosauische Deckentektonik, obwohl auch dort „Gosauablagerungen" anwesend sind.

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Hauptsächlich in den nördlichen Apuseniden bereitet die Korrelation der „Gosauschichten" in den, heute teilweise isolierten, Becken noch immer Schweirigkeiten. Infolge aber des bestehenden faunistischen un mikrofaunistischen Materials wovon auch die Rudistenassoziationen in Betracht genommen worden sind, gibt es die Möglichkeit eines Überblicks über die stratigraphische Gliederung sowohl wie die Gesammtentwicklung der Senon-Ablagerungen.

Im nördlichen Apuseni-Gebirge stellt das Roşia — Becken ein günstiges Gebiet dar, für das vergleichende stratigraphische Studium der Senonablagerungen die hier sowohl mergelige wie auch kalkige Rudistenführende Fazien mit denen die sich verzahnen bieten. Die mergelige Fazies liefert von unten nach oben *Marginotruncana angusticarinata lapparenti* Gandolfi, *G. lapparenti* Gandolfi wie auch *Gaudryceras mite* (Hauer) die ein Coniac-Alter nachweisen, dan *Inoceramus (Cordiceramus) mülleri recklingensis* Seitz., *I. (Cord.)*. sp. ex gr. *I. mülleri* Petrascheck, *I. (Cord.)* sp. ex gr. *I. platycephalus* Sornay, *I. (Cord.) bueltenensis* Seitz, *I. (Cord.)* sp. ex gr. *I. mülleri recklingensis* Seitz, *I. (Cord.)* sp. ex gr. *I. alpinus* Tsagareli, *I. (Selenoceramus) selenae* Seitz, *I. (Trochoceramus)* cf. *monticuli* Fugg&Kostn., *I. (Endocostea)* cf. *barabini* (Mort.), *I. (Platyceramus) Cycloides* n. sp. die dem Santon entsprechen.

In was die mit der erwähnten mergeligen Abfolge sich verzahnenden Rudistenkalk-Abfolge, betrifft, hier wurden drei Rudisten Assoziationen identifiziert (Lupu, 1976); die erste die aus *Vaccinites gosaviensis* Sonv., *Hippurites sarthacensis peroni* Douvillé, *H. praecessor* Douv., *H. Matheroni* Douv., *H. colliciatus* Woodw., *Gorianovicia paronai* (Wiontzek), *Lapeirouseia zitteli* Douv., *L. pervinquierei* (Toucas), *Sauvagesia tehnicostata* Polsak, *Sphaerulites boreaui* Toucas, *Praeradiolites caderensis* Toucas, *P. toucasianus* (d'Orb.), *P. sinuatus* (d'Orb.), *P. fissicostatus* (d'Orb.), *Biradiolites alatus* n. sp., *B. biplicatus* n. sp., *Plagioptychus paradoxus* Matheron die auf ein Santon Alter hinweisen; eine zweite Assoziation die aus *Vaccinites sulcatus* Defr., *Hippurites colliciatus* Woodw., *H. nabressinensis* Futt., *Radiolites subquamosus* Toucas, *R. gastaldianus* Pirona, *Neoradiolites matheroni* (Toucas), *Radiolites aurigerensis* Mun.-Chalm., *R. squamosus* Toucas die auf ein ungefähr ober Santon-unter Campan hinweist und eine dritte Assoziation: *Vaccinites sulcatus* Defr., *V. oppeli* Douv., *V. vredenburgi* Kühn, *V. archiaci* Mun.-Chalm., *Radiolites subsquamosus* Toucas, *Praeradiolites soubtoucasi* Toucas, *Klinghardtites musculosus* (Klingh.), *Radiolites angeoides* (Lapeirouse) die Campan Alter vertritt.

Im Boroder Becken erscheinen, in einigen Abfolgen, drei Rudisten Niveaus die folgende Assoziationen besitzen: das erste Niveau: *Vaccinatus sulcatus* Douv., *V. gosaviensis* Douv., *V. oppeli santoniensis* Kühn, *V. cornuvaccinum gaudryi* Munier-Chalmas, *V. gosaviensis acicularis* Lupu, *Hippurites praecessor* Douvillé, *Praeradiolites caderensis* Toucas, *Plagioptychus toucasi* Matheron, *P. maestrei* nov., sp., *P. paradoxus* Matheron, *P. borodense* Lupu, das Santon Alter besitzt, und das zweite; *Vaccinites sulcatus* Defr., *V. inaequicostatus* Münst., *V. archiaci* Mun.-Chalm., *V. oppeli* Douv., *V. vre-*

denburgi Kühn, *Praeradiolites aristidis* (Mun.-Chalm.), *Bournonia* aff. *africana* Douv. die auf ein Campan Alter hinweisen. Das dritte Rudistenniveau das manchmal direkt auf den Vorsenonen Untergrund lagert, besteht aus *Colveraia* und *Joufia* die in Assoziation mit *Clypeorbis mamillata* (Schlumberger), *Lepidorbitoides minor* (Schlumberger), *Simplorbites gensacicus* (Leymeire), *Siderolites* sp. deren Obercampan-Untermaastricht Alter bekannt ist.

Ein Äquivalent des dritten Rudisten-Niveaus befindet sich am oberen Teil einer Senonabfolge im Remeţi Becken wo es durch Exemplare von *Pseudopolyconites* vertreten ist. Das stratigraphische Niveau wo sich diese Rudisten befinden wurde als Untermaastricht bezeichnet, gemäss den Angaben aus der Jugoslavischen Litteratur (Milovanović, 1960), sowie infolge neuer Daten welche im Horehronie Gebiet, in den Slowakischen Karpaten (Lupu, 1976) *Pseudopolyconites* Exemplare im Zusammenhang mit *Globotruncana stuarti* (Lapp.), *Gl. linneana* (d'Orb.), *Gl. elevata* (Brotzen), *Rzehakina* sp., *Miliammina* sp., ein Obercampan-Unter Maastricht Alter beweisen.

Im südlichen Apuseni Gebirge sind Gosauablagerungen nur an der nördlichen Flanke der Senon-Abfolge representiert, dort wo sie transgresiv am krystallinen Untergrund lagern.

Die Senontransgression beginnt im westlichen Teil der Süd-Apuseniden mit dem Coniacien. In diesem Gebiet — das Drocea Gebirge — lagert am Krystallinen Untergrund eine detritische Serie die meistens von Sandsteinen und Mergel representiert ist, die aber auch Rudistenkalklinsen einschliesst.

Die Fauna der sandigmergeligen Schichtfolge besteht am unteren Teil von *Plagioptychus arnaudi* Douv. und *Reesideoceras* sp. die als Coniacien vertretend bezeichnet werden können. Die entsprechenden Rudistenkalkriffe besitzen *Hippurites-socialis* Douv., *Vaccinites oppeli felixi* Kühn., *V. giganteus* d'Hombr. Firm., *H. praerenensis* Touc.

Die Schichten mit *Inoceramus mülleri recklingensis* Seitz, *Baculites* aff. *asper* Marton besitzen keine entsprechenden Rudistenriffe im Drocea-Gebirge.

Die Gosautransgression weist in den südlichen Apuseniden einen diachronen Charakter indem sie den Zentral-nördlichen Teil des Gebietes im Ober-Santon-Campan erreicht. Manchmal lagern hier direkt am Krystallinen Untergrund Rudistenriffe mit *Hippurites nabresinensis* Futt., *V. oppeli* Douv., *Lapeirouseia pervinquerei* (Toucas), *Sphaerulites boreaui* Toucas, *Sphaerulites* sp., *Radiolites* sp. die als ungefähr der Santon/Campan Wende entsprechen nach dem in den darüber lagernden Mergel *Inoceramus balticus* Boehn, i. ex gr. *monticuli* Fug. a Kastn. gefunden worden sind, und welche ein Campan-Alter bezeichnen.

Im nördöstlichen Teil des südlichen Apuseni Gebietes lagert am Krystallinen Untergrund ein Kalk-Niveau das folgende Arten enthält: *Vaccinites gosaviensis* Douv., *V. sulcatus* Defr., *V. praesulcatus* Douv., *V. cornuvaccinum gaudry* Mun.-Chalm., *Praeradiolites* sp., *Durania* sp., *Radiolites* sp. Die darüber lagernde Sandsteinschicht hat einige Exemplare von *Echinocorys vulgaris*, Breynius gefördert. Das darüber lagernde zweite Rudistenkalk-Niveau besitzt *Vaccinites oppeli* S o u v., *Hippurites heberti* M u n.- C h a l m., *V. inoequicostatus* M ü n s t., *V. cornuvaccinum* B r o n n., *V. archiaci* M u n.-C h a l m., Die zwei Rudistenkalkriffe wurden als Santon und Untercampan bezeichnet.

Gegensätzlich zum nördlichen Apuseni-Gebirge erscheinen in den südlichen Apuseniden keine Rudistenkalke im unteren Maastricht, Zugleich weist die Gesamtenwicklung in den zwei Gebieten wesentliche Unterschiede, indem in den nördlichen Apuseniden die Senon ausschliesslich in „Gosau-Fazies'' entwickelt ist, während im südlichen Apuseni-Gebirge die Gosaufazies im inneren des Sedimentationsbecken zu Flysch übergeht und am oberen Teil, Campan-Maastricht, von Flysch bedekt wird.

Einige Bemerkungen über die Rudistenfauna

Infolge des Studiums der ziemlich reichen Rudistenfauna des Apuseni -Gebirges machen sich folgende Daten bemerkbar :

— Im Coniacien und Unteren Santon besteht die Fauna meistens aus zahlreichen Hippuritidenexemplaren, aber nicht reich in Arten. Zwischen den Rudistenarten ist *Vaccinites oppeli felix* nicht nur im Coniacien bekannt sondern auch im Santon, aber, im allgemeinen, sind die Assoziationen verschieden. Während des Ober-Santons — Unter Campans wird die reiche Anzahl der Exemplare durch eine Verreicherung der Arten ergänzt, hauptsächlich in was die Gattung *Hippurites* betrifft. Eine reichliche Entwicklung weisen auch die Arten der Gattung *Radiolites* und einigermassen auch diejenigen der Capriniden.

Die Am Ende Campan-Untermaastricht erscheinenden Rudisten, obwohl sie nicht in grosser Anzahl sind besitzen die Grösse der in der südmediterranen-Provinz bekannten Exemplare.

In was den stratigraphischen Wert der Rudistenassoziationen betrifft machen sich zwei Elemente bemerkbar :

— Die Rudistenassoziationen des Coniacs, Santons, Campans und Maastrichts sind im allgemeinen verschieden. Einige Arten die auch Unterarten besitzen, erscheinen in etwas grösseren Zeitspannen wie z. B. *Vaccinites oppeli* D o u v., dessen Unterart *V. oppeli felixi* K ü h n sowolh in Coniac wie in Untersanton erscheint, *V. oppeli santoniensis* K ü h n im Santon erscheint und *V. oppeli oppeli* in Campan typisch ist. Arten wie *Vaccinites sulcatus*, *V. gosaviensis* D o u v., *H. colliciatus* W o o d w., *V. cornuvaccinum* B r o n n, erscheinen sowohl in Santon wie im Campan. Dafür erscheinen andere Arten stratigraphisch begrenzt wie z.B. : *V. Giganteus* d'H o m b r e F i r m a s für Coniac, *Vaccinates chaperi* D o u v., für Santon, *V. boehmi* für Campan.

In mehreren Abfolgen wurde der stratigraphische Wert der Rudistenfauna mit anderen Faunen oder mit Mikrofaunen überprüft.

Trotzdem meinen die Autoren, das in der stratigraphischen Analyse der Senonablagerungen, nur die Rudisten-Assoziationen und nicht einzelne Exemplare in Betracht genommen werden können.

Wenn man die Rudistenfauna der Apuseni-Gebirges als allgemeines in Betracht nimmt, sind die folgenden Beobachtungen möglich :

— in was die Rudistenfauna des Coniacs betrifft erscheinen in den Ostalpen nur die Arten *Plagioptychus paradoxus* M a t h. und eventuell die Coniac Unterart des *V. oppeli*, *V. giganteus*, Sonst sind die, im südlichen Apuseni-Gebirge, vertretenden Rudistenarten in der südlichen Provnz bekannt.

Die Santon Rudistenfauna markiert eine grosse Entwicklung der Hippuritiden, Radiolitiden und Capriniden. Im Vergleich mit der bis jetzt bekannten Rudistenfauna der Ostalpen besitzt die Fauna des Apuseni-Gebirges eine reichere Gaztungen und Artenbeschaffenheit. In was die Unter-Campan-Rudisten betrifft, sind die bisher bekannten Arten ungefähr dieselben im Apuseni-Gebirge wie in der Ostalpinen Provinz: hauptsächlich viele gemeinsame Hippuritiden, etwas veniger Radiolitiden.

Aus dem Maastricht der Ostalpen ist bisher nur ein Exemplar von *Jouffia* (Lupu, 1977) aus Wittersdorf, in kärnthen bekannt. Vom selben Ort ist auch ein Exemplar von *Neoradiolites*, der von Dr. P. Beck Mannagetta zur Bearbeitung Übergeben wurde, bestimmt worden.

In manchen Fällen lässt sich eine Migration bemerkbar machen, wie der Fall von *H. colliciatus*, der im Apuseni-Gebirge, in den Ostalpen und in Südfrankreich im Obersanton-Untercampan bekannt ist und dafür im der Türkei (Carakabey, 1959) in Assoziation mit *Jouffia reticulata* Boehm, im Obersenon erscheint.

Ein besonderer Fall ist auch die Heterochronie der Gattung *Pseudopolyconites* Milovanović die in Jugoslavien für ihren stratigraphischen Wert in mittel und Obermaastricht bekannt ist, und welche in den slowakischen Karpaten im Obercampan (Lupu, 1976) und im Apuseni-Gebirge im Obercampan-Untermaastricht erscheint (Lupu, 1976). Dadurch könnte man eine interesante Folgerung erreichen das in diesem Fall die Migration eine in Richtung südwärts war.

Schlussbemerkungen

Im allgemeinen kann die „Gosau"-artige Schichtfolge des Apuseni-Gebirges hauptsächlich im nördlichen Teil als eine typische bezeichnet werden. In der Gesamtentwicklung der Schichtfolge können, in den nördlichen Apuseniden ingressive „Momente" im Obersanton-Untercampan und im Untermaastricht unterschieden werden.

Die Anzahl der Rudistenarten ist im Apuseni-Gebirge grösser als in den Ostalpen infolge eines ausgeprägten mediterranen Einflusses.

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QUESTIONS

A. P o l š a k . 1. Y-a-t-il d'autres faciès du Crétacé supérieur dans les Monts Apusen (par exemple, flysch) ? Peut-on parler d'une relation entre ces faciès ?

2. Quelle est la forme la plus fréquente des formations à Rudistes (biostromes, bioshermes ou barrière-récifs) ?

3. Je voudrais accentuer que le même développement du Crétacé supérieur et la même faune des Rudistes ont été observés dans la région de Bosnie septentrionale et dans la région entre Zagreb et Fruska gora, c'est-à-dire dans une zone entre la plate-forme Adriatique (Dinarides externes) et la plate-forme pannonienne (ou "Tisia" d'après K o v a c s). Une partie de cette zone se prolonge le plus vraisemblablement vers les Monts Apuseni.

Réponse : Oui, le "Gosau" des Monts Apuseni du Sud est recouvert par des dépôts de type flysch d'âge campanien supérieur-maestrichtien. On a aussi des données d'où il résulte que le flysch remplace à l'intérieur du bassin le "Gosau". Dans les Monts Apuseni du Nord, e "Gosau" est recouvert d'une série hémipélagique, parfois flyschoïde.

2. Plus fréquentes sont les biostromes. Le plus typique bioherme est celui de Valea Neagră — Monts Apuseni du Nord.

APERÇU SUR LA FAUNE DES ÉCHINIDES DANS LA RÉGION CARPATHO-BALKANIQUE EN YOUGOSLAVIE [1]

PAR

JOVANKA MITROVIĆ-PETROVIĆ [2]

La région des Carpatho-Balkanides yougoslaves coïncide avec la frontière de la Serbie de l'Est.

La faune des échinides la plus ancienne de la Serbie de l'Est est liée au Jurassique moyen, la plus riche est celle du Crétacé inférieur, un peu moins riche est celle du Crétacé supérieur pendant que dans le Paléogène elle n'existe point et dans les sédiments néogènes (miocènes) elle est aussi assez pauvre.

Toutes les espèces des échinides du Jurassique moyen (jusqu'à présent on n'en a déterminé que 5) proviennent d'une même localité, à savoir la localité de Staro Selo (entre la ville de Knjaževac et la frontière bulgare) et toutes témoignent du Bathonien (fig. 1).

Tous les étages du Crétacé inférieur abondent en faune des échinides ; certaines espèces sont liées strictement aux étages déterminés, d'autres à deux ou à plusieurs étages.

On a déterminé 4 espèces du Valanginien, 11 du Valanginien-Hauterivien ou Néocomien, 11 espèces de l'Hauterivien, 10 espèces du Barrémien, 10 espèces du Barrémien-Aptien, 10 espèces de l'Aptien, 10 espèces de l'Urgonien et une seule espèce de l'Albien.

On n'a pas observé jusqu'à présent des espèces cénomaniennes au cadre du Crétacé supérieur en Serbie de l'Est. Une seule espèce est liée au Turonien. On connaît 7 espèces du Sénonien et dans la plupart des cas il s'agit du Maestrichtien.

La faune complète des échinides du Néogène appartient au Miocène moyen et provient de deux localités : celle des environs de Golubac (8 espèces) et l'autre de Despotovac (3 espèces) (fig. 2).

On connaît jusqu'à présent 81 espèces en Serbie de l'Est, dont 34 appartiennent aux échinides réguliers, y compris les espèces déterminées uniquement sur la base des piquants et 47 irréguliers. Tous les échinides réguliers ont l'âge crétacé inférieur.

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[1] Note présentée au 12ème Congrès de l'Association Géologique Carpatho-Balkanique, 8 – 13 septembre 1981, Bucarest, Roumanie.

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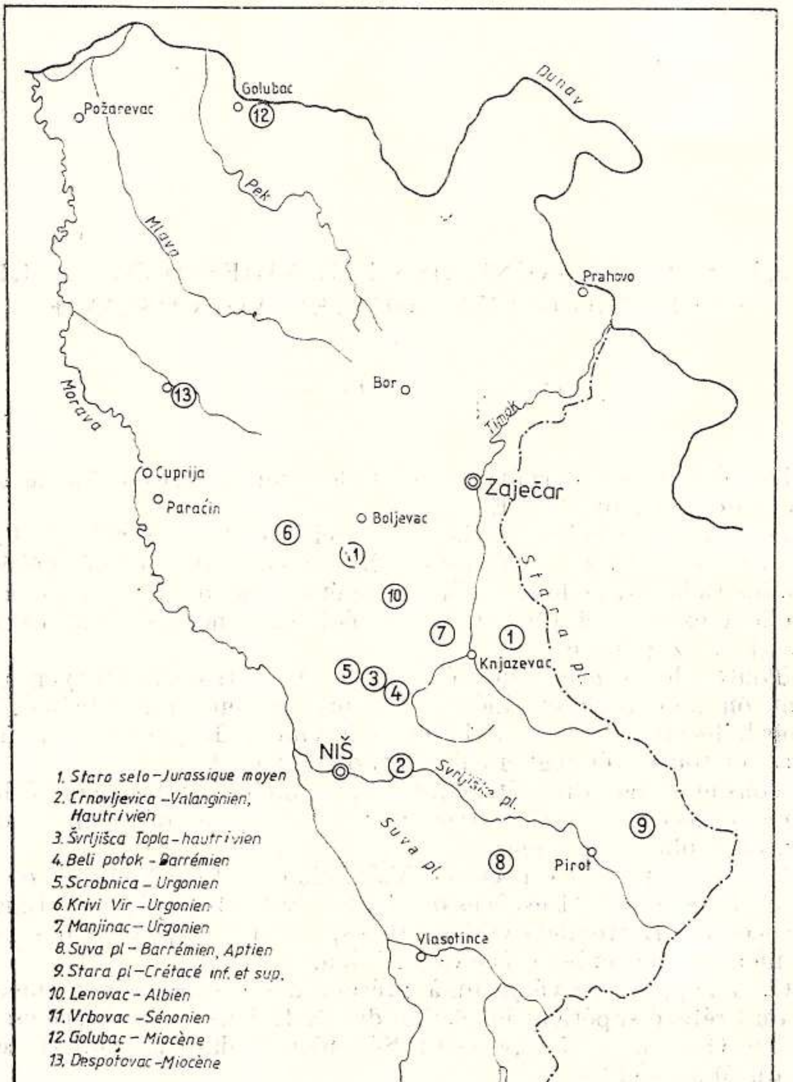

Fig. 1. – Carte de la Serbie orientale avec l'indication des localités fossilifères.

Examinée en totalité, la faune a une grande importance biostratigraphique et paléoécologique. En même temps les échinides réguliers sont au point de vue stratigraphique assez indifférents, d'où leur importance biostratigraphique est sensiblement moindre par rapport aux échinides irréguliers. Pour les interprétations paléoécologiques, pourtant, tous les deux groupes ont une même valeur.

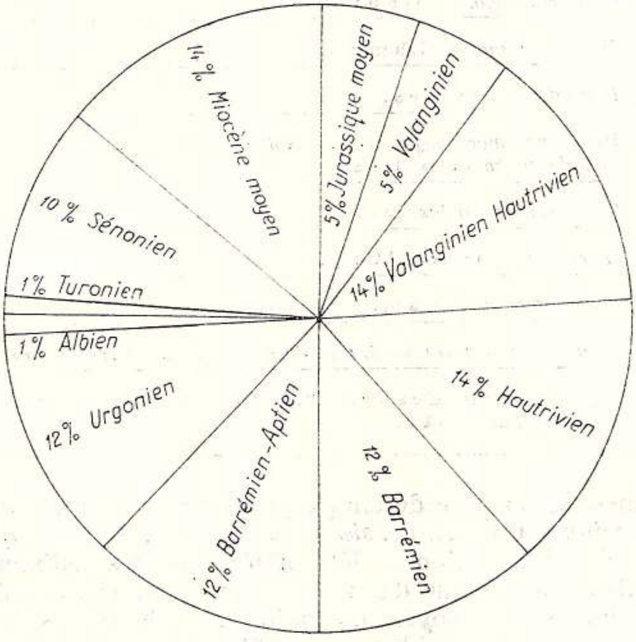

Fig. 2. – Le nombre des espèces dans le cadre de certains étages du Jurassique, du Crétacé et du Miocène en Serbie de l'Est.

Dans mes travaux précédents, j'ai déjà donné les listes des espèces et des associations les plus typiques pour chaque étage du Crétacé inférieur et supérieur et du Miocène (Mitrović-Petrović 1966, 1972, 1976, 1977); cette fois-ci je ne donnerai qu'un aperçu des espèces les plus importantes au point de vue biostratigraphique.

Cette fois-ci je veux indiquer les cas où les échinides ont joué le rôle décisif pour la détermination de l'âge des sédiments dans certaines localités de la Serbie de l'Est.

Le genre *Clypeus* est lié exclusivement aux sédiments jurassiques. Il apparaît en Bajocien, il atteint sa culmination en Bathonien et en Callovien, en Oxfordien il devient très rare (dans chacun de ces étages on ne connaît qu'une espèce du genre *Clypeus)* et il n'y en a aucune trace en Kimméridgien at en Tithonique.

Cette petite expension verticale d'une part et les caractéristiques morphologiques spécifiques de l'autre part, qui rendent possible la détermination générique d'après un seul fragment, lui donnent le statut d'un fossile très caractéristique et important.

Tableau synoptique des espèces les plus importantes au point de vue biostratigraphique

Miocène moyen	*Scutella subrotundata* Lam., *S. vindobonensis* Lbe., *Amphiope bioculata* Desm., *Schizaster desori* Wr., *Brissopsis genei* /Sism./Des.
Sénonien	*Echinocorys ovatus* Lam., *Clypeolampas ovatus* d'Orb. *Guettaria angladoi* Gauth.
Turonien	*Hemiaster sannio* Lamb.
Albien	*Discoidea conica* Des.
Urgonien	*Holectypus macropygus* Des., *Salenia grasi* Cott., *Codechinus rotundus* Des.
Aptien	*Codechinus rotundus* Des.
Barrémien	*Heteraster oblongus* d'Orb.
Hauterivien	*Toxaster retusus* Lam.
Valanginien	*Cidaris pretiosa* Des., *C. pustulosa* Gras., *Magnosia lens* Des.
Jur.moyen *Bathomien*	*Clypeus sinantus* Leske, *Cl. boblayei* Mich., *Cl. mülleri* Wrig., *Cl.davosianus* Cott.

C'est ainsi, sur la base de cinq espèces de ce genre, *Clypeus mülleri* Wr. Cl. *davosianus* Cott., *Cl. sinantus* Leske, *Cl. boblayei* Michž, et *Cl. ploti* Klein, qu'on a déterminé l'âge des sédiments à Staro Selo, (Bathonien), tandis que d'après tous les devanciers ils étaient indiqués comme jurassique moyen ou bajocien-bathonien sans possibilité d'une analyse plus précise (Sikosek, 1952).

Dans le cadre du Crétacé inférieur, en étudiant les échinides de Suva planina on a constaté que les sédiments antérieurement traités comme hauteriviens (Petković, 1930) étaient d'un âge plus jeune, barrémien-aptien, quoiqu'il y ait des espèces qui proviennent de l'Hauterivien. C'est le genre *Toxaster* qui a joué ici le rôle décisif avec plusieurs espèces et qui est découvert dans notre pays pour le première fois il y a quelques années sur les terrains de Suva planina.

Le genre *Echinocorys* a joué un rôle très important au Crétacé supérieur. Il apparaît en Turonien et disparaît en Paléocène. L'espèce la plus connue et la plus répandue de ce genre est *Echinocorys ovata* Leske. Elle indique toujours le Campanien, c'est pourquoi sa présence fréquente en Serbie de l'Est est très importante parce que l'âge des sédiments d'où elle provient, sur la base de l'autre macrofaune, est traité plus largement (comme sénonien) ou bien l'âge est déterminé pour la première fois sur la base de la macrofaune (le village Štrbac dans le synclinal Tupižnica-Knjaževac— Jankičević et Mitrović, 1973).

La faune des échinides du Miocène moyen de la Serbie de l'Est n'a pas une contribution importante pour les recherches biostratigraphiques, parce qu'on a trouvé la faune des mollusques beaucoup plus riche dans les localités d'où elle provient.

Sauf quelques caractéristiques paléoécologiques communes, on a remarqué que les conditions et le mode de vie des échinides sont totalement différents d'une époque géologique à d'autre.

Les espèces jurassiques sont peu nombreuses et liées au fond sableux de la région néritique peu profonde.

On a fait l'analyse des caractéristiques morphologiques de nombreux échinides crétacés et de la faune accompagnante ainsi que l'analyse des sédiments d'où la faune provenait et on a conclu :

— Les échinides crétacés peuplaient surtout l'eau peu profonde (la zone littorale et néritique) tandis qu'ils étaient plus rares dans les eaux plus profondes.

— Dans la région néritique ils peuplaient presque toutes les catégories du fond de la mer : le fond rocheux *(Cidaris, Acropeltis, Codechinus, Psammechinus, Salenia, Codiopsis, Orthopsis, Magnosia* etc.), les récifs *(Salenia, Psammechinus)*, les environs des récifs *(Holectypus Pyrina)*, le fond sableux *(Holectypus, Pygaulus, Discoides)*, le fond argileux *(Epiaster, Toxaster, Hemiaster)*.

Parmi les habitants du fond rocheux et des récifs, certains représentants supportaient bien les battements forts de l'eau *(Cidaris, Codechinus, Psammechinus, Salenia* etc.). L'adaptation à ce mode de vie se reflète soit dans le nombre augmenté des pieds ambulacraires sur la face orale à l'aide desquels ils s'attachaient étroitement aux roches *(Codechinus, Psammechinus)*, soit dans les carapaces basses et hémisphériques qui ne surpassaient pas beaucoup la roche *(Salenia)*, soit dans les piquants très forts comme moyen de défense *(Cidaris)*. Les autres représentants cherchaient l'eau calme pour leur existence et c'est pourquoi ils occupaient souvent les cavités et les crevasses sur les écueils saillants ou sur les récifs *(Codiopsis, Orthopsis, Magnosia)*. Ce qui caractérise ces genres se sont les hautes carapaces qui n'ont pas le nombre augmenté de pores ambulacraires de la face orale.

Les habitants les plus fréquents du fond sableux sont : *Holectypus, Pygaulus, Discoidea* etc. Ce sont les échinides irréguliers qui rampent lentement sur le fond sableux ou ils s'enfouient en partie.

Les genres *Epiaster* et *Hemiaster* sont les habitants des eaux plus profondes et du fond argileux. Leur adaptation à ce mode de vie se reflète sur les petites carapaces fragiles, le labrum bien développé, les ambulacres courts qui sont enfoncés profondément dans leurs carapaces et dans les fascioles bien dévelopées.

— Les échinides crétacés se trouvent le plus souvent en association avec les Polipiers, les Brachiopodes, les Bivalves (spécialement avec les Pachiodontes) et avec les Gastropodes. C'est une association typique pour les eaux peu profondes qu'on rencontre dans presque toutes les localités examinées. Ce n'est que rarement qu'on trouvait les échinides avec les Ammonites et les Belemnites (par exemple dans les sédiments valanginiens-hauteriviens de Stara planina).

— La température de l'eau au cours du Crétacé était assez élevée urtout pendant la formation des récifs au Crétacé inférieur ainsi qu'au Crétacé supérieur.

Les échinides miocènes de la Serbie de l'Est proviennent de deux localités et leurs associations sont tout à fait différentes, par suite de différents types des sédiments d'où ils proviennent.

La faune de Golubac est plus riche (3 genres avec 8 espèces). Le genre *Scutella* prédomine avec 5 espèces. *Amphiope* occupe la deuxième place avec deux espèces, tandis que le genre *Echinolampas* n'est représenté que par une seule espèce. Il est intéressant qu'on n'a constaté aucune espèce du genre *Clypeaster* qui est d'ailleurs très fréquent dans les sédiments du Miocène moyen non seulement dans les autres régions de la Serbie et de l'Yougoslavie, mais il est aussi un des genres prédominants dans le Miocène moyen de l'Europe tout entière. On peut facilement expliquer son absence par le caractère des sédiments. Dans les environs de Golubac (village Vojilovo) la faune des échinides est trouvée dans les conglomérats, tandis qu'on rencontre le *Clypeaster* le plus souvent sur les fonds sableux et sur les récifs.

Pour les genres *Scutella* et *Amphiope* sont caractéristiques les grandes carapaces basses et les faces orales tout à fait plates. Les sillons ambulacraires sur la face orale sont ramifiés ce qui indique un nombre augmenté des pieds ambulacraires. Toutes ces caractéristiques morphologiques indiquent la vie dans l'eau peu profonde et agitée. Les carapaces basses hémisphériques qui ne dépassent qu'à peine la surface des sédiments avaient résisté plus facilement aux battements des vagues que les carapaces hautes. Les surfaces basales plates et solides, le nombre augmenté des pores ambulacraires sur la face orale empêchaient le renversement des échinides sous l'action des vagues.

Sur la base de tout ce qui est dit on peut conclure que les sédiments des environs de Golubac étaient formés dans la zone littorale dont la profondeur ne dépaissait pas une dizaine de mètres, et que l'agitation de l'eau était forte. M o r t e n s e n (1948) et C o t t r e a u (1913) sont d'une même avis. D'après M o r t e n s e n les Scutellidae récents vivent dans la zone de flux et de reflux, de façon qu'elles peuvent restées sur le sec pendant la marée. Elles préfèrent le fond sableux et couvrent ses carapaces des grains de sable. C o t t r e a u cite aussi que les genres *Scutella* et *Amphiope* sont strictement littoraux et peuplent les plages avec les différents élements détritiques : les conglomérats, les sables aux grains petits et aux grains gros. Ici l'eau est très agitée et ne dépasse pas la profondeur de 5 — 6 m.

En communauté avec les échinides on a trouvé une faune des mollusques néritiques très riche.

L'association des échinides du Miocène moyen de Despotovac a un caractère tout à fait différent. On n'a déterminé que deux genres de cette localité : *Schizaster* et *Brissopsis* (le premier avec deux espèces, le second avec une seule). La faune provient des argiles gris-bleuâtres. *Schizaster* et *Brissopsis* appartiennent aux *Spatangoida*. Leurs carapaces sont proportionnellement petites, minces, le labrum bien dévelopé, les ambulacres très raccourcis et enfoncés dans le test, les fascioles nombreuses et bien développées.

Tout cela indique la vie sur le fond argileux où ils s'enfoncaient partiellement ou totalement et témoignent qu'ils sont limnivores (labrum bien développé). L'eau était relativement calme (la présence des fascioles,

les endroits où s'attachaient les clavules dont le rôle étaient de nettoyer le test des différents grains et impuretés.) Chez les habitants des eaux agitées les clavules ne sont pas nécessaires. Cette région était certainement plus profonde que la région de Golubac, mais pas bathyale, étant donné que dans les argiles avec des échinides on trouve les traces du charbon.

La faune complète des échinides du Miocène de la Serbie de l'Est a un caractère tropique-subtropique.

Sur les terrains de la Serbie de l'Est il n'y avait pas des possibilités pour les recherches taphonomiques.

Conclusions. La faune des échinides des Carpatho-Balkanides yougoslaves est très riche. On a déterminé jusqu'à présent 81 espèces (34 échinides réguliers et 47 échinides irréguliers).

La faune provient des sédiments jurassiques, crétacés et miocènes.

Son importance biostratigraphique est grande et grâce à cela l'âge des sédiments dans certaines localités est déterminé pour la première fois ou bien on a fait la révision des opinions précédantes concernant l'âge.

L'analyse morphofonctionnelle de la faune des échinides et de la faune accompagnante, ainsi que l'étude des types des sédiments d'où la faune provient ont rendu possible par la reconstruction de l'ancien byotope une reconstruction réussie des conditions et du mode de vie des échinides au cours du Jurassique, du Crétacé et du Miocène.

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DINOSAURIENS ORNITHOPODES DANS LES BAUXITES NÉOCOMIENNES DE L'UNITÉ DE BIHOR (MONTS APUSENI)[1]

PAR

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L'une des découvertes les plus remarquables faites dernièrement en territoire carpathique est l'identification d'ossements de dinosaures dans une bauxite néocomienne située au coeur même de l'édifice des Carpathes (fig. 1). Sauf les Monts Apuseni, le seul endroit des chaînes alpines péritéthysiennes où l'on ait encore signalé la présence de dinosaures dans une formation néocomienne se trouve au sud du Grand Caucase, sur le territoire du Bloc Géorgien. Mais dans ce dernier cas il s'agit seule-

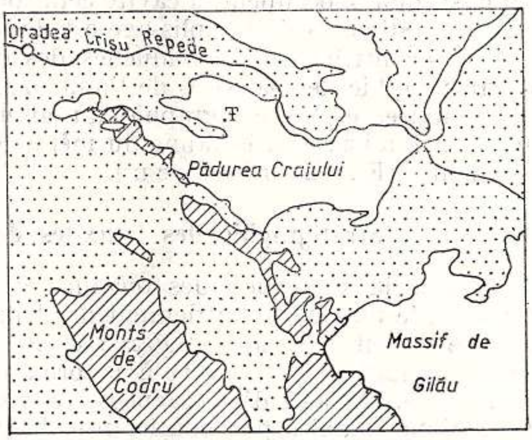

Fig. 1. – Emplacement du gisement à dinosaures de Brusturi (Monts Pădurea Craiului) ; en blanc- unité de Bihor ; hachures-système de nappes de Codru ; pointillé-couverture et formations intrusives post-nappe.

ment de traces de pas tridactyles *(Sathapliosaurus)*, imprimées sur un banc de dolomie vacuolaire (faciès lagunaire). Le gisement des Monts Apuseni se trouve localisé dans la partie centrale du plateau karstique de Pădurea Craiului, à l'ouest de Valea Mnierei (Fig. 1). On y accède par

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le système de galeries de la Mine Brusturi I. La bauxite à dinosaures se trouve logée dans une cavité profonde des calcaires récifoïdes tithoniques, qui y constitue le substratum d'une bauxite diasporique massive (lentille 204). Un premier inventaire de la faune a été dressé par Jurcsac et Popa (1978).

La coupe de la double lentille 204 comporte de bas en haut : 1, calcaire blanc massif du Tithonique (calcaire de Cornet) ; 2, calcite largement cristallisée (spathite) avec, au sommet, une zone décimétrique noirâtre, légèrement bitumineuse ; 3, bauxite boehmitique rouge-brun constituée par une alternance de couches arénitiques ou arénitiques — ruditiques et de couches pélitomorphes ; 4, couche centimétrique à submétrique d'argile kaolineuse, jaunâtre ou rougeâtre, localement avec un peu de boehmite ; 5, spathite ; 6, calcaire de Cornet ; 7, bauxite diasporique rouge-brun, arénitique, massive ; 8, calcaire micritique noirâtre à characées et ostracodes (lacustre) ; 9, calcaire gris à gastéropodes saumâtres ; 10, calcaire gris à pachyodontes (Barrémien).

Les calcaires du Tithonique y sont traversés par deux systèmes de fissures subverticales (35 — 55° et respectivement 130 — 150°) générées au cours des mouvements qui ont conduit finalement à l'émersion de la plate-forme carbonatée jurassique.

La coupe décrite plus haut met en évidence deux étages bien individualisés du karst néocomien. La cavité de l'étage inférieur est de forme discoïdale ; son diamètre atteint 45 m, la hauteur en est de 7 m au moins. Tant le plancher que la voûte sont tapissés par une couche épaisse de spathite. La bauxite boehmitique à ossements de dinosaure et l'argile qui la surmonte occupent la cavité centrale de cette énorme géode. L'épaisseur des sédiments de remplissage est de 3 m au moins.

Les contributions individuelles des auteurs de cet article concernent : la stratigraphie des bauxites de l'unité de Bihor (Patrulius), les traits sédimentaires et la pétrographie du gisement à dinosaures de la lentille 204 (A. Baltreș), la faune du même gisement et la mise en place des ossements (F. Marinescu).

Stratigraphie des bauxites de l'Unité de Bihor

Dans la succession des terrains mésozoïques de Pădurea Craiului, le niveau le plus bas où affleurent des bauxites se situe à la limite entre le calcaire massif récifoïde de Farcu et le calcaire oncolithique d'Albioara (faciès lagunaire) c'est-à-dire à la proximité de la limite Kimméridgien-Tithonique. Il s'agit d'occurrences très rares, centimétriques à décimétriques, d'une bauxite peu consistante, blanchâtre ou rose, caractérisée par une forte teneur en aluminium ($\pm$ 70% Al_2O_3). Toutefois on ne saurait préciser pour le moment s'il s'agit d'un dépôt précédant la sédimentation du calcaire d'Albioara, ou bien d'une infiltration post-jurassique de matériel alitique le long d'un contact de discontinuité lithologique, donc susceptible d'avoir été soumis à l'érosion karstique. Dans le territoire où le calcaire d'Albioara est substitué par le calcaire de Cornet ce dernier comporte à sa partie sommitale, sur quelques dizaines de mètres d'épaisseur, un système complexe de cavités karstiques, dont on distingue au moins trois générations, chacune étant caractérisée non seulement par une morphologie particulière, mais aussi par la nature de ses dépôts de remplissage.

Par sa faune, qui présente des affinités étroites avec celle de Stramberg, le calcaire de Cornet est daté comme tithonique. Pas loin de la lentille 204, à Cornet (au nord), ou à Secătura et Osoiu (à l'est de Valea Mnierei), le mur même des gisements de bauxite diasporique a livré *Phaneroptyxis staszycii* (Zeusch.) et d'autres nérinéidés, *Plegiocidaris cervicalis* (Agas), *P. blumenbachi* (Münst.), *Rhabdocidaris copeoides* (Agas.), *Sphaerotiaris quenstedti* (Merian), *Balanocrinus subteres* (Münst.), *Isocrinus astralis* (Quenst.), des espèces de *Millericrinus, Burdigalocrinus, Thiollericrinus,* des coraux très abondants, des bryozoaires (*Neopora),* de rares ammonites dont *Neoglochiceras pseudocarachteris* (Favre), espèce connue du Tithonique moyen. Ailleurs (Dealu Ana) les mêmes calcaires contiennent à leur sommet des *Ellipsactinia.* Le calcaire de Cornet est également représenté dans l'unité de Vălani (système des nappes de Codru), mais là au faciès graveleux et bioclastique, avec *Clypeina jurassica* Favre localement en abondance, se trouvent associés des calcaires pelsparitiques à *Calpionella alpina* Lorenz (Patrulius, 1971). Il en résulte que la sédimentation des bauxites logées dans le calcaire de Cornet a eu lieu après le Berriasien inférieur (Zone Euxina). En ce qui concerne les calcaires qui constituent le toit des bauxites diasporiques, leur âge reste conjectural. Un âge Barrémien inférieur pour la base „du calcaire à pachyodontes inférieur'' est indiqué par la présence de biostromes à *Requienia minor* Douv. À Cornet le premier niveau à *Requienia* se trouve situé à environ 10 m au dessus du contact avec les bauxites. Il n'est donc pas exclu que „le calcaire à characées'' et „le calcaire à gastéropodes'', qui reposent directement sur les bauxites, appartiennent au Néocomien.

Après l'émersion intra-ou post-berriasienne, qui a affecté la plateforme carbonatée de Bihor (y compris l'unité de Vălani), une première phase d'érosion karstique a généré des cavités superficielles largement béantes et à contour plus ou moins lobé, à parois souvent verticaux, parfois même surplombantes (fig. 2, A). Ces cavités ont éte comparées par Pop et Mîrza (1977) à celles du karst phréatique de Cuba, dénommées "caguanes''. Leur remplissage est constitué par de la bauxite massive arénitique, diasporique et hématitique, rarement avec un peu de boehmite à la partie sommitale (Papiu et al., 1970).

La deuxième génération du karst néocomien comporte de rares cavités profondes, discoïdales ou en forme d'entonnoir renversé, comparables dans une certaine mesure aux cavités de type Aston du karst phréatique (fig. 2, B). Leur remplissage est constitué de spathite (origine vadeuse), de bauxite litée boehmitique et hématitique, arénitique ou pélitomorphe (sédiments d'origine lacustre), d'argile kaolineuse blanche ou jaunâtre, souvent avec un peu de boehmite, de marnes à ostracodes de grande taille et de calcaire détritique ocre à débris remaniés d'échinodermes, par endroits aussi à gastéropodes d'eau douce (famille de Cyclophéridés). Les occurrences de roches carbonatées lacustres sont fort rares et minuscules (Dealu Brusturi, Dealu Secătura).

La dernière phase de karstification néocomienne est caractérisée par la formation d'un karst fissuraire, dont le remplissage comporte principalement des argiles kaolineuses blanchâtres ou roses, par endroits aussi boehmitiques et avec de petits granules disséminés de boehmite (fig. 2, C). Les argiles kaolineuses et boehmitiques de la troisième génération de sédiments karstiques se trouvent souvent infiltrées à la périphérie des

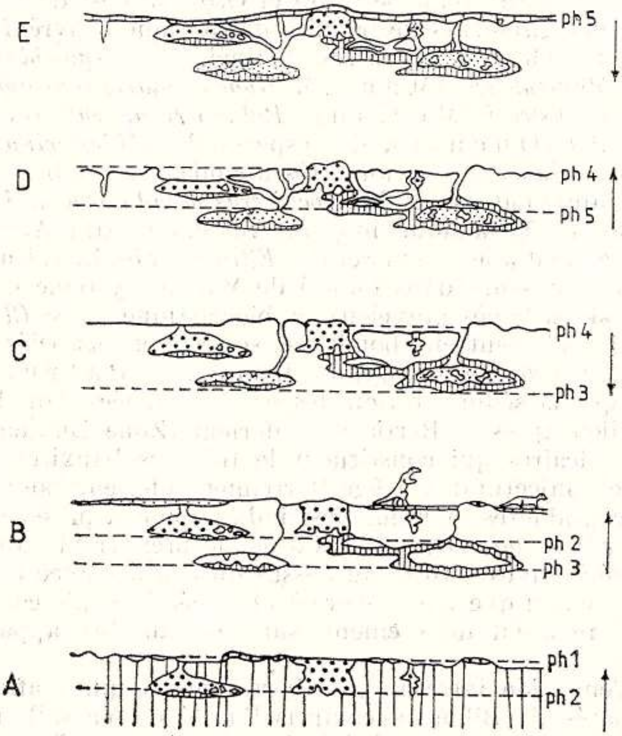

Fig. 2. – Modèle d'évolution du système karstique néocomien sur le territoire de l'unité de Bihor. A) 1ère génération de karst phréatique et 1ère génération de bauxite néocomienne : arénitique, massive, principalement diasporique et hématitique ; la cavité discoïdale à droite comporte des brèches calcaires provenant du toit ; B) 2ème génération de karst phréatique, ensuite vadeux : profond, à cavernes discoïdales tapissées de spathite ; pénétration des dinosaures sur le territoire de l'unité de Bihor, où il y a des lacs intermittents à vase bauxitique ; C) 2ème génération de bauxite néocomienne : arénitique et pélitomorphe, litée, boehmitique et hématitique à dépôts calcaires associés ; les ossements de dinosaures et les vases lacustres sont entrainés dans les cavernes, par les eaux de crues saisonnières ; D) 3ème génération de karst néocomien ; peu évolué, fissuraire ; E) 3ème génération de dépôts karstiques néocomiens : argiles kaolineuses, en partie aussi boehmitiques, infiltrées à la périphérie des corps de bauxite de la première et de la deuxième génération. Les flèches indiquent le mouvement relatif de la plate-forme carbonatée ; les traits interrompus, la position relative du niveau phréatique (ph.).

corps de bauxite massive diasporique (Patrulius in Ianovici et al., 1976). Elles constituent aussi le toit de la bauxite à ossements de dinosaures de la lentille 204, ou le remplissage des fissures qui ont affecté cette bauxite.

Une bauxite plus récente du territoire de l'unité de Bihor se trouve logée dans une crevasse très profonde des calcaires barrémiens à pachyodontes (Patrulius, Iosof, 1974).

Stratonomie, pétrographie et minéralogie [3] du gisement

La bauxite boehmitique de la lentille 204 Brusturi est formée par une alternance de couches arénitiques ou arénitiques-ruditiques granoclassées, épaisses de 4 — 10 cm, et de couches pélitomorphes ayant jusqu'à 30 cm d'épaisseur (Fig. 3). Le classement du matériel arénitique est

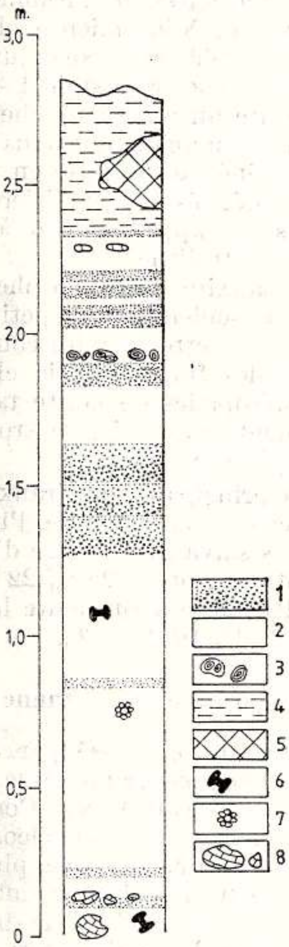

Fig. 3. — Séquence de la bauxite à dinosaures dans la lentille 204 de Brusturi (étage inférieur) : 1, bauxite arénitique-ruditique ; 2, bauxite pélitomorphe ; 3, gros lithoclastes de bauxite ; 4, argiles rouges et jaunes au toit des bauxites ; 5, lentille de spathite ; 6, ossements de dinosaures ; 7, Characés ; 8, lithoclastes de calcaires jurassiques.

soit simple et incomplet, car il n'y a pas de passage au matériel pélitomorphe du toit, soit multiple et récurrent représenté par une alternance de lits détritiques subcentimétriques et de bandes millimétriques pélitomorphes. Les surfaces de litage sont planes. On n'y observe aucune trace d'érosion interstratale. Le matériel arénitique est représenté principalement par des lithoclastes plus ou moins arrondis de bauxite boehmitique riche en fer. Les clastes à structure concentrique doivent leur zonation aux processus de diagenèse (pseudooïdes avec pellicule de boehmite ou de kaolinite enveloppant un noyau de hématite). De rares fragments de calcaires du Jurassique supérieur viennent s'ajouter au matériel bauxitique. Le liant des clastes est un ciment sparitique, ou une matrice de bauxite pélitomorphe. Le matériel ruditique comporte des clastes plus grossiers de bauxite boehmitique, des fragments d'os et une catégorie très spéciale de "galets", dont le diamètre est de 1 à 4 cm. Environ 90 % de ces "galets" sont des calcaires, le reste des roches cristallines (quartz, micaschiste, serniphite, "porphyre quartzifère"). Les éléments calcaires sont de trois sortes : calcarénites du Tithonique à bioclastes d'échinodermes ; calcaires micritiques du Néocomien à characées et ostracodes ; fragments plus ou moins bien roulés de fossiles jurassiques (coraux et bivalves). Le diamètre des "galets" calcaires est de 1 — 2,5 cm. Dans la plupart des cas leur surface présente un relief irrégulier, avec des bosses et des creux témoignant d'une corrosion acide, qui dans le cas des calcaires bioclastiques du Jurassique est pénétrative, mise en évidence par un réseau complexe de canalicules imprégnés de matériel ferrugineux. Selon toute vraisemblance, ces galets qui appartiennent à une catégorie granulométrique restreinte sont des gastrolithes.

La bauxite pélitomorphe présente une texture très uniforme. On y remarque seulement des petits amas diffus de clastes arénitiques riches en hématite. Cette bauxite contient en outre, à la partie inférieure de la séquence, des fragments de characées (principalement tiges) et de très rares gastéropodes de petite taille. Les ossements de dinosaures, souvent parfaitement conservés, se trouvent disséminés principalement dans les couches pélitomorphes.

Les principaux minéraux constitutifs des bauxites à ossements de dinosaures sont la boehmite, l'hématite et la kaolinite avec les proportions respectives suivantes (à titre d'exemple) : 69 %, 16 %, 15 % dans une variété arénitique ; 50 %, 28 %, 22 % dans une variété pélitomorphe. Dans la couche d'argile qui surmonte la bauxite le pourcentage des mêmes minéraux est 16 %, 10 %, 74 %.

Composition de la faune et mode de gisement des ossements

L'inventaire dressé par Jurcsak et Popa (1978) comporte, à coté d'un *Iguanodon* probable, de nombreux restes attribués à *Dryosaurus*, des dinosaures carnivores (Coelurosauridae et Carnosauridae), ainsi que des crocodiliens marins (Teleosauridae !). Cette liste est à reconsidérer de façon radicale. Quoique plusieurs milliers de pièces aient été récoltés, principalement vertèbres et métapodiums, il n'y a aucune dent ou phalange terminale (grifféale) de dinosaure carnivore, aucune plaque dermique ou vertèbre qu'on puisse attribuer avec certitude à un crocodilien. En jugeant d'après les dents, les vertèbres et les nombreux phalanges terminales plates provenant de la lentille 204 seuls des iguanodontidés s'y trouvent représentés. Les restes le mieux conservés indiquent la présence du genre *Camptosaurus*, ou d'une forme étroitement apparentée, à dents pourvues de crénelures fines. Le fait que les vertèbres sacrées qu'on y a trouvées sont moins pincées que ceux d'*Iguanodon* et habituellement non-soudées plaide également pour l'appartenance de la plupart de ces restes à *Camptosaurus*. On ne saurait toutefois exclure la présence d'*Iguanodon* à côté d'un iguanodontidé à caractère plus primitif. Il s'agit donc d'une association sinon monotypique, en tout cas oligotypique. En jugeant d'après les dimensions des métapodiums (phalanges unguéales de 4 — 5cm, métatarsiens ou métacarpiens de 8 — 10 cm), les plus grands spécimens ne devaient pas dépassér 3 m en longueur.

En ce qui concerne le mode de gisement des ossements, il faut spécialement souligner les particularités suivantes : 1, les ossements se trouvent pour la plupart disséminés dans la bauxite pélitomorphe ; ils sont très rares dans les couches arénito-ruditiques ; le grand axe des pièces est habituellement parallèle aux surfaces de litage ; 2, des os en connexion sont d'occurrence exceptionelle (dans un seul cas, deux vertèbres sacrées) ; 3, les dimensions des pièces individuelles ne dépassent pas généralement 10 cm (15 cm au maximum). Dans cette catégorie dimensionnelle rentrent quelques exemplaires fragmentaires d'os longs appartenant à des spécimens juveniles. Ils sont d'espect gracile, mais toutefois ne possèdent pas la structure pneumatique caractéristique des Coelurosauriens. Des os longs appartenant à des spécimens adultes on ne trouve que des épiphyses brisées. Il faut noter que non seulement les os longs entiers de grands spécimens (humérus ou tibia de 40 cm, radius ou humérus de 30 — 35 cm) manquent au tableau, mais aussi les pièces complètes de la ceinture pelvienne (iléon ou ischion de 40 cm de long) ; 4, bon nombre de pièces sont fragmentaires. Il ne s'agit pas seulement d'os de grande taille, mais aussi de corps vertébraux amputés de leur arc neural. Par contraste, à côté des vertèbres brisées, on trouve d'autres qui ont conservé leur arc neural complet, quoique les apophyses en soient fort délicates, surtout celles des vertèbres caudales. Dans le cas de beaucoup de pièces le cortex a été partiellement enlevé de sorte que la matrice bauxitique fait corps commun avec le remplissage du tissu spongieux ; 5, en jugeant d'après le nombre des vertèbres récoltées jusqu'à présent au moins 30 individus ont participé avec leurs pièces squelettiques à l'accumulation d'ossements de la lentille 204.

De ces faits d'observation il résulte en première instance que les ossements du gisement de Brusturi proviennent d'un „cimetière" situé en bordure d'un lac (lieu d'abreuvage probable) dont les sédiments étaient principalement constitués de boues bauxitiques ; que les squelettes étaient complètement dissociés avant que leurs pièces soient mélangées à ces boues et transportées ensemble dans la cavité sousterraine par des courants à faible pouvoir de traction, probablement au cours de crues saisonnières. Reste à expliquer pourquoi beaucoup d'os sont brisés et pourquoi les dimensions des pièces ne dépassent que très rarement 10cm. Une sélection des ossements peut s'expliquer par le faible pouvoir de traction des courants qui les ont entrainés, mais dans ce cas il faut faire appel à un autre mécanisme pour rendre compte du fait que beaucoup de pièces ont été brisées avant la mise en place finale. C'est à l'activité des reptiles nécrophages qu'on vient à penser en premier lieu, non nécessairement à des dinosaures carnosauriens, mais à des bêtes capables de briser les diaphyses à cortex épais des os longs pour en extraire la moelle et si friands de cartilages et de ligaments qu'ils ont finalement réussi à enlever une partie du cortex mince qui enveloppe le tissu spongieux des épiphyses, spécialement vertébrales. À part le fait qu'on n'a pas trouvé jusqu'à présent une telle bête, cette explication s'accorde mal avec le fait que dans le gisement en question il y aussi beaucoup de pièces fragmentaires de grands os plats qui ne semblent pas avoir représenté un aliment enviable.

Mais disons que dans la nature il y en a pour tous les goûts, et que notre nécrophage hypothétique brisait tout ce qui était recouvert de tissu comestible pour rejeter ensuite les débris inutilisables.

Le problème le plus ardu que pose la présence de dinosaures néocomiens sur le territoire de l'unité de Bihor, c'est-à-dire au coeur même de l'édifice des Carpathes, concerne la voie de migration que ces bêtes ont suivi pour y arriver.

À partir de la plate-forme nord-téthysienne les obstacles qui interdisaient l'accès au Néocomien étaient multiples : fosse vocontienne, fosse du flysch rhéno-danubien, fosse piénine, tout le territoire de la plate-forme moesienne. Des obstacles en creux pas moins formidables couvraient la marge sud-téthysienne : le(ou les) fosse(s) du flysch de Jebel Tisirène (Maroc) et de Monte Soro (Sicile), du flysch galestrino (Lucanie), les bassins à sédiments pélagiques de la dorsale calcaire (Maroc), des zones centrales de Sicile (Vicari, Campofiorito), de l'Ombrie et des Alpes Méridionales, des Dinarides-Hellenides (zones ionienne, pindique, du flysch bosniaque ou béotien, zone serbe).

Le seul territoire qui offre quelque perspective en vue d'une solution est situé à l'est, où une voie de migration à envisager, c'est par les pontides, l'ensemble Rhodopes-massif serbomacédonien-nappes gétique et supragétique, la marge orientale et l'extrémité nord, de la zone ophiolitique des Metalliferi, enfin le massif de Gilău.

[3] Les analyses minéralogiques et chimiques ont été effectuées par Florica Popescu et respectivement Alexandra Medeşan.

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THE PRE-ALBIAN CRETACEOUS SEDIMENTARY IN THE FORELAND OF THE ROMANIAN CARPATHIAN MOUNTAINS[1]

BY

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Introduction

During the last decades, the geological units situated outside the Romanian Carpathian massifs have been intensely investigated by drillings, in order to define their oil and gas potential. In the sedimentary cover, the pre-Albian Cretaceous deposits constitute a sequence with remarkable facies variations. They are developed in the Moesian Platform (Fig. 1) and the Pre-Dobrogean Depression with continuity along the western margin of the Moldavian Platform (Fig. 2).

1. The Moesian Platform

In this unit, the Lower Cretaceous is predominantly carbonate, of basinal (pelagic) type in its central-western part and of submerged platformic type at its extremities (Fig. 1).

1.1. The platform basinal domain comprises pelagic deposits up to about 250 m in thickness, made up of micritic limestones, marls and marly limestones, bearing ammonites, tintinnidae, hedbergellae, radiolaria, sponges, benthic foraminifera, etc. By the study of the ammonitic fauna (Muţiu, 1963), all the Lower Cretaceous subdivisions have been identified, from the Berriasian to the Lower Barremian, and they were equally attested by the microfaunal studies (Costea et al., 1978).

Supplimentary data regarding the succession, the fauna as well as the sedimentary conditions of the pre-Albian deposits can be found in the previous papers (Muţiu, 1963; Dragastan et al., 1973; Costea et al., 1978).

[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, 1981 September 8 – 13, Bucharest, Romania.

[2] Ministry of Petroleum, Bucharest, Romania.

1.2. The submerged platform domains cover the western and the central-eastern parts of the Moesian Platform and comprise a wide range of carbonate Lower Cretaceous deposits, locally associated with lagunary and continental deposits. Macrofossil remains certainly proved only the Barremian-Aptian age of the uppermost part of the sequence (M u ţ i u,

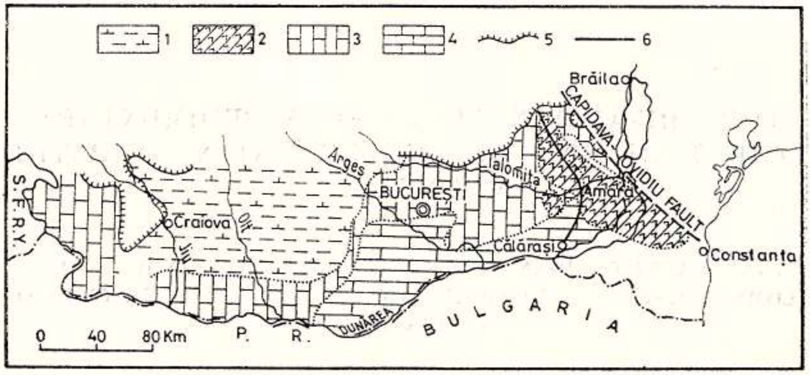

Fig. 1. — Moesian Platform. Spreading of the pre-Albian—Cretaceous deposits. 1, Pelagic ammonitic facies (Berriasian-Lower Barremian). Platformic facies : 2, Middle Berriasian (gypsiferous complex) developed only at the eastern end of the platform (rests on various Kimmeridgian-Tithon levels); 3, Upper Berriasian-Hauterivian (rests on Tithon and gypsiferous complex); 4, Barremian-Aptian (rests on Hauterivian and different levels of the Valanginian); 5, border of the sedimentary area ; 6, line of correlation in Plate — A.

aboratory diagnosis), but in the micropaleontological content several characteristic assemblages were separated, considered as defining al the Lower Cretaceous subdivisions (V i n o g r a d o v , D r a g a s t a n, 1975; C o s t e a et al., 1978). With some minor modifications, these assemblages are the following:

— the Favreina salevensis assemblage: Berriasian-Lower Valanginian;

— the Trocholina assemblage: Upper Valanginian;

— the Spirillina assemblage: Hauterivian;

— the Palorbitolina and Choffatella assemblage: Barremian-Aptian.

The pile of Lower Cretaceous deposits crossed by deep drillings, roughly corresponds to the schemes drawn out for the synchronous deposits, either outcropping or crossed by shallow drillings in South Dobrogea (B ă n c i l ă , 1973; C h i r i a c et al., 1977; N e a g u et al., 1977; D r a g a s t a n, 1978), some difference being related only in the manner of considering certain boundaries between the Lower Cretaceous subdivisions and between Jurassic and Cretaceous.

— The Favreina salevensis assemblage is included in a sequence of deposits from the central-eastern part of the Moesian Platform, which can be separated, lithologically and paleontologically, into three complexes:

— The lower complex (or the gypsiferous complex) is only developed in the eastern extremity of the Moesian Platform, north and south of the Ialomiţa River mouth, from where it extends beyond the Danube in the Cernavodă-Constanţa area, south of the Capidava-Ovidiu Fault (Fig. 1, Pl.—A). It may locally reach about 300 m in thickness (Amara) and consists of anhydrites and gypsum, with more or less thick interbeddings of limestones and marly limestones. North of the Ialomiţa River, *Favreina salevensis* (Paréjas) associated with Ostracodae was identified in some of these limestones (Patrulius 1964, Costea et al., 1978) and in the Amara area, a micritic limestone (about 35 m thick) with ostracodae has been noted at the base of the complex.

In South Dobrogea, the gypsiferous complex rests upon the Kimmeridgian and, for this reason, it has been regarded as Purbeckian (Băncilă, 1973, Chiriac et al., 1977, Neagu et al., 1977). In the Ialomiţa Valley area, it overlies, however, various Malm levels, from Lower Kimmeridgian to Upper Tithonian inclusively (Pl.—A) and contains, as already mentioned, *Favreina salevensis*; it is estimated that it should rather be regarded as a Lower Cretaceous term.

— The median complex (the limestones with *Anchispirocyclina lusitanica)* overlies the gypsiferous complex in the Ialomiţa Valley-Cernavodă area and the Upper Tithonian (with *Clypeina jurassica* and/or *Macroporella praturloni)* in the rest of the central-eastern platform. The complex, 70 — 200 m thick (increasing from north to south), consists of micritic limestones interbedded with pelletal, oosparitic and intraclastic limestones. The microfauna comprises numberless forams, associated with ostracods, decapods, anomurs and algas (Costea et al., 1978). *Favreina salevensis* (Paréjas) is quite widespread and explosively developed on the northern platform margin, where it makes up coprolitic limestones. In the micritic and pelletal limestones, next to *Favreina*, *Macroporella embergeri* (Bouroullec & Deloffre) is rather frequent and specific.

Out of foraminifera, peculiar mention should be made of *Anchispirocyclina lusitanica* (Egger), *Feurtillia frequens* Maync, *Everticyclammina virguliana* (Koechlin), *Rectocyclammina chouberty* (Hottinger), *Pseudocyclammina lituus* (Yokohama), *Torinosuella peneropliformis* (Yabe & Hanzawa), etc. In the Moesian Platform, these forms have never been encountered in the Jurassic; the deposits including them are considered to mark a post-Jurassic marine sedimentary cycle.

— The upper complex (the multicoloured or the variegated complex) overlies the median complex; it is 50 — 70 m thick and consists of micritic limestones interbedded with intrasparitic limestones, marly limestones and grey, greenish, chocolate-brown or red marls. Limestones are more frequent on the northern margin of the Platform, where they are usually ferruginous and from west to east they are progressively substituted by mottled clays.

Lituolidae and miliolidae, associated with ostracodae, favreinae and algae *(Macroporella embergeri)* have been commonly identified in limestones, while Ostracods and Characeans have been prevailingly encountered in clays.

The regressive tendencies are evident and it is appreciated that the complex closes the sedimentary cycle which starts with the *Anchispirocyclina lusitanica* limestones.

The multicoloured complex is covered by marine deposits containing a very rich microfauna, with an explosive development of Trocholinids: *Trocholina elongata* (Leupold), *T. alpina* (Leupold), *T. valdensis* (Reichel). This sequence marks a new sedimentary cycle and was differently dated — Upper Valanginian in the boreholes (Costea et al., 1978) and Upper Berriasian-Valanginian or Middle-Upper Valanginian in the outcrops (Chiriac et al., 1977; Neagu et al., 1977, respectively Dragastan, 1978) — but most of the data suggest an Upper Valanginian age and in this context, the underlying deposits have been considered as Berriasian-Lower Valanginian.

The chronostratigraphic value of the microfossils in these deposits is still questionable (it is still debated within the entire Mediterranean Basin — the papers of the Colloque of Lyon, 1973, on the Jurassic-Cretaceous boundary) and a precise boundary between the Valanginian and the Berriasian cannot be established yet. Until further knowledge, it has been provisorily situated between the *Anchispirocyclina lusitanica* limestones and the multicoloured complex.

As for the Jurassic-Lower Cretaceous boundary, it is obviously marked by the extinction of *Macroporella praturloni* and/or *Clypeina jurassica* and by the occurrence of *Favreina salevensis* (at the gypsiferous complex level) or of the *Anchispirocyclina lusitanica* and/or *Macroporella embergeri* (at the limestone level). On the other hand, between the Tithonian and the Lower Cretaceous, a sedimentary gap is obvious, defined by the gypsiferous complex deposition overlying various Kimmeridgian-Tithonian levels in the easternmost part of the platform (Pl.—A) and by the deposition of the *Anchispirocyclina lusitanica* limestones over the Upper Tithonian (without any gypsiferous complex in-between) in the rest of the eastern platform. Most likely, this gap includes, at least, a part of the Lower Berriasian and, in this frame, the Anchispirocyclina lusitanica complex could be tentatively regarded as Upper Berriasian and the gypsiferous complex as Middle-uppermost Lower Berriasian.

The limestones with trocholinae of Upper Valanginian age are overlain by a carbonate sequence comprising the Hauterivian (the Spirillina assemblage) and the Barremian-Aptian (the Palorbitolina and Choffatella assemblage). A detailed account on all these deposits has been given in some previous works (Chiriac et al., 1977; Neagu et al., 1977; Costea et al., 1978; Dragastan, 1978) and is not necessary to insist upon them.

2. The Pre-Dobrogean Depression and the Moldavian Platform

In these two units (Fig. 2, Pl. I — B), the pre-Albian Cretaceous transgressively overlies the Tithonian (the Pre-Dobrogean Depression) and the Paleozoic or the Riphean (the western margin of the Moldavian Platform), being comparable to that in the Moesian Platform.

The most complete succession has been identified in the Bîrlad Depression (the north-western end of the Pre-Dobrogean Depression); it is 400—700 m thick and comprises several complexes, clearly distinct lithologically and paleontologically.

— The limestones complex with *Anchispirocyclina lusitanica* (200 — 400 m thick) overlies the Upper Tithonian (with *Macroporella praturloni*) and according to the microfauna (foraminifera, corals and dasycladaceae or codiaceae algae) it is obviously the equivalent of the Anchispirocyclina lusitanica complex in the Moesian Platform. The gypsiferous complex is lacking in the Bîrlad Depression.

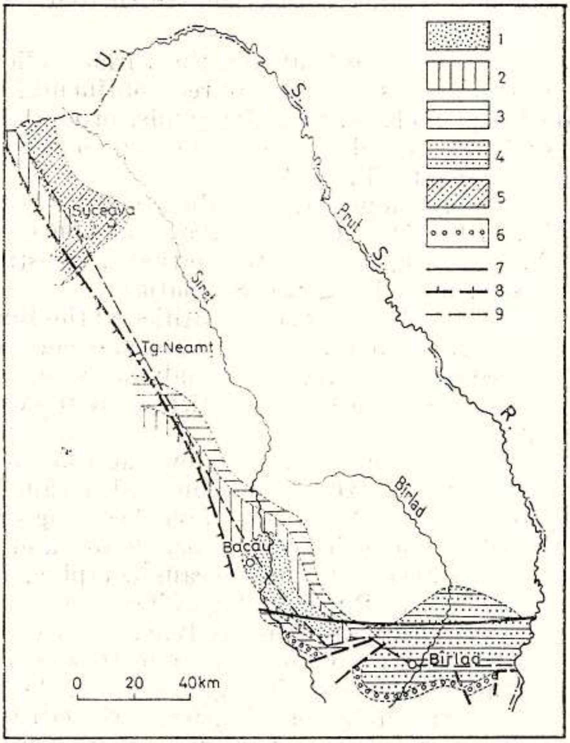

Fig. 2. — Bîrlad Depression and Moldavian Platform. The spreading of Lower Cretaceous deposits (according to the drilling data).
1, Palorbitolina and Choffatella complex (Barremian-Aptian); 2, Trocholina complex (Upper Valanginian); 3, oolitic sandstones complex (late Lower Valanginian); 4, red continental complex (late Lower Valanginian); 5, ostracods and chara complex (early Lower Valanginian); 6, Anchispirocyclina lusitanica and ostracods chara complexes (Upper Berriasian-early Lower Valanginian) in the Bîrlad Depression; 7, their northern border below the late Lower Valanginian; 8, front of the pre-Carpathian overthrust (at the surface); 9, line of the correlation presented in Plate — B.

The limestones have been encountered only between the Prut and the Siret valleys but one may suppose that they are also present on the deep western margin of the Moldavian Platform, still uninvestigated by drillings.

— The complex with ostracodae and characeae, which overlies the *Anchispirocyclina* limestones, consists of red-brown or grey-greenish clays at the base (about 50 m thick), followed by anhydrites (the Glăvănești Anhydrite — about 50—70 m thick) with thin intercalations of grey clays and dolomites. The ostracods and the characeans (to which favreinids are added at certain levels) are similar to those in the variegated complex in the Moesian Platform. Regression was stronger here, even favouring anhydrite formation.

In a partly different facies, the complex was also encounterred on the north-western end of the Moldavian Platform (Suceava area — Fig. 2, Pl. I—B), in the direct southern prolongation of the Lwow Depression from the USSR. It reaches here 350 m in thickness and rests upon the Paleozoic. It consists of brick-red sandstones, brown clays and ferruginous marls with more or less thick intercalations of dolomicrites, dolomicrosparites, micritic limestones and anhydrites. There were identified the same ostracods and characeans as in the Bîrlad Depression and a connection could be presumed through the western, sunken extremity of the Moldavian Platform, still uninvestigated by drillings.

Between the Prut and the Bîrlad valleys, the ostracods and characean complex is overlain by red continental deposits (up to 300 m thick), consisting of clays and gritty sands, often thinly interbedded with gypsum, anhydrites and dolomites. The sequence is transgressively overlain by the Badenian (Pl. —B).

Similar deposits, exceeding 500 m in thickness, have also been identified in the Danube Delta area (the south-eastern extremity of the Pre-Dobrogean Depression), where they are resting upon the Upper or Middle Jurassic, underlying the Sarmatian.

No fossils have been identified in the Bîrlad Valley, but in the Danube Delta region a palynoflora has been encountered, which points to a Lower Cretaceous age (B a l t e ș — laboratory diagnoses) and, as a whole, this sequence is estimated to constitute the top of the ostracods and characeans complex.

West of the Bîrlad Valley, the Glăvănești Anhydrite is overlain by grey-violet quartzitic sandstones with thin beds of leptochloritic or calcareous oolites. No fossils have been signalled, but the sandstones are covered by deposits with trocholinids and can be regarded as the top of the ostracods and characeans complex, in a different facies from the red one in the Bîrlad Valley. These two facies could be separated by a threshold (still less known) lying west of the Bîrlad Valley (Pl.—B).

In the Siret Valley, the sandstones exceed 300 m in thickness, overpass the anhydrites, directly overlying the Tithonian, and extend northwards on the Moldavian Platform western border, up to the Tîrgu-Neamț locality. Here they rest upon the Riphean or Paleozoic and display tidal features with subcontinental episodic. The anhydritic cement of some sandstones suggests a sedimentation of the Sabkha type.

— The complex with trocholinids and miliolids is only known in the Bacău town and in the Suceava town areas (Fig. 2), displaying a sandy intraclastic calcareous facies in the first area, and an intrapelsparitic calcareous facies in the second one (sited in the southern prolongation of the Lwow Depression in the USSR). The deposits, over 100 m in thickness, include a rich assemblage of foraminiferal, algal, and equinidal remains, in which the explosion of trocholinids allows a certain correlation with the trocholinids complex in the Moesian Platform.

— The Palorbitolina and Choffatella complex follows after a sedimentary gap corresponding to the Hauterivian and is fairly developed south of the Bacău town. Here it can reach a thickness of about 150 m and consists of quartzitic and lithic sandstones, in places with interbedded conglomerates and oolitic limestones. A rich content of forams, briozoans, corals, ostracods, characeans has been identified, out of which *Palorbitolina conoidea* (G r a s), *P. lenticularis* (B l u m e n b a c h) and *Choffatella decipiens* (S c h l u m b e r g e r) are very common, as in the Moesian Platform, indicating a Barremian-Aptian age.

In a single well only, a rest of the complex which escaped to erosion was found in the Suceava area. It is about 30 m thick, the fossil content is the same, but the lithofacies is quite different (calcitic sandstones, bioclasts and gritty glauconitic marls) suggesting certain relationships with the Lwow Depression in the USSR.

Inferences

The pre-Albian Cretaceous sedimentary sequence crossed by drillings in the Foreland of the Romanian Carpathians is predominantly developed in a facies of platform type, excepting the central-western part of the Moesian Platform, where it exhibits a pelagic ammonitic facies.

The study of the macro- and especially of the microfauna has allowed the identification of all the Lower Cretaceous subdivisions, from Berriasian to Aptian, but in detail, the boundaries between these subdivisions can be still questionable.

In the platform facies, the presence of some lagunary and/or continental deposits has allowed to separate sedimentary cycles, whose depositional manner shows that the sedimentation has not begun always with the same deposits and that it was not throughout continuous.

The oldest Cretaceous deposits (probably late Lower Berriasian) are lagunary and marine, rest upon different terms of the Kimmeridgian and Tithonian and are only developed north and south of the Ialomița River mouth with extension into South Dobrogea.

Furthermore, the sedimentation area extended over the entire Moesian Platform and into the Bîrlad Depression, with carbonate marine deposits at the beginning (Upper Berriasian) and then continental and/or lagunary ones (Lower Valanginian). At this last level, the sedimentation area also comprised the north-western margin of the Moldavian Platform.

From the Upper Valanginian, the sedimentation has continued up to the Aptian with some stratigraphical gaps, regional (Hauterivian in the Bîrlad Depression and the Moldavian Platform) or local (Moesian Platform).

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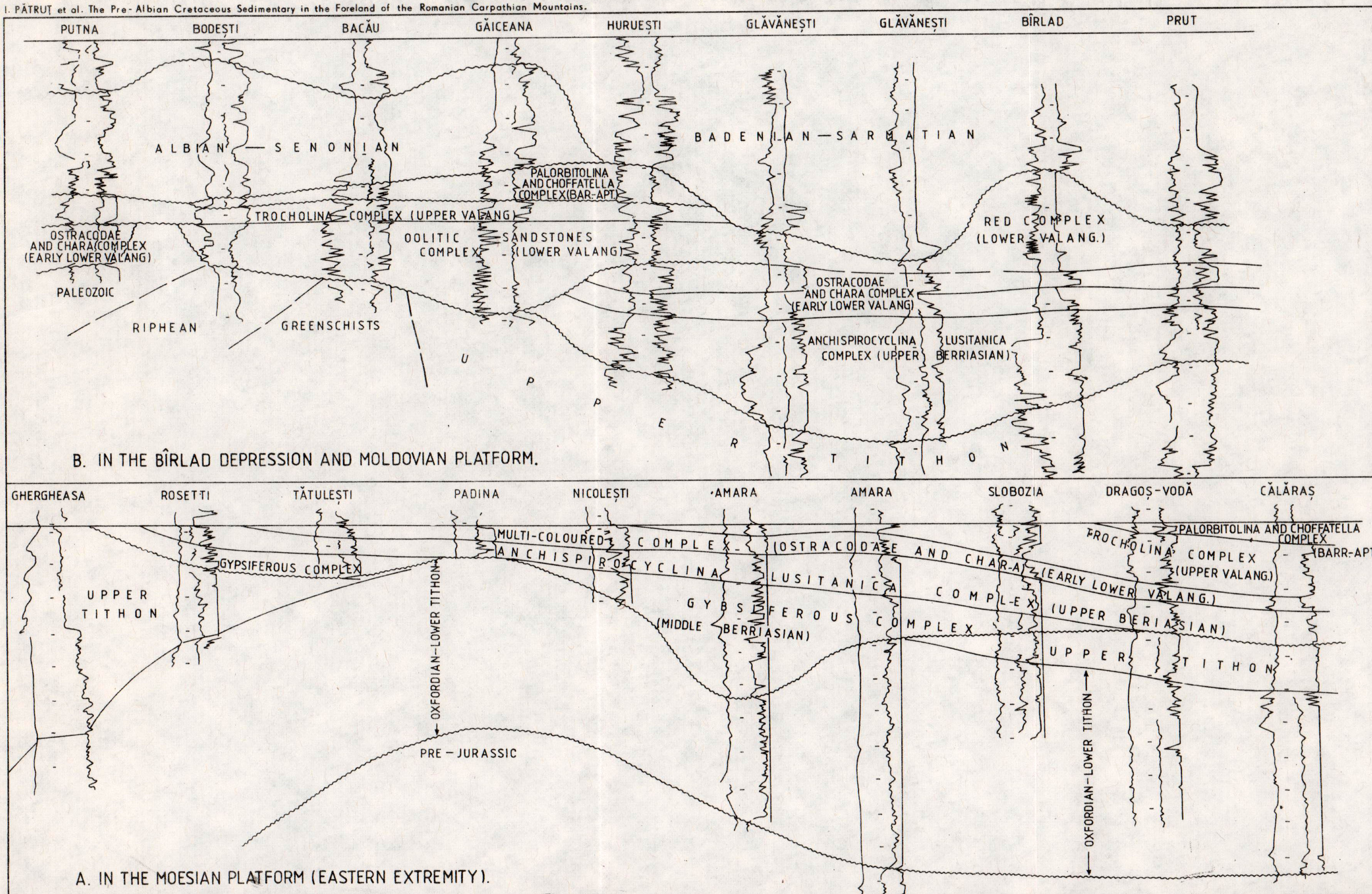
CORRELATION OF THE LOWER CRETACEOUS DEPOSITS



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LES FORMATIONS ROUGES SITUÉES À LA LIMITE JURASSIQUE-CRÉTACÉ ET DE L'APTIEN DES UNITÉS D'AVANT-PAYS DES CARPATHES ROUMAINES [1]

PAR

CONSTANTIN VINOGRADOV [2]

Les formations rouges de la limite Jurassique-Crétacé et de l'Aptien ont été détectées en majorité grâce aux forages, à savoir dans le nord-ouest de la Plate-forme moldave, dans les Dépressions de Bîrlad et pré-dobrogéenne, dans le nord-est de la Plate-forme moesienne et dans la Dobrogea méridionale (fig. 1, 2).

La genèse continentale-lacustre ou continentale-lagunaire des formations rouges pose des problèmes en ce qui concerne la datation chronostratigraphique exacte. Afin de les résoudre convenablement, les séquences marines fossilifères qui entourent ou qui s'intercalent dans les formations rouges y ont une contribution essentielle.

Les formations rouges situées à la limite Jurassique-Crétacé

Dans les forages peu profonds (150 — 200 m) effectués dans la Dobrogea méridionale (Cernavodă, Saligny, Nazarcea) on a rencontré, au-dessous des calcaires d'âge berriasien supérieur, la succession suivante : marnes et argiles polycolores, calcaires micritiques, gypses, anhydrites et dolomicrites lagunaires. Seules les marnes et les argiles polycolores contiennent une association de Charophytes et d'Ostracodes : *Clavator pecki* Mädler, *Nodosoclavator nodulosus* (Peck.), *Cypridea dololabrata* Martin, *Cypridea alata* Wall, *Iliocypris* cf. *jurassica* Martin (Neagu et al., 1977). Dans les calcaires d'âge berriasien supérieur qui recouvrent les marnes et les argiles polycolores, les mêmes auteurs ont déterminé : *Anchispirocyclina maynci* (Hott.), *Trocholina valdensis* (Reichel), *Salpingoporella annulata* Carozzi, *S. steinhauseri* Conrad, Prat. & Radoič., *Likanella bartheli* Bernier, Pachiodontes, Éponges, Gastropodes. Tenant compte du contexte régional, au-dessous du complexe laguno-continental il y a des calcaires cor-algals

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[1] Note présentée au 12 ème Congrès de l'Association Géologique Carpatho-Balkanique, 8 — 13 septembre 1981, Bucarest, Roumanie.

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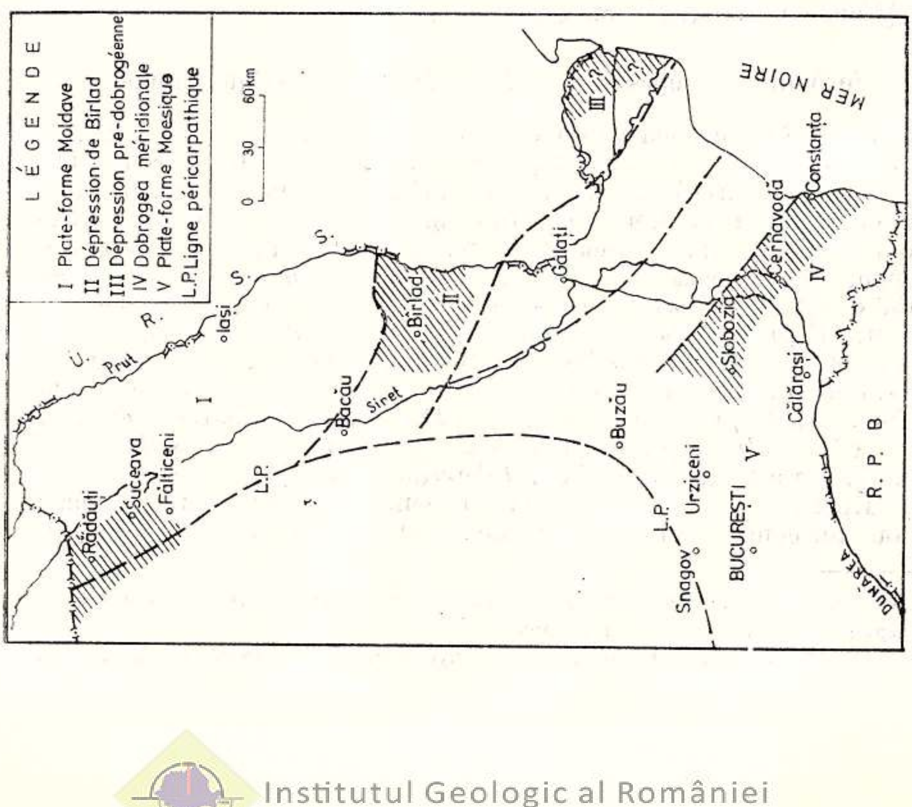

Fig. 1. — Distribution des formations rouges purbecko-wealdiennes sur l'avant-pays des Carpathes roumaines.

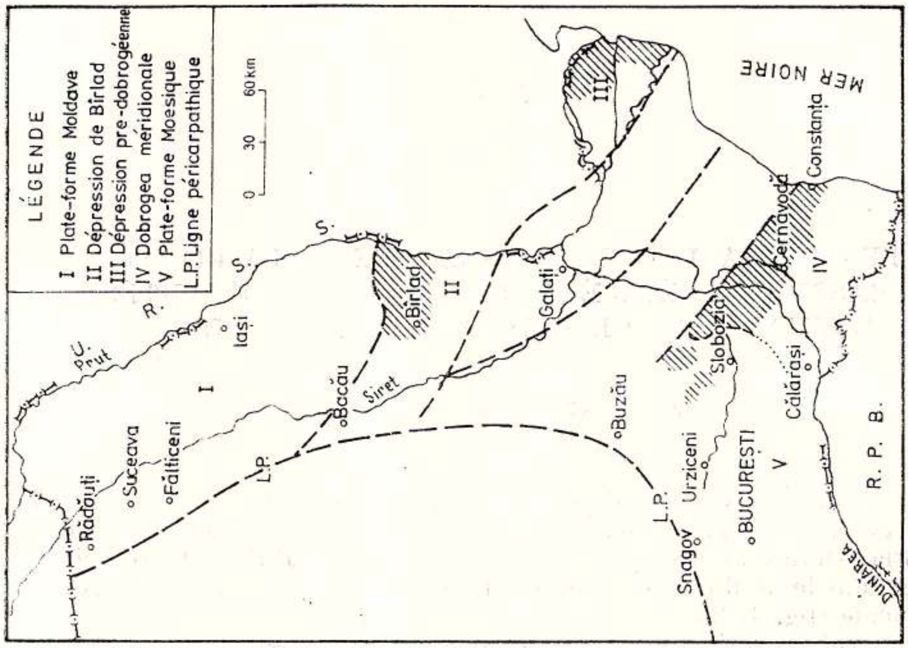

Fig. 2. — Distribution des formations rouges aptiennes sur l'avant-pays des Carpathes roumaines.

et des dolomies kimméridgiennes (C h i r i a c et al., 1977). B ă n c i l ă (1973) a attribué à la formation rouge terrigène-évaporitique l'âge purbecko-wealdien ; ultérieurement, N e a g u et al. (1977), la considèrent exclusivement purbeckienne, correspondant à l'intervalle Tithonique-Berriasien moyen.

Poursuivie vers l'ouest (par les forages du nord-est de la Plateforme moesienne, le secteur Slobozia, Amara, Padina, Jugureanu), la formation rouge s'effiloche progressivement ; des marnes, marnocalcaires, argiles polycolores et évaporites alternant avec des calcaires (pélétaux, micritiques, oolithiques, microoncolithiques) à microfaune et microflore caractéristiques pour un milieu marin côtier (le domaine inter- et supratidal). C o s t e a et al. (1978) ont déterminé dans ces calcaires des coprolithes : *Favreina salevensis* (P a r é j a s), des foraminifères : *Anchispirocyclina lusitanica* (E g g e r), *Trocholina elongata* (L e u p o l d), *Feurtillia frequens* M a y n c, *Pfenderina neocomiensis* (P f e n d e r), des algues : *Macroporella embergeri* (B o u r o u l l e c & D e l o f f r e), *Actinoporella podolica* (A l t h.), *Radoiciciella subtilis* D r a g a s t a n. Les marnes et les argiles polycolores contiennent, tout comme dans le forage de Cernavodă, d'Ostracodes et de Charophytes. A partir du contenu fossile ci-dessus, on a attribué à ces dépôts l'âge berriasien-valanginien inférieur. Au-dessous de ceux-ci il y a des dolomies, des calcaires à *Clypeina jurassica* F a v r e aux intercalations gypsifères et anhydritiques, donc un faciès marin-lagunaire d'âge tithonique supérieur (V i n o g r a d o v et al., 1978).

Plus loin vers l'ouest (le secteur Urziceni, Snagov, Potigrafu) les dépôts berriasiens-valanginiens inférieurs se développent en exclusivité dans le faciès marin littoral abondant en calcaires pélétaux et coprolithiques à *Favreina salevensis* (P a r é j a s). Sous ces dépôts il y a des calcaires intrasparitiques et des biolithites algales à *Clypeina jurassica* F a v r e, *Cayeuxia moldavica* F r o l l o, des Coraux et des Bryozoaires d'âge tithonique supérieur. Surmontant les dépôts berriasiens-valanginiens inférieurs, il y a des calcaires intrasparitiques et micritiques à *Pfenderina neocomiensis* (P f e n d e r), *Trocholina elongata* L e u p o l d et *Pseudotextulariella salevensis* C h a r o l l., B r ö n n., Z a n i n d'âge valanginien supérieur.

Dans la Dépression de Bîrlad, la formation rouge ayant une épaisseur stratigraphique de 200 — 300 m, est constituée de dépôts terrigènes bariolés fins (marnes et argiles ferrugineuses à Ostracodes et Charophytes, grès fins ferrugineux), de roches carbonatées (micrites, pelsparites, oosparites, dolomicrites aux variétés argileuses et ferrugineuses) et évaporites (anhydrites).

Au-dessous de la formation rouge terrigène-lagunaire reposent des calcaires récifaux cor-algals (par endroits dolomitisés et aux nids d'anhydrite) d'âge tithonique supérieur, à *Conicospirillina basiliensis* M o h l e r, *Clypeina jurassica* F a v r e, *Macroporella praturloni* D r a g a s t a n et des microoncolithes. La partie sommitale de la formation récifale cor-algale contient des foraminifères : *Anchispirocyclina lusitanica* (E g g e r), *Pseudocyclammina lituus* (Y o k o), *Everticyclammina virguliana* (K o e c h l i n) et des algues : *Thaumathophorella parvovesiculifera*

(R a i n e r i), *Actinoporella podolica* (A l t h.) qui indiquent l'âge berriasien pour ce segment stratigraphique.

Dans le sud de la Dépression de Bîrlad (Polocin et Aldești) les forages ont traversé des dépots hauteriviens, localement développés en faciès pélagique (calcaires micritiques siltiques, marnes à Spirillinidae et Hedbergellinae).

On pourrait donc conclure sur l'âge valanginien, peut-être berriasien supérieur, de la formation rouge dans la Dépression de Bîrlad.Le caractère de la sédimentation est mixte, des séquences marines épicontinentales (calcaires pélétaux, oolithiques et récifaux) se développant également sur le fond continental-évaporitique.

Dans la Dépression pré-dobrogéenne (les forages de Stipoc, Letea, C. A. Rosetti, L. Roșu dans le Delta du Danube) on a rencontré des argiles ferrugineuses dolomitiques, de grès fins et de siltites au cément calcitique-ferrugineux, dolomitique-gypsifère ou à matrice argileuse-séricitique, qui s'étendent sur une épaisseur stratigraphique de 300 — 500 m. La datation stratigraphique de ces dépôts est incertaine, car on n'y a pu déterminer que quelques espèces palynologiques [3] : *Trilobosporites apiverucatus* C o u p e r , *Clavifera triplex*, *Bolchovitina*, *Pilosisporites trichopapillosus* (T h i e r g a r t) D e l c o u r t & S p r u m o n t qui indiqueraient le Crétacé inférieur. L'épaisseur considérable de cette formation, aussi bien que la présence sous-jacente d'une formation récifogène d'âge tithonique inférieur [4] (à microoncolithes, très rarement *Saccocoma*, fragments d'échinides, brachiopodes, gastéropodes, bryozoaires) permettent de supposer un âge compréhensif purbecko-wealdien, même aptien pour la formation rouge de la Dépression pré-dobrogéenne.

Dans le nord-ouest de la Plate-forme moldave, la formation rouge épaisse d'environ 200 m se compose de dépôts terrigènes barriolés (marnes et argiles ferrugineuses, grès fins à cément dolomitique-ferrugineux, conglomérats polymictiques), roches carbonatées (micrites et microsparites brunes, dolomicrites ferrugineuses, calcarénites) et évaporites (anhydrites). On y a récolté de diverses espèces d'Ostracodes et de Characeae, *Favreina salevensis* (P a r é j a s) et *Aeolissacus*. Au-dessus de la formation rouge reposent des calcaires biointramicritiques, biopelsparitiques et micritiques à foraminifères : *Pfenderina neocomiensis* (P f e n d e r), *Trocholina alpina* (L e u p o l d), *Trocholina elongata* (L e u p o l d), *Neotrocholina valdensis* R e i c h e l , *Haplophragmoides joukowskyi* C h a r o l l., B r ö n n., Z a n i n , *Quinqueloculina* div. sp. et des algues : *Salpingoporella annulata* C a r o z z i, ce qui atteste leur âge valanginien moyen-supérieur. Les forages effectués jusqu'à présent n'ont pas mis en évidence la présence des dépôts jurassiques supérieurs dans le soubassement de la formation rouge qui, soit manquent dans ce secteur, soit se trouvent un peu vers l'ouest. Admettant donc un âge berriasien-valanginien inférieur pour la formation rouge du nord-ouest de la Plate-forme moldave, on ne pourrait faire abstraction de la présence possible du Purbeckien dans la partie basale d'un secteur plus occidental.

Les formations rouges d'âge aptien

Les dépôts aptiens dans la Dobrogea méridionale sont représentés par deux faciès distincts : continental-lacustre (la formation rouge) et marin. Chiriac et al., 1977, font préciser l'étendue du faciès continental-lacustre : dans le secteur compris entre la vallée de Carasu (la parallèle Cernavodă) et la ligne du chevauchement Capidava-Ovidiu, aussi bien que son âge (bédoulien inférieur et gargasien). Il y a donc deux niveaux pour la formation rouge entre lesquels on trouve une intercalation lentilliforme de calcaires micritiques bédouliens supérieurs à *Cheloniceras* sp., *Deshayesites* sp., *Palorbitolina lenticularis* (Blum.). Le Clansayesien (la zone à *Nodosocostatum* et *Bigoureti*) constitué de sables et grès glauconieux, graviers ou microconglomérats à phosphates recouvre transgressivement la formation rouge. A la constitution de la formation rouge participent : sables siliceux, graviers, conglomérats, argiles sableuses et kaolineuses multicolores. La nature de ces dépôts est terrigène-détritique, car ils sont le résultat du remaniement d'un matériel résiduel provenu par l'altération de l'éruptif nord-dobrogéen dans un milieu sous-aérien (Popescu et al., 1969). L'épaisseur stratigraphique de ces deux niveaux de formation rouge est de 1 — 20 m (Bédoulien inférieur), respectivement de 2 — 30 m (Gargasien).

On a également rencontré la formation rouge aptienne au moyen des forages dans la partie orientale de la Plate-forme moesienne (le secteur Slobozia-Padina-Jugureanu) où elle recouvre transgressivement la formation rouge du Berriasien-Valanginien inférieur. Les lithotopes terrigènes caractéristiques: argiles, marnes, sables, grès et conglomérats bariolés (polycolores) dominent dans l'est et le nord du secteur (Jugureanu, Odăeni, Ştefan Gheorghiu, Roşiori) ; vers l'ouest (Brăgăreasa-Colelia-Călăraşi) il y a des calcaires biomicritiques et oolithiques à *Palorbitolina lenticularis* (Blum.), *P. conoidea* (Gras.), *Choffatella decipiens* (Schlum.) qui prédominent. Les pélites polycolores contiennent de diverses espèces d'Ostracodes et de Characées. L'épaisseur stratigraphique de la formation rouge aptienne (les séquences calcaires marines littorales y comprises) est de 10 — 25 m allant jusqu'à quelques dizaines de mètres dans les secteurs dépressionaires.

Dans la Dépression de Bîrlad, la formation rouge berriasienne-valanginienne est recouverte d'une manière discordante par un deuxième niveau de formation rouge terrigène-continentale qu'on a attribué, faute de matériel paléontologique, à partir de la corrélation des diagraphies électriques, à l'Aptien. L'épaisseur des dépôts aptiens rouges décroît de l'ouest (Bîrlad, 250 m) vers l'est (Prut, 110 m) ; des dépôts marins badéniens les recouvrent d'une manière discordante. Du point de vue lithologique, elles se caractérisent par des marnes et des argiles ferrugineuses bariolées, argiles dolomitiques, sables, grès, dolomies calcaires gréseuses. Elles diffèrent de la formation rouge inférieure (berriasienne-valanginienne) par l'absence des anhydrites et des intercalations calcaires d'origine marine. Dans la partie occidentale de la Dépression de Bîrlad (le secteur Găiceana-Hurueşti), l'Aptien se développe en faciès marin-littoral à grès siliceux, grès lithiques, sables, graviers et argiles contenant *Palorbitolina conoidea* (Gras.), *P. lenticularis* (Blum.), *Choffatella decipiens* (Schl u m.), *Trocholina aptiensis* I o v c e v a. Plus vers le nord (le secteur Bacău), l'Aptien à un contenu microfaunique similaire, se compose de calcaires oosparitiques, de grès siliceux et d'intercalations conglomératiques.

A l'égard d'une possible existence des dépôts aptiens dans le toit de la formation rouge dans la Dépression pré-dobrogéenne on a fait certaines référances dans le paragraphe ci-dessus.

Conclusions

A la limite Jurassique/Crétacé et dans l'Aptien, les secteurs périphériques des unités d'avant-pays des Carpathes roumaines sont colmatés avec des produits résiduels provenus de l'écorce d'altération jurassique et crétacée inférieure formée sur le compte de l'éruptif nord-dobrogéen. C'est le cas de la Dobrogea méridionale, de l'extrêmité nord-est de la Plate-forme moesienne, la Dépression pré-dobrogéenne et la Dépression de Bîrlad. Leur emplacement dans la proximité de la région dénudée, aussi bien que la présence des éléments de l'éruptif altéré, des feldspaths potassiques, des métaquartzites et du zircon dans les grès lithiques et dans les conglomérats indiquent comme source l'éruptif nord-dobrogéen. A partir des considérations ci-dessus, les formations rouges d'âge jurassique supérieur-crétacé inférieur autour de l'éruptif nord-dobrogéen constituent une province pétrologique sédimentaire. La formation rouge du nord-ouest de la Plate-forme moldave (le secteur Rădăuți) appartient à une autre province pétrologique sédimentaire; la source probable du matériel résiduel pourrait être les calcaires paléozoïques de la Plate-forme moldave. L'idée est soutenue aussi par de nombreux fragments de biosparites à pigment ferrugineux qui entrent dans la composition des calcirudites et des calcarénites intercalées dans la formation rouge.

Le caractère continental-lacustre ou continental-lagunaire diminue avec la croissance de la distance de la zone dénudée; ainsi se produit une transition graduelle vers les faciès mixtes (continentaux et marins-littoraux) et finalement vers les faciès marins épicontinentaux typiques.

Les formations rouges sont précédées temporellement par des formations récifales cor-algales (Kimméridgian-Tithonique inférieur qui par endroits persistent également au Tithonique supérieur et même au Berriasien), ayant une vaste étendue régionale, accompagnées par de processus très actifs de dolomitisation syngénétique et du développement des faciès évaporitiques.

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[3] N. B a l t e ș – communication verbale.

[4] O. D r a g a s t a n – communication verbale.

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QUESTIONS

E. Dudich. Vinogradov a évoqué la destruction d'une ancienne croûte d'altération comme source principale du matériel des couches rouges étudiées. Est-ce qu'il y a des données chimiques et/ou minéralogiques sur leur composition ? Cela présente un intérêt particulier par rapport à la présence éventuelle des minéraux d'alumine libre, donc bauxitique, d'origine latéritique.

Réponse: L'existence des anciennes croûtes latéritiques aux alentours des formations rouges constitue une hypothèse plausible qui n'a pas été confirmée dans les forages jusqu'à présent. Par conséquent nous n'avons pas disposé du matériel pour les analyses requises.

NANNOPLANCTON CALCAIRE À LA LIMITE CRÉTACÉ-TERTIAIRE DANS LE FLYSCH DES CARPATHES ORIENTALES ROUMAINES[1]

PAR

IOAN COSTEA[2], PETRE IONESCU[2]

Le problème de la limite Crétacé-Paléogène dans les Carpathes orientales roumaines a intéressé depuis longtemps de nombreux chercheurs. Faute de macrofossiles, leurs conclusions ont été exclusivement fondées sur des arguments d'ordre micropaléontologique (microfaune, palynologie). Parmi les chercheurs qui ont étudié, au cours des dernières 15 — 20 années, la séquence stratigraphique mentionnée, on pourrait rappeler : Vinogradov, 1960; Tocorjescu, 1960, 1963; Săndulescu, Săndulescu, Kusko, 1962; Grigoraş, Pauliuc, Costea, 1962, 1964, 1966, 1970; Dumitrescu, 1963; Joja, Cosma, Dumitrescu, 1963; Costea, Dima, 1965; Agheorghiesei, Băncilă, Costea, Roza, 1967; Băncilă, Marinescu, Costea, 1968; Costea, Saraiman, Comşa, 1969; Bratu, Alexandrescu, 1970; Dimian, 1970; Bratu, Gheţa, 1972, 1974; Tocorjescu, Ionesi, 1974; Bratu, 1975; Balteş, Cosma, Lebenzon, Munteanu, 1975; Ionesi, 1975; Micu, 1977 etc.

Que les travaux accomplis jusqu'à présent aient réalisé des biozonations plus larges ou plus détaillées, tous montrent que dans les Carpathes orientales, les dépôts crétacés se trouvent en continuité avec les dépôts paléogènes et que le segment de la limite indique l'existence du Paléocène inférieur et supérieur. La continuité de la sédimentation est évidente, se dessinant dans toutes les unités majeures du flysch carpathique où il y a des dépôts crétacés aussi bien que paléogènes.

On n'a pas encore essayé, en Roumanie, d'élucider le problème de la limite Crétacé-Paléogène dans les Carpathes orientales à partir de l'étude du nannoplancton calcaire. Par conséquent, l'ouvrage ci-présent vient de s'assumer cette responsabilité pour la première fois.

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[1] Note présentée au 12 ème Congrès de l'Association Géologique Carpatho-Balkanique, 8 — 13 septembre 1981, Bucarest, Roumanie.

[2] Ministère du Petrole, Bucarest, Roumanie.

Source du matériel d'étude. Pour le travail ci-présent on a prélevé 352 échantillions géologiques dans les affleurements les plus représentatifs du flysch crétacé-paléogène et on a choisi 396 échantillons de puits des forages qui ont traversé la succession comprise entre le Campanien supérieur et l'Eocène inférieur. Selon leur distribution sur les unités structurales (d'après l'opinion de B ă n c i l ă, 1958), ces dernières sont :

— l'unité ouest-interne ; Cetățeni-Lăicăi (le bassin de la Dîmbovița) et les forages de Fieni ;

— l'unité est-interne : Neagra et Ticoș (le bassin du Bicaz) ;

—l'unité médio-interne (Audia) ; les forages de Siriu et de Vulpea-Bertea ;

— l'unité médio-marginale.

A) La sous-unité médiane : vallée de Sadău (le bassin de la Suceava), Tașca (le bassin du Bicaz) et les forages Vatra Moldovița, Ața Brateș, Asău, vallée d'Uzului, Apa Neagră, Hilib, Ojdula, Ghelința et Bîsca Cernat ;

B) La sous-unité intermédiaire : vallée de Voroneț (le bassin de la Moldova), vallée de Pluton (le bassin du Neamț), vallée de Pîngărați (le bassin de la Bistrița) et les forages de Gura Humorului, Frasin, Suha, Pluton, Pipirig et Pîngărați ;

C) La sous-unité marginale : vallée de Sucevița (le bassin de la Suceava), vallée de Secuța (le bassin du Neamț) et les forages de Putna Dumesnic, Geamăna, Chilii, Zemeș, Tașbuga, Moinești, Supan, Lapoș, Dărmănești, Păcurița, Cerdac, Brețcu, Oituz, Lepșa, Zăbala et vallée de Gardului.

D) L'unité externe : vallée de Runcu (le bassin de la Bistrița), vallée de Putna-Lepșa (le bassin de la Putna) et les forages Straja, Geamăna,Zemeș, Tazlău, Tașbuga, Moinești, Dărmănești, Oituz et Ojdula.

Biozones de nannoplancton dans le segment stratigraphique étudié. Suivant la distribution des espèces de nannoplancton dans l'intervalle compris entre le Vraconien et l'Eocène inférieur, on a pu séparer 7 zones de nannoplancton calcaire.

Les biozones établies ont été correlées avec les zones proposées par : C e p e k - H a y, 1969 ; M a n i v i t, 1971 ; R o t h - T h i e r s t e i n, 1972 ; R o t h, 1973 ; P e r c h - N i e l s e n, 1977 ; V e r b e e k, 1977 et C o s t e a , C o m ș a , 1979, pour les dépôts crétacés et M a r t i n i, 1970 et H a y, 1970, pour les dépôts paléocènes. Pour permettre la réalisation des corrélations détaillées, régionales et extrarégionales, on a fait des références aux schémas stratigraphiques basées sur des foraminifères et surtout sur les foraminifères planctoniques.

Les zones individualisées sont :

1. *La zone Ceratolithoides aculeus.* Définition : l'intervalle stratigraphique compris entre la première apparition de l'espèce *Ceratolithoides aculeus* (S t r a d n e r, 1961) P r i n s, S i s s i n g h, 1977 et celle de l'espèce *Tetralithus trifidus* (S t r a d n e r , 1961) B u k r y , 1973.

La zone individualisée dans le Campanien inférieur dans le faciès des couches de Hangu de l'unité moyenne-marginale est corrélable avec la zone microfaunique Globotruncana elevata et G. stuartiformis.

En même temps que l'apparition de l'espèce *Ceratolithoides aculeus* (Stradner, 1961) Prins et Sissingh, 1977, sur le fond fossile de l'association santonienne, les espèces nouvelles suivantes viennent de s'ajouter, tout en caractérisant des différents niveaux stratigraphiques: *Arkangelskiella cymbiformis* Vekshina, 1959; *A. specillata* Vekshina, 1959; *Broinsonia parca* (Stradner, 1963) Bukry, 1969; *Kamptnerius magnificus* Deflandre, 1959; *K. tabulatus* Perch-Nielsen, 1968; *Biscutum constants* (Gorka, 1957) Black, 1967; *Cribrosphaerella ehrenbergi* (Arkhangelsky, 1912) Deflandre, 1952; *Reinhardites mirabilis* Perch-Nielsen, 1968; *Cylindralithus serratus* Bramlette, Martini, 1964; *Chiastozygus anceps* (Gorka, 1957) Shumenko, 1976; *C. cuneatus* (Liulieva, 1967) Cepek, Hay, 1969; *Tranolithus manifestus* Stover, 1966; *T. orionatus* (Reinhardt, 1966) Reinhardt, 1966; *Vagalapilla compacta* Bukry, 1969; *Vekshinella stradneri* Rood, Hay § Barnard, 1971; *Zygodiscus deflandrei* Bukry, 1969; *Z. theta* (Black, 1959) Bukry, 1969; *Ahmuellerella octoradiata* (Gorka, 1957) Reinhardt, 1966; *Micula concava* (Stradner, 1960) Bukry, 1969; *Rucinolithus hayi* Stover, 1966; *Tetralithus gothicus* Deflandre, 1959; *T. obscurus* Deflandre, 1959; *T. pyramidus* Gardet, 1955 et *Marthasterites furcatus* (Deflandre, 1954) Deflandre, 1959.

La partie supérieure de la séquence caractérisée par l'association à *Ceratolithoides aculeus* est marquée par la disparition de deux de ses espèces, précisément *Zygodiscus theta* (Black, 1959) Bukry, 1969 et *Rucinolithus hayi* Stover, 1966.

2. *La zone Tetralithus trifidus*. Définition: l'intervalle stratigraphique compris entre la première apparition de l'espèce *Tetralithus trifidus* (Stradner, 1961) Bukry, 1973 et l'apparition de l'espèce *Lithraphidites quadratus* Bramlette, Martini, 1964.

La zone a été mise en évidence par les unités ouest-interne, moyenne-marginale et externe du flysch des Carpathes orientales de la Roumanie et correspond à la séquence du Campanien supérieur et du Maestrichtien inférieur développée dans les marnes rouges de Gura Beliei, les couches de Hangu et les couches de Caşin, corrélable à la zone de foraminifères Globotruncanella stuarti et Globotruncana contusa.

Association riche qualitativement et quantitativement, elle se fait distinguer par 20 espèces supplémentaires au niveau inférieur du segment stratigraphique, également caractérisé par la disparition de 29 espèces du niveau supérieur. Le tableau de distribution annexé présente les espèces supplémentaires, tout comme celles qui disparaissent.

Au niveau inférieur du segment, caractérisé par ces espèces, s'ajoutent: *Biscutum melaniae* (Gorka, 1957) Reinhardt, 1969; *Markalius reinhardtii* Perch-Nielsen, 1968; *Repagulum parvidentatum* (Deflandre et Fert, 1954) Forchheimar, 1972; *Cretarhabdus crenulatus* Bramlette & Martini; *C. ingens* (Gorka, 1957) Reinhardt, Gorka, 1967; *Reinhardtites anthophorus* (Deflandre, 1959) Perch-Nielsen, 1968; *Chiastozygus quadriperforatus* Gartner, 1968; *Eiffellithus gorkae* Reinhardt, 1965; *Vekshinella bochothnicae* (Gorka, 1957) Verbeek, 1977; *Zygodiscus acanthus* (Reinhardt, 1965) Reinhardt, 1966; *Z. ponticulus* (Deflandre, 1954) Reinhardt, 1966; *Z. sigmoides* Bramlette, Sullivan, 1961; *Z. spiralis* Bramlette, Martini, 1964; *Braarudosphaera discula* Bramlette, Riedel, 1954; *Lithraphidites carniolensis* Deflandre, 1963; *Microrhabdulus decoratus* Deflandre, 1959; *Lucianorhabdus cayeuxi* Deflandre, 1959; *Rhomboaster cuspis* Bramlette, Sullivan, 1961; *Tetralithus ovalis* Stradner, 1963 et *T. trifidus* (Stradner, 1961) Bukry, 1973.

Les espèces qui disparaissent sont: *Broinsonia parca* (Stradner, 1963) Bukry, 1969; *Kamptnerius tabulatus* Perch-Nielsen, 1968; *Biscutum constans* (Gorka, 1957) Black, 1967; *Markalius circumradiatus* (Stover, 1966) Perch-Nielsen, 1968; *Repagulum parvidentatum* (Deflandre, Fert, 1954) Forchheimer, 1972; *Cretarhabdus actinosus* (Stover, 1966) Forchheimer, 1972; *C. loriei* Gartner, 1968; *C. surirellus* (Deflandre, 1954) Reinhardt, 1970; *Cribrocorona gallica* (Stradner, 1963) Perch-Nielsen, 1973; *Parhabdolithus asper* (Stradner, 1963) Manivit, 1971, *P. splendens* (Deflandre, 1953) Noël, 1969; *Corollithion exiguum* Stradner, 1961; *Lithastrinus floralis* Stradner, 1962; *Chiastozygus cuneatus* (Liulieva, 1967) Cepek, Hay, 1969; *C. quadriperforatus* Gartner, 1968; *Eiffellithus eximinus* (Stover, 1966) Perch-Nielsen, 1968; *Tranolithus exiguus* Stover, 1966; *T. manifestus* Stover, 1966; *Vagalapilla compacta* Bukry, 1969; *Vekshinella bochotnicae* (Gorka, 1957) Verbeek, 1977; *V. stradneri* Rood, Hay, Barnard, 1973; *Zygodiscus acanthus* (Reinhardt, 1965) Reinhardt, 1966; *Z. deflandrei* Bukry, 1969; *Z. ponticulus* (Deflandre, 1954) Reinhardt, 1966; *Lithraphidites carniolensis* Deflandre, 1963; *Ceratolithoides aculeus* (Stradner, 1961) Prins, Sissingh, 1977; *Lucianorhabdus maleformis* Reinhardt, 1966; *Rucinolithus hayi* Stover, 1966; *Tetralithus gothicus* Deflandre, 1959; *T. trifidus* (Stradner, 1961) Bukry, 1973; *T. ovalis* Stradner, 1963 et *Marthasterites furcatus* (Deflandre, 1954) Deflandre, 1959.

3. *La zone Lithraphidites quadratus*. Définition: l'intervalle stratigraphique compris entre l'apparition des espèces *Lithraphidites quadratus* Bramlette, Martini, 1964 et *Tetralithus quadratus* Stradner, 1961 et l'apparition de l'espèce *Micula murus* (Martini, 1961) Bukry, 1973.

La zone, mise en évidence par les unités ouest-interne, moyenne-marginale et externe, caractérise le Maestrichtien moyen et on peut la corréler à la biozone Globotruncana gansseri et à la partie inférieure de la biozone Abathomphalus mayaroensis.

A la partie supérieure du segment caracterisé par la zone Lithraphidites quadratus, 12 espèces dont l'évolution commence au Crétacé inférieur et finit en Turonien, disparaissent (planche); *Watznaueria communis* Reinhardt, 1964; *W. deflandrei* (Noël, 1965) Reinhardt, 1971; *Cretarhabdus conicus* Bramlette, Martini, 1964; *Podorhabdus* cf. *decorus* (Deflandre, 1954) Thierstein, 1972; *Reinhard-*

tites anthophorus (Deflandre, 1954) Perch-Nielsen, 1968; *Cylindralithus serratus* Bramlette, Martini, 1964; *Chiastozygus amphipons* (Bramlette, Martini, 1964) Gartner, 1968; *C. litterarius* (Gorka, 1957) Manivit, 1971; *Eiffellithus gorkae* Reinhardt, 1965; *Tranolithus orionatus* (Reinhardt, 1966) Reinhardt, 1966; *Zygodiscus diplogrammus* (Deflandre, 1954) Gartner, 1968; *Tetralithus obscurus* Deflandre, 1959.

4. *La zone Micula murus*. Définition : l'intervalle stratigraphique compris entre la première apparition de l'espèce *Micula murus* (Martini, 1961) Bukry, 1973, et la première apparition de l'espèce *Markalius astroporus* (Stradner, 1961) Hay et Mohler, 1967.

En même temps que l'espèce *Micula murus* de l'association, individualisé dans les mêmes unités que la zone précédente, apparaissent aussi les espèces : *Biscutum notaculum* Wind, Wise, 1977; *Coccolithus pelagicus* (Wallich, 1870) Schiller, 1930; *Neococcolithes dubius* (Deflandre, 1954) Black, 1967; *Ceratolithoides kamptneri* Bramlette, Martini, 1964 et *Dictyolithus quadratus* Gorka, 1957.

Les espèces disparaissant au niveau de la limite supérieure de l'intervalle stratigraphique caractérisé par la zone, définissent, en nombre considérable, un seuil bionomique important et mettent en évidence les modifications importantes des conditions de milieu au niveau de la limite Crétacé-Paléogène; *Arkhangelskiella cymbiformis* Vekshina, 1959; *A. specillata* Vekshina, 1959; *Gartnerago obliquum* (Stradner, 1963) Noël, 1970; *Kamptnerius magnificus* Deflandre, 1959; *Biscutum ignotum* (Gorka, 1957) Reinhardt, Gorka, 1967; *B. melaniae* (Gorka, 1957) Reinhardt, 1969; *Manivitella pemmatoidea* (Deflandre, Manivit, 1965) Thierstein, 1971; *Watznaueria barnesae* (Black, 1959) Perch-Nielsen, 1968; *Cretarhabdus crenulatus* Bramlette, Martini, 1964; *C. ingens* (Gorka, 1957) Reinhardt, Gorka, 1967; *Cribrosphaerella ehrenbergi* (Arkhangelsky, 1912) Deflandre, 1952; *Parhabdolithus angustus* (Stradner, 1963) Stradner, Adamiker, Maresch, 1968; *P. embergeri* (Noël, 1958) Stradner, 1963; *Prediscosphaera cretacea* (Arkhangelsky, 1912) Gartner, 1968; *P. spinosa* (Bramlette, Martini, 1964) Gartner, 1968; *Reinhardtites mirabilis* Perch-Nielsen, 1968; *Stephanolithion laffittei* Noël, 1956; *Chiastozygus anceps* (Gorka, 1957) Shumenko, 1976; *Eiffellithus turriseiffeli* (Deflandre, 1954) Reinhardt, 1965; *Ahmuelerella octoradiata* (Gorka, 1957) Reinhardt, 1966; *Lithraphidites quadratus* Bramlette, Martini, 1964; *Microrhabdulus decoratus* Deflandre, 1959; *Ceratolithoides kamptneri* Bramlette, Martini, 1964; *Dictyolithus quadratus* Gorka, 1957; *Lucianorhabdus cayeuxi* Deflandre, 1959; *Micula concava* (Stradner, 1960) Bukry, 1969; *M. decussata* Vekshina, 1959; *M. murus* (Martini, 1961) Bukry, 1973; *Tetralithus descriptus* Martini, 1961; *T. pyramidus* Gardet, 1955 et *T. quadratus* Stradner, 1961.

5. *La zone Markalius astroporus*. Définition : l'intervalle stratigraphique compris entre la première apparition de l'espèce *Markalius astro-*

porus (Stradner, 1961) Hay, Mohler, 1967 et l'apparition de l'espèce *Cruciplacolithus tenuis* (Stradner, 1961) Hay, Mohler, 1967.

La zone s'est individualisée dans les unités ouest-interne, moyenne-marginale et externe. L'association de la zone est limitée comme nombre d'espèces et d'exemplaires vis-à-vis des biozones précedentes, contenant, en dehors de la fossile dénominante, les espèces suivantes : *Biscutum notaculum* Wind, Wise, 1977 ; *Coccolithus crassus* Bramlette, Sullivan, 1961 ; *C. pelagicus* (Wallich, 1870) Schiller ; 1930 ; *Prinsius bisulcus* (Stradner, 1963) Hay, Mohler, 1967 ; *Markalius inversus* (Deflandre, 1954) Bramlette, Martini, 1964 ; *M. reinhardtii* Perch-Nielsen, 1968 ; *Neococcolithes dubius* (Deflandre, 1954) Black, 1967 ; *Zygodiscus sigmoides* Bramlette, Sullivan, 1961 ; *Z. spiralis* Bramlette, Martini, 1964 ; *Braarudosphaera bigelowi* (Gran, Braarud, 1935) Deflandre, 1947 ; *B. discula* Bramlette, Riedel, 1954 ; *Biantholithus sparsus* Bramlette, Martini, 1964 et *Rhomboaster cuspis* Bramlette, Sullivan, 1961.

L'association correspond à la zone NP_1 (Martini, 1960), peut se corréler à la zone Globoconusa daubjergensis et à la partie inférieure de la zone Subbotina triloculinoides/Subbotina pseudobulloides et caractérise le premier segment du Paléocène inférieur c'est-à-dire le Danien inférieur.

6. *La zone Cruciplacolithus tenuis. Définition :* l'intervalle stratigraphique compris entre la première apparition de l'espèce *Cruciplacolithus tenuis* (Stradner, 1961) Hay, Mohler, 1967 et la première apparition de l'espèce *Chiasmolithus danicus* (Bramlette, Martini, 1964) Hay, Mohler, 1967.

A la partie inférieure du segment stratigraphique qui y corespond, les espèces suivantes viennent de s'ajouter sur le fond de l'association antérieure : *Chiasmolithus bidens* (Bramlette, Sullivan, 1961) Hay, Mohler, Wade, 1966 et *Coccolithus eopelagicus* (Bramlette & Riedel, 1954) Bramlette, Sullivan, 1961, tandis qu'à la partie supérieure, l'espèce *Biscutum notaculum* Wind, Wise, 1977, cesse son évolution.

L'association, toujours pauvre, individualisée dans les unités ouest-interne et moyenne-marginale, est corrélable avec celle de la zone NP_2 (Martini, 1970) et avec la biozone de foraminifères Subbotina triloculonoides/Subbotina pseudobulloides et caractérise le Paléocène inférieur, c'est-à-dire le Danien supérieur.

7. *La zone Chiasmolithus danicus*. Définition : l'intervalle stratigraphique compris entre l'apparition de l'espèce *Chiasmolithus danicus* (Bramlette, Martini, 1964) Hay, Mohler, 1967 et la disparition de l'espèce *Markalius reinhardtii* Perch-Nielsen, 1968.

L'association de la zone Chiasmolithus danicus, pratiquement similaire à celle adiacente inférieure *(Cruciplacolithus tenuis)* en dehors du fait que son niveau apporte aussi l'espèce dénominante, corrélable avec la zone NP_3 au même nom (Martini, 1970) correspond à la partie terminale de la biozone Subbotina triloculinoides/Subbotina pseudobulloides et à la partie inférieure de la zone Acarinina inconstans, parmi les biozones de foraminifères. L'association caractérise l'intervalle stratigraphique qui correspond au Paléocène inférieur et a été identifiée dans les unités ouest-interne, moyenne-marginale et externe.

Conclusions. L'étude de la répartition stratigraphique du nannoplancton calcaire du niveau de la limite entre le Crétacé et le Tertiaire, avait démontré que la séquence analysée se fait remarquer par une continuité manifeste de sédimentation entre les dépôts des deux périodes. Les 7 biozones séparées en sont la preuve.

En général, on a réussi à mettre en corrélation les zones du nannoplancton calcaire séparées par les subdivisions définies, pour les Carpathes orientales à partir des foraminifères.

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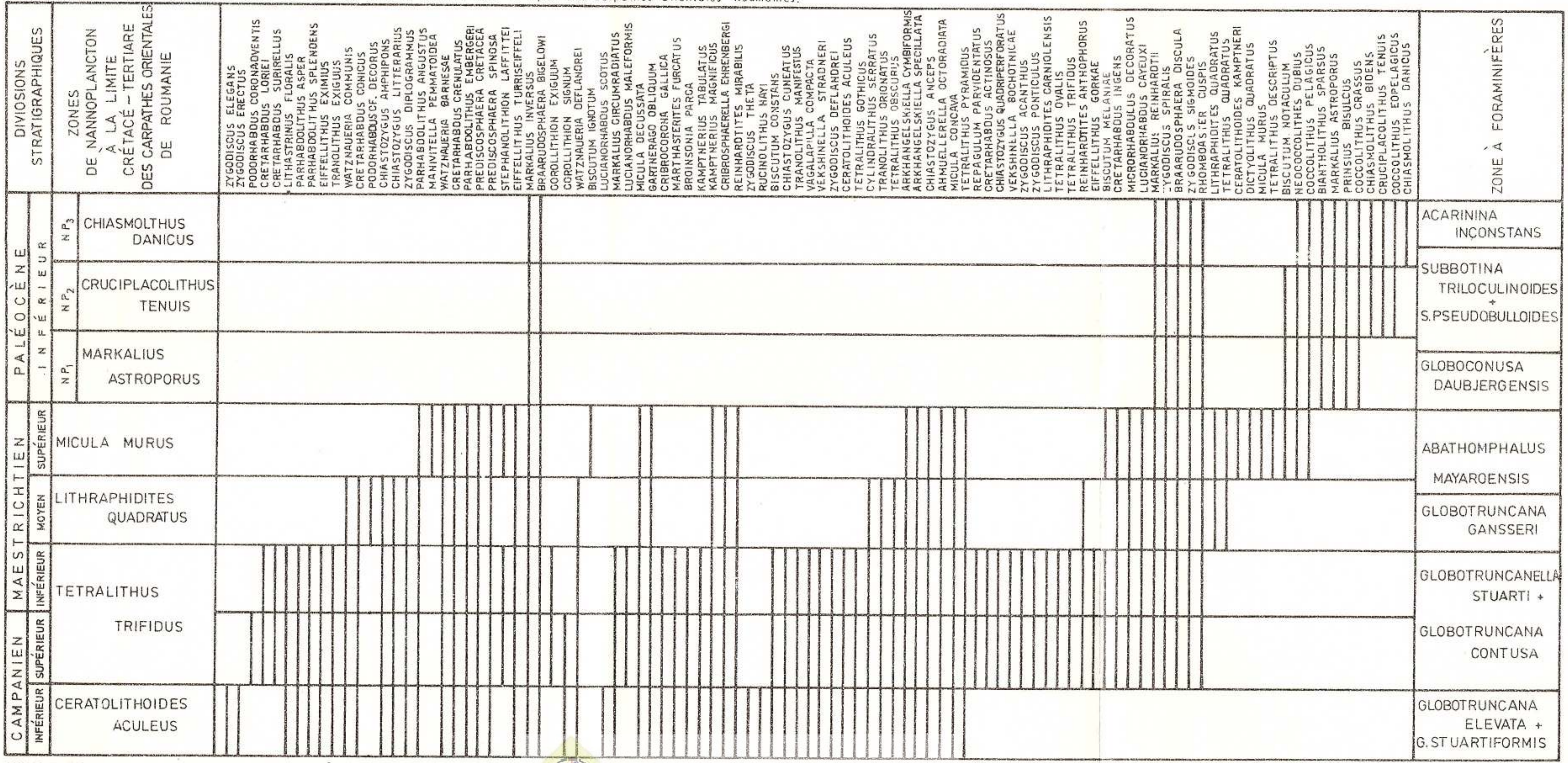
DISTRIBUTION STRATIGRAPHIQUE DU NANNOPLANCTON CALCAIRE AU NIVEAU DE LA LIMITE CRÉTACÉ-TERTIARE

I. COSTEA, P. IONESCU. Nannoplancton calcaire à la limite Crétacé-Tertiaire dans le flysch des Carpathes Orientales Roumaines.



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CORRELATION OF THE PLIOCENE AND LOWER PLEISTOCENE DEPOSITS IN THE DACIC AND EUXINIC BASINS [1]

BY

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1. Introduction

This paper represents the results of the researches carried out within a scientific cooperation theme between the Romanian and the Soviet specialists. The object proposed was the elaboration of a scheme of biostratigraphic correlation of the Pliocene and Pleistocene deposits in the Dacic and Euxinic basins.

With this aim in view, complex researches were carried out using biostratigraphic, sedimentologic, paleomagnetic, and absolute age dating methods. A general correlation table of the two basins has been elaborated (Plate II).

2. Correlation of the Pliocene Deposits in the Dacic and Euxinic Basins

Pontian

The term Pontian, proposed by Le Play (1842), was redefined by Barbot de Marny (1869) and then by Andrusov (1897, 1912, 1923, etc.) as representing a chronostratigraphic unit of the upper Neogene in the Paratethys.

The Pontian stage can be defined as the time period during which, in the Paratethys, there were deposited the beds lying between the Upper Meotian deposits and the Lower Dacian (= Kimmerian) formations. In biostratigraphic respect, the Pontian is mainly characterized by the appearance and development, among mollusca, of the Limnocardiids and Dreissenides.

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The study of the faunal content of the Pontian deposits in the Euxinic Basin, where the stratotype of the stage exists, allowed Andrusov (1923) to separate two substages within the Pontian: Novorossian and Bosphorian.

Several horizons have been distinguished by different authors within the Novorossian: Odessa Beds (Mihailovskii, 1909); Eupatoria Beds (Davitaşvili, 1933); Congeria subrhomboidea Beds (Eberzin, 1967); C. rhomboidea Horizon (Celidze, 1953), etc. Stefanovič (1951) has proposed the Portaferrian substage for the C. rhomboidea Beds and Macarovici et al. (1965) suggested the Odessian substage for the Odessa Beds with Eupatoria Beds in the basement. Eberzin et al. (1967) consider that the Portaferrian in the Euxinic Basin corresponds to the C. subrhomboidea Beds, supposed to be an equivalent of the C. rhomboidea Beds in the Dacic and Pannonian basins. The making evident of the C. rhomboidea Beds in the west of the Georgian Soviet Socialist Republic (Celidze, 1953) and the specification of their stratigraphic position (Taktakisvilii, 1971; Celidze, 1974) led to the conclusion that in the Euxinic Basin the Portaferrian includes the C. subrhomboidea Beds at the lower part and the C. rhomboidea Beds at the upper part. In this case horizon IV from Kamyshburun, assigned by Andrusov (1923) to the Bosphorian, is to come back to the Portaferrian.

At present, in the Euxinic Basin, the Pontian is subdivided into Novorossian, Portaferrian and Bosphorian, the meaning of the first substage being obviously different from that proposed by Andrusov (1923).

In the Dacic Basin, the Pontian is subdivided like in the Euxinic Basin, the only difference being that for the Lower Pontian the term Odessian is used, which corresponds to the beds with *Eupatorina littoralis* and *Congeria rumana*. The Portaferrian includes the beds with *C. subrhomboidea*, *C. rhomboidea* and the lower part of the beds with *Prosodacnomya sturi sabbae*, and the Bosphorian comprises the upper part of the beds with *P. sturi sabbae*, the beds with *Phyllocardium planum planum*, *Parapachydacna*, *Pontalmyra*, *Dreisenomya*, etc. (Andreescu, 1975; 1977; 1979; Motaş and Papaianopol, 1977; Marinescu, 1975, etc.).

In spite of the diversification of the Pontian lithofacies, the correlation of the Pontian, as well as of its three substages in the Dacic and Euxinic basins is quite easy due to the similar faunal content as a result of the permanent paleogeographic connections between the two basins.

The Pontian deposits studied in paleomagnetic respect, in the section of the Sărățelu Valley (Berca) in the north-eastern part of the Dacic Basin, pointed out only reversed polarities; it made Cepalîga (1980) assign the entire sequence to the lower part of the Gilbert epoch, ranging from 4.9 to 5.2 m.y (Plate I). Reversed polarities were also rendered evident by the paleomagnetic analyses carried out by Pevzner and Cikovani (1978) on the Pontian deposits in the Kerci-Taman region. Semenenko and Pevzner (1979) assigned the above-mentioned interval to epoch 6 of reversed magnetism (about 6.3 – 5.7 m.y.).

In these circumstances we consider that the paleomagnetic results obtained up to now on the Pontian deposits of the Dacic Basin have to be regarded as preliminary, further details will clarify the discrepancies concerning their calibration.

Dacian

The Dacian stage, proposed by Teisseyre (1907), redefined by Andreescu (1972), includes beds with fauna of transition from the Pontian to the Romanian. In the Dacic Basin, during the Dacian the evolutive lines of the limnocardiids and congerias end; it is the time interval when the beds with *Pachydacna* (= Lower Dacian = Getian) (Macarovici et al., 1965; redef. Eberzin et al., 1966) and the beds with *Pachyprionopleura* (= "*Psilodon*") ex gr. *haueri* and *P.* ex gr. *neumayri* (= Upper Dacian = Parscovian) (Andreescu, 1972) have been deposited.

The Lower Dacian fossil assemblages are generally unitary in the Dacic Basin. Within them there predominate the limnocardiids (*Pachydacna mirabilis*, *P. natella*, *Parapachydacna cobalcescui*, *P. serena*, *Prosodacna* ex gr. *longiuscula*, *P. semisulcata*, *Stylodacna heberti*, *Zamphiridacna cucestiensis*, *Limnodacna ukrainica*, *Dacicardium rumanum*, *Tauricardium olteniae*, *Pachyprionopleura munieri*, etc.), beside which congerias and fresh-water mollusca are to be found. Among the fresh-water mollusca an important part is played by unionids (*Rumanunio rumanus*, *Psilunio* ex gr. *neumayri*, *Potomida* ex gr. *psilodonta*, etc.) and viviparids (*Viviparus argesiensis*, *V.* ex gr. *rumanus*, *V. pannonicus*, etc.). All the mentioned assemblages belong to the Pachydacna mirabilis Zone (Andreescu, 1978, 1979).

The Parscovian faunal assemblages are dominated, in the area of the Dacian stratotype (Slănicu de Buzău Valley, in the south-eastern part of the Dacic Basin), by "psilodonts" evolved from the haueri and neumayri groups, occurring beside *Horiodacna rumana*, *Prosodacnomya sturi sturi*, *P. stenopleura*, *Zamphiridacna zamphiri*, *Limnodacna rumana*, *Congeria mirabilis*, *C. turgidopsis*, *C. caucasica*, *C. prahovensis*, *Dreissena* div. sp., *Rumanunio rumanus*, *Viviparus rumanus*, etc.

In the western part of the Dacic Basin the Upper Dacian faunal complexes are characterized by the participation of numerous fresh-water mollusca (*Psilunio neumayri*, *Potomida recurvus*, *P. saratae*, *R. rumanus*, *Viviparus rumanus*, *V. suessi*, etc.) beside which there occur: *Congeria turgidopsis*, *C. caucasica*, *C.* ex gr. *rhodonica*, *Horiodacna rumana*, *P. stenopleura*, *P. sturi sturi*, *Pseudocatillus donacoides*, *Parapachydacna* div. sp., etc. The Parscovian fossil assemblages belong to the P. haueri-P. neumayri zone (Andreescu, 1979).

Kimmerian

In the Euxinic Basin, the Kimmerian proposed by Andrusov (1907) occupies a time interval comparable with the Dacian stage, occurring between the Pontian and the Kujalnikian (=Pakveşian) or between the Pontian and the Egrissian (Taktakisvili, 1978).

At present, the Kimmerian is subdivided into three substages: Azovian, Kamyshburunian and Panticapean; however, recent researches

(Semenenko, 1975; Semenenko, Liuleva, 1976; Taktakisvili, 1978) indicate that the Panticapean is an equivalent of the Kamyshburunian.

The direct correlation of the Dacian and the Kimmerian is more difficult particularly as regards the upper part, where there occur several differentiations in the Parscovian and the Kamyshburunian faunal assemblages as a result of the periods of breaking of the connections between the Dacic and the Euxinic basins. Consequently, endemic faunas occur, e.g. the highly specialized *Prosodacna* belonging to the P. macrodon groups, typical of the Kamyshburunian-Panticapean, have not reached the Dacic Basin, and the evolved "psilodonts" have not been found in the Euxinic Basin.

There are, however, common faunal elements which make it possible to correlate the Parscovian with the Kamyshburunian-Panticapean. Thus, in Preazovia, Eberzin and Semenenko (1966) have detected the species *Horiodacna rumana*, and the Duab Beds, in the west of the Georgian Soviet Socialist Republic, contain several congerias (*C. caucasica*, *C. mirabilis*, etc.), limnocardiids (*P. donacoides*) and unionids, occurring in the Upper Dacian, as well.

With a view to rendering evident some new possibilities of correlation the Dacian deposits have been investigated in paleomagnetic respect on some profiles in the east of the Dacic Basin. The analyses have indicated that: the Romanian-Dacian boundary (= between the Viviparus bifarcinatus and P. haueri-P. neumayri zones) coincides with the Cochiti event (about 3.8 m.y.) in the Gilbert epoch; the Upper Dacian-Lower Dacian boundary is at the upper part of another episode of normal polarity, possibly Nunivak (about 4.05 m.y.) in the Gilbert epoch; the lower boundary of the Dacian corresponds to an interval of normal polarity considered by Cepalîga (1980) as the Tvera event (about 4.85 m.y.) in the lower part of the Gilbert epoch (Plate II).

The paleomagnetic analysis of the Kimmerian deposits in the Kerci-Taman region led Semenenko and Pevzner (1979) to the conclusion that the Kimmerian corresponds to the Gilbert epoch and to epoch 5, being included in a time interval (3.4 — 5.8 m.y.) wider than that considered by Cepalîga (1980).

Romanian

The Romanian stage has been proposed by Krejci-Graf (1932) and redefined by Mihăilă (1969) and Andreescu (1972). In the Dacic Basin, this stage represents the chronostratigraphic unit delimited by the Dacian stage at the lower part and the Lower Pleistocene at the upper part.

At the stratotype, located on the Slănicu de Buzău Valley, in the east of the Dacic Basin, the Romanian includes the beds with smooth unionids and *Viviparus bifarcinatus*, the beds with sculptured unionids, as well as a pile of nonfossiliferous deposits representing a part of the Cîndești Beds.

In the same stratigraphic area the Romanian deposits have been assigned, from the lithostratigraphic viewpoint, to the Rîmna Formation and the Cîndești Formation (G h e n e a et al., 1981).

A n d r e e s c u (1978, 1979) has proposed, on the basis of a rich fauna of fresh-water mollusca, a biozonation of the Romanian, separating three zones : Viviparus bifarcinatus (with two subzones : Jazkoa sturdzae and Psilunio sibinensis), Rugunio lenticularis (with four subzones : Rytia brandzae, Pristinunio pristinus, Rytia bielzi, and Cuneopsidea iconomiamus) and Ebersininaia milcovensis-Rugunio riphaei (with four subzones : E. milcovensis, E. geometrica, Unio kijalnicensis, and Bogatschevia tamanensis, R. riphaei).

The Romanian formations in the Dacic Basin, as well as their equivalents in the Euxinic Basin, contain numerous remains of fossil mammals which allowed the correlation with other deposits of the same age in Europe.

Thus, in the north-eastern part of the Dacic Basin, V. bifarcinatus Zone corresponds to the deposits with mammal fauna of Berești and Mălușteni (= Berești Beds, G h e n e a, 1968), equivalent with the lower part of the "Moldavian Complex" (Kuchiurgan) in the south of the Soviet Union (G h e n e a et al., 1981). In the interval of the R. lenticularis zone, in different sectors of the Dacic Basin remains of fossil mammals have been reported from the Cîndești Formation at Tulucești, Cernătești, Covrigi, etc., correlable with the upper part (Kotlovina) of the "Moldavian Complex" in the south of the Soviet Union and with the Lower Villafranchian in the south-western Europe (G h e n e a et al., 1981).

Corresponding to E. milcovensis – R. riphaei Zone, mammal faunas have been reported (F e r u et al., 1978) from Milcovu din Vale, Slatina, Cherlești, equivalent with the Haprovian Complex in the south of the Soviet Union and with the Middle Villafranchian (Roccaneyra,St. Vallier, Olivola, etc.) in the south-east of Europe (G h e n e a et al., 1981).

The data referring to the biostratigraphy of the Romanian point out that this stage corresponds almost entirely to the Akchagylian stage, most of the zones and subzones distinguished in the Dacic Basin (A n d r e e s c u, 1979) occurring also in the Euxinic-Caspian Realm.

The paleomagnetic analyses carried out on samples from the stratotype section of the Romanian as well as from other sections (Table 1), indicated that on the Slănicu de Buzău Valley the base of the Romanian is situated in the upper part of the Gilbert epoch, corresponding to the Cochiti event ($\approx$ 3.8 m.y.); the Lower Romanian-Middle Romanian boundary coincides with the limit between Gilbert epoch and Gauss epoch ($\approx$ 3.4 m.y.) (G h e n e a et al., 1981). The Middle Romanian-Upper Romanian boundary, that is the limit between R. lenticularis and E. milcovensis – R. riphaei zones in the profile on the Olt Valley, at Slatina, is located in the upper part of the Gauss epoch, for which an age of about 2.7 – 2.6 m.y. is estimated (A n d r e e s c u et al., 1981). In the same profile, the limit between the last zone of the Romanian and the first zone of the Quaternary – *Unio apscheronicus* – corresponds to an interval of normal polarity of the Matuyama epoch, possibly the Olduvai event (about 1.7 – 1.9 m.y) (A n d r e e s c u et al., 1981).

From the above-mentioned data there results that the Romanian stage is a little older than the Akchagylian stage ; in this case the lower boundary of the Romanian would correspond to the Kujalnikian base (= Pakveshian). The upper limit of the Romanian corresponds to the Akchagylian-Apsheronian boundary in the south of the Soviet Union, marked by the same zone — U. apscheronicus (C e p a l î g a, 1972 ; N i k i f o r o v a et al., 1976).

3. Correlation of the Upper Pliocene and Lower Pleistocene Lithostratigraphic Units

The Cîndeşti Formation (= Cîndeşti Beds ; M r a z e c and T e i s s e y r e , 1901). In the Dacic Basin a large area of the Carpathian Foredeep, starting from the upper part of the Middle Romanian, there developed a continental formation with molasse character, represented by alternations of pebbles, conglomerates, sands and clays. The sedimentological characters are determined by the prevalence of the pebble and conglomerate banks (Argeş, Dîmboviţa, Prahova, Buzău, Rîmnic valleys) at the upper part. Towards the lower part there is a gradual decrease of the thickness of the pebble banks concomitantly with the reducing of the diameter of the pebbles. The thicknesses frequently reach hundreds of meters, and in the Carpathian Bend Zone they can exceed 1,000 m. The stratigraphic position of the Cîndeşti Formation can hardly be estimated because of the extremely reduced paleontologic content. However, the stratigraphic bulk has been estimated on the basis of paleomagnetic analyses on certain sections characterized by continuity of sedimentation in the Romanian-Lower Pleistocene time span. Thus, at Beceni and Pleşcoi the coarse intercalations constituting the Cîndeşti Formation are individualized beginning with the level of 2.7 — 2.5 m.y. (Plate I).

As regards the comparison with the Euxinic Basin, the time interval corresponding to the deposition of the Cîndeşti Formation corresponds to a great extent to the Middle and Upper Akchagylian (Table 2).

The Frăteşti Formation. In the Lower Pleistocene, in the central and southern part of the Dacic Basin there deposited alluvial accumulations consisting of a sequence of pebbles and sands, separated by clayey intercalations. In the western part (west of the Argeş River), such deposits are found in the constitution of the interfluves, pebbles and rolled boulders, with thicknesses of tens of meters. In the eastern part of the Dacic Basin, the processes of subsidence determined the sinking of these deposits under more recent formations. Numerous drillings in the area of Bucharest indicated the presence of such deposits at depths varying between 100 — 300 m, with greater thicknesses (100 — 200 m), and disposed in three banks of pebbles separated by clayey intercalations.

The Frăteşti Formation was first described under the name of "Frăteşti Beds" (L i t e a n u, 1953), the author referring to the lithologic constitution and the paleontologic content. In respect of the paleontologic content, it is to be mentioned the frequency of remains of *Archidiskodon meridionalis* reported from several points of the Dacic Basin. There are also reported remains of *Libralces gallicus* and *Dicerorhinus etruscus*, on the basis of which the period of alluviation of the paleofluve, which generated the Frătești Formation, was assigned to the Lower Pleistocene.

The Lower Pleistocene in this region occurs better characterized by the existence of rich faunas of mammals. The best-known deposit is to be found in the Olteț Valley, at Tetoiu (Bugiulești). The mammal assemblage is characterized by the presence of a single elephant *(Archidiskodon meridionalis)* beside *Paradolicophitecus geticus*, *Allohipus stenonis*, *Dicerorhinus etruscus*, *Nyctereutes*, *Trogontherium boisvilleti*, *Megalovis*, *Leptobos*, *Praealces gallicus*, *Canis etruscus* (Samson and Rădulescu, 1972). The mentioned fauna assemblage can be correlated with the Odessian Complex in the Euxinic-Caspian Realm and the Unionid biozones in the Olt Valley allow their equivalence with the lower part of the Lower Apsheronian (Plate II).

The Uzunu Beds. In some points in the southwestern area of the Dacic Basin (Uzunu, Stoenești), at the upper part of the "Frătești Formation", one can observe sandy deposits with marly clayey intercalations with a molluscan fauna, represented by: *Bogatschevia sturi*, *Corbicula fluminalis*, *Pisidium amnicum*, *Sphaeridium corneum*, *Viviparus craiovensis*, *V. diluvianus*, *V. romaloi*, *V. geticus*, *Litoglyphus naticoides*, *Planorbis planorbis*, *Anisius (Spiralina) vortex*, etc.

The Uzunu Beds with *Bogatschevia sturi* correspond to the Boshernitian horizon in the Euxinic Basin (Lower Pleistocene).

The Coconi Beds (Marly Complex). In the eastern area of the Dacic Basin, a pile of marls and clays with thin intercalations of fine sands overlies the Frătești Formation, and in some sectors lies concordantly over the Uzunu Beds. They have been described first in the Dacic Basin under the denomination of "Marly Complex" (Liteanu, 1953). This horizon has been found in all the drillings carried out in the area of Bucharest and in other points towards the eastern part of the Dacic Basin. They outcrop on the Mostiștei Valley as a sequence of grey marls with rare lenticular intercalations of marly fine sands.

The paleontologic content is characterized by the presence of species of *Corbicula* cf. *fluminalis*, *Valvata piscinalis*, *V. sibinensis sibinensis*, *Pisidium priscum*, *P. amnicum*, *P. clessini*, *Sphaeridium rivicola*, *Planorbis planorbis*, *P. corneus*, *Viviparus diluvianus*, etc.

4. Conclusions

The biostratigraphic data known up to now in the two realms of sedimentation — Dacic and Euxinic — have been correlated for the first time with magnetostratigraphic researches. A more precise dating and correlation of the Pliocene and Pleistocene formations have been achieved.

For the Upper Pliocene and the Lower Pleistocene the results have been in agreement with the general image on the stratigraphic volume of the mentioned intervals; however, the evidence referring to the Lower Pliocene seems to be surprising.

Thus, if the boundary between the Phyllocardium plannum plannum Zone (Upper Pontian) and the Pachydacna mirabilis Zone (Lower Dacian — Lower Kimmerian) coincides with the Tvera event (~4,9 m. y.) of the Gilbert epoch and the Pontian sequence indicates even for the levels only reversed polarity (assigned to the Gilbert epoch) a very large volume of sediments (the Pontian) have been deposited during a time interval ranging approximately between 4.9 — 5.2 m.y.

There are necessary minute studies on profiles with continuity of sedimentation in the Meotian-Dacian.

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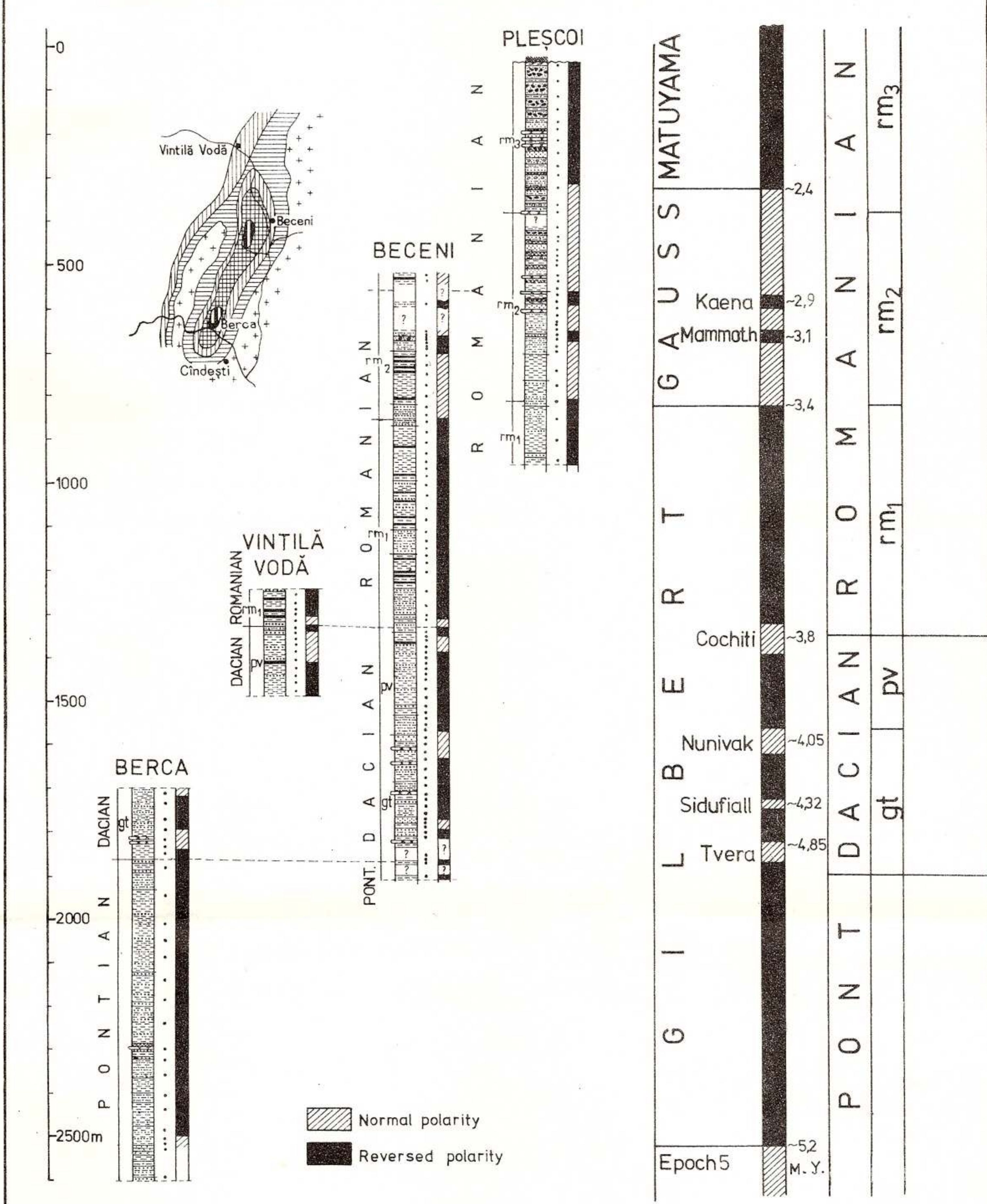
QUESTIONS

P. Stevanović. The title of the paper must be modified as it deals only with the late Pliocene, the beginning of the Dacian, not with the whole Pliocene.

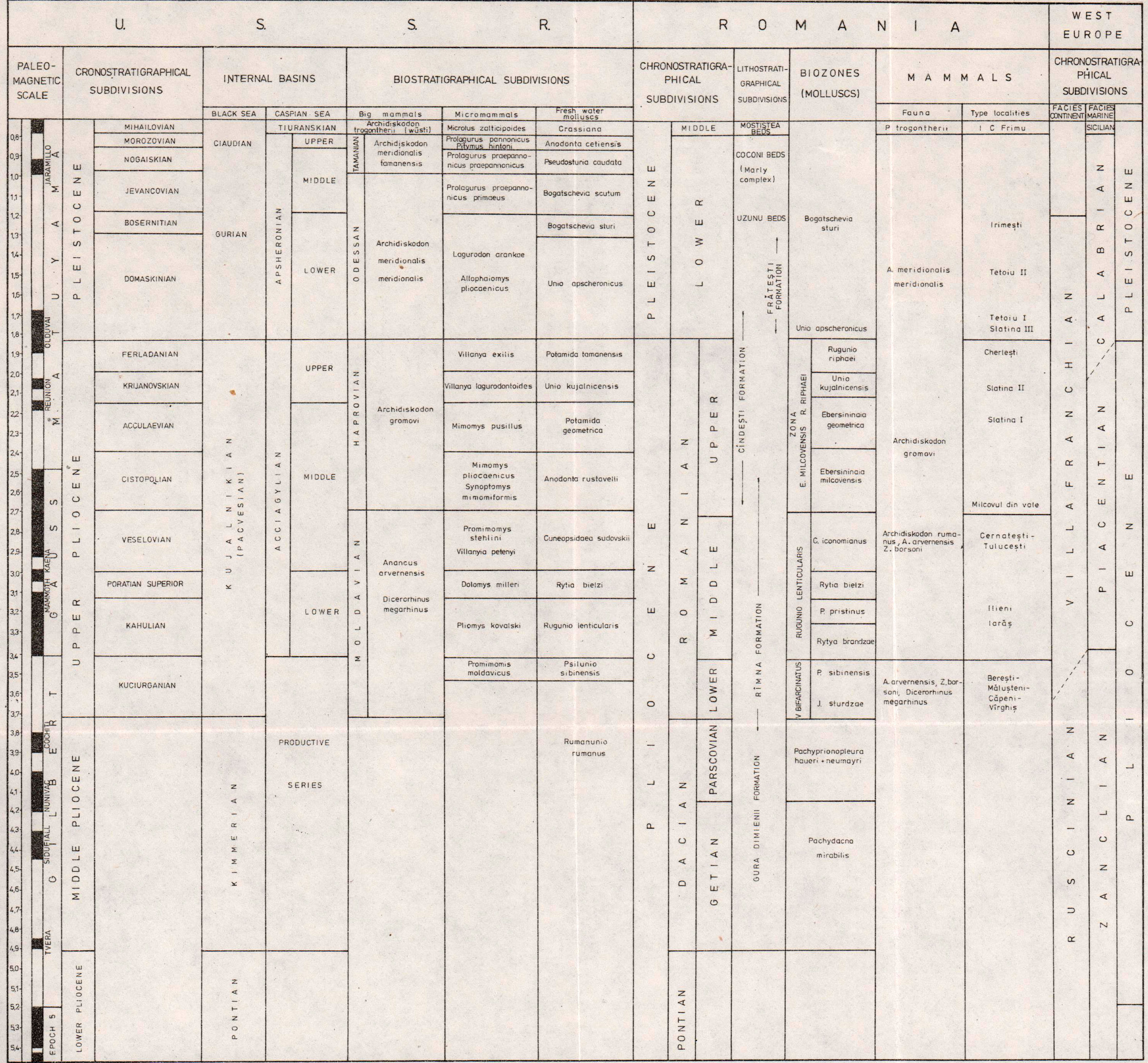
Answer: The title of the paper must not be changed as we consider that the Pliocene starts concomitantly with the Dacian, not with the Pontian or the Meotian as previously accepted.

SELECTED SECTIONS PALEOMAGNETICALLY INVESTIGATED



Imprim. Atel. Inst. Geol. Geof.

CORRELATION DIAGRAM OF THE PLIOCENE AND PLEISTOCENE DEPOSITS FROM THE SOUTHERN PART OF THE SOWIET UNION AND ROMANIA



Imprim. Atel. Inst. Geol. Geof.

BIOCHRONOLOGY AND CHRONOSTRATIGRAPHY OF THE UPPER PLIOCENE AND LOWER PLEISTOCENE IN THE DACIC BASIN [1]

BY

ION ANDREESCU [2]

A biochronologic scale based on mollusca has been proposed (Andreescu, 1978, 1979) for the Sarmatian-Lower Pleistocene interval in the Dacic Basin. This scale includes 14 biozones (12 for the Upper Neogene and 2 for the Lower Pleistocene) marked $NSM_1 - NSM_{12}$ (NSM = Upper Neogene Mollusca) and QM_1, QM_2 (Quaternary Mollusca).

Zones NSM_1 — Abra reflexa, NSM_2 — Cryptomactra, NSM_3 — Sarmatimactra podolica — S. fabreana, and NSM_4 — S. bulgarica belong to the Sarmatian: zones NSM_5 — Abra tellinoides and NSM_6 — Congeria panticapaea tournoueri-Andrusoviconcha novorossica navicula are referred to the Meotian; zone NSM_7 — Eupatorina littoralis and the first two subzones (NSM_{8a} and NSM_{8b}) of zones NSM_8 — Prosodacnomya sturi sabbae characterize the Pontian; subzone NSM_{8c} — Pachydacna mirabilis and zone NSM_9 — Pachyprionopleura haueri — P. neumayri belong to the Dacian, and the Romanian includes zones NSM_{10} — Viviparus bifarcinatus, NSM_{11} — Rugunio lenticularis and NSM_{12} — Ebersininaia milcovensis-Rugunio riphaei.

General Considerations

For the making up of the Romanian biochronologic scale we have referred to the main fossil groups which may provide undoubtful biostratigraphic markers. These fossil groups are first of all represented by fresh-water molluscan faunas, among which unionids and viviparids are to be mentioned.

In the calibration of the molluscan zones we have taken into account the paleomagnetic data, the result of our collaboration with geologists from the Institute of Geology of the Academy of Sciences of the USSR-Moskow (Andreescu et al., 1981; Ghenea et al., 1982).

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[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, 1981 September 8 – 12, Bucharest, Romania.

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The study of the Romanian fresh-water molluscan assemblages pointed out the possibility to outline three main stages in their evolution and succession.

The first stage coincides with the time interval formerly called Lower Romanian (Andreescu, 1972). It includes a fauna of unionids and viviparids relatively poor in new genera and species.

The second stage is marked by the outburst of the "Levantine fauna" of unionids and sculptured viviparids. The succession of different taxa is rapid, spectacular occurrences and extinctions being recorded. The proliferation of fresh-water molluscan fauna occurs concurrently with the development and considerable spreading of Lower Villafranchian mammal fauna (Perrier-Etouaires type), unlike the previous stage in which the mammal faunas preserve numerous elements of the advanced Ruscinian fauna. In the Villafranchian, among the large mammals mention should be made of the appearance and spreading (in Europa) of *Archidiskodon rumanus*, genus *Equus*, genus *Bos* (= *Leptobos*), namely the group ELE (Tobien, 1970). Concurrently, among the small mammals the first appearance of genus *Lemmus* within the assemblage with *Mimomys polonicus* is recorded in Central and South Europe (Chaline, 1977).

The third stage of evolution of the Romanian fresh-water molluscan faunas is characterized by the extinction of many genera and species of the previous stage concurrently with the appearance of new genera and species; however, one may say that on the whole the faunas become poorer. In this interval the boreal elements often alternate with the termophile ones. At the same time in Europe, within the Middle Villafranchian mammal faunas, several new species appear *(Archidiskodon gromovi, Croizetoceros ramosus, Pliomys kretzoi, Mimomys rex, M. pliocaenicus,* etc.) beside *Anancus arvernensis, Equus stenonis, Dicerorhinus etruscus, Leptobos etruscus*, etc. (Tobien, 1970; Ballesio et al., 1973; Kowalski, 1973; Chaline, 1977, etc.).

Succession of the Biozones

The three above-mentioned stages of evolution of the Romanian molluscan faunas mostly correspond to three biozones which may be regarded as superzones (Plate I).

Zone NSM_{10} — Viviparus bifarcinatus (Andreescu, 1978) is defined by the occurrence interval of index species, limited at the upper part by the development of sculptured unionids. The fossil assemblage of the zone is dominated by smooth-shelled unionids, dreissenes, carenate vivipars, melanopsides, neritines, etc. In the Buzău—Milcov area, in the eastern part of the Dacic Basin, in the lower part of the zone NSM_{10}, species *Prosodacnomya sturi sturi* (Cob.) is found, endured from zone NSM_9.

Considering the succession of some marker taxa, zone NSM_{10} has been subdivided into two subzones:

NSM_{10a} — *Jazkoa sturdzae*

NSM_{10b} — *Psilunio sibiensis*

Zone NSM_{10} has a wide spreading within the Paratethys Realm; it can be delimited both in the Pannonian Basin (V. bifarcinatus + V. stricturatus zones) and in the whole Dacic Basin. In the Euxinic Basin subzone $NSM0_{10a}$ has its equivalent in the upper part of the Duab Formation. Subzone NSM_{10b} is to be found, in the northern part of the Euxinic Basin, in the lower part of the Poratian (= P. sibinensis — Potomida sandbergeri Zone) (Nikiforova et al., 1976).

Zone NSM_{10} spans the time interval corresponding to the Proto-Villafranchian mammal faunas from Berești and Mălușteni, equivalent with the upper part of the "Moldavian (Kuchurgan) Complex" in the south of the USSR and with the mammal deposits at Wéze, Ivanovce, Csarnota, Wölfersheim, Perpignan, Sète, etc. in Central and Western Europe (= zone MN_{15}, Mein, 1975) (Plate I).

Zone NSM_{11} — Rugunio lenticularis (Andreescu, 1978) corresponds to the interval during which, in the Paratethys, the sculptured unionids and the ornamented vivipares develop luxuriantly; beside them smooth unionids can also be observed. Most of the taxa of this zone have a range equal to that of zone NSM_{11}. Other taxa, particularly representatives of the genera *Potomida, Psilunio, Cyclopotomida*, some vivipars, etc., are also to be found in zone NSM_{10}. Numerous species of unionids and vivipars proper to zone NSM_{11} have a reduced range that allowed the separation of four subzones (Andreescu, 1979) (Plate I):

NSM_{11a} — *Rytia brandzae*
NSM_{11b} — *Pristinunio pristinus*
NSM_{11c} — *Rytia bielzi*
NSM_{11d} — *Cuneopsidea iconomianus*

Subzone NSM_{11a} corresponds to Viviparus nothus Zone and possibly to the lower part of V. sturi Zone in the Pannonian Basin and to Rugunio lenticularis (partim) Zone in the Euxino-Caspian Realm, located in the upper part of the Lower Poratian (Nikiforova et al., 1976; Andreescu, 1979).

In the Pannonian Basin the equivalent of subzone NSM_{11b} is represented by the middle part, possibly the upper part too, of V. sturi-V. ornatus Zone. In the Euxinic Basin subzone NSM_{11b} characterizes the upper part of the Poratian; it lies between R. lenticularis and R. bielzi zones (Tchepalyga, 1972; Nikiforova et al., 1976).

Subzone NSM_{11c} is characterized by the scarcity of species found in the previous subzones, the appearance of new species or the predominance of taxa with a previous reduced frequency (Andreescu, 1981; Andreescu et al., 1981). The faunal assemblage of subzone NSM_{11c} points to a possible correlation with the lower part of Viviparus hoernesi Zone, eventually with the final part of V. sturi-V. ornatus Zone in the Pannonian Basin and R. bielzi Zone in the Euxino-Caspian Realm.

Subzone NSM_{11d} marks the time-interval in which most species of unionids and viviparids, typical of the former subzones, occur sporadically or become extinct. It is to be mentioned that even the zone marker species *(R. lenticularis)* is recorded only in places. On the other hand the representatives of boreal unionids of the Unio group occur more frequently. The same process of extinction affects most of the sculptured viviparids.

On the whole, zone NSM_{11} can be regarded as an equivalent of the R. lenticularis + R. bielzi + Cuneopsidea sudovskii zones separated by T c h e p a l y g a (1972) within the Upper Pliocene unionid fauna in the USSR.

In the interval of zone NSM_{11}, in different areas of the Dacic Basin, fossil mammal remains have been recorded from the Cindești Formation. The best known mammal faunas are those from Tulucești and Cernătești, which have been correlated (S a m s o n, R ă d u l e s c u, 1973 ; S a m - s o n, 1976) with the Upper "Moldavian Complex" in the south of the USSR, corresponding to the Vialètte faunas, recently redated at 3.2 — — 2.6 m.y. (=Lower Villafranchian) (B a n d e t et al., 1978). Consequently one may admit the equivalence of zone NSM_{11} with zone MN_{16} (M e i n, 1975, 1979), within which the start of the Mimomys polonicus — M. pliocaenicus phyletic-line is recorded simultaneously with the beginning of the Villafranchian (C h a l i n e, 1977) (Plate I).

Zone NSM_{12} — Ebersininaia milcovensis-Rugunio riphaei (A n d r e- e s c u, 1978) is defined by the occurrence of the two marker species : the former prevails in the lower part and the latter—in the upper part. The successions of molluscan assemblages allowed the separation of four subzones (A n d r e e s c u , 1978, 1979) within the zone NSM_{12} (Table 1) :

NSM_{12a} — *Ebersininaia milcovensis-Viviparus turritus*
NSM_{12b} — *Ebersininaia geometrica-Bogatschevia pretamanensis*
NSM_{12c} — *Unio kujalnicensis*
NSM_{12d} — *Rugunio riphaei-Bogatschevia tamanensis*

Subzone NSM_{12a} has its equivalent in the Cistopol, Simbughino horizons (=Anodonta rustavellii Zone) of the Lower Akchagylian (Upper part)— Middle Akchagylian (lower part) (N i k i f o r o v a et al., 1976; I a h i m o v i c i et al., 1977 ; A n d r e e s c u, 1979).

To a large extent (?probably entirely) subzone NSM_{12b} correlates with E. geometrica Zone (= Sultanaevo Complex) of the Middle Akchagylian (= Akkuleaevo Horizon) in the Euxino-Caspian Realm (N i k i- f o r o v a et al., 1976). At present it is difficult to correlate subzone NSM_{12b} with a certain zone of the "Paludina Beds" in the Pannonian Basin. However, a comparison with the terminal part of the Viviparus hoernesi Zone and the lower part of the V. zelebori Zone is not out of question.

Subzone NSM_{12c} (T c h e p a l y g a , 1972) is defined by the interval during which the faunal assemblages become poorer as compared to the former subzone, pointing out the covering of a period with a cold climate. The assemblage is dominated by boreal unionids (*Unio kujalnicensis* M a n g., *Unio* sp.), smooth vivipars (*Viviparus subconcinnus, V. fasciatus)*, as well as small gastropoda of the genera *Fagotia, Lithoglyphus*, *Melanopsis*, *Theodoxus*, etc.

The abundance of specimens of *U. kujalnicensis*, beside the occurrence of species *Dolomys milleri* (= $Slatina_1$ level, F e r u et al., 1978) made it possible to equate subzone NSM_{12c} to U. kujalnicensis Zone (= Krijanovka Complex, upper part =Liventzovka Lower Horizon = = lower part of the Haprovian Complex) of the Kujalnik at Odessa (= Upper Akchagylian) (N i k i f o r o v a et al., 1976 ; F e r u et al. 1979 ; A n d r e e s c u et al., 1981).

Subzone NSM_{12d} corresponds to the occurrence interval of the index species, limited at the upper part by the faunal assemblage of zone QM_1.

The interval corresponding to zone NSM_{12} overlaps the interval including the species *Archidiskodon gromovi*, mostly corresponding to the Haprovian Complex in the south of the USSR (=Upper Akchagylian) and Middle Villafranchian in Central and South-West Europe (Beremend$_5$, St. Vallier, Le Coupet, etc.) (= zone MN_{17} part., Mimomys pliocaenicus) (M e i n, 1975, 1979; F e r u et al., 1978, 1979) (Plate I).

Zone QM_1 — Unio apscheronicus (T c h e p a l y g a, in N i k i f o r o v a et al., 1976) is defined by the occurrence interval of the index species limited at the upper part by the occurrence of the species *Bogatschevia sturi*.

The faunal assemblage, dominated by reofile (Unio pictorum type) and stagnofile (Euphrata type) boreal unionids, indicates a severe cooling of the climate, possibly corresponding to the Donau glaciation. The molluscan assemblage allows the correlation of zone QM_1 with the Domashkino Complex (Lower Apsheronian) as well as with the mammal Odessan Complex, lately considered in the base of the Pleistocene (N i k i f o r o v a et al., 1976).

Zone QM_2 — Bogatschevia sturi (T c h e p a l y g a, 1972) corresponds to the occurrence interval of the index species which, in the Dacic Basin, in the Euxino-Caspian Realm and in the Pannonian Basin seems to have a limited range (about 1.57 — 1.3 m.y.). Meanwhile, in the Dacic Basin, the upper limit of the zone has not been established yet.

Magnetostratigraphy

At the stratotype of the Dacian and Romanian stages, on the Slănicu de Buzău Valley, village of Beceni, Buzău District, the boundary between the two mentioned stages coincides with the limit between zone NSM_9 and NSM_{10}; it is situated at the level of normal polarity Cochiti event (3.7 — 3.8 m.y.) of the Gilbert paleomagnetic epoch (A n d r e e s c u, 1979; G h e n e a, et al., 1982) (Plates I, II).

In the section on Slănicu de Buzău Valley, the boundary between zones NSM_{10} and NSM_{11} coincides with the Gilbert-Gauss limit (3.32 m.y.) (Tables 1, 2). At Podari (Jiu Valley) subzones NSM_{11a} and NSM_{11b} are situated in the lower part of the Gauss epoch (Plate II). For both subzones a time span of cca 0.3 m.y. can be estimated.

In the type section of the Romanian the base of subzone NSM_{11d} coincides with the Kaena event of the Gauss epoch (cca 2.8 — 2.9 m.y.). (Plate II). (A n d r e e s c u, 1979; G h e n e a et al., 1982). By extrapolation it may be inferred that subzone NSM_{11c} might approximately correspond to the Mammoth event and to the normal polarity interval prior to the Kaena event (cca 0.15 — 0.18 m.y. from the Gauss epoch).

In the Slatina and Milcovu din Vale (Olt Valley) sections subzone NSM_{12a} is located in an interval of normal polarity which, by correlation with the Simbughino Complex (I a h i m o v i c i et al., 1977), might correspond to the upper part of the Gauss epoch; therefore, the beginning of zone NSM_{12} may be estimated at about 2.6 — 2.7 m.y. (A n d r e e s c u et al., 1981) (Plates I, II). Subzones NSM_{12b-d} are found in the lower part of the Matuyama epoch of reversed polarity. It is possible that the Reunion events (normal polarity), which have not been pointed out in the Slatina section, might correspond to a sedimentary gap in the base of subzone NSM_{12d} (Table 2). Subzone NSM_{12d}, by analogy with its equivalent in the Upper Akchagylian (Nikiforova et al., 1976), begins at about 2.0 m.y., and zone QM_1 in the base of the Apsheronian corresponds to the Olduvai event (cca 1.8 – 1.6 m.y.) of the Matuyama epoch (Andreescu et al., 1981) (Plate II).

Chronostratigraphy

The Romanian has been previously (Andreescu, 1972) subdivided into two parts: the Lower Romanian – corresponding to the "beds with smooth unionids" – and the Upper Romanian – including the "beds with sculptured unionids".

Considering the evolution of different groups of organisms (mollusca, mammals, etc.) on the one hand, and the evolution of the sedimentary basin, on the other hand, we have subdivided subsequently the Romanian into three substages (Andreescu, 1979, 1981; Andreescu, et al., 1981) (Plate I).

The Lower Romanian maintains the same sense as defined in 1972, corresponding now to zone NSM_{10}. As the classic area of formations development and of fauna occurrence, on the basis of which the substage has been delimited, is found at the Romanian stratotype, we have proposed (Andreescu, 1979) the name of Siensian (from the Daco-Getic tribes of the Siens, who lived in the East Carpathians Bend Zone) for the Lower Romanian.

Taking into account the mammal faunas of Berești, Mălușteni, Kuchurgan, etc. there results that the Siensian substage corresponds to the lower part of the Moldavian Complex (zone MN_{15}; Mein, 1975, 1979), belonging to the Upper Ruscinian (= "Proto-Villafranchian" auct.) (Plate I).

The Middle Romanian is defined by the interval of maximum proliferation of the "Levantine fauna" of sculptured unionids and ornamented viviparids, within which the occurrence of zone NSM_{11} is pointed out.

The most representative deposits yielding the unionid fauna of zone NSM_{11}, are to be found in the west of the Dacic Basin, the Jiu hydrographic basin, at Bucovăț, Podari, Crețești, Bîlta, etc., in the outskirts of Craiova. Consequently, we have proposed the name of Pelendavian (from the Dacic town Pelendava situated in this region) for the Middle Romanian substage (Andreescu, 1979).

In the Euxino-Caspian Realm, the Middle Romanian roughly coresponds to the Poration (Lower Akchagylian) and in the South-West European Realm it is mostly equivalent to the Lower Villafranchian (Plate I).

The Upper Romanian (for which the name Wallachian was proposed in 1979) is characterized by faunal assemblages clearly differentiated from those of the Middle Romanian. We may say that, at this level, the fresh-water molluscan faunas point to significant Euxino-Caspian affinities, unlike those of the Middle and Lower Romanian which are found mostly in the Pannonian Basin. From the climate point of view, the Wallachian is characterized by important oscillations, when the cold intervals alternate with the warmer ones (Plate I).

We have mentioned that the lower boundary of the Romanian is situated at cca 3.7—3.8 m.y. and the upper boundary, corresponding to the limit between zones NSM_{12} and QM_1 (Pliocene/Pleistocene boundary in the present acceptance), coincides with the Olduvai event (1.6 — 1.8 m.y.). Consequently the Romanian spans an interval of cca 20. m.y., out of which cca 0.5 m.y. belong to the Siensian, 0.7 m.y. to the Pelendavian, and 0.8 m.y. to the Wallachian.

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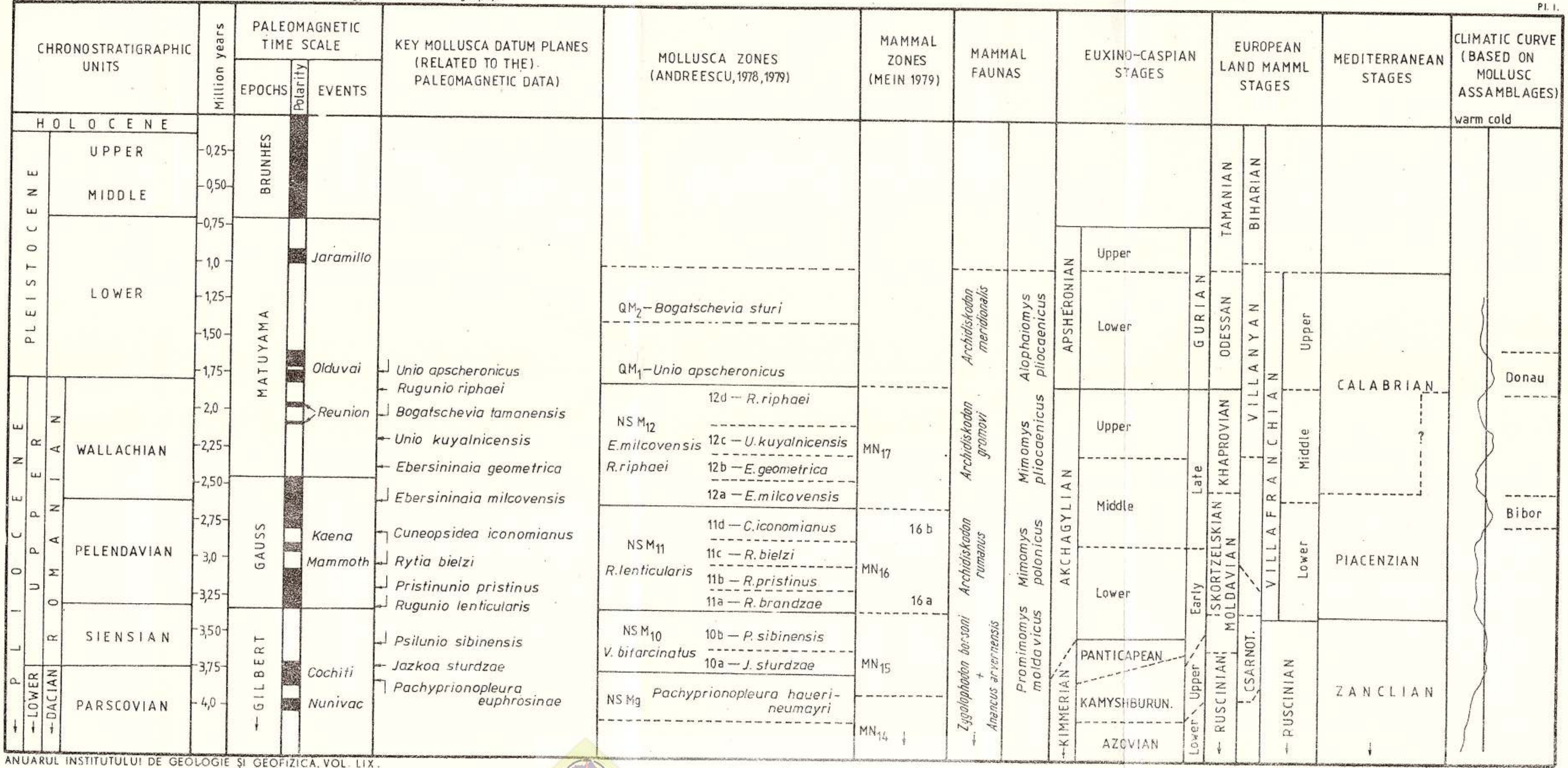
QUESTIONS

Krstić. Where from did you take the paleomagnetic samples – from outcrops or from boreholes ?

Answer: All the samples were taken from outcrops.

STRATIGRAPHIC SCHEME OF THE UPPER PLIOCENE AND LOWER PLEISTOCENE OF THE DACIC BASIN



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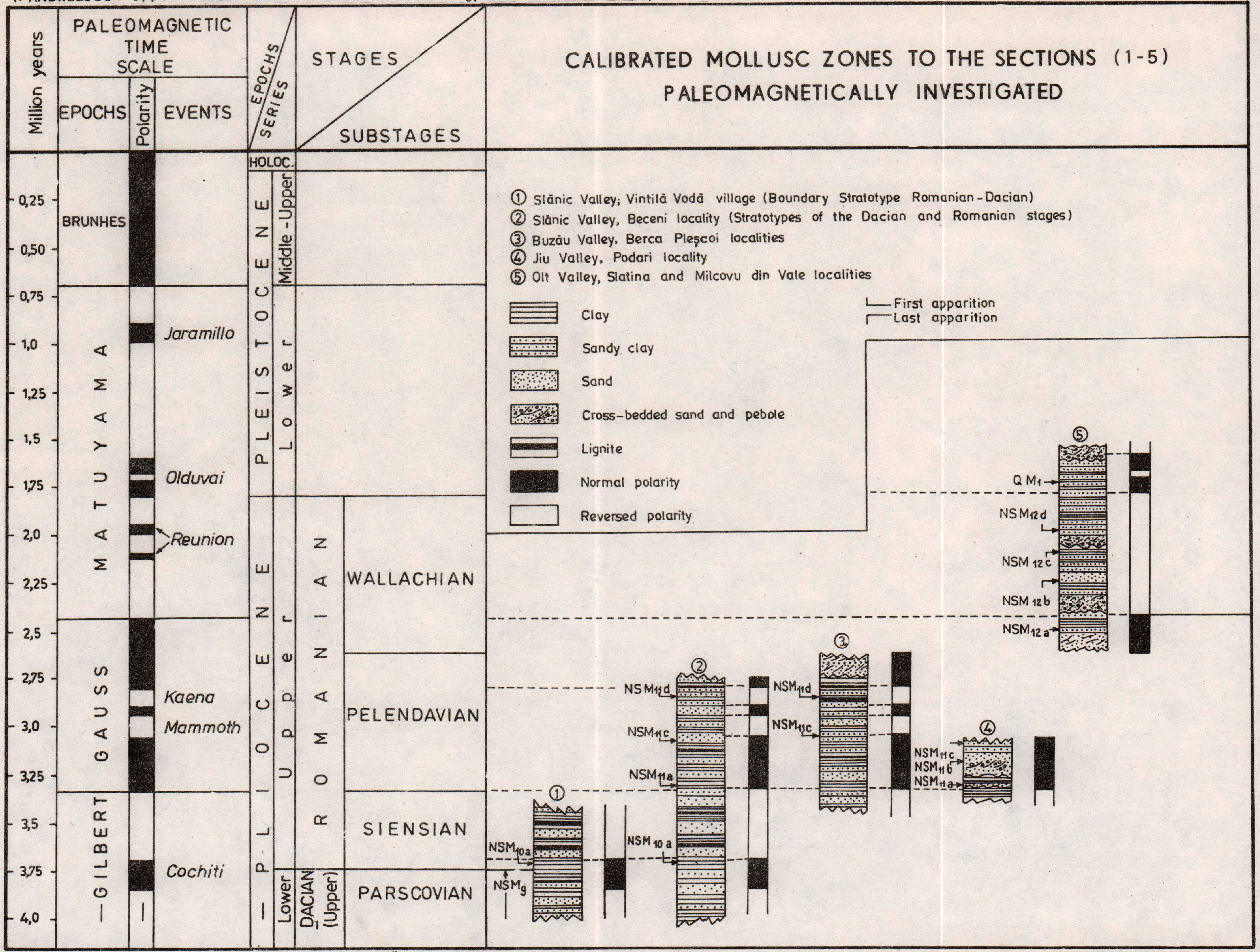

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SUCCESSION DES MAMMIFÈRES PLIO-PLÉISTOCÈNES DANS LE BASSIN DACIQUE (ROUMANIE)[1]

PAR

MIRCEA ULPIU FERU[2], CONSTANTIN RĂDULESCU[3], PETRE SAMSON[3]

Les recherches que les auteurs ont entreprises sur les formations fluvio-lacustres du Bassin Dacique depuis 1958, basées autant sur des nids fossilifères fouillés systématiquement que sur des pièces éparses dont la position stratigraphique a été repérée, permettent de fixer quelques moments importants dans le développement des principaux ensembles de Mammifères plio-pléistocènes en Roumanie.

Des associations de gros Mammifères ont été signalées dans la moyenne vallée de l'Olteț, à Tetoiu (= Bugiulești) et à Irimești (Necrasov et al., 1961; Samson, Rădulescu, 1973), ainsi que dans la zone Craiova-Filiași (Schovert et al., 1963) et dans l'interfluve Jiu-Motru (Feru et al., 1965). La connaissance des Micromammifères est plus récente, des associations de Rongeurs et d'Insectivores étant mises en évidence premièrement à Slatina, Cherleștii-Moșteni et Izvoru (Feru et al., 1978), ensuite à Podari, Milcovu din Vale et Drăgănești-Olt (Feru et al., 1979). Ajoutons qu'à Slatina, Milcovu din Vale et Izvoru, les recherches concernant les Mollusques associés aux Mammifères, ainsi que les datations paléomagnétiques ont apporté des précisions complémentaires (Andreesco et al., 1981).

D'une manière générale, nous avons choisi pour caractériser chaque association de Mammifères une localité type — la plus riche en espèces significatives — à laquelle nous avons réuni les découvertes équivalentes de la même zone (figure). Les principales corrélations que nous avons établies (tableau) se rapportent aux schémas de l'Union Soviétique (Nikiforova et al., 1976; Tchépalyga, 1980) des Pays Bas (Zagwijn, 1974), de la France (Heintz, 1970), de l'Italie (Ambrosetti et al., 1972; Azzaroli, 1977) et du Néogène méditerranéen (Mein, 1975).

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[2] Entreprise géologique de Prospections, Caransebeș 1, Bucarest.

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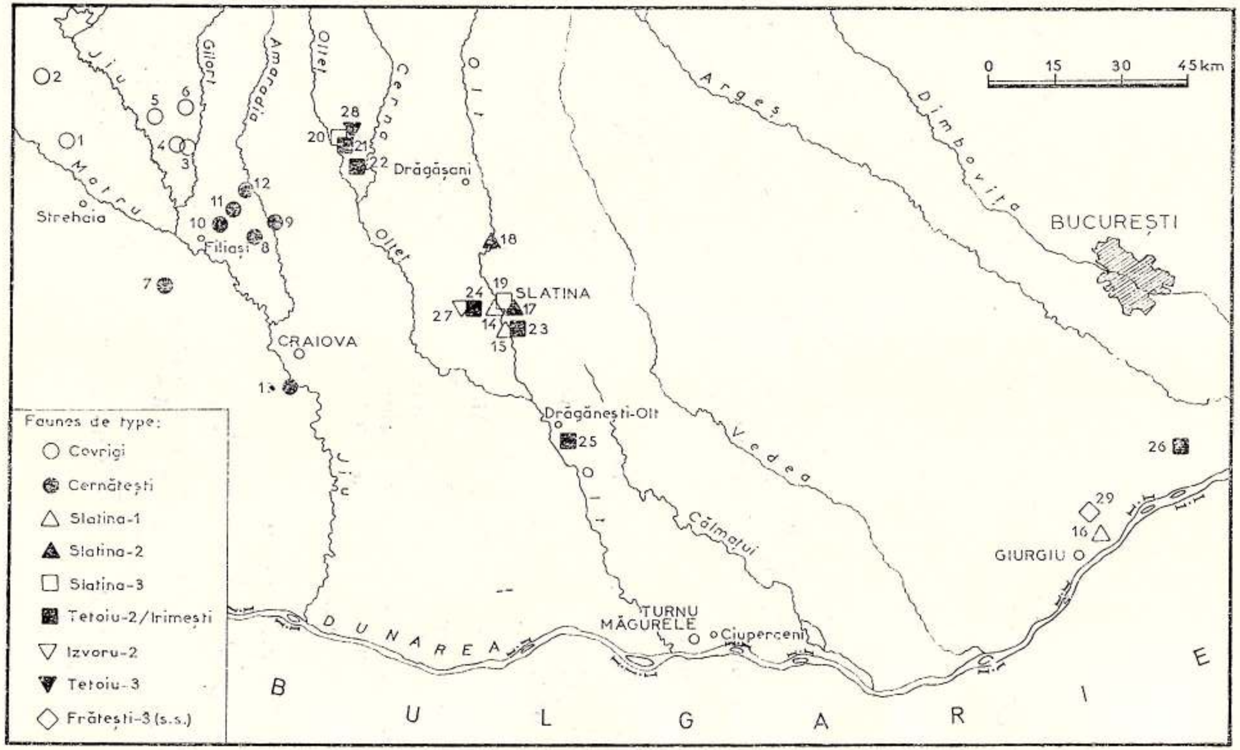

Carte des gisements fossilifères du Bassin Dacique ; 1, Covrigi ; 2, Mătăsari-Brădeț ; 3, Groșerea ; 4, Costești ; 5, Piscurile ; 6, Săulești ; 7, Cernătești ; 8, Valea Muierii ; 9, Amărăști ; 10, Fratoștița ; 11, Măiag (= Valea Boului) ; 12, Urda de Sus ; 13, Podari ; 14, Slatina-1 ; 15, Milcovu-1 ; 16, Frătești-1 ; 17, Slatina-2 ; 18, Cherleștii-Moșteni ; 19, Slatina-3 ; 20, Tetoiu-1 ; 21, Tetoiu-2 ; 22, Irimești ; 23, Milcovu-2 ; 24, Izvoru-1 ; 25, Drăgănești-Olt ; 26, Frătești-2 (Prundu) ; 27, Izvoru-2 ; 28, Tetoiu-3 ; 29, Frătești-3 (*s.s.*).

Faune de Covrigi (partie inférieure de l'horizon sablo-graveleux situé entre les couches XII et XV de charbon) : *Anancus arvernensis*, *Dicerorhinus* cf. *jeanvireti*, *D.* cf. *etruscus*, *Metacervocerus pardinensis*.

Autres gisements — Faune de Groşerea (même horizon) : *Zygolophodon borsoni*, *Anancus arvernensis*, *Dicerorhinus* cf. *jeanvireti*, *Equus* cf. *simionescui*, *Metacervocerus pardinensis*, *Cervus* cf. *perrieri*.

Equivalents (Dépression de Braşov) — Faune de Iarăş-2 : *Dicerorhinus* cf. *jeanvireti*, *D.* cf. *etruscus*, *Cervus* sp. (taille grande). — Faune de Ilieni : *Anancus arvernensis*, *Tapirus arvernensis*, *Dicerorhinus* cf. *jeanvireti*, *Cervus* sp. (taille petite), *Gazella* sp., *Ursus minimus*, *Hystrix* cf. *refossa*, *Castor* sp.

Paléomagnétisme : normal à Iarăş-2, époque de Gauss avant l'événement de Kaena (G h e n e a et al., 1981).

Caractéristiques de la faune : association de deux Mastodontes ; Eléphant absent ; deux formes du genre *Dicerorhinus* de taille différente ; Cervidés archaïques ; Equidé monodactyle primitif.

Corrélations : sommet de l'horizon kagoulien du complexe faunique moldave, partie supérieure de la zone des Etouaires (~ 3 MA), Triversa, zone MN 16.

Faune de Cernăteşti (partie moyenne de l'horizon sablo-graveleux identifié à l'horizon moyen du "Levantin") : *Zygolophodon borsoni*, *Anancus arvernensis*, "*Archidiskodon*" *rumanus*, *Dicerorhinus* cf. *jeanvireti*, *D.* cf. *etruscus*, *Metacervocerus pardinensis*.

Autres gisements — Faune de Podari (partie sommitale de l'horizon moyen, sablo-graveleux du "Levantin") : *Desmana kormosi*, *Trogontherium* sp. (taille petite), *Pliomys hungaricus*, *Dolomys milleri* (archaïque).

Paléomagnétisme : normal, époque de Gauss (détermination par V. M. T r o u b i k h i n e, communiquée par A. L. T c h é p a l y g a).

Equivalents (Moldavie du sud) — Faune de Tuluceşti : *Z. borsoni*, *A. arvernensis*, "*Archidiskodon*" *rumanus*, *Allohippus major euxinicus*, *Paracamelus alutensis*, *Cervus* cf. *perrieri*.

Caractéristiques de la faune : apparition d'un Eléphant très primitif ("*A*". *rumanus*) ; Chevaux sténoniens de très forte taille ; *Dolomys* du groupe *milleri* archaïque.

Corrélations : complexe faunique skortselskien (= Poratien supérieur) (A l e x e e v a, 1977), Montopoli où l'Eléphant, assimilé à *A. gromovi*, représente une forme plus primitive.

Les faunes de type Covrigi et Cernăteşti se rangent dans le Romanien moyen.

Faune de Slatina-1 (couche 14, argilo-limoneuse) : *Desmana kormosi*, *Apodemus* sp., *Dolomys milleri* ssp., *Mimomys minor*.

Paléomagnétisme : inverse, début de l'époque de Matuyama.

Caractéristiques de la faune : dominance de *Dolomys* dont le degré d'évolution est nettement plus avancé qu'à Podari.

Corrélations : horizon kryjanovkien du complexe faunique khaprovien, partie moyenne du Prétiglien, moitié inférieure de la zone de St. Vallier et de la zone MN 17.

Faune de Milcovu-1 (coupe de Milcovu din Vale, couche 3, sablo-limoneuse) : *Trogontherium* sp. (taille petite), *Dolomys milleri* ssp., *Mimomys* cf. *minor*. Bien que les raccords stratigraphiques placent le niveau de Milcovu-1 au-dessous de celui de Slatina-1, nous n'avons pas séparé les deux faunes, étant donné leurs étroites affinités ; les Mollusques associés, caractéristiques de l'horizon tchistopolien situent le niveau de Milcovu-1 antérieurement à l'horizon kryjanovkien.

Faune de Slatina-2 (couche 17, sablo-graveleuse, ravinante) : *Desmana nehringi*, *Talpa fossilis*, *Beremendia fissidens*, Léporidé cf. *Hypolagus brachygnathus*, *Apodemus* sp., *Dolomys milleri milleri*, *Mimomys minor*, *Eucladoceros* sp. et très probablement *Archidiskodon gromovi*.

Paléomagnétisme : inverse, époque de Matuyama.

Autres gisements — Faune de Cherleştii-Moşteni (couche 3, argileuse) : *Desmana nehringi*, *Trogontherium* sp. (taille petite), *Allactaga ucrainica*, *Apodemus* sp., *Dolomys milleri milleri*, *Borsodia* sp., *Mimomys* gr. *polonicus-pliocaenicus*, *M. minor*. La position stratigraphique et la faune situent le gisement de Cherleştii-Moşteni immédiatement au-dessus de celui de Slatina-2. — Faune de Frăteşti-1 (zone de Frăteşti-Giurgiu, horizon inférieur,sablo-graveleux) : l'association de Mammifères, connue dès la fin du siècle dernier, renferme à côté de *Anancus*, un Eléphant proche, peut-être, de *A. gromovi* ; ainsi, la faune se rangerait près de celle de Slatina-2.

Caractéristiques de la faune : diminution sensible de *Dolomys* qui est plus évolué que dans le niveau de Slatina-1 ; dominance du groupe *Mimomys* ; présence de *Desmana nehringi* ; apparition, à Cherleşti, de la lignée *M. polonicus-pliocaenicus*, de *Borsodia* et de *Allactaga* indiquant le début d'une première détérioration climatique (continentalisation) enregistrée par les Mammifères.

Corrélations : horizon ferladanien du complexe faunique khaprovien, phase fraîche du Tiglien, deuxième moitié de la zone de St. Vallier (antérieurement à St. Georges-d'Aurac, daté de 1,92 MA) et de la zone MN 17.

Les faunes comprises entre le niveau de Milcovu-1 et celui de Cherleşti se situent dans le Romanien supérieur.

Faune de Slatina-3 (couche 37, sablo-limoneuse) : *Trogontherium boisvilletti dacicum*, *Mimomys* sp.

Paléomagnétisme : normal, événement d'Olduvai (couches 37—40).

Autres gisements — Faune de Tetoiu-1 (vers la base de l'horizon inférieur à dominante sablo-limoneuse) : *Archidiskodon meridionalis*, *Dicerorhinus etruscus*, *Allohippus athanasiui*, *Eucladoceros* sp., *Mitilanotherium inexspectatum*, *Pliotragus ardeus*, *Nyctereutes megamastoides*, *Ursus etruscus*, *Homotherium crenatidens*, *Megantereon megantereon*, *Lynx issiodorensis*, *Beremendia* cf. *fissidens*, *Hystrix refossa*, *Castor plicidens*, *Trogontherium boisvilletti dacicum*, *Hypolagus brachygnathus*, *Paradolichopithecus geticus*.

Caractéristiques de la faune : *Archidiskodon meridionalis*, seul ; chevaux sténoniens de grande taille ; Cervidés de type *Eucladoceros* ; présence de *Trogontherium* de taille moyenne, du Girafidé *Mitilanotherium* et du Primate *Paradolichopithecus geticus*.

Corrélations : début de l'horizon domachkinien du complexe faunique odessan, fin du Tiglien, début de la zone de Senèze (antérieurement au gisement type).

Limite Pliocène/Pléistocène — Des données convergentes, concernant les Mammifères, les Mollusques et le paléomagnétisme, indiquent, dans le profil de Slatina, la situation de la limite plio-pléistocène au niveau de Slatina -3 (Andreescu et al., 1981; Feru et al., 1978). La même limite a été tracée par Samson et Rădulescu (1973) à la base de Tetoiu-1, ce qui vient d'être renforcé par les résultats obtenus à Slatina.

Faune de Tetoiu-2/Irimeşti (vers le milieu de l'horizon moyen à dominante sablo-graveleuse); *Archidiskodon meridionalis, Dicerorhinus etruscus, Allohippus stenonis mitilanensis, Allohippus* gr. *major, Sus strozzi, Mitilanotherium inexspectatum, Eucladoceros* sp., *Allocaenelaphus* sp., *Cervus nesti, Praealces gallicus, Megalovis latifrons, Leptobos* sp., *Canis etruscus, Pachycrocuta brevirostris, Castor plicidens, Trogontherium boisvilletti boisvilletti.*

Autres gisements — Faune de Milcovu-2 (coupe de Milcovu din Vale/Olt, couches 26—29, sablo-graveleuses) : *A. meridionalis, Paracamelus alutensis.* — Faune de Izvoru-1 (couche 1, sablo-graveleuse) : *A. meridionalis, Allocaenelaphus* sp. — Faune de Frăteşti-2 (Prundu, horizon moyen sablo-graveleux): *A.meridionalis, Praealces gallicus.* — Faune de Drăgăneşti—Olt (couche 2, sablo-graveleuse) : *Desmana* gr. *nehringi, Trogontherium* sp., *Allactaga* sp., *Parapodemus* sp., *Mimomys pliocaenicus, M.* cf. *coelodus, M. reidi, M. newtoni, Clethrionomys kretzoii, Borsodia* cf. *lagurodontoides.* Il est difficile de préciser la position chronologique de la faune de Drăgăneşti par rapport aux niveaux de Tetoiu-1 et de Tetoiu-2; très probablement, elle est plus proche du second.

Caractéristiques de la faune : association de deux Equidés sténoniens, de taille différente; apparition de *Trogontherium* de taille grande, de *Praealces*, de *Megalovis* et des Mégacérins primitifs du groupe *Allocaenelaphus*; ces dernières espèces ainsi que *Paracamelus, Allactaga, Borsodia, Clethrionomys kretzoii, Mimomys newtoni* et la fréquence accrue de *M. pliocaenicus* indiquent une détérioration climatique.

Corrélations : partie supérieure de l'horizon domachkinien du complexe faunique odessan, début de l'Eburonien, zone de Senèze (la localité type), Tasso et Casa Frata.

Faune de Izvoru-2 (couche 9, limono-sableuse) : *Desmana nehringi, Petenyia hungarica, Beremendia fissidens,* Léporidé cf. *Hypolagus brachygnathus, Citellus primigenius, Trogontherium boisvilletti boisvilletti, Borsodia* cf. *hungarica, Mimomys* gr. *reidi-pusillus, M. coelodus, M. pitymyoides, Kislangia rex.*

Caractéristiques de la faune : Insectivores plus nombreux; présence de *Citellus, M. pitymyoides* et *Kislangia*; l'ensemble suggère une amélioration du climat.

Corrélations : début de l'horizon bochernitsien du complexe faunique odessan, phase plus chaude de l'Eburonien (antérieurement à Brielle), sommet de la zone de Senèze.

Faune de Tetoiu-3 (horizon supérieur à dominante graveleuse) : *Archidiskodon meridionalis* ssp. (évolué), "*Allohippus*" cf. *suessenbornensis*, "*A*". cf. *marxi*, *Equus aluticus*, *Soergelia* cf. *elisabethae*, *Bison* cf. *schoetensacki*, *Trogontherium boisvilletti boisvilletti*.

Equivalents (Dépression de Brașov) — Faune de Rotbav-Silvestru (horizon sablo-graveleux inférieur) : *A. meridionalis*, *Dicerorhinus etruscus*, "*Allohippus*" cf. *suessenbornensis*, "*A*" cf. *marxi*, "*A*". cf. *altidens*, *Equus aluticus*, *Allocaenelaphus arambourgi*.

Caractéristiques de la faune : présence de formes évoluées du groupe "*Allohippus*", de *E. aluticus* et de *A. arambourgi* ; l'association indique une détérioration climatique.

Corrélations : horizon kaïrien du complexe faunique tamanien, épisode plus frais du Waalien (B).

Les faunes de Tetoiu-1 (Slatina-3) à Tetoiu-3 représentent trois phases successives dans le Pléistocène inférieur.

Faune de Frătești-3 (Frătești s.s.) (horizon supérieur sablo-graveleux) : *Archidiskodon meridionalis* (évolué), *Praemegaceros verticornis*.

Caractéristiques de la faune : présence de *P. verticornis* ; forme évoluée de *A. meridionalis*.

Corrélations : horizon mikhaïlovkien du complexe faunique tiraspolien, début du "Complexe Cromérien".

La faune de Frătești-3 s'inscrit dans la partie inférieure du Pléistocène moyen.

En guise de conclusion, nous devons relever que deux notions, celles de "couches de Cîndești" et "couches de Frătești", sans être formulées explicitement, sont restées sous-entendues à notre exposé. L'espace ne nous permet pas d'entrer dans les détails de ce problème ; d'ailleurs nous les avons discutés récemment (F e r u et al., 1979). Rappelons, toutefois, que les faunes se situant entre les niveaux de Covrigi et de Cherleștii-Moșteni ou même de Milcovu-2 proviendraient des sédiments attribués généralement aux "couches de Cîndești" et que les associations comprises entre les horizons de Tetoiu-1 et de Frătești-3 appartiendraient plutôt aux „couches de Frătești". Mais, il est évident, qu'à partir au moins du Romanien moyen, des niveaux chronologiquement équivalents se trouvent aussi bien dans l'une que dans l'autre des deux formations ; ainsi, "couches de Cîndești" et "couches de Frătești" ne représenteraient — quand elles auront une définition lithologique adéquate — que tout au plus deux faciès différents d'un même processus de remblaiement du Bassin Dacique. Et, à la fin, mentionnons que des découvertes récentes, faites près du Danube, à Ciuperceni (T e r z e a, B o r o n e a n ț, 1979) — donc dans le domaine des "couches de Frătești" — laissent entrevoir un niveau plus ancien encore, proche de celui de Mălușteni-Berești, ce qui vient de renforcer nos vues antérieures.

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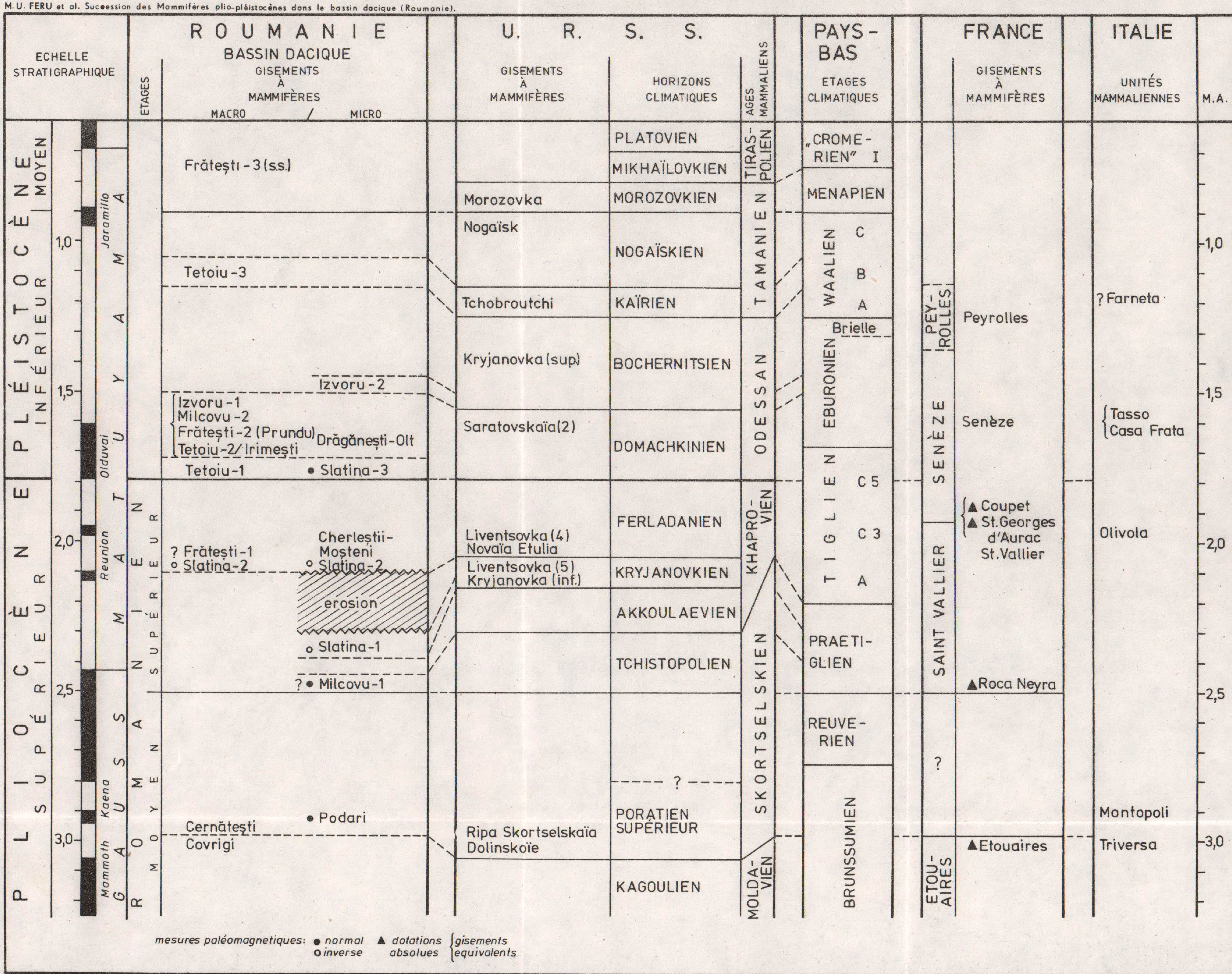

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REPRESENTATIVES OF THE GENUS BORELIS IN THE BADENIAN AND LOWER SARMATIAN OF YUGOSLAVIA[1]

BY

NADEŽDA GAGIĆ[2]

The localities where genus *Borelis* occurs in the Badenian and Sarmatian of Yugoslavia, and of some neighbouring countries of the Central and Eastern Paratethys, are outlined in the present paper.

Representatives of the genus *Borelis*, generally and mostly the species *Borelis melo* (Ficht., Moll), are frequently encountered in the Badenian (auct. "Tortonian") of Yugoslavia, especially in Serbia where these sediments are quite widespread. They are contained in limestone, marl, marly clay and sand, subordinately in sand of oolitic habitus.

The "Tortonian" formations are referred to in this paper as the Badenian, based on the results and conclusions of the Working Group for the Paratethys of the Regional Committee on Mediterranean Neogene Stratigraphy, and Project No. 25 of the International Geological Correlation Programme.

Borelis melo and *Borelis* cf. *haueri* are found in western Serbia in the Korenite-Jadar Basin (Petrović, 1963/67; Gagić, 1965), in Lajtovac Limestone of Middle and Upper Badenian (loc. 1), in Koceljevo environs, in boreholes to a depth of some thirty metres; representatives of *Borelis melo* (Gagić, 1977; loc. 2) are encountered in marly clay in Ammonia beccarii zones of the Upper Badenian.

On the territory of Beograd, the species *Borelis melo* is found, according to Spajić and Džodžo-Tomić (1973), in sandy clay of boreholes at Veliki Mokri Lug (91.4 m), Kumodraž (146 – 149 m, together with *B. haueri*), and Leštani (195 – 201 m), and in reef limestone with *Lithothamnium* below Kalemegdan, all in Bolivina dilatata Zone of the Upper Badenian, while the species *Borelis haueri* is identified in the same limestone of borehole B-19/74 (35 – 48 m) in the Railwaymen Hospital, Beograd (Gagić, 1974; loc. 3).

South of Beograd, representatives of *Borelis melo* are found in sandy clay of borehole B-1/73 (191 – 211 m; Gagić, 1973), north of the village of Venčane; based on microfossil community with large *Elphidium*

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[2] Geological Institute, Beograd, Yugoslavia.

and *Cibicides* (*Heterolepa*), these correspond to upper Badenian formations (loc. 4).

In the province of Vojvodina, *Borelis melo* is found in carbonate sediments (limestone and marly sandstone) and clay in the Upper Badenian, (Ammonia beccarii Zone) recorded in deep boreholes Marković, 1956—1981 ; Gagić, 1959 ; Bo-18 at depth interval 1241 — 1247 ; loc. 5).

Džodžo mentions (1964) *Borelis melo* from marly clay with *Planulina wuellestorfi* (Schw.) near Golubac (loc. 6), eastern Serbia, from sandy clay of Bulimina-Bolivina Zone of the Upper Badenian near Štubik and Šarkamen (same author, 1959 and 1979 ; loc. 7), and from the Jagnjilski Potok (same author, 1963). This species is also encountered in clay belonging to lower lagenid zone, the zone known in the Lower Badenian Vienna Basin, in borehole B-25/63 (22 — 24 m) at Vrška Čuka (Popović and Gagić, 1969 ; loc. 8) ; it is very abundant in highly marly clay of the upper lagenid zone in Braćevačka Reka (Gagić, 1971 ; loc. 9). It is encountered in argillaceous-arenaceous sediments of the Slatina Basin in the Ammonia beccarii Zone (Gagić, in Popović, 1968 ; loc. 10).

The species *Borelis melo* is found in northern Bosnia (Muldini-Mamužić, 1955) in marl from borehole Ravna Trešnja 1 (521 — 642m), north of Tuzla (loc. 11), corresponding to the upper lagenid zone of the type Badenian. Abundant *Borelis melo* is mentioned (M. Eremija, 1970 ; loc. 12) in the Ammonia beccarii Zone in Upper Badenian marly sandstone of the Prnjavor Basin. Also in Bosnia, it is identified in sediments of northeastern slopes of Kozara, in sandy clays and clay-calcareous sandstones of the same zone (Petrović, 1969).

Only individual specimens of *Borelis melo* have been found in Upper Badenian sediments of Croatia and Slovenia according to oral communication of Muldini-Mamužić (1981) and Rijavec (1981).

This is a short survey of the localities where the genus *Borelis* was recorded in Yugoslavia (Fig. 1), in Middle Miocene (Badenian) sediments, not covering all data of the Geozavod Documentation Fund, but sufficient to illustrate the occurrence of the mentioned species, particularly on the territory of Serbia.

Studying Miocene marine microfossil associations I have come upon some new data concerning the range of the *Borelis melo*.

More than ten years ago I noted a frequent occurrence of *Borelis melo* in some sections near Beograd and in Pomoravlje in Lower Sarmatian (Volhynian) limestones, marly clay and sand, in Elphidium antoninum Zone, or its equivalent zone with small forms of *Elphidium macellum* and *Elphidium crispum*. It was a new knowledge of this species' vertical distribution in the Paratethys.

The frequent occurrence was not of individual specimens, though it also was possible, but of an abundance of individuals in a microfossil association, both adult and juvenile ones. It took time, however, for this species occurrence in Lower Sarmatian sediments to be accepted as a true range of *Borelis melo* and recognized, instead of being associated with resedimentation from Badenian sediments.

Besides, a large number of specimens had to be found in the lowermost Lower Sarmatian, in Elphidium reginum Zone, what was also obtained in marly clay of borehole BM-1/74 (Zabrdje, 150 m; loc. 13) and limestone of Duboki Potok (east of loc. 13), in Pomoravlje. They were identified in limestones on the territory of Beograd (G a g i ć, 1974, boreholes; loc. 14).

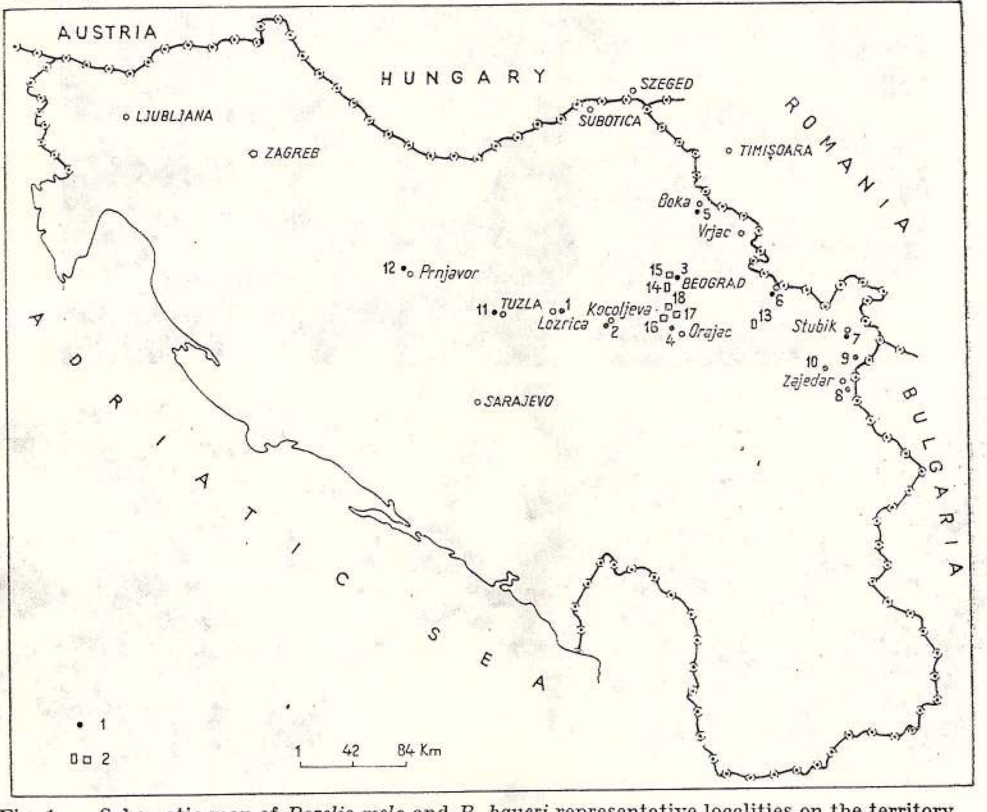

Fig. 1. — Schematic map of *Borelis melo* and *B. haueri* representative localities on the territory of Yugoslavia.

1, Badenian, loc. 1 — 12; 2, Lower Sarmatian, loc. 13 — 18 (only *B. melo*).

These discoveries documented to a certain point the range of this species in Lower Sarmatian brackish sea of the Pannonian Basin in Yugoslavia, and called for a correlation with those in adjacent regions of the Paratethys, hence also the communication of the species within the Lower Sarmatian sea in general.

Borelis melo in upper Lower Sarmatian, in Elphidium antoninum Zone, on Beograd territory (loc. 15) is encountered in oolitic limestone at the depth interval roughly between 30 m (Svetozara Marcovića street; G a g i ć, 1980) and 60 m (Kronštatska Street; D ž o d ž o, 1975).

Near Beograd, *Borelis melo* was found in the same zone, in the pseudo-oolitic limestone of the Seona Stream (G a g i ć, 1972; loc. 16) (Fig. 2) in the sandy limestone of the village of Slatina (same author;

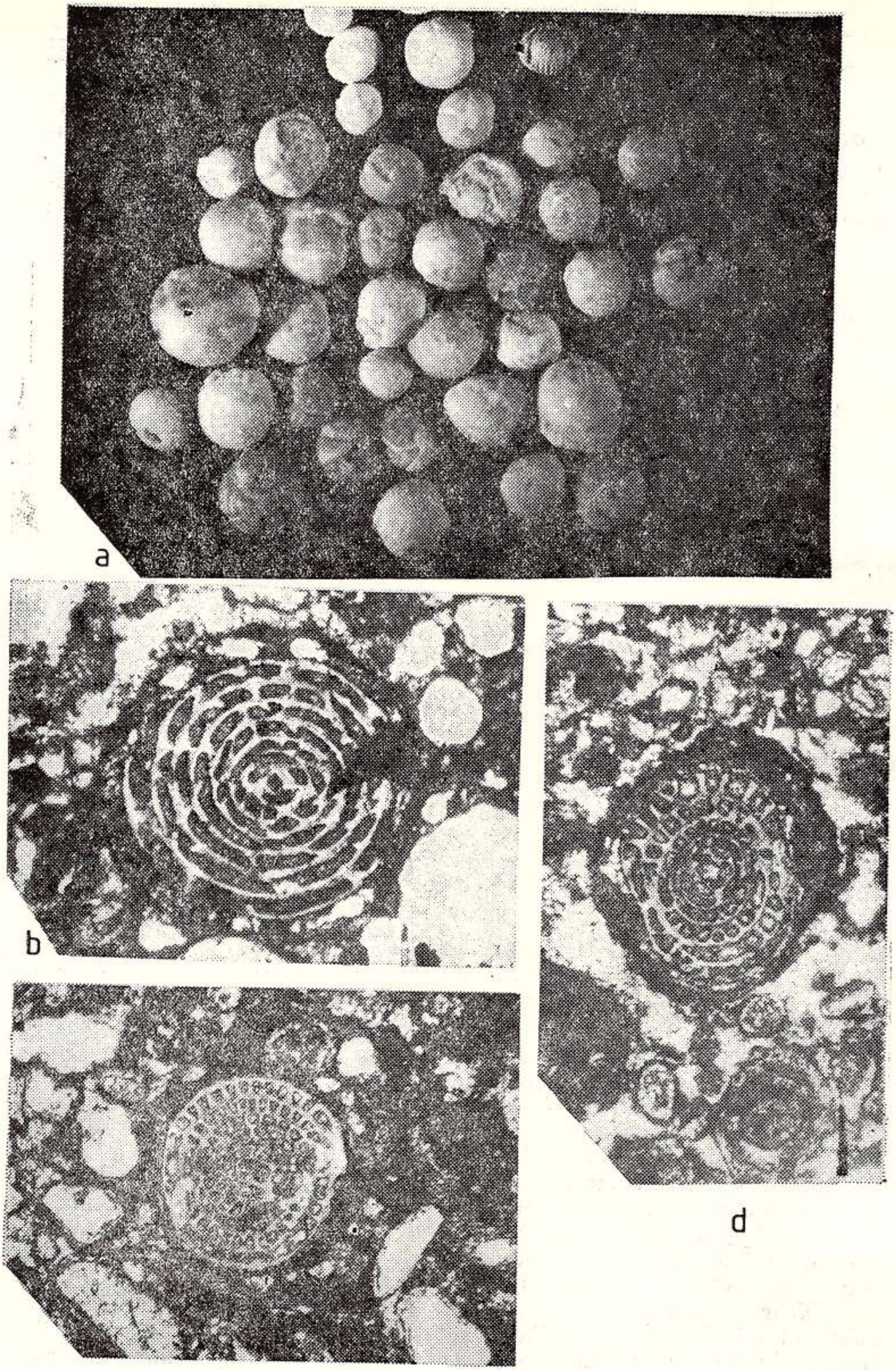

Fig. 2. — *Borelis melo* (F i c h t e l and M o l l) from Seona Stream, loc. 16, Lower Sarmatian. a, wash-out specimens, magn. + 21 ; b, equatorial section with visible postseptal canals. Magn. × 52 ; c, d, axial sections, test slightly damaged. Magn. × 52.

loc. 17) (Fig. 3), and in the most instructive section abunding with *Borelis melo* in banks and sides of the Mišljevac Stream, the village of Guberevac (G a g i ć, 1974 ; loc. 18) (Fig. 4).

Macrofaunal data and a general description of sediments in this section are given by S t e v a n o v i ć, 1949.

Some small additions will be made in the list of microfossils from sediments in this stream in addition to the lithostratigraphic column (Plate). Microfossils encountered in bed 5 are *Elphidium* (d 'O r b .), *Elphidium crispum* (L i n n é)— small forms, *Elphidium* sp., *Ammonia beccarii* (L.), *Cytheridea hungarica* Z a l ., *Aurila mehesi* (Z a l .), *Miocyprideis sarmatica* (Z a l .), *Hemicyprideis* cf. *dacicus* (H.), ; in bed 6 prevail ostracods ; *Cytheridea hungarica*, *Aurila mehesi*, *Hemicyprideis* cf. *dacicus*, *Xestoleberis* cf. *lunaris* R y b i n a . Besides ostracods as in bed 6, occasional *Xestoleberis* is encountered in bed 7.

The Elphidium reginum Zone, i.e. basal Sarmatian, terminates with bed 12 which contains the last encountered representatives of *Cytheridea hungarica*. The first representatives of *Aurila notata* in bed 14 already belong to the lowermost upper zone of the Volhynian substage with *Elphidium antoninum*. Upward follows a set of beds, 14 through 24, which contain more or less *Aurila notata* ; bed 15 includes minute miliolids, elphidiums, *Halicoryne morelleti* (P o k .) ; bed 16 has sparse *Leptocythere plana* (S c h n .), and bed 17 only *Xestoleberis* sp. Only rare *Nonion* and *Elphidium* are contained in bed 24.

The group of beds from 25 to 31 includes limestones, one of which (28) contains *Borelis melo* where it was first found in medium number.

Thus with beds 32 and 33 ends the accessible part of the section, on the left side of the Mišljevac Stream, and Sarmatian sediments continue from bed 34 to the right side of the stream but nearly only in calcareous facies. *Borelis melo* is more frequent in bed 35 and is abundant in bed 37.

The next member upward in the section is an oolitic limestone (bed 38) with algae and infrequent nubeculariids (adhesive forms), and a limestone with ostreae as the closing upper Lower Sarmatian bed.

The next 4.30 metres (bed 40) are lower Middle Sarmatian limestones with an abundance of peneroplids, and some olitic limestones.

Thus, following the lithological section, two quite big sets of strata could be separated.

The lower set of beds 1 — 12, which contain relatively abundant microfauna, represent the old littoral shallow-water region with clay-sand bottom.

A similar situation, with greater or smaller inflow of terrigenous constituents, and with higher or lower proportion of carbonate solution, when limestones (intercalations to thin beds) were formed, may have prevailed until the formation of compact limestones (bed 30).

Limestone occurrence is frequent from this bed to the end of the section (bed 40), with the exception of bed 34. It begins with compact limestones, to be succeeded by sandy, oolitic, vuggy, chalky limestones. Occasionally, they have a reef character of form back reef shoal series or fore reef shoals. The same could be inferred by the microfaunal associations.

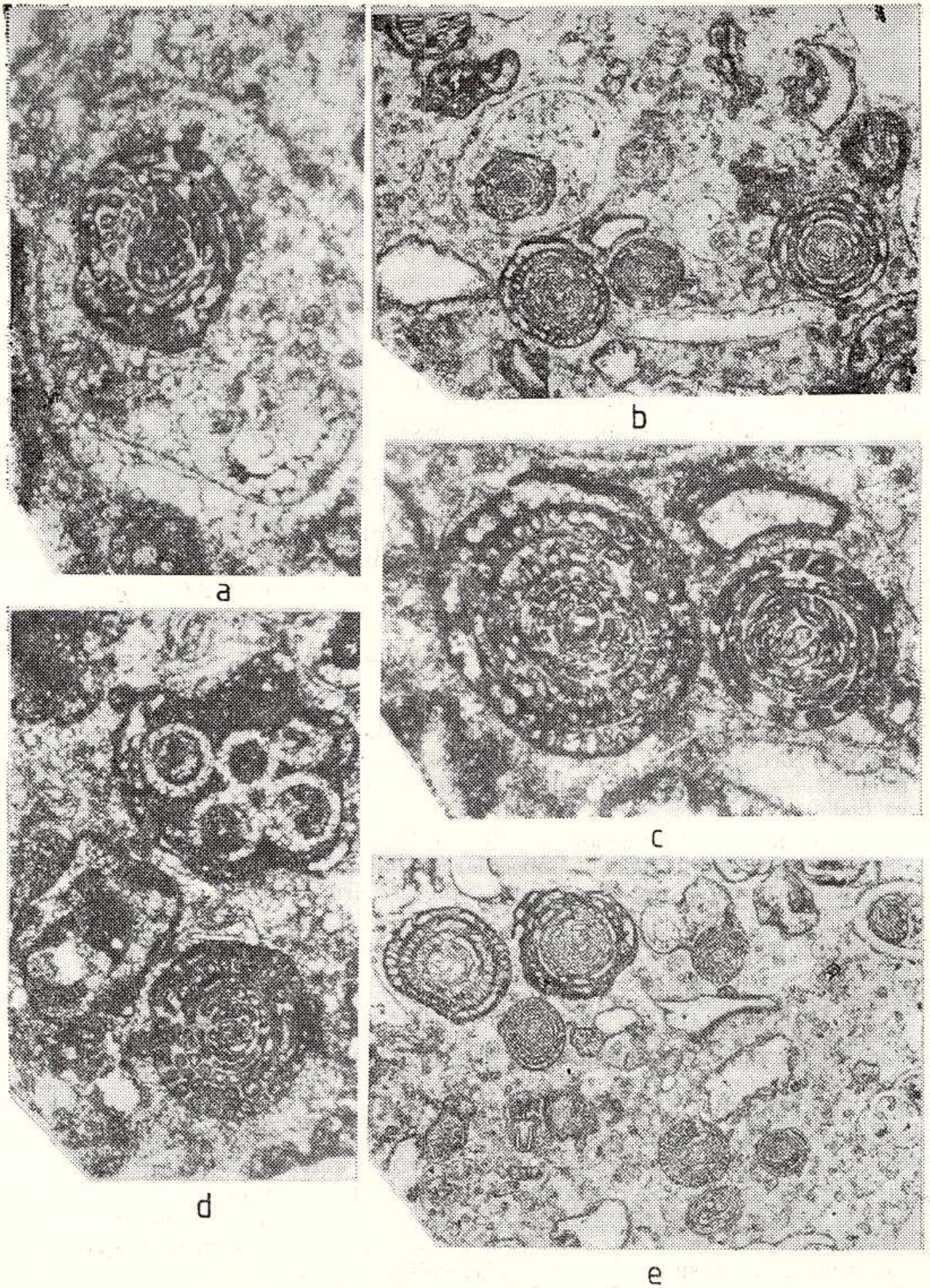

Fig. 3. — *Borelis melo* (F i c h t e l, M o l l) from sandy limestone of Slatina village, loc. 17, Lower Sarmatian.

a — c, axial and tangential sections ; figs. 1 (× 60) and 3 (× 52) details of fig. b (× 21) ; d, axial section, somewhat elongated form ; *Halicoryne morelleti* (P o k .), besides *borelis*. Magn. × 52 ; e, axial and tangential sections ; some forms slightly deformed. Magn. × 21.

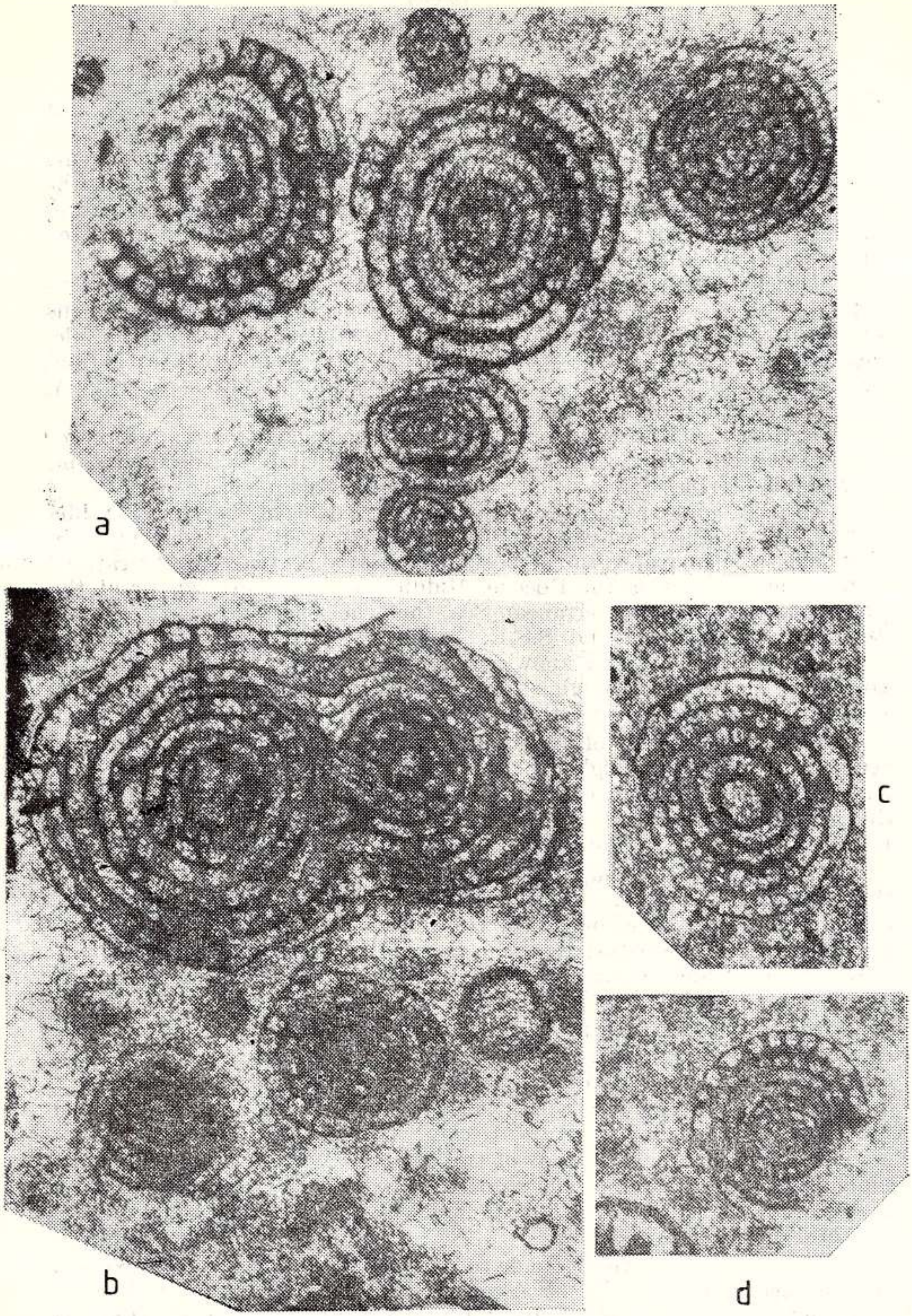

Fig. 4. – *Borelis melo* (Fichtel, Moll) from Lower Sarmatian sediments of Mišljevac Stream, loc. 18.

a – c, axial sections of specimens from bed 35. The older and stronger individuals mechanically cover the younger ones preventing their growth, developing themselves. The phenomenon is possible among a multitude of specimens in a small area (fig. b) Magn. × 52 ; d, axial section bed 28. Magn. × 65.

Oolitic limestones contain in the nuclei some of terrigene components in addition to microfauna and algal fragments. They are formed along the coast line exposed to high waves, i.e. in the high-energy zone; the oolites are large-developed, in places deformed, when the salinity increased probably to as much as $25^0/_{00}$.

In view of present knowledge of the environment in which representatives of *Borelis* have lived, as stenohaline forms in shallow-water regions of warm seas, with the prevailling temperatures from 25° to 30°, in mobile, clear water rich in oxygen, bivalent calcium and magnesium, at normal or higher salinity rate of 35 – 50‰ some new information is contributed concerning the Sarmatian Sea.

Recently some new information is obtained about Peneroplid representatives from the lower Middle Sarmatian of Ukraina, Moldavia and Yugoslavia (Serbia) which will be briefly commented upon.

If an abundance of peneroplids was recognized in lower Middle Sarmatian on the territory of Serbia (Gagić, 1974), knowing that they live only in sea basins of normal or nearly normal salinity rate as it was the case during the Eocene, Middle Miocene, Maeotian or at the present time in the Mediterranean Sea, then their presence in Yugoslavia and in south Ukraina and S.S.R. Moldavia (Didkovskiy, 1959) during Sarmatian is a new knowledge of the peneroplid occurrence. They could have adapted, though stenohaline forms, to the new conditions of much lower salinity and thereby prove their ecological validity.

Using the criterion of lower salinity (probable to $20^0/_{00}$) and not high temperature (also not higher than 20°), representatives of *Borelis melo* should also perhaps be accepted as possible indicators of ecological validity, in which case their correlation in the areas of both Paratethyses during the Lower Sarmatian could be very instructive.

It also should be supposed that *Peneroplis* and *Borelis* faunas in the Sarmatian Sea might have indicated a possible intermittent communication between this and open sea. However, we know today that they are not typical representatives of the discovered sea basins, but are important for their wide ecological validity. Besides, the association of Sarmatian microfossils does not include members of open sea basin.

I should emphasize that, according to the Sarmatian subdivision in the areas of the Central Paratethys (Jiriček, 1972), beds 1 – 12 in the considered section correspond to Lower Sarmatian *sensu stricto* ("a"), beds 13 – 39 to the Middle Sarmatian *sensu stricto* ("b" + "c"), and bed 40 to the Upper Sarmatian *sensu stricto* ("d").

However, I take that beds of this section can be separated, by the similarity of their microfossils associations those of the Volhynian and Bessarabian in south Ukraina and Moldavia, as the Lower Sarmatian and lower Middle Sarmatian.

A brief reference will be made to some of the published data concerning the occurrence of *Borelis* species in Middle Miocene sediments of neighbouring countries as schematically represented in Fig. 5 which is copied from the article by Bignot et Guernet (1976).

It is mentioned in Hungary in *Bulimina* and *Spiroplectammina* horizons of the Upper Badenian in boreholes Vilyvitányj 6 (60 – 62 m), Füzérhajata 2 (715 — 716 m) according to I. L. K o r e c z n é (1973; loc. V), and the Lower Badenian Lajtovac Limestone on Mecsek Mountains (same author, 1968).

In Austria, in Badenian holostratotype at Baden-Soos clay-pit, in sandy lens, *Borelis melo* is encountered among the microfauna (P a p p, 1978) and in faciostratotype of Stiefingtal (K o l l m a n n - R ö g l, 1978).

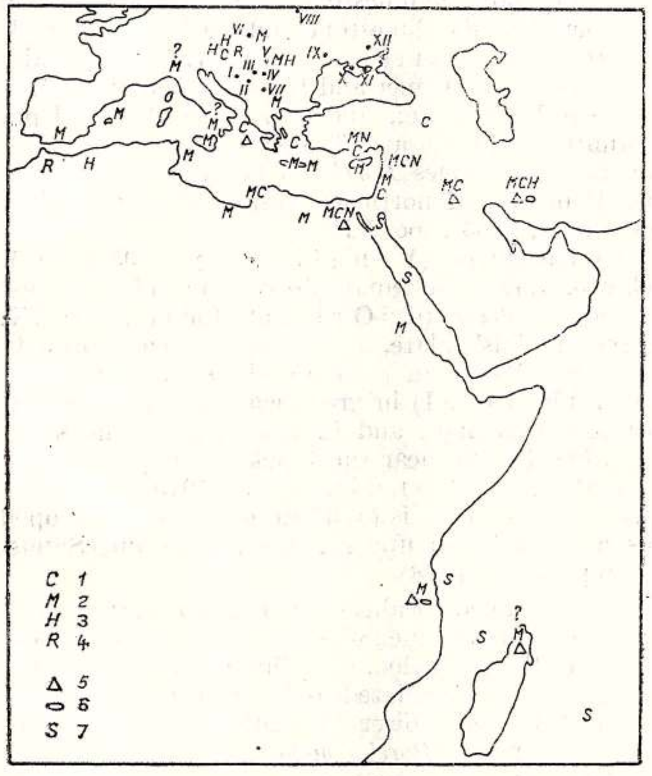

Fig. 5. — *Borelis* deposits in the Miocene of the peri-Mediterranean regions of the Paratethys and of the Middle East, after different authors (G. B i g n o t and C. G u e r n e t, 1976).

Besides the mentioned ones, some of the localities quoted in the present paper are also marked; I, Serbia; II, Bosnia; III, Vojvodina — Yugoslavia; IV, Romania (Intra-Carpathian depression); V, Hungary; VI, Poland; VII, Bulgaria; VIII, Volino-Podolsk plate; IX, Cis-Carpathians; X, Stepnoi Crimea; XI, Kerch Peninsula; XII, The Black Sea littoral.

1, *Borelis curdica*; 2, *B. melo*; 3, *B. haueri*; 4, *B? rotellus*; 5, Miogypsines in association; 6, Lepidocyclines with *Borelis*; 7, biotopes actual with *B. schlumbergeri*.

In Poland, *Borelis melo* is identified near Bochnia (loc. VI) in Chodenice Beds of Lower Kossovian, and in Grabowice Beds of Upper Kossovian (Ł u c z k o w s k a, 1955). It is also found in Lower Badenian of Benczyn near Wadovice, together with *B. haueri* and *B. rotellus* (Łuczkowska, 1957).

In Romania, in the western part of Intracarpathian depression, *Borelis melo* is encountered in Volhynian calcareous-marly sediments of Zarand Basin (Ionesi, Sagatovici, 1970; loc. IV), and in Caransebeș-Mehadia Basin (Popescu, Marinescu, 1978) in Upper Badenian organogenic limestone. A biozone with *Borelis melo* from borehole in Upper Badenian limestone and marl is mentioned northeast of the Moesian Platform (Vasilescu et al., 1971). It is also mentioned (Popescu, 1979) in the Upper and Lower Badenian of Mehadia Basin, Maramureș, Transylvania, etc., and *Borelis rotellus* in Upper Badenian of Oltenia, Muntenia, Banat and Transylvania.

In Bulgaria, the species *Borelis melo* and *Borelis haueri* are found in the Upper Badenian of northeast Bulgaria, in the village of Ohrid (Tzankov et al., 1965), loc. VII.

In trans-Carpathians (Venglinskiy, 1958, 1975) in sandy-micaceous clay of Lower Badenian Tresvenian suite on Solotvin slope, *Borelis melo* and *B. rotellus* (d'Orb.) are found in loc. IX.

In Volhyno-Podolsk plate, *Borelis melo* is encountered in Ternopol Beds of the Upper Badenian (Venglinskiy, 1975) in loc. VIII; on Kerch Peninsula (loc. XI) in green clay and marl of Konka Horizon of the uppermost Badenian, and in limestone of the same horizon in Stepnoy Krim, loc. X, and near the Black Sea depression, loc. XII (Goretskiy et al., 1974; Venginskiy, 1975).

The species *Borelis melo* is found in many well-developed specimens, mentioned as noteworthy in upper lagenid zone and Sandschalen zone localities (Papp et al., 1978).

Of all the mentioned localities in neighbouring countries only one is in Lower Sarmatian limestone; it is the Zarand Basin of western Intracarpathian Basin of Romania (loc. IV). This *Borelis*, while represented by sparse specimens, can be correlated, only attention should be given when microfauna is studied from Lower Sarmatian sediments for determining the presence of the species *Borelis melo*.

Borelis melo and *Borelis haueri* have been found in Yugoslavia in Badenian formations and *Borelis melo* in Lower Sarmatian sediments, too. Each of these species in Central and Eastern Paratethys is mentioned to have come from Badenian sediments.

In the Mediterranean domain, representatives of *Borelis* occur somewhat before that of *Orbulina*, in the early Langhian, and disappear in the late Tortonian, while the above mentioned authors maintain the idea that *Borelis melo* continues to exist into the recent age and is found at present time as *Borelis pulchra* (d'Orb.).

The discovery of *Borelis melo* in the Lower Sarmatian of the Central Paratethys allows a wider correlation with those in the Mediterranean, i.e. zones of Blow (1969). Thus the Lower Sarmatian of the Paratethys could generally correspond to the terminating part of zone N13, to whole of zones N14 and N15 and, finally, to the lower part of zone N16.

The lower Middle Sarmatian also contains instructive microfossil data, in scope of regional correlation of the Western and Eastern Paratethys, but also both of these realms and the Mediterranean. This will be discussed in the following publications of the present author; in the meantime, we hope that *Borelis melo* representatives will be found in the Lower Sarmatian and in other regions of the Central and Eastern Paratethys.

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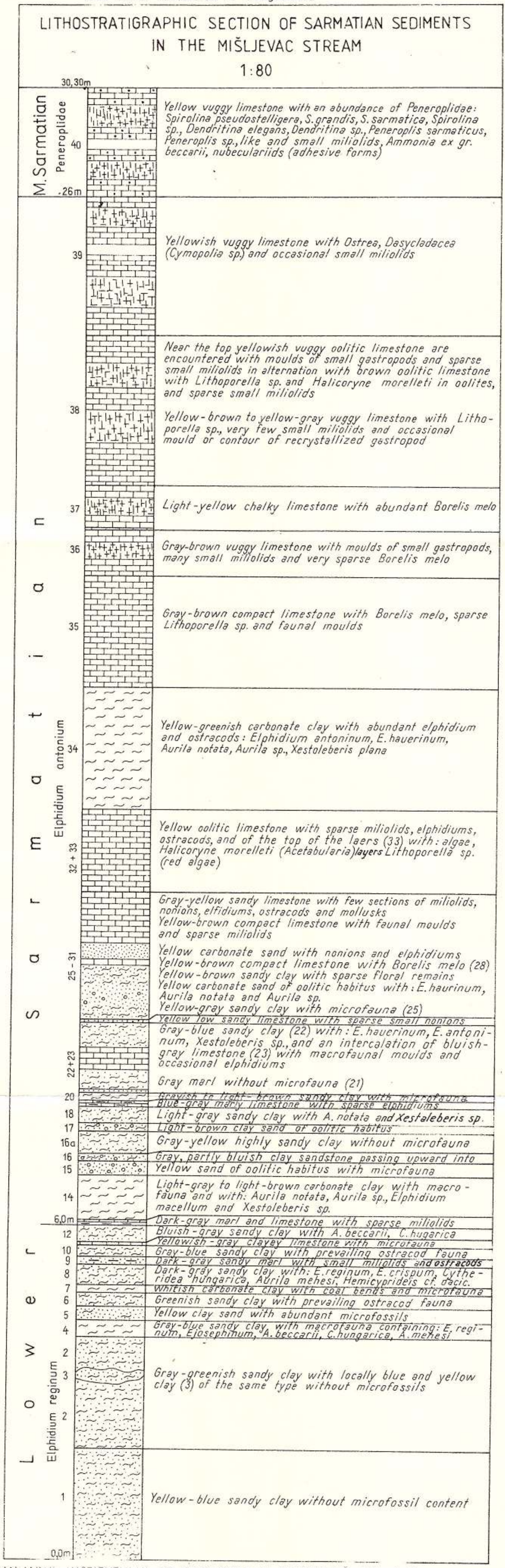
LITHOSTRATIGRAPHIC SECTION OF SARMATIAN SEDIMENTS IN THE MIŠLJEVAC STREAM
1:80
30,30m
M. Sarmatian
Peneroplidae
40
Yellow vuggy limestone with an abundance of Peneroplidae: Spirolina pseudostelligera, S. grandis, S. sarmatica, Spirolina sp., Dendritina elegans, Dendritina sp., Peneroplis sarmaticus, Peneroplis sp., like and small miliolids, Ammonia ex gr. beccarii, nubeculariids (adhesive forms)
26m
39
Yellowish vuggy limestone with Ostrea, Dasycladacea (Cymopolia sp.) and occasional small miliolids
38
Near the top yellowish vuggy oolitic limestone are encountered with moulds of small gastropods and sparse small miliolids in alternation with brown oolitic limestone with Lithoporella sp. and Halicoryne morelleti in oolites, and sparse small miliolids
Yellow-brown to yellow-gray vuggy limestone with Lithoporella sp., very few small miliolids and occasional mould or contour of recrystallized gastropod
37
Light-yellow chalky limestone with abundant Borelis melo
36
Gray-brown vuggy limestone with moulds of small gastropods, many small miliolids and very sparse Borelis melo
35
Gray-brown compact limestone with Borelis melo, sparse Lithoporella sp. and faunal moulds
Elphidium antonium
34
Yellow-greenish carbonate clay with abundant elphidium and ostracods: Elphidium antoninum, E. hauerinum, Aurila notata, Aurila sp., Xestoleberis plana
32 + 33
Yellow oolitic limestone with sparse miliolids, elphidiums, ostracods, and of the top of the laers (33) with: algae, Halicoryne morelleti (Acetabularia) layers Lithoporella sp. (red algae)
Gray-yellow sandy limestone with few sections of miliolids, nonions, elfidiums, ostracods and mollusks
Yellow-brown compact limestone with faunal moulds and sparse miliolids
25 - 31
Yellow carbonate sand with nonions and elphidiums
Yellow-brown compact limestone with Borelis melo (28)
Yellow-brown sandy clay with sparse floral remains
Yellow carbonate sand of oolitic habitus with: E. haurinum, Aurila notata and Aurila sp.
Yellow-gray sandy clay with microfauna (25)
Yellow low sandy limestone with sparse small nonions
22+23
Gray-blue sandy clay (22) with: E. hauerinum, E. antoninum, Xestoleberis sp., and an intercalation of bluish-gray limestone (23) with macrofaunal moulds and occasional elphidiums
Gray marl without microfauna (21)
20
Grayish to light-brown sandy clay with microfauna
Blue-gray marly limestone with sparse elphidiums
18
Light-gray sandy clay with A. notata and Xestaleberis sp.
17
Light-brown clay sand of oolitic habitus
16a
Gray-yellow highly sandy clay without microfauna
16
Gray, partly bluish clay sandstone passing upward into
15
Yellow sand of oolitic habitus with microfauna
14
Light-gray to light-brown carbonate clay with macrofauna and with: Aurila notata, Aurila sp., Elphidium macellum and Xestoleberis sp.
6,0m
Dark-gray marl and limestone with sparse miliolids
12
Bluish-gray sandy clay with A. beccarii, C. hugarica
Yellowish-gray clayey limestone with microfauna
10
Gray-blue sandy clay with prevailing ostracod fauna
9
Dark-gray sandy marl with small miliolids and ostracods
8
Dark-gray sandy clay with: E. reginum, E. crispum, Cytheridea hungarica, Aurila mehesi, Hemicyprideis cf. dacic.
7
Whitish carbonate clay with coal bends and microfauna
6
Greenish sandy clay with prevailing ostracod fauna
5
Yellow clay sand with abundant microfossils
4
Gray-blue sandy clay with macrofauna containing: E. reginum, E. josephinum, A. beccarii, C. hungarica, A. mehesi
2
3
Gray-greenish sandy clay with locally blue and yellow clay (3) of the same type without microfossils
2
Elphidium reginum
1
Yellow-blue sandy clay without microfossil content
0,0m
Lower Sarmatian

BIOCHRONOLOGICAL SIGNIFICANCE OF THE NEOGENE MACROFLORAS IN ROMANIA [1]

BY

RĂZVAN GIVULESCU [2], NICOLAE ŢICLEANU [3]

1. Introduction

The use of Neogene fossil plants for biochronological purposes has constituted a constant concern of the Romanian palaeobotanists the more so as in Romania there are many deposits which can be dated only on the basis of the macroflora content.

The first method used for dating the deposits with fossil plants was the comparison with palaeofloras whose age was specified on macrofaunistical criteria (P o p , 1936).

Another method that can be applied only to rich fossil floras, based on the ratio between the exotic and the native elements, was intoduced by S z a f e r (1946) and used by G i v u l e s c u (1957, 1961, 1969).

Lately, G i v u l e s c u (1979) studying the distribution in time of the fossil alga *Cystoseirites partschi* has ascertained that in Romania it reaches its maximum development in the Lower Sarmatian; thus he succeeded to date the flora at Tîmpa.

The intensification of the palaeobotanic investigations led to the determination of fossil plant associations characteristic of certain time spans (G i v u l e s c u , 1967 a; Ţ i c l e a n u , M i c u , 1980).

In the last decade the progress of the palaeobotanic researches in Romania led, on the one hand, to the finding out of new fossil floras and the completion of the palaeoflora synopsis and, on the other hand, to the revision of known palaeofloras. Concomitantly, the stratigraphic studies specified the age of some floras and thus a more correct interpretation of their evolution was made and a new attitude towards the biochronologic significance of the fossil plants was taken up.

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2. Analysis of the Biostratigraphic Value of the Taxa

With a view to analyzing the biostratigraphic value of the taxa, in the present paper we have presented only the taxa described on the basis of foliar or fruit imprints, both from the papers printed up to now (Givulescu, 1960, 1966, 1973, 1978; Semaka, Givulescu, 1965; Givulescu, Ţicleanu, 1977) and from some papers which are in press or unpublished data.

The Neogene flora synopsis in Romania includes 579 taxa specifically determined, out of which the seventh part may be analysed in biostratigraphic respect. It is because most of the taxa known till now — — 382 taxa — have been found only in one or two fossiliferous sites and other 97 are not certain. The taxa with obvious common characters — — *Typha latissima*, *Phragmites oeningensis*, etc. — have not been discussed. Some taxa, although mentioned once in the Miocene fauna, are of special significance as they continue in the Oligocene flora.

Figure 1, including the selected taxa, shows 5 main groups of taxa, developed in the following intervals[4]: Oligocene-Lower Burdigalian, Oligocene-Pannonian E, Aquitanian-Pliocene, Sarmatian-Pliocene, Pliocene-Lower Pleistocene.

There is only one taxon — *C. partschi* — that occurs only in the Volhynian-Lower Bessarabian, which it characterizes.This fact is due to its nature, this taxon being a marine alga.

2.1. *Oligocene-Lower Burdigalian Taxa*

Although in a small number, the fossil plants of this group *(Lygodium kaulfusii, O. lignitum, C. furcinervis, B. miocenicum)* have biostratigraphic significance because, in association with elements which begin their evolution in the Aquitanian, they characterize the Aquitanian-Lower Burdigalian.

2.2. *Oligocene-Pannonian E Taxa*

The number of the taxa typical of this group is considerable : *S. abietina*, *S. sternbergii*, *T. dubium*, *L. salicornioides*, *A. gaudinii*, *E. orsbergensis*, *M. lignitum*, *C. bilinica*, *B. prisca*, *C. decheni*, *D. cinnamomeum*, *P. princeps*, *P. braunii*, *O. herii*, *B. antiqum*, *L. primigenia*, *L. pseudoprinceps*, *P. platanifolia*, *S. falcifolius*, *Z. zizyphoides*, *R. rossmasleri*, *A. protogea*, *G. lyelliana*, *G. knorii*, Palmae div. sp. These palaeoflora elements characterize the Miocene flora in Romania, differentiating it from the Pliocene one by the large number and frequency of the Lauraceae, the presence of exotic Juglandaceae and Fagaceae, especially *E. orsbergensis* and *C. decheni*, as well as the palmtrees, their last specimen — — possibly relic — being described from the flora at Băiţa (Givulescu, Rüffle, 1971).

Some taxa exceed little the Pannonian E, they being relics since this interval or even earlier. Exceptionally taxa of this group have been mentioned in the Pliocene and Lower Pleistocene but most of them seem to be errors of determination.

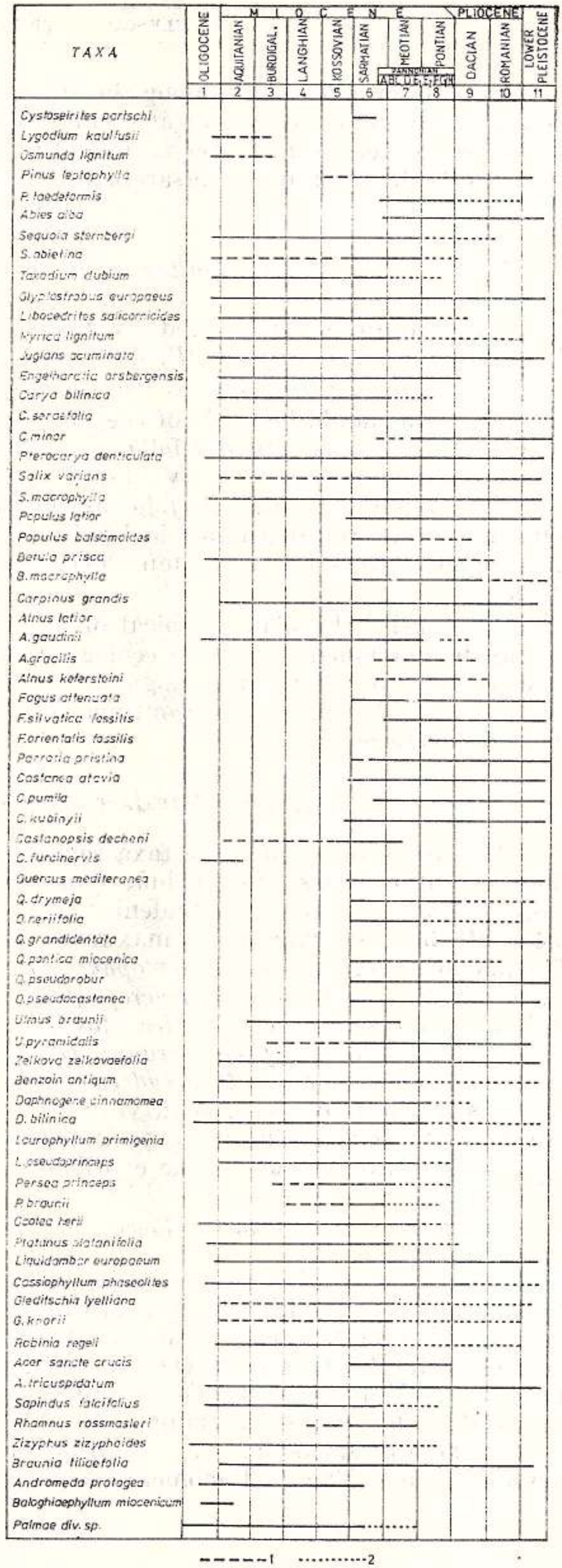

Fig. 1. – Stratigraphic occurrence of the main Neogene taxa in Romania.

It is noteworthy that among the Miocene taxa there are some which have a maximum development at certain stratigraphic intervals : *L. salicornioides* as well as *E. orsbergensis* and *G. lyelliana* are frequent in the terminal Badenian-Lower Bessarabian and *P. platanifolia* in the Bessarabian.

2.3. *Aquitanian-Pliocene Taxa*

This group is represented by the following taxa: *G. europaeus*, *J. acuminata*, *C. sereaefolia*, *P. denticulata*, *S. macrophylla*, *C. grandis*, *L. europaeum*, *C. phaseolites*, *A. tricuspidatum*, and *B. tiliaefolia*.

Mention should be made of the taxa specific to the coal-forming bog facies : *G. europaeus*, *B. tiliaefolia*, *L. europaeum*, and *A. tricuspidatum*. In our previous papers (Givulescu, 1967 b) we have pointed out that *G. europaeum* + *B. tiliaefolia* association presents maximum development at several stratigraphic levels, beginning with the Sarmatian. *L. europaeum* has its maximum development in the Upper Pontian-Dacian.

Obviously, the plants typical of this group are very long—ranged; it may be explained by a wide ecologic valency and especially by adaptations to temperature, an ecologic factor which, in spite of oscillations, had a general background of continuous decrease from the Oligocene up to the Pleistocene.

2.4. *Sarmatian-Lower Pleistocene Taxa*

The appearance of these taxa represents a decisive moment in the floral evolution. These taxa include types which appear in the Sarmatian, more rarely in the terminal Badenian, and reach the Lower Pleistocene, most of them showing their maximum development in the Pliocene. The taxa of this group are: *P. leptophylla*, *P. taediformis*, *A. alba*, *C. minor*, *P. latior*, *P. balsamoides*, *B. macrophylla*, *A. latior*, *A. gracilis*, *A. kefersteini*, *F. attenuata*, *F. silvatica fossilis*, *Parrotia pristina*, *C. atavia*, *C. pumila*, *C. kubinyi*, *Q. mediteranea*, *Q. drymeja*, *Q. neriifolia*, *Q. pontica miocenica*, *Q. pseudorobur*, *Q. pseudocastanea*, *A. sancte crucis*.

As one may observe, with very few exceptions, all these elements are Arcto-Tertiary, being in contrast to the group of elements characterizing the Miocene, where the exotic taxa were prevailing.

2.5. *Pliocene-Lower Pleistocene Taxa*

The taxa of this group begin their evolution in the Pannonian G/H, concurrently with the flora of Chiuzbaia, when the first species of *Quercus* of roburoid type occur (Givulescu, 1980). The maximum development of the species *Quercus* with a lobate leaf, beside that of Betulaceae, Ulmaceae and *Fagus*, constitutes the characteristic feature of the Pliocene flora in Romania.

Generally speaking, the evolution of the Neogene flora has three main moments. The first moment, in the Aquitanian, characterized by the appearance of many new types, which will constitute the Miocene flora, some of them reaching the Pliocene. The second moment is located in the basal Sarmatian when several new types appear which will coexist with old Miocene elements up to the Pannonian E. Hence a highly important concurrence zone resulted, superposed to the Sarmatian-Pannonian E interval. The third moment coincides with the Chiuzbaia flora — Pannonian G/H — when most elements constituting the Pliocene flora appear.

Following the extension in time of the main taxa of the Pliocene flora one may observe the persistence of most of them on large stratigraphic intervals, generally including several stages; it reduces the possibility of their use in the definition of certain biostratigraphic units characterized by a wide areal distribution within a short time interval. In spite of this fact some biozones, characterizing shorter intervals can be delimited. Thus, the concomitant existence of the taxa *C. furcinervis*, *L. kaulfussi* and of Arcto-Tertiary elements represents a concurrence situated in the Aquitanian, possibly the Burdigalian base. Close to the size of a stage is also the concurrence zone between *A. alba*, which occurs in Pannonian B/C, and the last appearance of the species *E. orsbergensis*, found in the Pannonian E_1 of Delureni.

In case of the three species of ***Fagus*** (acc. to Givulescu, unpublished data), although their occurrence does not present necessarily distinct biostratigraphic limits, two interval biozones can be delimited: the former only with *F. attenuata*, corresponding to the Sarmatian (*sensu* Suess), and the latter, when *F. attenuata* coexists beside *F. silvatica fossilis*, covering the Upper Sarmatian-Lower Pontian interval.

Although the limits of an acme zone may be subjective, in case of the taxon *C. partschi* there is such a zone in the Volhynian-Lower Bessarabian when, probably due to ecologic optimum conditions, the taxon is developed on a large area from Hungary up to the Caucasus.

3. Stratigraphic Value of the Fossil Plant Associations

In order to obtain an image of the succession in time of the different fossil plant associations we used first of all the evidence given by palaeofloras with numerous plants (e.g. Coruş, Valea de Criş, Chiuzbaia) and then the information from several synchronous fossiliferous sites with poorer floras, thus establishing associations based on the frequency of the elements.

We have taken into account the possibility that two floral provinces would have existed simultaneously on the Romanian territory. Thus, for the Volhynian-Lower Bessarabian interval (possibly Middle Bessarabian), a discordance occurs between the fossil plants in the north of the country and those in the central and southern parts. The northern province is characterized by the predomination of the betulaceous type and scarcity of the Lauraceae and Fabaceae. It can be compared with the associations presented by Knobloch (1973) from Hungary and Czechoslovakia. In the central-southern province, the Lauraceae and Fabaceae are found beside Betulaceae which are less frequent, as well as palmtrees. The two provinces have been climatically conditioned, the southern one having a climate with dry influences.

Some palaeophytocenoses are generally represented by one vegetal association (e.g. the Aquitanian lauracee association), others, especially beginning with the Pannonian E, correspond to at least two vegetal associations : one of them being forest moor with *G. europaeus* and the other one with mesophytic forests with the predominance of the Fagaceae, Ulmaceae and Betulaceae.

The succession of the fossil associations (Plate) allows us to ascertain the existence of at least three types. The first association is of Lower Miocene age, within which the exotic Lauraceae and Fagaceae prevail, the Arcto-Tertiary elements being found sporadically. The second association is found in the Badenian-Pannonian E_2 interval and is represented by Lauraceae, Fabaceae, Juglandaceae *(E. orsbergensis)*, *Myrica*, *Sapindus*, as well as a few Arcto-Tertiary elements. Concurrently with the Chiuzbaia flora (Pannonian G/H) a clear domination of the Fagaceae, Betulaceae and Ulmaceae is observed.

The reconstitution of the plant associations and the study of their succession in time may bring important contributions to the Neogene biostratigraphy; such a succession of the main plant associations has already been outlined (Givulescu, 1980).

In the present stage of palaeobotanic research we consider that almost all fossil plant associations — except the Lower Burdigalian and Lower Badenian ones which are less known — can be carefully used in biostratigraphic correlations (see Plate).

4. Ratio between Exotic and Native Elements

At present, the ratio between the exotic and native elements constitutes one of the most certain methods of age determination (Givulescu, 1957, 1961, 1968).

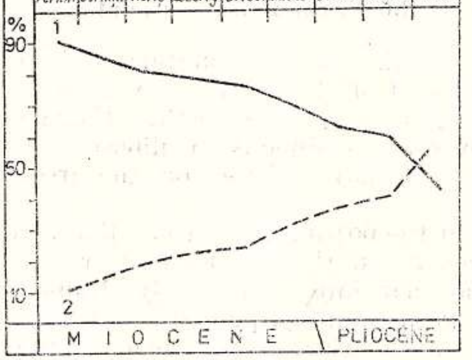

Fig. 2. — Aspect of the curves of exotic elements (1) and native elements (2) for the Upper Miocene floras of Romania and Europe.

The diagram in Figure 2 points out that the age determination is more easily made for floras whose percentage value for exotic and native elements can be calculated.

Although this method has a high degree of precision it is limited first by the scarcity of floras with a sufficient number of taxa which allow the establishing of a correct ratio between the exotic and native elements, and second by the inherent difficulties to establish living equivalents of the fossil taxa. The latter inconvenience can be eliminated by the intensification of the palaeobotanic researches.

5. Conclusions

The use of fossil plants for biostratigraphic correlations for Neogene deposits still requires caution : first of all due to the insuficient knowledge of the fossil flora and secondly because of the reduced number of palaeoflora with enough taxa for the qualitative and quantitative statistic studies. There are also difficulties in connection with the large number of long-ranged taxa some of them sometimes ranging from the terminal Oligocene to the Lower Pleistocene. The facies plant associations (the moor with *G. europaeus*, etc.) persist, without significant qualitative variations, during long time intervals. We should take into account the possibility of the existence of palaeofloristic provinces.

In spite of all these difficulties, at present the age determination on the basis of fossil plants using individual taxa, groups of taxa, fossil associations and, obviously, the combination of them, is possible, at least for some time intervals.

[4] The stratigraphic divisions used in this paper are according to Moisescu, Popescu (1980) and Andreescu (in press).

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AGE				FOSSILIFEROUS SITES	FOSSIL PLANT ASSOCIATION
PLEISTOC.	LOWER PLEISTOC.			Biborțeni Baraolt, Miclușoara, Doboșeni	*Q. robur, Q. pseudosuber, Quercus aff. infectoria, Quercus aff. trojana, Acer monspesulanum, F. silvatica, F. orientalis fossilis, Z. zelkovaefolia, L. europaeum, Carpinus betulus, C. orientalis*
PLIOCENE	ROMANIAN			Borsec	*F. attenuata, Z. zelkovaefolia, Castanea vesca, C. kubinyii, Carya minor, Q. roburoides, cf. Cassia phaseolites, Populus latior, Pterocarya denticulata*
PLIOCENE	DACIAN			Timișani, Dedovița	*1. G. europaeus, B. tiliaefolia, Salix div. sp., A. tricuspidatum, Alnus sp., B. macrophylla L. europaea* *2. Quercus roburoides, Carya sereaefolia, C. grandis*
PLIOCENE	PONTIAN	PANNONIAN	G/H	Chiuzbaia	*F. attenuata, Z. zelkovaefolia, Quercus div. sp. (lobate leaves), L. europaeum, Carya minor, Vitis strictum, C. grandis, Acer integerrimum, U. pyramidalis, Betula prisca, G. europaeus*
MIOCENE	PONTIAN	PANNONIAN	F	Sinersig, Vișag, Derna, Sărmășag	*1. G. europaeus, B. tiliaefolia, A. cecropiaefolia, B. macrophylla, Salix macrophylla, S. varians, A. tricuspidatum; 2. F. attenuata, F. silvatica, U. pyramidalis, C. grandis, Betula verrucosa, C. berenices*
MIOCENE	PONTIAN	PANNONIAN	E_2	Băița	*1. G. europaeus, Taxodium dubium, Alnus cecropiaefolia, Braunia tiliaefolia, Betula macrophylla* *2. Carpinus grandis, U. pyramidalis, Hovenia dulcis fossilis;* last occurrence of the species *Laurophyllum pseudoprinceps* and of the palm trees
MIOCENE	MEOTIAN	PANNONIAN	E_1	Delureni	*Lauraceae div. sp, Daphnogene div. sp, M. lignitum, G. europaeus, Pinus div. sp., Z. zelkovaefolia, C. kubinyii, Quercus div. sp, L. salicornioides*
MIOCENE	MEOTIAN	PANNONIAN	D		unknown flora
MIOCENE	MEOTIAN	PANNONIAN	C/B	Valea de Criș	*L. salicornioides, M. lignitum, E. orsbergensis, Thuja af. occidentalis, Tsuga europaea, Pseudotsuga cf. taxifolia; Castanopsis decheni, Q. drymeja, Z. zelkovaefolia, Fabaceae div. sp., D. cinnamomeifolia, C. kubinyii, S. falcifolius*
MIOCENE	MEOTIAN	PANNONIAN	C/B	Cornițel	*Glyptostrobus europaeus, L. europaeum, Pinus div. sp. (binae), D. bilinica, D. cinnamomeu, Juglans cinerea fossilis, C. kubinyii, C. decheni, P. platanifolia, Populus latior, Alnus kefersteini, Z. zelkovaefolia, Fabaceae div. sp.*
MIOCENE		PANNONIAN	A		unknown deposits
MIOCENE	SARMATIAN	SARMATIAN (S. Suess)		Racșa, Cavnic, Corni, Baia Sprie Luncșoara, Daia, Sacadat, Feleac, Fizeș, Tîmpa, Deva, Slătioara, Aușeu	Northern province: *Z. zelkovaefolia, Quercus pseudocastanea, Liquidambar europaeum, J. acuminata, Fagus attenuata, M. lignitum, Parrotia pristina, U. pyramidalis, Acer tricuspidatum* Southern province: *C. kubinyii, S. falcifolius, Platanus platanifolia, Pinus div. sp. (binae), L. salicornioides, Lauraceae div. sp. Fabaceae div. sp. E. orsbergensis, M. lignitum, B. dryadum, Z. zelkovaefolia*
MIOCENE	KOSSOV.			Ciocadia, Pîrlage, Hășdate	*E. orsbergensis, Castanea kubinyii, Pinus sp. (binae), Libocedrites salicornioides, Gleditschia lyelliana, C. berenices, M. lignitum, Betula dryadum, Sapindus falcifolius, D. cinnamomeifolia*
MIOCENE	LANGH.			Căstău-Orăștie and other sites	*Juglandaceae (Juglans, Carya), Lauraceae (L. primigenia, Neolitsea magnifica, D. bilinica, Persea princeps), Pinus div. sp.*
MIOCENE	BURDIGALIAN	UPPER			unknown floras
MIOCENE	BURDIGALIAN	LOWER		Tihău	*Daphnogene cinnamomeifolia, D. bilinica, Litsea sp., M. lignitum, Juglans acuminata, Carya denticulata, C. mirabilis, Alnus cf. nepalensis, Carpinus cuspidens, U. pyramidalis, U. braunii, Z. zelkovaefolia, Cassiophyllum berenices*
MIOCENE	AQUITANIAN			Coruș	*Daphnogene div. sp. (D. bilinica, D. cinnamomeum etc.) Laurophyllum div. sp. Engelhardia orsbergensis, Castanopsis furcinervis, Myrica lignitum, Lygodium kaulfussii, L. gaudinii, Rhus neoggerathi, Steinhaurea subglobosa, Ulmus pyramidalis, Palmae div. sp.*

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ICHTHYOFAUNA CHANGES IN THE TERTIARY OF THE CARPATHIANS AND OF THE CAUCASUS [1]

BY

ANNA JERZMANSKA [2], JANUSZ KOTLARCZYK [3]

Well-preserved imprints of fish skeletons of the Teleostei group were reported since many years to occur in the Menilite-Krosno Series of the Carpathians (detailed bibliography is given by Kotlarczyk and Jerzmańska, in press) and in the Maikop Series of the Caucasus and adjacent areas (Daniltshenko, 1960; Jafarova, 1964; Fedotov, 1976; Daniltshenko, Endelman, 1977). Fairly long time interval of accumulation of these deposits (from Lower Oligocene to Burdigalian, inclusively, i.e. cca. 20 m.y.) suggests the possibility of variation of the Ichthyofauna in time. In fact, as follows from detailed studies, there are distinct analogies between successions of some ecological assemblages and simultaneous appearance of some genera in both regions.

Consequently, the present authors attempted a correlation of the Carpathian and Caucasian fish assemblages. The study consisted, first of all, in estimating to what degree the ichthyozones of local ecostratigraphic subdivision for the Carpathian region (Kotlarczyk, Jermańska, 1980) confirmed by otolith examinations (Brzobohaty, 1981) can be applied in the Caucasian Basin. The obtained biostratigraphic data, allowing to make such an attempt, can be summarized, as follows:

In the Carpathians — during the deposition of the Oligocene-Lower Miocene Menilite-Krosno Series there were distinguished (Kotlarczyk, Jermańska, 1976: lower bathypelagic assemblage (IPM 1 zone) younger neritic-sublittoral assemblage (IPM 2 zone) and upper bathypelagic assemblage (IPM 3 to IPM 6 zones). Ecological character of the youngest, still poorly examined, Lower Miocene assemblage

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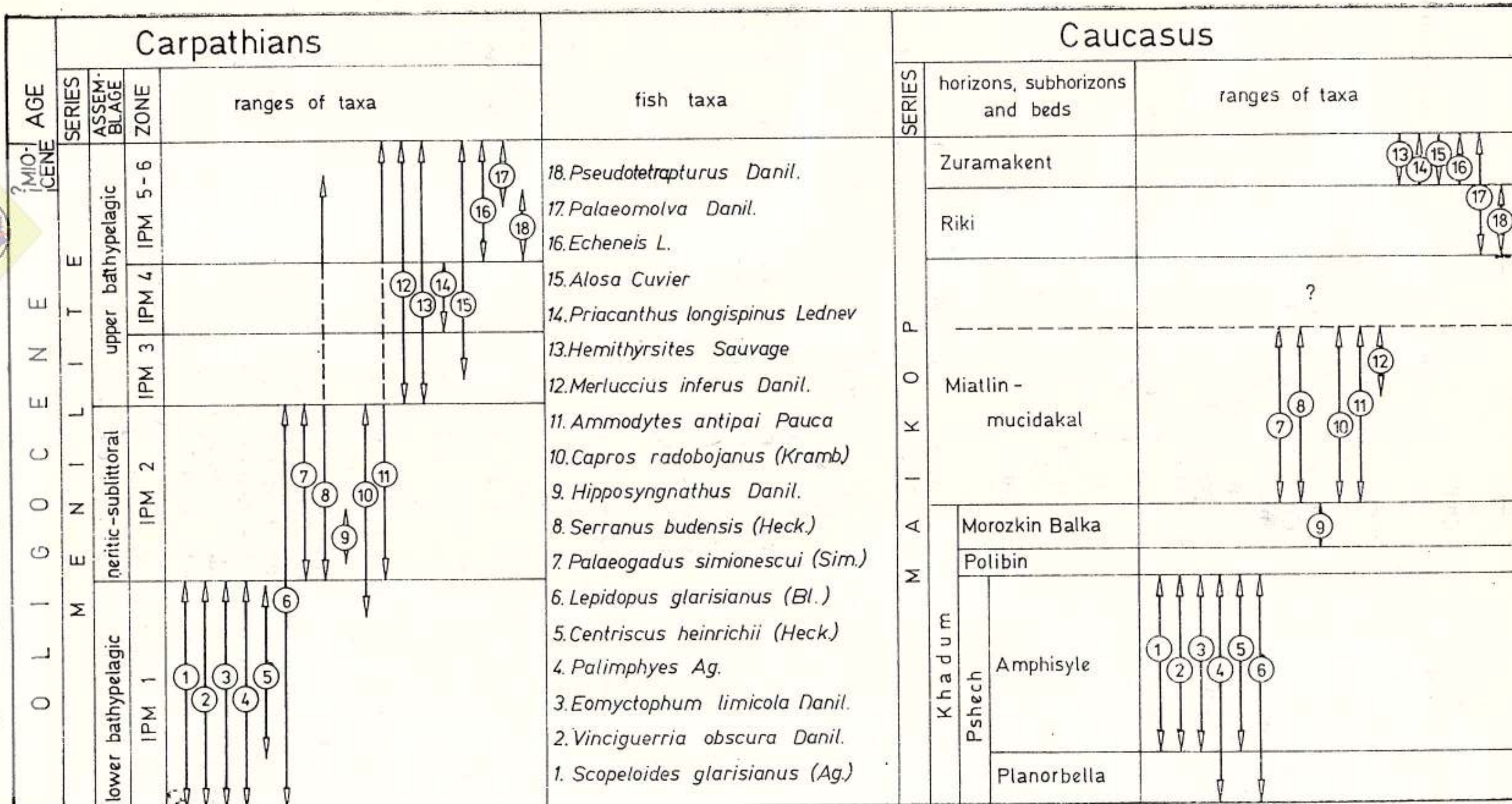
(INM O zone) (Jerzmańska, Kotlarczyk, 1981) cannot yet be established precisely.

In the Caucasus – in the Maikop Series Daniltshenko (1960) established the occurrence of different Ichthyofaunal assemblages in stratigraphic sequence. The oldest deep-water fish assemblage of the Pshech Subhorizon display different taxon composition in its lower and upper parts. This refers first of all to lack of fishes with photophores (Gonostomatidae, Sternoptychidae, Myctophidae) in lower parts (Planorbella Bed) and their appearance only in the upper one (Amphisyle Bed). On this basis, Daniltshenko (1960, p. 175) evaluated the depth of this basin to be at least 1000 m. In the overlying sediments (Polibin and Morozkin Balka subhorizons) the fish remnants are very scarce. Worth mentioning is the occurrence of shallow-water *Hipposyngnatus convexus* Danil. in the Morozkin Balka Subhorizon (Daniltshenko, 1960, p. 85). In the upper horizons (Miatlin-Mucidakal, Riki and Zuramakent), there appears a fish assemblage which, despite some differences in temporal occurrence of taxons, is characterized by lack of bathypelagic fishes. Daniltshenko (1960) explains this fact by changes of depth of the upper boundary of hydrogen sulphide zone in the basin. Consequently, according to the fish data, he does not estimate the depth of the basin but that of water layer free of this poisoning gas. During the deposition of the Miatlin-Mucidakal Horizon sediments, this layer would be not more than 300 m thick, whilest during sedimentation of those of the Zuramakent Horizon – even 400 – 500 m (Daniltshenko, 1960, pp. 176, 179). Later discovery of benthic Pleuronectiformes in the Zuramakent Horizon (Jafarova, 1966) and the development of abundant benthopelagic Gadidae in all the horizons of the Maikop Series (Fedotov, 1976) impairs from paleontologic viewpoint Daniltshenko's hypothesis on the variations of depth of the H_2S layer as the cause of ecological changes of Ichthyofaunal assemblages in the Caucasian Basin.

The present state of taxonomic elaborations of individual Teleostei families from the Oligocene deposits of the Caucasus and the Carpathians is far from being complete. It seems, therefore, that the number of species, described till now for each of these areas, cannot reflect the original diversity and endemism of Ichthyofauna in parent basins, representing the same biogeographical province. This conclusion results e.g. from the fact that apart from comparatively small amount of common fish species, corresponding to cca. 20 percent of all of them, much more numerous are those "endemic" for each of the areas in question, even for their different parts. Successive revisions indicate that these are only synonyms of earlier estimated forms. It seems that monographic elaboration of individual families should increase the amount of taxons common for these areas. This can be exemplified by Gadidae, being one of the better examined families in the last decade. In the Miatlin-Mucidakal Horizon of the Caucasian Paleogene deposits, Fedotov (1976) established the occurrence of *Palaeogadus simionescui* (Sim.), being the index species in the IPM 2 zone of the Carpathians. Besides, our recent studies have shown the presence of the genera *Palaeomolva* Danil. (Gadidae) and *Pseudotetrapturus* Danil. (Palaeorhynchidae) (Jerzmańska –



Tentative correlation of the Carpathian ichthyozones with the Caucasian fish horizons.

unpublished data) in the upper part of the Menilite-Krosno Series. These genera were considered to be endemic for the Caucasus (D a n i l t s h e n k o, 1960 ; F e d o t o v, 1976).

The range of this paper does not allow for a detailed taxonomical discussion. Nevertheless, it is necesary to emphasize an urgent need of revision of other Teleostei families on the ground of comparative studies on the collections from the Caucasus and the Carpathians. The estimation of endemism and heterogeneity of Ichthyofauna in these areas is not possible without strict cooperation of palaeoichthyologists.

As follows from the presented comments, the parallelization of fish assemblages in the Caucasus and in the Carpathians will be based both on the occurrence of species and of genera. The latter ones are used only in these cases when the form cited occurs only in one type of ecological assemblage (e.g. deep-water).

The similarity of taxonomic composition of fish assemblages of the Caucasus and the Carpathians in individual parts of the porfiles is presented in Figure. The following conclusions can be deduced from this comparative study.

1. The time range of occurrence of pelagic fishes of the *Palimphyes* A g . genus and *Lepidopus glarisianus* (B l .) is the linking factor of the IPM 1 zone with the Pshech Subhorizon in the Caucasus. Other species of the IPM 1 zone occur in the Caucasus exclusively in the Amphysile Bed. In our opinion, this heterogeneity of fish assemblage of the Pshech Subhorizon is ecological in character, being caused by some differences in the depth of basin, in which the occurrence of deep-water fishes with photophores (*Scopeloides glarisianus* A g., *Vinciguerrua obscura* D a n i l., *Eomyctophum limicola* D a n i l.) was possible during the sedimentation of the Amphysile Bed only. It seems, however, that the Planorbella Bed was not deposited in a fairly shallow basin since apart form the *Palimphyes* genus there occurs here *Gephoroberyx robustus* (B o g .), close to contemporaneus *Gephyroberyx darwini* (J o h n s o n) (D a n i l t s c h e n k o P . G., p. 87). The latter species is a pelagic form, living at depths from 200 to cca. 500 m (N i e l s e n, 1973). Other fossil genera (*Palaeotroctes* D a n i l., *Protobrotula*/D a n i l. /, *Holostheus* A g.) from the Planorbella Bed, defined by D a n i l t s h e n k o P. G. (1960) as deep-water ones, cannot be taken into account in estimating the depth of basin, since taxonomical revision in this case is necessary (A r a m b o u r g, 1967 ; J e r z m a n' s k a , 1979).

On the other hand, very interesting and apparently not comprehensible is the occurrence of *Centriscus heinrichi* (H e c k.) in some deep-water deposits of the Carpathian and Caucasian seas. This species, considered to be a shallow-water form (D a n i l t s h e n k o, 1960), could be able to live also in surface ocean waters (J e r z m a ń s k a, 1968). The first scarce individuals of this species appear in the lower part of the Amphysile Bed and occur abundantly only in the middle (D a n i l t s h e n k o, 1960, p. 77) and upper part of this bed (F e d o t o v, 1976, p. 55). In the Carpathians this species, as well as two others, belonging to the *Centriscus* L. genus, occur only in the middle part of the IPM 1 zone and are very abundant (K o t l a r c z y k, J e r z m a n s k a, 1976). It is evident that the lack of *Centriscus heinrichi* in the lowermost part of the IPM 1 zone and in Planorbella Bed could not depend on the depth of basin. In the present authors' opinion, the appearance of this species and its comparatively short duration within the area of the Caucasian and Carpathian basins had to be conditioned by higher temperature of surface waters in this period. Such interpretation is consistent with warming episods between 35 and 32 m.y. B.P., postulated by Haq et al. (1977).

Besides, on the ground of available data, it is not clear why in the Amphysile Bed there occur several genera which are known in the Carpathians only in the upper bathypelagic assemblage. They are represented by: *Argyropelecus cosmovicii* Cosm. and Pauca, *Eomyctophum menneri* Danil., *Eomyctophum koraense* Danil., *Bregmaceros filamentosus* (Priem.). In the case of *Argyropelecus cosmovicii* species, known only from one specimen of old Grossheim collection (Daniltshenko, 1960, p. 32), the possibility of erroneous localization of sampling is not excluded. The occurrence of the latter taxa can be explained by their earlier appearance in the Caucasian Basin than in the Carpathian one. Another possible explanation consists in admitting that the shallowing of the former basin took place later i.e. in the period when fish assemblage known from the upper bathypelagic assemblage already started to develop. However, the estimation of vertical ranges of these genera in the Caucasus and in the Carpathians calls for further studies.

2. Distinct correlation concerning both the composition of taxa and ecological character of assemblages is observed when comparing the IPM 2 zone with the Miatlin-Mucidakal Horizon. Even in the very poor in fish remnants Morozkin-Balka Subhorizon, the presence of shallow-water genus *Hipposyngnatus* Danil., common to both Carpathians and Caucasus, was noted.

3. The appearance of *Merluccius inferus* Danil. in the upper part of the Miatlin-Mucidakal Horizon is connected in this part of the profile with the IPM 3 zone of the Carpathians, where we note the first occurrence of this genus.

4. Lack of deep-water fishes with photophores in the Riki and Zuramakent horizons renders difficult their parallelization with the zones of upper bathypelagic assemblage. However, the estimation of representatives of the *Palaeomolva* Danil. Gadidae genus and of pelagic *Pseudotetrapturus* Danil. (Palaeorhynchidae) and *Echeneis* L. (Echeneidae) genera in the IPM 5 and IPM 6 zones (Jerzmańska — non-published data) allows to compare these beds with the Riki and Zuramakent horizons. Besides, this conclusion is consistent with mass appearance of drifting assemblage of brown algae and associated mass occurrence of shallow-water fishes of the genus *Syngnathus L.* both in the Carpathians (Jerzmańska, Kotlarczyk, 1976a) and in the Caucasus (Hekker, Merklin, 1946). It should also be noted that the appearance of the genus *Echeneis* L. in both the basins can be connected with Late Oligocene warming of oceanic surface waters, which begun 28 m.y. B.P. (Haq et al., 1977).

The differences in the succession of ecological fish assemblages in the area under consideration can be explained by diversely proceeding changes of depth in various parts of the Tethian Basin. Therefore, it is no use to accept the thesis of Daniltshenko (1960) and Fedotov (1976) on the variation of depth of the upper boundary of H_2S-saturated zone of sea-water.

At the beginning of the period in question, the deepest parts of the basin were situated in western part i.e. in the Carpathians. This deepening moved afterwards toward the east, and during the sedimentation of the upper part of the IPM 1 zone, the depth of the basin was the same in the whole region under consideration. Later shallowing of the basin was, most probably, simultaneous in the Carpathians and the Caucasus. In the initial stage of this process, the Caucasian Basin waters were subjected to a freshening process (Polibin and Morozkin Balka subhorizons — Veselov, 1979; Veselov, Luleva, 1980). Subsequent return to normal salinity is marked by the appearance of neritic-sublittoral fish assemblage in the whole basin. Repeated considerable deepening of the Carpathian Basin, which resulted in the appearance of the upper bathypelagic assemblage, did not reach the region of the Caucasus. Though in the latter region there was some deepening marked by the Riki and Zuramakent horizons, the increasing depth was not sufficient for the development of bathypelagic fishes. Such conditions started to govern in the Caucasus region but later in the Tarkhan Horizon, in which bathypelagic forms appear (Daniltshenko, 1960).

The above-presented tentative parallelization of the Caucasian and the Carpathian fish assemblages (mainly pelagic) allows us to accept some definite stratigraphic conclusions.

1) The lower boundary of the Menilite Beds corresponds in age to that boundary of the Pshech Subhorizon.

2) Lattorfian age of the IPM 1 zone suggests the same age of the Pshech Subhorizon. Both conclusions are consistent with the opinions of the Soviet geologists (e.g. Veselov, 1979).

3) The age of both the IPM 2 zone and of the Polibin Subhorizon cannot be determined. Some premises from the Carpathian region indicate that the beginning of shallowing should not be younger than the nannoplankton NP-22 zone in the Martini zonation (Jerzmańska, Kotlarczyk, 1981). Taking into account the data from the Caucasus (Veselov, Luleva, 1980) it should be dated as the NP-23 zone. These results are not contradictory if we assume gradual migration of shallowing toward the east.

4) The correlation of Ichthyofauna of the Riki and Zuramakent horizons with that of the IPM 5 and IPM 6 zones suggests Upper Oligocene or Egerian age of both Caucasus horizons. This conclusion is closer to the opinion of these authors who assign to the Miocene only the latter horizon (Daniltshenko, 1960, Fedotov, 1976) than that of Veselov (1979) who considers both horizons to be Miocene in age. Because of lack of convincing paleontological data for the above mentioned Maikop sequence, our estimation, based on Ichthyofaunal data, does not contradict with any facts known till now.

The presented parallelization and stratigraphic conclusions should be considered as preliminary proposal and presentation of the problem. Its proper and definite solution will be possible only on the ground of more detailed studies in both regions under consideration.

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QUESTIONS

T. Nalbant. It is very difficult to make a comparison between two or more paleogeographical regions if the most, if not all, taxa are not identified at the correct generic and specific level. For instance, you mention only the genera *Alosa* in your list. In my opinion, the majority of species referred by authors (Cosmovici, Paucă, a.o.) to the genera *Clupea* and *Alosa* does not belong to these genera but to other genera, still undescribed.Think that some of them are possibly closely related to the actual genera *Clupeonella*, *Harengula*, rather than to the actual *Clupea* and *Alosa*. Perhaps only a few large specimens appear to be true *Alosa* on the basis of skull bones, vertebral number, fine ray formula, etc. I have examined a large number of remains from the Carpathian Oligocene (Romania) and I have found a very few specimens which can be included in *Clupea* or *Alosa*. Therefore the fossil shads need a careful revision on a worldwide basis. On the other hand, I suspect that the remains actually included in the modern genus *Serranus* represent another error. I hope that a true cooperation between all ichthyologists and paleoichthyologists will solve these problems.

Answer : In an article of this length we cannot discuss the present state of taxonomic knowledge of fossil Teleostei from the Carpathians and the Caucasus. However, some families have been discussed elsewhere (for instance Jerzmańska 1967, 1968, 1979). I would like to inform you that my fellow-worker Dr. W. Szymczyk has been working on osteology and taxonomy of the Paleogene clupeids and myctophids. She is visiting the American Museum of Natural History in New York where she will compare the osteology of fossil and living fishes of these two families. After her studies some taxonomical changes will be inevitable.

Nevertheless, the great resemblance between the fossil fish assemblages from the Carpathians and the Caucasus is very remarkable even on the basis of the present knowledge of the two ichthyofaunas. This resemblance concerns the temporal span of many taxa in the continuous stratigraphic sections, as well. It seems to me that your observation on the fossil clupeids from Romania agrees with the geological range of the *Alosa*. As far as I know the major part of paleoichthyological collections in Bucharest consists of fishes found in the middle of the Menilite Beds where the genus *Alosa* has been unknown.

Referring to the *Serranus*, I would like to emphasize that *Serranus budensis* from the Carpathians and the Caucasus may belong to another genus. According to Oberhlova (1975) this species differs from the living genus *Serranus* in some osteological characters. In spite of the fact that I have a large collection of *S. budensis*, I cannot concentrate on this taxonomical problem because I am now working on a detailed osteological and taxonomical analysis of other families (Sternoptychidae, Echeneidae, Trachichthyidae).

It is hoped that our tentative correlation will stimulate a more intensive study of both those regions and international cooperation among paleoichthyologist as well.

THE KISCELLIAN STAGE (OLIGOCENE). FACIOSTRATOTYPES AT NOSZVAJ (BÜKK MOUNTAINS, HUNGARY) [1]

BY

MIKLÓS KÁZMÉR [2], PÉTER VARGA [2]

The Kiscellian Stage

The Oligocene stages (Lattorfian, Rupelian and Chattian) have been set up in Western Europe. Their stratotype faunas belong to the Boreal bioprovince, therefore their correlation with the Oligocene faunas in the Carpathian Basin brings up several problems. According to B á l d i, (1980) the latter contain much Mediterranean and Indopacific elements. The southern connections and partial endemism of the faunas make the introduction of regional stages inevitable. It is to be hoped that the new stages will be correlated with Western European (Boreal), Mediterranean and other ones in due course.

The necessity for regional stages in the Miocene had been realized in the 1950s. The Regional Committee on Mediterranean Neogene Stratigraphy then established them for the Central Paratethys and the stages have been described in the volumes of the series "Chronostratigraphie und Neostratotypen".

The first regional Oligocene stage, the Egerian, has been established and described in connection with the work on Miocene stages (B á l d i, S e n e š, 1975). A new stage, Kiscellian (pronounced as kesh-tzall-ian) corresponding to "Lattorfian" (*sensu* M a r t i n i, 1969) plus Rupelian has been suggested in an early paper of B á l d i (1966). This proposal was renewed in an official form at the Mediterranean Neogene Congress in Athens (B á l d i, 1979a).

The essence of his proposition is as follows: As the Paratethys has been separated from the Tethys in early Oligocene time (B á l d i, 1980) it is reasonable to introduce the stage Kiscellian for Lower and Middle

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Oligocene strata. The indefiniteness of the stratigraphic position of the Boreal Oligocene stages, especially of the Lattorfian (either if it belongs to the Oligocene or to the Eocene or rather to both series as a transitional stage), makes the introduction of Kiscellian a living question. This stage represents the interval between Priabonian and Egerian. Its lower boundary coincides with the Eocene-Oligocene boundary.

Báldi (1979 a) designated a provisional stratotype for the Kiscellian. The drilling R–8/3 in Budapest contains the lower boundary of the new stage close to the bottom of the Tard Clay Formation. The upper boundary of Kiscellian is defined by the lower boundary of Egerian, at the type locality in Eger.

Faciostratotypes of the Kiscellian Stage

SÍKFŐKÚT-quarry at Noszvaj, 10 km to the east of Eger, Northern Hungary (Fig. 1). The lowermost strata of the Kiscellian Stage are exposed in the quarry and in the nearby ravines.

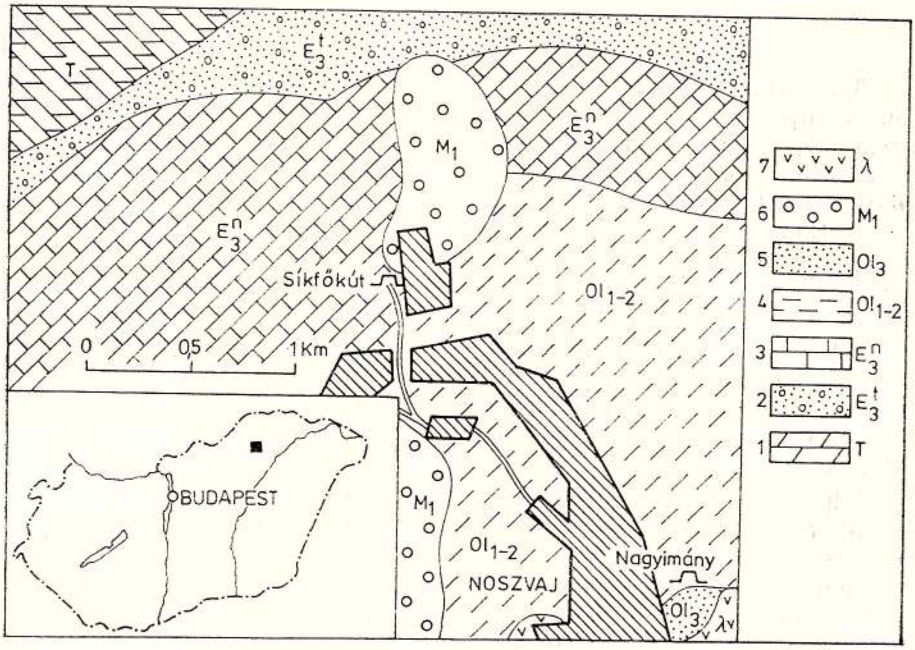

Fig. 1. – Geological map of Noszvaj.
1, Triassic dolomite; 2, Eocene terrestrial conglomerate; 3, Upper Eocene nummulitic limestone and marl; 4, Lower-Middle Oligocene Kiscell Clay Formation; 6, Lower Miocene terrestrial gravel; 7, Lower Miocene rhyolite tuff.

The underlying limestone, calcareous marl and marly beds are characterised by *Nummulites fabianii* Prever, *N. incrassatus* de la Harpe, *Spiroclypeus carpaticus* (Uhlig), *S. granulosus* Boussac, *Grzybowskaia multifida* Bieda and *G. reticulata* (Rütimeyer) (see

Z i l a h y, 1967). The Upper Eocene, Priabonian age of the sediments is proved.

The alternating white marl and yellow-green, glauconitic limestone beds (Fig. 3) of the Sikfőkút-quarry lie conformably on the Priabonian

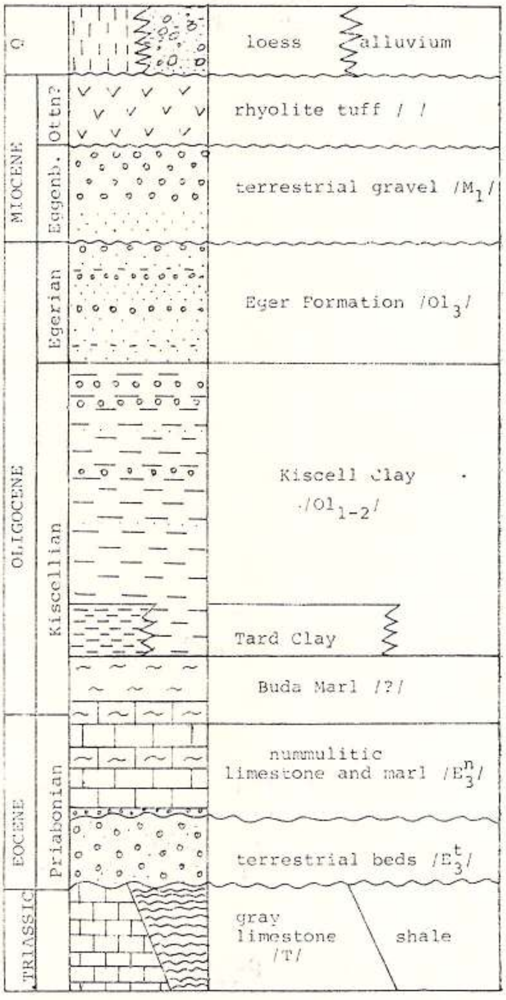

Fig. 2. – Ideal lithological section of the Noszvaj area.

limestone. The topmost layer of the section is Lower Miocene gravel lying unconformably on the Kiscellian.

The prevailing marly sequence contains 20–80 cm thick allodapic limestone beds (M e i s c h n e r, 1964). The sedimentological features of the allodapic beds at Sikfőkút are, as follows :

— sharp contact between marl and limestone at the bottom of the limestone beds ;

— transitional contact between limestone and marly beds at the top of the limestone beds;

— limestone beds containing marly pebbles (occasionally several centimeters in diameter) ripped from the marly substrate;

— the limestone beds consist of well-sorted, well-rounded micro-bioclastic grains, less than 1 mm in diameter: fragments of corallinacean algas, echinoids, nummulitids and other foraminifers, bryozoans, molluscs and worm tubes, altogether shallow sublittoral faunal elements;

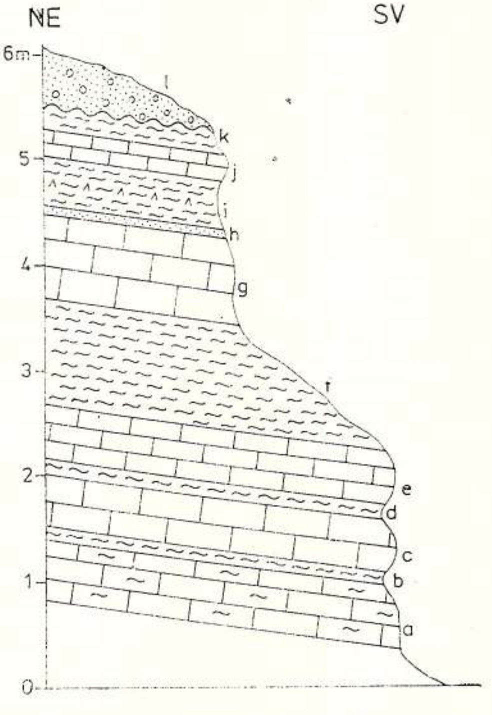

Fig. 3. — Profile of the Sikfökút-quarry.
a, sandy, calcareous marl; b, marl; c, allodapic limestone bed; d, marl; e, allodapic bed; f, marl; g, allodapic bed; *h*, fine sand; i, tuffitic white marl; j, allodapic limestone; k, marl; 1, Miocene gravel.

— in contradiction to the fauna of the limestone beds, the white marl beds contain a deep sublittoral — shallow bathyal foraminifer fauna.

The last two points are especially characteristic of allodapic limestones, i.e. the alternation of beds contanining shallow-water and deeper water fauna.

Some loose layers in the allodapic beds contain a *Nummulites* fauna of little diversity. It is predominated by *Nummulites incrassatus* de la Harpe, but *N. budensis* Hankten and *N. bouillei* de la Harpe occur, too. All of the three species are characteristic of the Upper Eocene formations, but their taxon-range-zones extend beyond the Eocene-Oligocene boundary. On the other hand, the conspicuous absence of *Nummulites fabianii* Prever — whose presence is widespread in the Upper Eocene of the Bükk Mts — indicates Early Oligocene age.

The lower marly layers of the Sikfökút-quarry contain a Bolivina —Globigerina association (M. Horváth, personal communication). The characteristic species are *Bolivina antegressa*, *Bulimina sculptilis*, *B. truncana*, *Uvigerina eocaena* and *Globigerina eocaena*. This association is probably of early Kiscellian age.

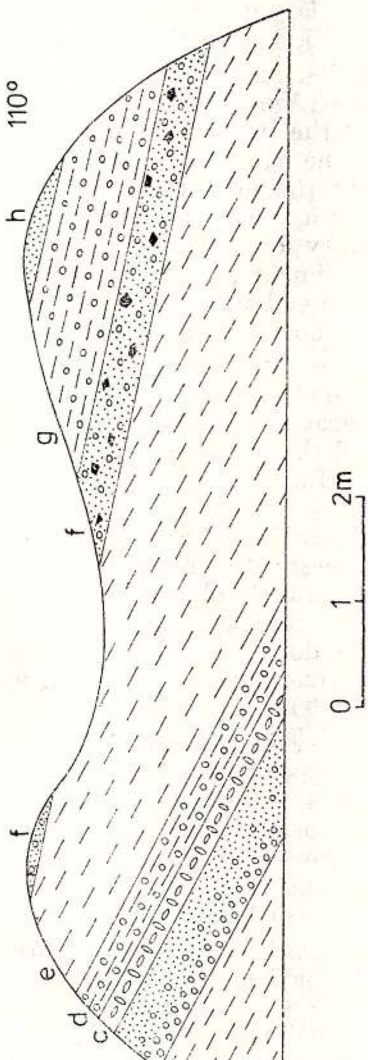

Fig. 4. – The Nagyimány section.

a, Kiscell Clay; b, graded gravel – sandstone with Pectinidae; c, clay with *Saxolucina* in life position; d, pebbly mudstone; e, typical Kiscell Clay; f, sandy conglomerate with angular mud pebbles; g, pebbly mudstone; h, sand.

The upper marly beds in the quarry contain a Bulimina sculptilis – Uvigerina eocaena association which must belong to the Lower Kiscellian.

The nannoflora of the upper marly beds has been examined by A. Nagymarosy (personal communication). These beds belong to the NP 21 – 22 *(Ericsonia subdisticha – Helicopontosphaera reticulata)* zone (*sensu* Martini), i.e. to the Lower Oligocene.

To summarize the results, on the basis of small foraminifera and nannoplankton investigations (*Nummulites* examinations could not indicate the definite position of the beds), the limy-marly sequence of the Sikfökut-quarry is of Oligocene age and belongs to the Kiscellian Stage.

We intend to mark out the Sikfőkút-quarry for one of the faciostratotypes (and possibly for a boundary stratotype) of the Kiscellian Stage, as it provides possibility for biostratigraphical correlation between shallow- and deeper-water sediments. As the excavation work continues it is very likely to find the lower boundary of the Kiscellian stage, which coincides with the Eocene-Oligocene boundary of worldwide importance.

The other faciostratotype of the Kiscellian lies in the northern slope of Nagyimány hill at Noszvaj (Fig. 1). The Kiscell Clay Formation exposed at this locality contains gravel beds (Fig. 4). Earlier authors (Schréter, 1939) have not realized that this sequence violates Walther's Law; they considered it as an interfingering of the Kiscell Clay and Hárshegy Sandstone formations.

The Kiscell Clay at Noszvaj (see Fig. 2) slightly differs from that of the locus typicus in Budapest; it is rather clayey silt (Fig. 4a and e), which contains coarse sand, gravel (b) and conglomerate beds (f). The sand displays graded bedding and contains shallow marine Pectinidae (b). The c clay layer contains double valves of *Saxolucina* sp., embedded in life position. This layer is overlain by a clay bed with centimeter-size, isolated quartzite pebbles (d). This sedimentological feature is the pebbly mudstone of Crowell (1957). The following bed is typical Kiscell Clay(e).

The most important feature of the Nagymány section is the repetition of the above sequence. The second rhythm begins with a coarse clastic (sandy conglomerate) bed (f), which contains angular pebbles of the Kiscell Clay; then follows pebbly mudstone again (q). Here the layer with saxolucinas is missing. The topmost bed (h) of the exposure is coarse sand again : it is the basal layer of the next rhythm.

Báldi (1979 b) recognised that the two rhythms establish an excellent example of fluxoturbidite, defined by Dzulynski et al. (1959). This sequence exhibits several diagnostic characteristics of turbidites : graded bedding, mud pebbles and resedimented shallow marine fossils. Other features of turbidites, like sole marks and convolute bedding are missing. On the other hand, pebbly mudstone and sandy conglomerate are characteristics of sediment slumping. As a combination of features, characteristic of turbidites and slumping as well, are present in the Nagyimány section, it is considered as fluxo-turbidite.

The Kiscell Clay itself has been deposited in a deep sublittoral environment, as megafauna and foraminifera indicate. The shallow marine mollusc fauna of the coarse clastic (sand and conglomerate) beds suffered transportation from its original habitat to the much deeper basin of the Kiscell Clay.

The Nagyimány section represents the uppermost part of the Kiscellian Stage. The Kiscell Clay contains a very well preserved Amphistegina choctawensis association. Characteristic species are: *Neoeponides schreibersii* (Orbigny), *Uvigerina* cf. *gallowayi* Cushman, *Asterigerinata falcilocularis* (Subbotina), *Turborotalia brevispira* (Subbotina), *T.* cf. *obesa* (Bolli). These species indicate a deep sublittoral environment. On the basis of nannoplankton investigations of A. Nagymarosy (personal communication) the sequence belongs to the NP 24 (*Sphenolithus distentus*) zone.

Tibor Kecskeméti has helped to determine the *Nummulites* fauna. Mária Horváth has determined small foraminifera and András Nagymarosy has investigated nannoplankton. Prof. Tamás Báldi has directed and supervised our work. Their help is greatly appreciated by the authors.

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QUESTIONS

Krstić. Tell us some more in fluxoturbidites : width of the area, depth of the sedimentation, etc.

Answer : The Noszvaj Member of the Kiscell Clay Formation is separated by its fluxoturbiditic character. The member lies on the southern flanks of the Bükk Mts, covering an area of several hundreds square kilometres, with a minimal thickness of 100 km. The sand and gravel beds contain hydrocarbons. The paleogeographical situation is the following : a river coming from the north, from among the mountains of the Bükk carried sand and gravel into the Kiscell Sea. Occasionally slumping occurred on the submarine delta fan. The slumping could not develop into a turbidity current because of the relatively shallow (less than 1 km deep) sea. Consequently, an intermediate feature – the fluxoturbidite – came into being, bearing some of the characteristics of a turbidite and some of a slumping. Benthic foraminifers ndicate a sedimentation depth of more than 600 metres (M. Honáth, pers. comm.).

DISCUSSIONS

Em. Kojumdgieva : The stages are separated after their fauna : the regional stages after their regional fauna. The Kiscellian *s.l.* includes three successive faunas (inferior – – marine, middle – with reduced salinity, with *Cardium lipaldi*, superior – marine). That is why we shall have to distinguish three stages, not one.

The Hungarian geologists worked hard and they have the evidence necessary for this division.

A. Rusu : Báldi's interpretation of the Kiscellian in 1979 differs entirely from his initial definition of this stage (Báldi, 1969). Considering Kiscellian the whole interval between the Priabonian and the Egerian, the author includes here an inferior part–marine–with faunas belonging to the Mediterranean Realm, anda superior part – brackish+marine–with faunas typical of the Trans-European province temporarily isolated.

In our opinion a regional stage would be useful only for the interval including the upper part of the Tard Clay + Kiscell Clay, an interval during which took place a first isolation of the realm which later on, definitely isolated, become the Paratethys.

LES COMMUNAUTÉS DE MOLLUSQUES DE L'OLIGOCÈNE DU BASSIN DE LA HAUTE THRACE (BULGARIE MÉRIDIONALE) ET LEUR IMPORTANCE STRATIGRAPHIQUE [1]

PAR

EMILIA KOJUMDGIEVA [2], V. SAPUNDGIEVA [3]

Le bassin (ou la dépression) de Haute Thrace occupe les parties supérieure et moyenne de la vallée de Maritza (Bulgarie méridionale). Les sédiments oligocènes du bassin débutent par les marnes et les argilites de la formation d'Ezerovo (Kojumdgieva, Dragomanov, 1979) succédées par les dépôts lacustres, fluviatiles et continentaux des formations de Dragoinovo (Kojumdgieva, Dragomanov, 1979) et de Maritza (Panov, 1962). La faune provient presque entièrement des forages. Nous exprimons notre reconnaissance aux géologues qui ont récolté cette faune — Dragomanov, Brunkin, Panov, Evstatiev, Levski, — aussi bien qu'aux géologues défunts Merklin et Popov de l'Institut Paléontologique de Moscou.

Les parties inférieures des marnes de la formation d'Ezerovo ne contiennent que des Foraminifères marins d'âge éocène terminal et oligocène inférieur (données non publiées de Dikova). En même temps dans les Rhodopes orientales se forment des calcaires marins contenant une riche faune de Foraminifères, Mollusques, Echinides, etc. (Belmustacov, 1968). Ils faisaient partie du détroit, qui à travers la région de Burgas (Bețov, Dikova, 1969), le bassin de Haute Thrace, les Rhodopes orientales et le bassin de Bas Thrace, reliaient la Paratéthys orientale avec la Téthys.

Les parties moyennes de la formation d'Ezerovo contiennent une communauté de Mollusques à *Loxocardium lipoldi* (Rolle) et *Nucula comta* Goldfuss prédominants, accompagnés par *Janschinella garetzkii* (Merklin) ou *Janschinella melitopolitana* (Nossovsky) et parfois par des rares *Parvicardium scobinula* (Merian), *Lenticorbula sokolovi slussarevi* (Merklin) et *Cerastoderma (Bessia) merklini* sp. n. L'espèce *Nucula comta* est commune dans l'Oligocène moyen et supérieur du Bassin

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[1] Note présentée au 12ème Congrès de L'Association Géologique Carpatho-Balkanique, 8 – 13 septembre 1981, Bucarest, Roumanie.

[2] Bulgarie.

du Nord et de la Paratéthys, *Parvicardium scobinula* a la même distribution et *Loxocardium lipoldi* est connu dans la Paratéthys centrale et fait partie d'une association, qui (au moins en Hongrie) est accompagnée par Nannoplancton de Zone NP-23 (Baldi, 1980). Les autres membres de l'association (*Janschinella abondantes* et rares *Lenticorbula*) correspondent approximativement à celle de l'horizon Polbinien (= couches molochaniennes = couches à Ostracodes = couches à *Ergenica cimlanica*) de la Paratéthys orientale, contenant aussi Nannoplancton de Zone NP-23 (Veselov, 1979; Veselov, Liuleva, 1980).

Le caractère endémique de cette association prouve que pendant l'Oligocène moyen, dans la zone de la Paratéthys, existait déjà un bassin isolé à salinité probablement réduite. Báldi (1979, 1980) propose d'utiliser le nom de la Paratéthys même pour le bassin oligocène, ce qui est raisonable. Le bassin de Haute Thrace en faisait partie, comme golfe profond de la Paratéthys orientale (figure)[3]. Sa faune endémique révèle que pendant l'Oligocène moyen le détroit reliant la Paratéthys orientale à la Téthys n'existait pas. Les Rhodopes orientales s'étaient élevées et séparaient le bassin de Haute Thrace et de la Téthys.

L'association suivante, à *Lenticorbula sokolovi*, est très abondante dans les parties supérieures de la formation d'Ezerovo. C'est ici que prédominent *Lenticorbula sokolovi sokolovi* (Karlov) ou *Lenticorbula sokolovi slussarevi* (Merklin), parfois accompagnées de *Parvicardium popovi* sp. n., *Cerastoderma serogosicum* Nossovsky, *Cerastoderma chersonensis* Nossovsky, *Cerastoderma* (Bessia) *pseudosarmaticum pseudosarmaticum* sp. n., *Peronaea nysti* (Deshayes), *Siliqua nysti* Deshayes, *S. asulcata* Hölzl, *Lenticorbula samodurovi* (Merklin), *Lenticorbula helmerseni transylvanica* (Moisescu), *Janschinella garetzkii* (Merklin), *Janschinella melitopolitana* (Nossovsky).

L'association à *Lenticorbula sokolovi* est voisine de celle de l'horizon sérogosien (= couches morozkiniennes = couches supraostracodes) de l'Ukraine méridionale (Nossovsky, 1962) et de ses analogues dans la région transcaspienne et en Géorgie (Merklin, 1974), ordinairement considérée d'âge oligocène moyen (Merklin, 1964; Nossovsky et Pasicin, 1965; Merklin, 1974; Veselov, 1979).

Les horizons polbinien et sérogosien de la Paratéthys orientale sont semblables, souvent difficiles à séparer d'après leur faune. A coté de quelques formes boréales euryhalines (*Nucula*, *Peronaea*, *Siliqua*, peut-être *Parvicardium* et *Losocardium*) y prédominent les espèces et même les genres endémiques (*Janschinella*, *Lenticorbula*, les représentants de *Cerastoderma*), qui prouvent qu'il s'agit d'un bassin isolé à salinité plus ou moins réduite. On pourrait penser à leur séparation comme étage régional pour la Paratéthys orientale (ils ont été nommés de par Jijcenko, 1965, 1973, étage Belozérien et par Merklin, 1964, Merklin, 1974, horizon solénovien). Nous nous abstenons pour le moment d'utiliser ces noms afin d'éviter les malentendus.

Les couches à *Loxocardium lipoldi* d'Hongrie ne correspondent qu'à l'horizon polbinien de la Paratéthys orientale. Elles sont recouvertes par les couches kiscelliennes marines, contenant du Nannoplancton de la zone NP-24 (Báldi, 1979, 1980), qui sont, probablement, entièrement ou partiellement synchrones à l'horizon sérogosien. Ce fait montre que les différences de l'évolution paléogéographique de la Paratéthys centrale et orientale existaient à partir de l'Oligocène. Les argiles kiscelliennes sont d'âge oligocène moyen- rupélien (S t e i n i n g e r et al., 1976).

L'extension stratigraphique des couches à *Lenticorbula* en Transylvanie est objet de discussion (M o i s e s c u, 1972, 1975, 1977 ; R u s u, 1972, 1977), mais est nettement plus grande que celle des couches à *Loxocar-*

TABLEAU

<table>
<tr><th colspan="3"></th><th colspan="2">Hongrie</th><th>Haute Thrace</th><th colspan="2">Paratéthys Orientale</th></tr>
<tr><td rowspan="4">O L I G O C È N E</td><td>supérieur</td><td>NP-25</td><td colspan="2">Egerien (marin)</td><td>ass. à Polymesoda convexa brongniarti (salinité basse)
ass. à Lent. helmerseni (sal. réduite)</td><td colspan="2">couches ascaniennes ou baigubekiennes à Lenticorbula helmerseni (marines)</td></tr>
<tr><td rowspan="2">moyen</td><td>NP-24</td><td colspan="2">couches de Kiscell (marinesi)</td><td>ass. à Lenticorbula sokolovi (salin. réduite)</td><td>horizon serogosien (sal. réduite)</td><td rowspan="2">Belozerien (Solenovien)</td></tr>
<tr><td>NP-23</td><td rowspan="2">couches de Tard</td><td>ass. à Lox lipoldi (sal. red.)</td><td>ass. à Lox lipoldi et Nucula comta (salin. réduite)</td><td>horizon polbinien (sal. réduite)</td></tr>
<tr><td>inférieur</td><td>NP-21/22</td><td>Foraminifères marins</td><td>Foraminifères marins</td><td colspan="2">Hadumien s.s. (marin)</td></tr>
</table>

dium lipoldi d'Hongrie et probablement correspond aux horizons polbien et sérogosien.

La partie occidentale du bassin de Haute Thrace (arr. de Plovdiv) est dépourvue des associations de Mollusques plus récents que celle à *Lenticorbula sokolovi*, mais en Maritza-ouest on rencontre au-dessus de la dernière, dans les parties terminales de la formation d'Ezerovo, une association à *Lenticorbula helmerseni helmerseni* (M i k h a i l o v s k y). On rencontre aussi *Cerastoderma (Bessia) pseudosarmaticum pizensis* ssp. n., *Abra bosqueti* (S e m p e r). L'espèce *Lenticorbula helmerseni* est connue dans les couches baygubekiennes (Oligocène supérieur marin) de la région transcaspienne (M e r k l i n, 1964, M e r k l i n, 1974), mais préfère les faciès à salinité réduite. L'association bulgare est composée des formes euryhalines et avait habité un bassin à salinité réduite, mais d'âge oligocène supérieur.

Les trois associations citées avaient habitées des eaux à salinité réduite (probablement brachyhalines) parce que les formes marines euryhalines sont constamment présentes. Toutes les trois sont accompagnées d'ostracodes abondantes, mais pas encore étudiées taxonomiquement.

Une autre association, composée des formes saumâtres et dulcicoles à prédominance de *Polymesoda (Pseudocyrena) convexa brogniarti* (B a s t e r o t), est rencontrée alternant avec celle à *Lenticorbula helmerseni* dans les parties terminales de la formation d'Ezerovo et rarement dans les parties basales de la formation de Maritza. Outre l'espèce dominante apparaissent également *Congeria (Mytilopsis) kochi* Andrusov, *Congeria (Andrussoviconcha) euchroma* (Oppenheim), *Theodoxus (Calvertia) crenulatus* Klein, *Melanopsis (Lyrcaea) impressa hantkeni* Hofmann. L'association (surtout la dernière espèce) est d'âge oligocène supérieur (confirmée par sa position stratigraphique) et habitait des eaux douces ou très adoucies.

Les associations bulgares et leur corrélation avec les associations d'Hongrie et de la Paratéthys orientale sont résumées dans le tableau. La subdivision de l'Oligocène en inférieur, moyen et supérieur est préliminaire, parce qu'il n'existe pas un schéma universellement reconnu et même ses limites inférieure et supérieure sont discutables. La corrélation (quoique indirecte) avec les zones nannoplanctoniques va permettre de trouver la place des sédiments saumâtres de la Paratéthys dans un schéma global futur.

[3] La figure-texte n'a pas été remise à la rédaction.

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DISCUSSIONS

M. Nosovsky: La découverte dans le bassin thrace supérieur de la faune saumâtre, dans les couches de Serogaz à éléments sténohaliniens de la faune de mollusques ne doit pas nous étonner, vu que des facteurs analogues existent aussi dans la région euxino-caspienne, à savoir : dans le Caucase du Nord (sur le ruisseau Ţrandon), à Engheneh et dans le sud de l'Ukraine (arrondissement Anostolovsk et couches de Serogaz *Glycimeris pilosus*). Cet état de choses est bien intéressant.

L'unification des couches de Polbinsk (Maloceansk) et de Serogaz en un seul étage est possible, mais si on redige des cartes géologiques, il faut que ces couches soient étudiées à part.

P. Stevanović. L'auteur prend en considération la définition initiale de la Paratéthys (Laskarev, 1924). Il envisage que, ce bassin s'est formé jusqu'au moment de la phase savienne. Quelles sont les limites entre la Téthys et la Paratéthys? L'extrémité du bassin est, par exemple, l'Oligocène de la Thrace, le graben de Rainsk, de la Macédoine, etc. Généralement, la délimitation de ces deux bassins est, selon notre opinion, encore confuse. En ce qui concerne le présent article, je peux dire qu'il soulève bien des questions ; leur réponse, je le pense, est prématurée.

E. Kojumdgieva. Dans la Téthys (en Espagne), la faune ruscinienne continue de vivre en Piacencien inférieur. Probablement, le changement de la faune mammalienne n'est pas synchrone ; en Roumanie elle est arrivée plus tôt que dans la région méditerranéenne.

THE PALEOCENE-EOCENE IN THE SOUTH-WEST OF THE MOESIAN PLATFORM (ROMANIA) [1]

BY

MIRCEA LEU [2], STEFAN GARTNER [3], IOAN COSTEA [2]

Introduction

After 1952, while discovering certain significant hydrocarbon deposits on the Romanian territory of the Moesian Platform, the above-mentioned structural unit became of a prime economic concern and a whole set of complex geological works have been carried out all over it. Out of these, the drilling works contributed to better knowing the stratigraphy and tectonics of this unit, as its geological structure is wrapped into a loess cover.

Biostratigraphical, especially micropaleontological studies have been achieved together with the drilling works. Thus, in the south-west of this unit, south of Craiova (Figure), underlying the Badenian deposits, within the crossed marl and marly-limestone packages, the presence of the Eocene has been revealed. Initially, microfaunistic and palyno analyses indicated the Upper Eocene (Costea, Balteş, 1962), then the microfauna pointed to the Middle-Upper Eocene, as well (Comşa, Costea, 1966).

Meanwhile, the oil wells disclosed the presence of the Paleogene in further drillings and made it possible the outlining of its distribution area. In order to shape a clear-cut image on the Moesian Platform stratigraphy and paleogeographic history during the Paleogene, calcareous nannoplankton study has been resorted to, applied both to the initial samples and to those collected subsequently.

Stratigraphic Considerations

The Paleogene calcareous nannoplankton study in the south-west of the Moesian Platform relies on the analysis of 17 samples. These have been prepared in the laboratories of the Oil and Gas Research and Design

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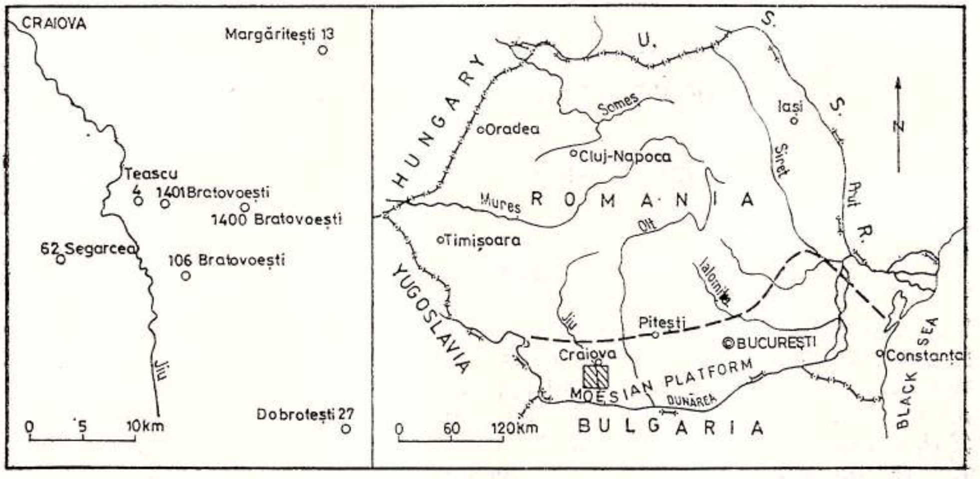
Institute, Bucharest, and, part of them, at College Station, Texas A & M University, U.S.A. (1977). The analyzed material has been included in a lithologic column, achieved due to the electric log correlation of the wells collecting the samples in question (Plate).

After studying the nannoplankton content, the analyzed samples have proved to include rich calcareous nannofossil associations. The



Detail sketch with the location of the wells with investigated samples (left).
Sketch map including the investigated area (right).

oldest elements — relatively scarce — are of a Lower Cretaceous age. More frequent are the Upper Cretaceous nannofossils, present in variable amounts in each sample, in certain cases reaching half of the number of species composing the encountered associations. Paleogene reworked coccoliths have also been witnessed for the younger associations within the same stage.

For calcareous nannofossil zones distribution and correlation, D. Bukry's (1978) standard zonation has been employed, as considered more recent and detailed than E. Martini's zonation (1971).

The lithofacies and the nannofossil content in the investigated samples from the south-western part of the Moesian Platform, presented for each well in Plate, are the following :

— Well 27 Dobrotești, depth 275 m. The sample is a grey-light greenish-marly-limestone, containing an extremely rich association of reworked Cretaceous calcareous nannofossils, accompanied by Paleocene species of low frequency. The association is likely to characterize the Discoaster multiradiatus Zone, the Chiasmolithus bidens subzone, Bukry (1978), i.e. NP9 and is mostly represented by small, primitive specimens. The Paleocene species typical of this interval are lacking, the association being dominated by small and medium size forms of the Coccolithus pelagicus group.

The microfossil preservation is quite the same for the Cretaceous reworked species and for the Paleocene indigenous ones. The age of the determined association is Upper Paleocene.

— The samples in the wells 1401 Bratovoeşti, depth 398 m and 106 Bratovoeşti, depth 350 m are represented by grey-greenish marls and contain an association in which both *Discoasteroides araneus* and *Rhomboaster calcitrapa* are present. This association could be correlated to Gartner's (1970), specific to the Upper Paleocene of the Blake Plateau Region, in a portion assigned to the upper part of the Discoaster multiradiatus Zone. These samples are, highly probably, belonging to an interval below the Paleocene-Eocene boundary.

The described zone has been called Upper Discoaster multiradiatus, correlates to the zone bearing the same name in Bukry's standard zonation (1978), the Campylosphaera eodela subzone, i.e. the upper part of E. Martini's NP 9 Zone (1971) and corresponds to the uppermost Paleocene.

— The samples in the wells 4 Teascu, depth 403 m and 1401 Bratovoeşti, depth 375 m, are represented by fairly calcareous, light grey marls with slight yellowish shades. They include a Paleocene and Lower Eocene nannofossil association, next to Upper Cretaceous forms.

This association integrates into the Tribrachiatus contortus subzone, is the base of the Discoaster diastypus Zone in D. Bukry's zonation (1978), i.e. the NP10 Zone (Martini, 1971), and corresponds to the lowermost Eocene.

The assignment of the Lower Eocene age relies on the presence of the species *Discoaster diastypus*, *Chiasmolithus eograndis* and of the complex Tribrachiatus contortus-Tribrachiatus bramlettei, dominated by *Tribrachiatus contortus*. The lack of the *Discoaster binodosus*, *Discoaster lodoensis*, *Chiasmolithus grandis* and *Sphenolithus radians* specimens equally pleads for assigning the above-mentioned age.

— In the wells 1400 Bratovoeşti, depth 400 m, 1401 Bratovoeşti, depth 348 m and 62 Segarcea 500 m, the samples are lithologically represented by fairly calcareous, light grey, compact, slightly micaferous marls, with pale greenish shades. They contain a nannofossil association dominated by Lower Eocene species, but where reworked Paleocene and Cretaceous elements are also found. Most specimens are of small or middle size, as the big-sized ones occur only in restricted number. Almost all specimens, especially the large ones, are degraded in a manner which suggests mechanical breakage rather than chemical action. The coccoliths preservation state indicates either transport, or diagenetic effects, such as the water action upon the test borders or the calcite precipitation over the nannofossil surface. The content of the association places the described samples in the Discoaster binodosus subzone, namely in the uppermost Discoaster diastypus Zone in D. Bukry's standard zonation (1978) (Martini's NP11 Zone-1971) and defines the Lower Eocene. The assignment of this age is based on the presence of the specimens *Tribrachiatus orthostylus*, *Chiasmolithus eograndis* and *Discoaster binodosus* and on the absence of the species *Discoaster lodoensis* and *Chiasmolithus grandis*.

Conformably overlying the above-described deposits, there comes the grey, slightly sandy, finely micaceous marl with slight olive shades, in the well 106 Bratovoești, depth 256 m. The sample contains a nannoplankton association with comparatively few older, reworked forms. Fossil preservation is generally good, although certain mechanical alterations are visible, as in the case of the delicate *Discoaster lodoensis*. The encountered association belongs to the Tribrachiatus orthostylus Zone, bearing the same name in D. Bukry's zonation (1971), and corresponds to the Lower Eocene. The stratigraphic location of the sample in the well 106 Bratovoești, depth 256 m has taken into account the simultaneous occurrence of the species *Discoaster lodoensis*, *Tribrachiatus orthostylus*, as well as the absence of any *Helicosphaera* specimens and of the species *Discoaster sublodoensis*, *Nannotetrina quadrata*, etc., characteristic of the Middle Eocene, or of younger deposits. The presence of the specimens *Sphenolithus radians* and *Chiasmolithus grandis* support the assignment of this age as the most appropriate.

— In the well 1401 Bratovoești, depth 323 m, there has been encountered a grey, compact marl with greenish shades, breaking irregularly, with fine fragments of degraded macrofauna. Its nannoplankton content is extremely rich, prevailing in the species *Chiasmolithus gigas*, *Discoasteroides kuepperi*, *Discoaster sublodoensis*. The presence of these species, as well as that of *Discoaster barbadiensis* and *Neococcolithes dubius* allows the assertion that the association belongs to the *Chiasmolithus gigas*, i.e. the Nannotetrina quadrata Zone in D. Bukry's zonation (1978), the NP15 Zone, respectively (Martini, 1971), and feature the Eocene.

Overlying these deposits, there is a grey-violet, compact, slightly sandy marl with grey-whitish marly enclaves, with fine micaceous spangles.

The sample contains a peculiar rich nannofossil association, out of which *Discoaster distinctus*, *Lanternithus minutus*, *Discoaster tani nodifer* are worth mentioning. Most forms in this sample are related to *Nannotetrina quadrata*, a species which lends its name to the zone the association is part of. The latter is located in the Coccolithus staurion subzone, in D. Bukry's standard zonation (1978), i.e. the Upper NP15 Zone in Martini's zonation (1971).

In the well 4 Teascu, at 272 m, a calcareous, light grey, fine marl with greenish-yellowish shades is encountered. The sample displays a very rich association of prevailingly Middle Eocene nannofossils. Among the fossils contained in the sample, *Nannotetrina quadrata* appears and fairly develops.

The nannoplankton association belongs to the Discoaster bifax subzone, the *Reticulofenestra umbilica* correlates to the identically called subzone in D. Bukry's standard zonation (1978), i.e. the NP16 Zone (Martini, 1971), and belongs to the Middle-Upper Eocene.

The age assigned to the sample relies on the presence of certain big-sized, well-developed *Reticulofenestra umbilica* and *Discoaster bifax* specimens.

In the well 1400 Bratovoești, depth 360 m and 62 Segarcea, depth 400 m, in equally calcareous, still yellowish, yellow-reddish in spots, compact, finely micaceous, marls, the nannoplankton content is characterized by a relatively low species diversity, the association prevailing in *Reticulofenestra umbilica*. The dimensions of the specimens in this marker-species are highly variable, measuring between 5 — 15 μ in diameter. Other typical nannoplankton species are: *Chiasmolithus solitus, Helicosphaera heezeni, Neococcolithus dubius, Discoaster saipanensis*. Reworked forms are extremely rare and seem to be restricted to the Cretaceous representatives.

The association in these two samples belongs to the Discoaster saipanensis subzone, correlates to the same subzone in Bukry's standard (1978) and to the NP17 Zone in Martini's zonation (1971) and corresponds to the Middle-Upper Eocene.

On most large surfaces of the *Reticulofenestra umbilica* and *Chiasmolithus grandis* specimens, there appear corosion aspects produced during deposition. It is likely that, alternatively, the geochemical conditions at the sediment-water interface were hostile to calcium carbonate, probably because of depth currents or vertical currents.

— Well 1401 Bratovoești, depth 248 m. The sample is a grey, brown-reddish, compact, finely micaferous marl, with friction lenses. It shows a very rich nannofossil content, dominated by the species *Chiasmolithus oamarnensis*, associated to *Reticulofenestra umbilica, Dictyococcites bisectus, Discoaster saipanensis, Discoaster barbadiensis, Neococcolithes dubius*. The described association, *Chiasmolithus oamaruensis*, correlates to the subzone with the same name, i.e. the Discoaster barbadiensis Zone in D. Bukry's zonation (1978) and to the NP18 Zone in Martini's zonation (1971) and corresponds to the Upper Eocene.

The samples in the wells 1400 Bratovoești, depth 327 m and 13 Mărgăritești, 513 m, are represented by a fairly calcareous marl, with slight greenish shades, with scarce calcareous sand films. They contain a very rich calcareous nannofossil association, dominated by Eocene species, although Paleocene or Cretaceous (mainly Upper Cretaceous) forms are equally present.

This association integrates into the Isthmolithus recurvus subzone, the Discoaster barbadiensis Zone, correlates to the subzone bearing the same name in Bukry's standard zonation (1978) and the NP20 Zone (Martini, 1971) and corresponds to the uppermost Eocene.

The Upper Eocene age is established due to the presence of the species *Chiasmolithus oamaruensis* and *Isthmolithus recurvus* (each characterizing the Upper Eocene sediments) and to the absence of any species usually characterizing post-Eocene depositions (e.g. the Sphenolithus predistentus-Sphenolithus distentus group).

Nannofossil preservation is generally good, although in the solution the maximal percentages belong to the youngest compartments of the association. It is, however, likely for the older forms while carried by solution to have been destroyed during reworking.

The following conclusions can be drawn after describing the lithofacial aspect and the nannoplankton content of the samples collected in the south-west of the Romanian Moesian Platform:

— The featured stratigraphic interval lies between the Upper Paleocene and the Upper Eocene.

— Correlation of electric logs with nannoplankton content helped towards drawing out a stratigraphic type-column for the investigated interval.

— Taking into account the well-location in the south-west of the Moesian Platform, the deposits are noticed to grow ever younger from south (27 Dobrotești) to north (13 Mărgăritești).

— The nannoplankton content is specific and allows a detailed stratigraphic biozonation, as 6 zones and 10 subzones are separated within the Paleocene-Eocene succession. In this way, the presence of the Upper Paleocene and Lower Eocene deposits, unnoticed microfaunistically, is revealed.

— The absence of the calcareous nannoplankton associations corresponding to the Lower Paleocene should be regarded as caused either by its non-deposition or by non-recoverable sampling within this formation.

Paleogeographic Considerations

The region of the investigated samples was obviously related during the Cretaceous to the Lom Depression in Bulgaria. These relationships also persist at the very next Paleocene and Eocene levels and were testified by the correlation according to their foraminiferal content.

The marly-calcareous lithofacies, the great number of benthic and planktonic foraminifera, as well as the uncommon abundance of calcareous nannoplankton representatives come to demonstrate the existence of an epicontinental, open sea of a sublittoral zone, of a normal salinity and moderate temperature.

The progressing line of the Paleocene-Eocene waters is from north to south and is stressed by the deposition of nannoplankton associations, ever younger from Dobrotești to Mărgăritești. While advancing, the waters cover the raised relief of Cretaceous rocks with ever newer rocks. In this way, the deposits in each substage are partly conformable and partly unconformable and contain, besides the autochthonous nannoplakton elements, fossil forms of the broken geological formations, drawn into the basin and resedimented (basically the deposits in the Upper Cretaceous substratum).

The sedimentation taking place during the Paleogene stops at the Upper Eocene level, and it is only after a considerable gap that it re-occurs in the Badenian.

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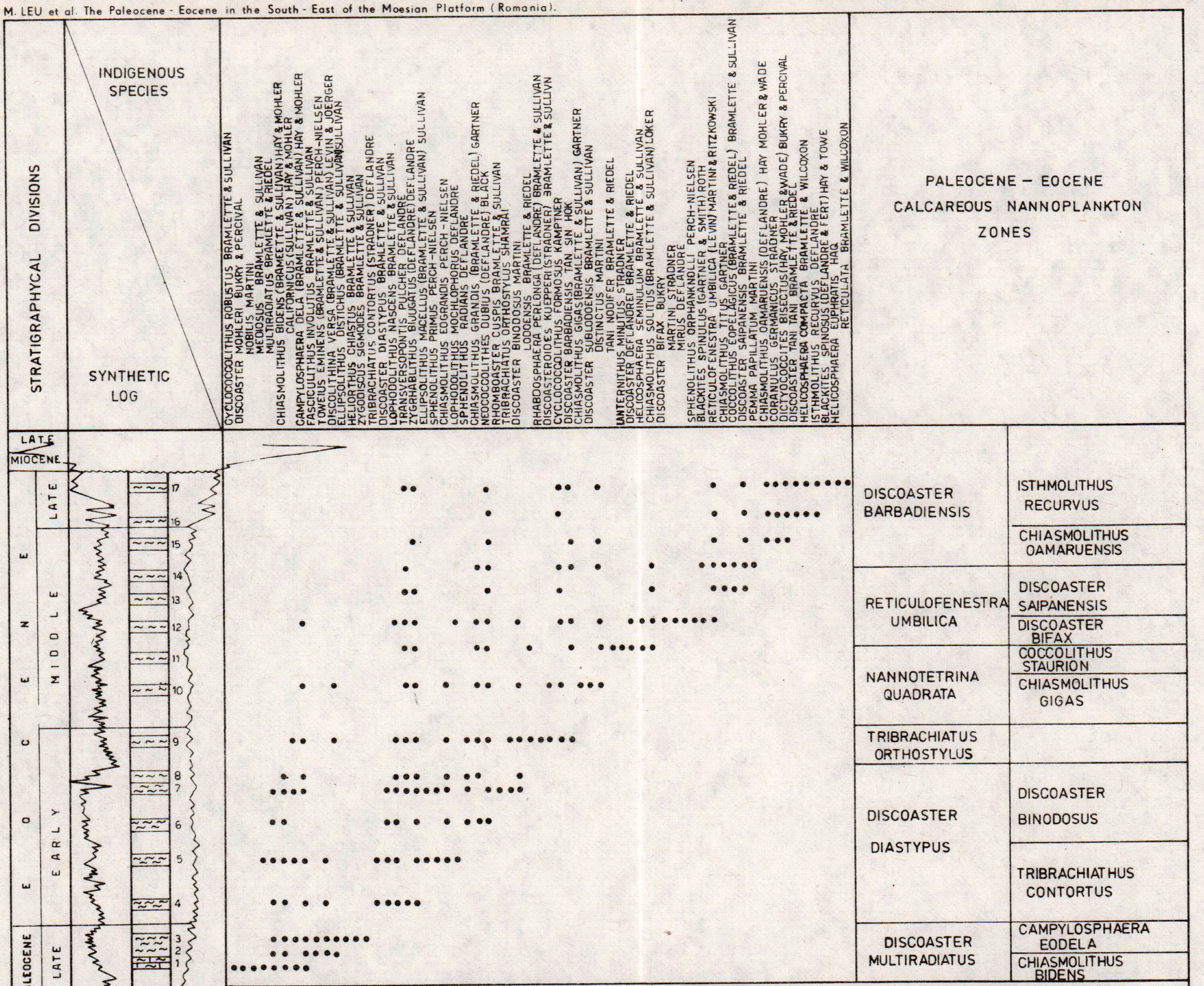

DISTRIBUTION OF CALCAREOUS NANNOFOSSILS IN THE PALEOCENE – EOCENE IN THE SOUTH-WEST OF THE MOESIAN PLATFORM

Imprim. Atel. Inst. Geol. Geof.

ON THE EVOLUTION OF LIMNOCARDIIDS AND VIVIPARIDS IN THE UPPER NEOGENE OF THE DACIC BASIN [1]

BY

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The more complete knowledge of large fossil molluscan groups — such as bivalves and gastropods — was and is still a requirement in the minute biostratigraphic investigations carried out on the Neogene deposits. These elements supply the first field evidence on the age of the deposits wherein they appear, first information referring to the age of the Neogene deposits intercepted by drillings and, later on, their study leads to obtain important data both for the correlation of remote Neogene deposits and for paleogeographic reconstructions.

The researches carried out up to now on some important groups of fossil mollusca — e.g. limnocardiids (among bivalves) and viviparids (among gastropods) — led to accumulation of data which allow us to follow their evolution tendencies in the Upper Neogene of the Dacic Basin. The more so as both limnocardiids and viviparids are more frequently found in the Upper Neogene of the Dacic Basin, yielding elements of utmost significance for the clearing out of the Upper Neogene biostratigraphy in this sedimentation area.

Further on we shall present the evolution tendencies of some genera and species of limnocardiids occurring in the post-Meotian deposits of the Dacic Basin. It is to be mentioned that we shall deal, on the one hand, with those genera and subgenera which, appearing in a large time span, make it possible to observe both the evolution tendencies and the maximums of flowering and diversification and, on the other hand, those supraspecific taxa whose degree of knowledge allow the tackling of such a subject.

Thus, subgenus *Euxinicardium* can be found in the Lower Pontian of the Dacic Basin, being represented by the taxa *Limnocardium (Euxinicardium) subodessae* (S i n z o w) and *L. (E.) subodessae ursina* E b e r -

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sin. In the Middle Pontian (Portaferrian) one may observe a remarkable proliferation of subgenus *Euxinicardium* (Papaianopol, 1981, in press), whereas in the Bosphorian it is in an obvious decline. Within the Dacian interval forms of *Euxinicardium* are more frequently found in the Upper Dacian (Parscovian) than in the Lower Dacian (Getian). During the Pontian and the Dacian in the Dacic Basin the evolution of *Euxinicardium* followed the following directions :

— The reduction of the shell sizes and its convexity. In the Dacic Basin no specimens of large-sized *Euxinicardium*, like those of *Limnocardium (Euxinicardium) esperanzae* Andrusov type of the Kimmerian in the Euxinic Basin, have been found.

— The constant increase of the rib number of the external surface, concomitantly with the stress of their asymmetry. Except few forms, such as *Limnocardium (Euxinicardium) inlongaevum* Ebersin, the Pontian forms have generally less than 14 — 15 ribs on the anterior field, whereas the Dacian species have generally more than 18 ribs on the anterior field, the ribs being often strongly asymmetrical in the anterior and posterior part of the anterior field, as in the case of the taxa *Limnocardium (Euxinicardium) moskoni* Papaianopol, *L. (E.) orolesi* Papaianopol, *L. (E.) peregrinum* Papaianopol.

— Referring to the hinge, it is to point out that, unlike the Pontian species, numerous enough Upper Dacian taxa have a posterior lateral tooth on the left valve, more or less obvious, like at *Limnocardium (Euxinicardium) orolesi* Papaianopol and *L.(E.) eximium* Papaianopol. Moreover, in case of the Dacian taxa the upper lateral tooth on the left valve appears much more frequently than at the Pontian species.

Subgenus *Tauricardium* is found in the Dacic Basin since the Middle Pontian, represented by *Limnocardium (Tauricardium) petersi* (Hörnes). In the Upper Pontian (Bosphorian), besides such specimens, there occurs species *Limnocardium (Tauricardium) praesquamulosum* Ebersin, whose ribs have a triangular cross section. The Lower Dacian (Getian) is characterized by the form *Limnocardium (Tauricardium) olteniae* (Ionescu-Argetoaia), which appeared in the Upper Pontian. It is characterized by a large-size, obvious convexity of the shell, triangular ribs on their whole length and a prominent umbonal zone. Therefore, it may be admitted that in the Dacic Basin forms of this subgenus evoluted in the sense of the size increase and shell convexity, sharpening of the ribs of the external surface and the prominence of the umbonal zone.

Genus *Chartoconcha* had an interesting evolution in the Dacic Basin. In the Middle Pontian (Portaferrian) it is characterized by species with a ribbed external surface of the shell, e.g. *Chartoconcha asaphiopsis* (Brusina) and *Ch. candida* Papaianopol. Beside them, in Portaferrian there also occurs *Chartoconcha bayerni* (R. Hoernes), which sometimes shows the existence of a fine ribbing on a small surface, around the umbo. In the Upper Pontian the existence of ribbed forms on the whole external surface is no longer observed; there are to be found only species whose ribbing appears only around the umbo zone, e.g. *Chartoconcha bayerni* (R. Hoernes), *Ch. minuta* Papaianopol, *Ch. rumana* (Wenz). At some Dacian taxa — e.g. *Chartoconcha ovata* Papaianopol — the external surface presents no traces of radial ribbing. Therefore, referring to the ribbing, it may be ascertained that there is a transition from species with the whole external surface ribbed — e.g. Portaferrian species — to forms with a less obvious ribbing only in the zone around the umbo (in the Bosphorian), and then to entirely smooth forms (in the Dacian) (Papaianopol, 1975).

As regards *Prosodacna*, specimens belonging to this genus are found in the Dacic Basin, in the Upper Pontian (Bosphorian). As mentioned on another occasion (Papaianopol, 1977), as compared to the type of the genus (*Cardium macrodon* Deshayes), at most of the Pontian species several features occurring at the Dacian and Kimmerian *Prosodacna* — e.g. hypertrophy on either valve of the anterior lateral tooth, strong thickening of the shell in its anterior part, and marked shifting of the umbo in the anterior part of the valve — are no longer observed or they are found in an incipient form at the Pontian *Prosodacna*. Also in the Upper Pontian, in the Dacic Basin there occur the first specimens of subgenus *Psilodon*, which will develop strongly in the Dacian. In the Lower Dacian (Getian) both species of the subgenus *Prosodacna* (with numerous, smooth and flat ribs) and forms of the subgenus *Psilodon* are apt to be found. In the Upper Dacian (Parscovian) subgenus *Psilodon* is widely developed, its evolution being in the sense of the hypertrophy of the anterior lateral teeth on either valve, the decrease of the rib number concomitantly with the increase of their height; the tendency of reduction of the cardinal tooth and of the posterior lateral one on the left valve can also be observed.

Genus *Zamphiridacna*, with its first occurrences in the Middle Pontian (Portaferrian) in the Dacic Basin, is widely developed and becomes typical of the Dacian deposits. Taking into account the morphologic characters of the Pontian and Dacian *Zamphiridacna* it may be ascertained (Papaianopol, 1976) that the species of this genus have evoluted constantly towards the increase of the size and convexity of the valves, the total reduction of the cardinal teeth (found at *Zamphiridacna portaferrica* Papaianopol of the Middle Pontian) and the disappearing of the ribbing of the posterior field and the anterior extremity of the anterior field. As regards the ribbing, both the Pontian and the Dacian species have the first anterior ribs rounded, which at some Dacian taxa — *Zamphiridacna orientalis* (Sabba) and *Z. zamphiri* (Cobălcescu) — become triangular or even carenate. It is worth mentioning, at some specimens of Upper Dacian *Zamphiridacna*, the existence of a posterior lateral tooth on the left valve (Papaianopol, 1976), element of the hinge which is completely lacking at most of the Pontian and Dacian *Zamphiridacna*.

Genus *Dacicardium*, which occurred in the Middle Pontian, is very frequently found in the Dacian. Species *Dacicardium vetustum* Papaianopol, found both in the Middle Pontian (Portaferrian) and the Upper Pontian (Bosphorian), which has a more complete hinge (five teeth on the right valve and three teeth on the left one), yields *Dacicardium rumanum* (Fontannes) of the Dacian; it usually has only one anterior tooth on the right valve, a large size and a less obvious keel of the last ribs of the anterior field. *Dacicardium rumanum* (Fontannes) gave rise, by the increase of the site and elongation of the shell, to *Dacicardium dacianum* (Papaianopol) and the latter provided *Dacicardium validus* Papaianopol et Agapia Popescu, with a rounded, more slightly inequilateral shell and a more marked umbo zone.

Genus *Plagiodacna*, also with an interesting evolution, is frequently found beginning with the Middle Pontian (Portaferrian) — represented by *Plagiodacna carinata* (Deshayes), species often found in the Upper Pontian (Bosphorian) and also in the Lower Dacian (Getian). In the Pontian-Dacian of the Dacic Basin the evolution tendency of *Plagiodacna* was towards the reduction of the hinge, the development of the subapical plate and diminish of the carena (Papaianopol, 1977 a); this evolution mostly coincides with that of the *Plagiodacna* from the Euxinic Basin (Ahvlediani, 1966).

Referring to Viviparids, it is worth mentioning that they are widespread in the Dacic Basin, *Viviparus moldavicus* Wenz, endemic species proper to the Meotian in the Dacic Basin, being found both in the Lower Meotian (Oltenian) and in the Upper Meotian (Moldavian). In the East Carpathian Bend Zone in the Upper Meotian deposits one could recognize species *Viviparus incertus* Macarovici, also reported by Roșka (1973) from the Upper Meotian in the south of the Moldavian Soviet Socialist Republic.

In the Lower Pontian (Odessian) *Viviparus incertus* Macarovici is found, a species described and illustrated for the first time by Macarovici (1940) as *Viviparus neumayri incertus*. This taxon has been subsequently reviewed by Taboiakova (1964), who recognized its affinities with *Viviparus achatinoides*, considering it a variety of the latter.

Viviparus incertus, frequently found in the Dacic Basin in the Odessian base, is a species with a fragile shell and a large size, related to *Viviparus achatinoides*; however, it maintains its individuality as particular species. *Viviparus incertus* Macarovici may be recognized both in the footwall and in the hanging wall of some coal seams in the Boteni area (Argeș district) as well as in different places in the Getic and Wallachian depressions. *Viviparus incertus* Macarovici may be encountered both in the Lower Pontian (Odessian) deposits and in the Middle Pontian (Portaferrian). In certain cases, *Viviparus incertus* Macarovici appears in association with *Viviparus botenicus* Lubenescu, characterized by a pronounced elongation of the coiles. Another endemic species — *Viviparus moskoni* Lubenescu — is encountered also in the Middle Pontian of the Dacic Basin, especially in the Congeria rhomboidea Beds.

Genus *Viviparus* has a peculiar proliferation in the Upper Pontian (Bosphorian). Since 1838, Deshayes has described the form *Viviparus achatinoides* in the Kerci Peninsula, in deposits considered now of Bosphorian age. As compared to the holotype — described probably after a young specimen — the specimens from the Dacic Basin are well developed, strong, some of them with endemic characters; it allowed us to separate some varieties, such as *Viviparus achatinoides pseudoneumayri* Lubenescu and *V. achatinoides sabbae* Lubenescu.

From the "achatinoide" trunk most of the Dacian and Romanian Viviparids developed. Also in the Upper Pontian (Bosphorian) one may observe the appearance of the first shells of *Viviparus* with a visible carena, or an obvious settling of the coiles, as in case of *Viviparus motruensis* (Sabba), *V. glogovensis* (Sabba), *V. papaianopoli* Lubenescu.

At the end of the Bosphorian and the beginning of the Lower Dacian (Getian) one may observe, on the one hand, the extinction of old species of Viviparids and, on the other hand, the appearance of new taxa. The sequence of Getian viviparids is highly numerous and diversified. Besides *Viviparus achatinoides* (Deshayes) — which may be found until the Getian — a lot of new species, proper to the Dacic Basin, occur at this stratigraphic level. New taxa — *Viviparus getianus* Lubenescu, *V. carenatus* Lubenescu, *V. macarovicii* Lubenescu — have been identified in the Getian beside forms pointed out since the last century — *Viviparus argesiensis* (Sabba), *V. monasterialis* (Fontannes).

The appearance, at some forms, of a strong carena, as well as of the spiral cords determined Sabba Ştefănescu (1896) to assign them to the subgenus *Tylotoma*. The existence of transition elements between carenate and noncarenate species points to a close relation between the two types of shells; however, no clear delimitation of a possible subgenus *Tylotoma* as against the nominative subgenus *Viviparus* can be made. That is why we do not consider it necessary to separate the subgenus *Tylotoma* within the genus *Viviparus*.

Within the Getian viviparids mention should be made of the predominance of the Romanian faunas ancestors. Thus, *Viviparus praecraiovensis* Lubenescu, ancestor of *Viviparus craiovensis* (Tournouër) from the Romanian, is found since the Upper Pontian. Similarly, *Viviparus berbestiensis* Lubenescu (occurring at the level of the Pachydacna Beds in the Getian base), coming from *Viviparus argesiensis* (Sabba), is an ancestor of *Viviparus bifarcinatus* (Bielz) and *Viviparus stricturatus* (Neumayr), for which it has often been mistaken. Likewise, *Viviparus muscelensis* Lubenescu shows many similarities with *Viviparus turgidus* (Bielz), the former being found between and above the Dacian coal seams in Oltenia and the Argeş District. *Viviparus cucestiensis* Lubenescu is likely to be related to *Viviparus rudis* (Neumayr).

In the Lower Dacian besides sculptured viviparids there are also observed smooth specimens and specimens with a relative high convexity — e.g. *Viviparus dacianus* Lubenescu, very often mistaken for *Viviparus rumanus* Tournouër.

The Upper Dacian (Parscovian) is characterized by the prevalence of smooth or sculptured, large-sized viviparids. Among the numerous species, mostly found out by Cobălcescu (1883), mention should be made of *Viviparus rumanus* Tournouër, *V. murgescui* Cobălcescu, *V. alexandrieni* Cobălcescu, *V. herberti* Cobălcescu, and *V. berbestiensis grandis* Lubenescu.

In the Romanian there is developed a fauna of strongly sculptured viviparids, with a size smaller than that of the Parscovian forms. It is only at this stage that we find species common to those occurring in the "Paludine Beds" of the Pannonian Realm, e.g. *Viviparus bifarcinatus* (Bielz), *V. stricturatus* (Neumayr), *V. rudis* (Neumayr).

At the same time we establish, in the Romanian of the Dacic Basin, the presence of forms similar to certain Slavonic species which, however, have particular characteristics; it made us to consider them local varieties of the latter. Among them, mention should be made of *Viviparus dezmanianus dacicus* Lubenescu and *V. pilari olteniae* Lubenescu. Beside them there also appear endemic species, typical of the Romanian in the Dacic Basin, e.g. *Viviparus turgidus* (Bielz), *V. craiovensis* (Sabba), *V. bergeroni* (Sabba), *V. falconensis* Lubenescu.

It is worth mentioning the existence of deposits including highly sculptured viviparids, similar to the forms found in the Upper Paludine Beds: *Viviparus novskaensis* Penecke and *Viviparus sturi* (Neumayr). Such specimens have been reported from Sopîrlița (Olteț Valley) and the Igozălău Valley (Gorj District). As known, the comparisons between the Dacic and the Pannonian basins relying on viviparids were made up to the level of the lower and middle Paludine Beds. Wenz (1943) pointed out that even in the Dacic Basin the evolution of this gastropoda group ended before its maximum development. The identification of the two deposits with strongly sculptured viviparids makes us infer that, in certain zones of the Dacic Basin, viviparids continued their evolution. It is possible that these zones might have been more extended but erosion removed the necessary evidences.

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QUESTIONS

P. Stevanovič : 1. It is obvious that none of the mentioned species of Viviparids exist in Romania and the Pannonian Realm. It is likely that the collections have not been compared. There had to be forms in common and a revision had to be made.

2. Is there the species *Pterodacna* in the Romanian Pontian?

Answers : 1. The study of Viviparids in the Dacic Basin, carried out by one of the authors of the paper (V. Lubenescu), pointed out that at the Pontian level there existed local faunas of Viviparids, well established, which might represent a barrier for other forms which could have come from the west (the Pannonian Basin). The presence of some common forms of Viviparids in the Middle Pontian (Portaferian) deposits in the Pannonian and Dacic basins has not, however, been out of question, but the confirmation or not of this hypothesis supposes the possibility of carrying out of a comparative study of the Pontian Viviparids in the two sedimentary basins.

2. The existence of the genus *Pterodacna* was pointed out by I. Papaianopol in the Upper Pontian deposits in the east of Muntenia (eastern part of the Dacic Basin).

REMARKS ON OLIGOCENE CHRONO- AND BIOSTRATIGRAPHY IN TRANSYLVANIA (ROMANIA)[1]

BY

ANATOL RUSU[2]

In the last decade several remarkable biostratigraphic results on the Oligocene in NW Transylvania were obtained. The large number of published papers, different interpretation of results, as well as the difficulty of using an adequate chronostratigraphic scale made more laborious the obtaining of a clear image on the Oligocene stratigraphy in the study zone. The present paper tries to synthetize the existing biostratigraphic data – including it into a logic and unitary scheme – and to clear up certain apparent contradictions.

Three events have revolutioned the modern view on the stratigraphy of the Oligocene deposits in Transylvania. In their chronologic order they are, as follows:

– establishing of planktonic foraminiferal zones in the Vima Beds (Popescu, 1972) and, relying on them, the specification of the Oligocene/Miocene boundary within this formation;

– discovery of the *Nucula comta* level (Rusu, 1972), with a typical Rupelian *s.s.* fauna, in the base of the Ileanda Beds;

– determination of the Lattorfian position in the Transylvania Basin on the basis of the calcareous nannoplankton (Martini & Moisescu, 1974).

Several relevant papers on foraminifera, ostracoda, nannoplankton, mollusca and fossil flora completed the known biostratigraphic scheme, allowing a more precise determination of the content of the Oligocene stages in the epicontinental deposits of Transylvania.

Eocene/Oligocene Boundary. According to the evolution of the researches concerning the Transylvanian Paleogene, it may be ascertained that the possibility to establish this boundary, as it is accepted at present, appears since the first study on the nannoplankton of the Brebi Marls.

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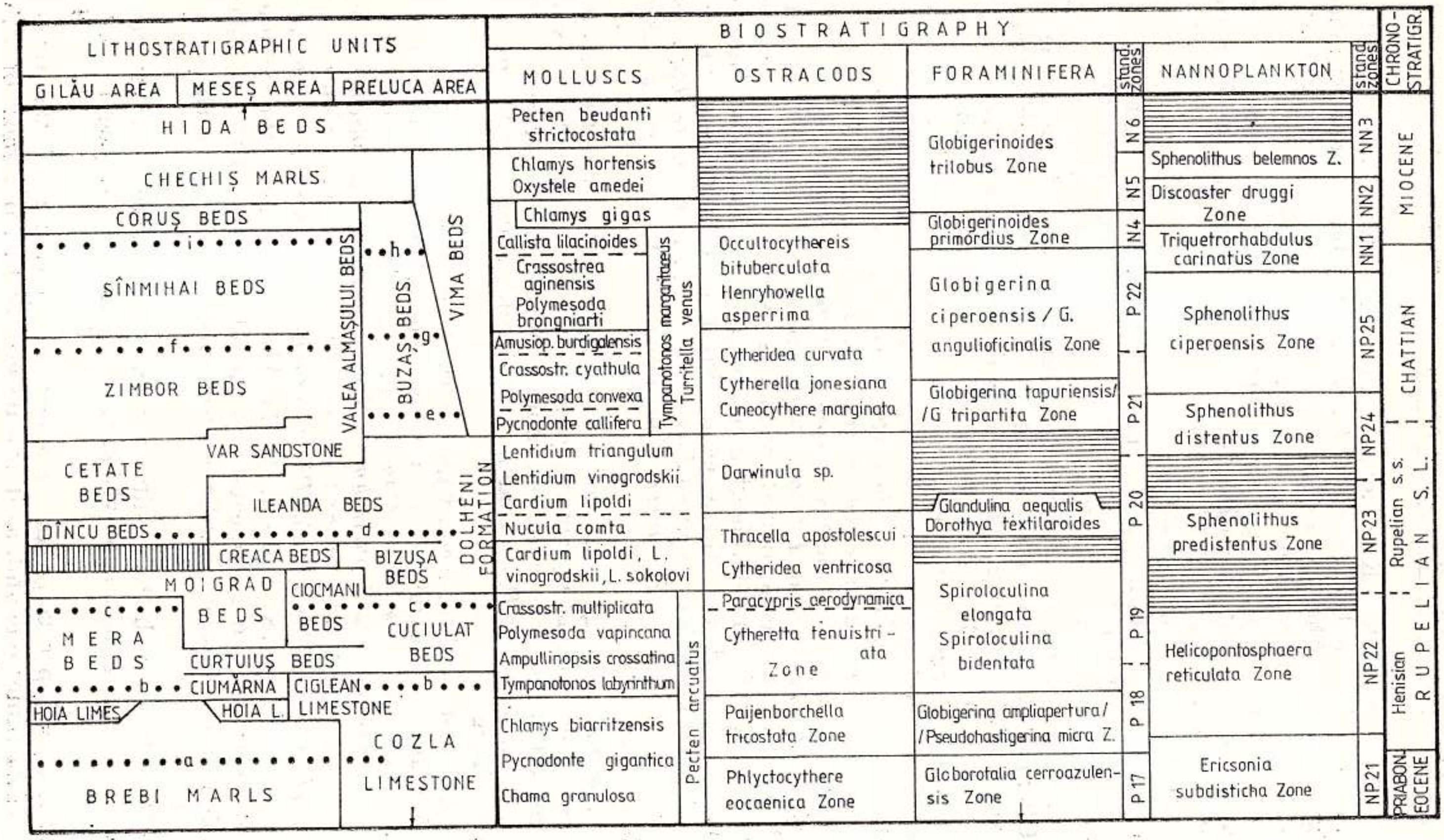
(=Bryozoa Marls) in which Popescu and Gheța (1972, p. 132) pointed out: "At the level of sample 2761 (Turea) there occurs a modification by the extinction of species *Cyclococcolithus formosus* and more frequent occurrence of the discolithines". Using other biozonation schemes, the above—mentioned authors do not point out that it represents also the limit between Martini's (1969) standard nannoplankton zones NP21 (Lattorfian or Lower Tongrian) and NP 22 (Upper Tongrian =Henisian), which was pointed out in Transylvania, too (Martini and Moisescu, 1974) (see Figure). Meantime Mészáros et al. (1973, 1974) separate the same two nannoplankton zones, but situated in other intervals of the lithologic succession. Considering erroneously the base of zone NP21 above the *Pycnodonte gigantica* level, the authors delimited, fortunately well, the Eocene/Oligocene boundary in the Mera profile (Mészáros et al., 1974, Fig. 1).

The zonation achieved by Martini and Moisescu (1974) indicated, for the first time correctly, the position of the Lattorfian in Transylvania, a stage including the extension of the nannoplankton standard zone NP21 (Ericsonia subdisticha zone) and which was regarded, according to the West German authors, as Oligocene. Implicitly it was considered that the Eocene/Oligocene boundary — corresponding to the threshold between zone NP 20 and zone NP 21 — would be situated towards the base of the Brebi Marls. As regards Transylvania's stratigraphy a situation not accepted in the mesogean realm is reached, that is to establish the boundary within the *Nummulites fabianii* typic zone, nearby the epibolic level of the index species.

The results obtained by Martini and Moisescu (1974) were reinterpreted by us (Rusu, 1974), when it was established that the boundary discussed should be located between the standard zones NP 21 and NP22, that is in the interval between the *Pycnodonte gigantica* marker level and the top of the Brebi Marls (see also Rusu and Drăgănescu, 1976, Rusu, 1977). The determination of the base of the Helicopontosphaera reticulata zone (NP 22) made only after the extinction of specimens of *Cyclococcolithus formosus* is approximatively enough, especially that in the Brebi Marls overlying the *Pycnodonte gigantica* level, this form is found only sporadically (Gheța, oral information). Unfortunately in the Transylvania Basin species *Cyclococcolithus margaritae* Roth & Hay is unknown, whose occurrence (used in biozonation) would give more certainty as to the establishing of the boundary between zones NP 21 and NP 22.

A bionomic threshold is pointed out also in the ostracod fauna approximately in the interval of the *Pycnodonte gigantica* level. Thus, the distribution table given by Olteanu and Popescu (1973) shows that at this level several species become extinct (*Cythereis dadayi* Mehes, *Paijenborchella tritacosta* Olteanu, *Loxoconcha parva* Pietrzeniuk, *Cytherura oedelemensis* Keij, *S emicytherura forestensis* (Keij) and new species appear, e.g. *Cytheridea helvetica* (Lnks), *C. henisensis* Keij, *C. strigosa* Pietrzeniuk and *Krithe pernoides* (Bornemann). În 1977, Olteanu separated, within the Brebi Marls, even two local zones of ostracoda, having the extension of the index forms:



Correlation of the fossil assemblages and biozones for the Oligocene in Transylvania

a, *Pycnodonte gigantica* Level; b, *Scutella* Lower Level; c, *Scutella* Upper Level; d, *Nucula comta* Level; e, *Pycnodonte callifera* Level; f, *Crassostrea cyathula* Level; g, *Amusiopecten burdigalensis* Level; h, *Callista lilacinoides* Level; i, *Crassostrea aginensis* Level.

Phlyctocythere eocaenica Keij zone, below the *Pycnodonte gigantica* level, and Paijenborchella tricostata (Lnks)zone, above the mentioned level (Figure). The author considered the first zone of Eocene age and the second one of Eocene-Oligocene age. Lately, Olteanu (1980) has established the Eocene/Oligocene boundary at the *Pycnodonte gigantica* level, marked by the explosive development of the species *Bathocypris arcuata* (Münst.) and *Pterigocythereis fimbriata* (Münst.) and the appearance of new species in the basin.

The planktonic foraminifera fauna brings the best arguments in favour of the drawing of the Eocene/Oligocene boundary in the Transylvania Basin. Thus, within the Brebi Marls, Iva and Rusu (1982) distinguished two planktonic foraminifera zones typical of the mesogean realm: Globorotalia cerroazulensis zone, the extension of which approximately corresponds to the planktonic foraminifera standard zones P 16 (partim) and P 17 (Blow, 1969) and Globigerina ampliapertura/ Pseudohastigerina micra zone, equivalent to zone P 18 (Figure). The bionomic threshold between the two mentioned zones — established on the extinction of the forms *Globorotalia cerroazulensis cerroazulensis* (Cole) and *G. cerroazulensis cunialensis* Toumarkine & Bolli and the occurrence of the species *Globigerina ampliapertura* Bolli — is situated at the base of the *Pycnodonte gigantica* level and corresponds to the Priabonian/Oligocene boundary in the reference sections of the Paleogene in Italy. In the Globigerina ampliapertura /Pseudohastigerina micra zone in Transylvania there occur also typical Oligocene species such as *Pseudohastigerina naguewichiensis* (Mjatliuk), *Globigerina liverovskoe* (Bykova), *Chiloguembelina gracilima* Andrae, etc. (see Iva & Rusu, 1982).

Other groups of organisms do not bring specifications as regards the Eocene/Oligocene boundary; modifications within the fauna of molluscs, echinids, nummulits can be observed more upwards in the stratigraphic column (at the level of the Hoia Beds), partially determined by the paleogeographic changes.

The Mera Beds and their equivalent in the Preluca area — the Cuciulat Beds (Figure) — include a typical Oligocene mollusca fauna, represented by *Ampullinopsis crassatina* (Lmk.), *Turritella biarritzensis* (Bouss.), *Tympanotonos labyrinthum* (Nyst), *Pirenella plicata galeotti* (Nyst), *Polymesoda convexa vapincana* (d'Orb.), *Crassostrea cyathula multiplicata* (Tourn.), *Pitar verneuili* (d' Arch.), etc. (see Moisescu, 1972, 1978; Rusu, 1977). Also at this level the echinid *Scutella subtrigona* Koch appears, considered (Rusu, 1977, p. 164) a geographic subspecies of *Scutella subtetragona* Grat. of the early Oligocene in Aquitania.

The Mera Beds, with their unitary macrofauna, have been assigned to the Henisian, considered a subdivision of the Rupelian *s.l.* (see Rusu, 1977).

Henisian (=Upper Tongrian) includes: the upper part of the Brebi Marls, the Hoia Limestone and, as it will be seen further on, most of the Mera Beds and their lateral equivalents (Figure).

Henisian/Rupelian s.s. Boundary. As the biostratigraphic results obtained up to now are not conclusive for the specification of this limit, the question is still open for discussion. However, we may approximately locate the Henisian/Rupelian boundary within the Mera Beds.

As known, the Henisian has the extinction of the nannoplankton standard zone NP 22 (Helicopontosphaera reticulata zone). In the Transylvania Basin this zone is considered to be spread up to the base of the Mera Beds from the type profile (M é s z á r o s et al., 1974) and beyond the Scutella Lower Level (M a r t i n i , M o i s e s c u, 1974). According to M a r t i n i and M o i s e s c u (1974) the upper limit of zone NP 22 is poorly defined, the next zone (NP 23) being represented on tables by a question mark. Our observations on two important marker species of the Paleogene — *Isthmolithus recurvus* D e f l . and particularly *Lanternithus minutus* S t r a d n e r (see M a r t i n i and M o i s e s c u, 1974, Fig. 3) — made us include the whole interval of the Mera Beds to the zone NP 22 (R u s u , 1977, p. 165).

The ostracod faunas of the Cuciulat Beds (O l t e a n u, 1980) bring new elements for the establishing of the Henisian/Rupelian *s.s.* boundary. They indicate a bionomic threshold between the Scutella Upper Level and a packet of grey clays, marked by the extinction of some taxa, such as *Leguminocythereis decipiens* (L n k s .), *Cytherella gracilis* L n k s., *Triebelina punctata* D e l t e l, *Occultocythereis gradata* P r t z k., *Hermanites memorans* M o o s, *Paracytheridea tuberosa* L n k s., *Xestoleberis muelleriana* L n k s., *X. subglobosa* (B o s q.), etc. and the appearance of new species : *Paracypris aerodynamica* O e r t l i, *Cytheretta* aff. *ramosa* (L n k s.), *C.* aff. *schoelleri* (K e i j), *C.* aff. *gutzwileri* O e r t l i, *C.* aff. *fallens* O e r t l i and *Bairdia* aff. *subdeltoidea* (M ü n s t). Therefore, we admit as Henisian/Rupelian *s.s.* boundary (considered substages of the Rupelian *s. l.*) the limit given by ostracods, located towards the terminal part of the Cuciulat Beds or the Mera Beds, immediately over the *Scutella* Upper Level (Figure).

A typical Rupelian (*s.s.*) fauna has been reported from two members : the Bizuşa and Ileanda beds, as well as their littoral equivalents in the basin (Creaca, Dîncu and Cetate beds), all assigned to the Dolheni Formation (R u s u , 1977, p. 122). Here there occurs a particular malacofauna, specific to the Rupelian in the Transeuropean domain, represented by *Corbula* ex gr. *sokolovi* (K a r l o v), *Corbula* ex gr. *helmerseni* M i c h a j l o v s c h i, *Lentidium triangulum* (N y s t.), *L.* ex gr. *nitidum* (S o w.), *L. vinogrodskii* M e r k l i n, *Polymesoda convexa* B r g t., *Cardium serogosicum* N o s s., *C. lipoldi* R o l l e, *Congeria brardi* (B r g t.), *C. nysti* (d'O r b .), *C. tenuissima* M o i s e s c u , *Rzehakia cimlanica* P o p o v, etc. (M o i s e s c u , 1972 ; R u s u , 1977). The ostracod fauna is dominated by the species *Thracella apostolescui* S ö n m e z, *Cytheridea ventricosa* G o e r l i c h, *Cytherossa spathacea* (L n k s .) (O l t e a n u, 1980).

The mentioned invertebrate assemblages belong to a pliohaline brackish sea (salinity $9 - 16.5^{0}/_{00}$), extending from the Carpathian zone up to the Aral Lake region, isolated by Tethys and the northern realm. It represents the first temporary separation of the territory which later on — definitively isolated — will become the Paratethys Realm. In the author's opinion it is only at this level that the Kiscellian might begin, a regional stage older than the Egerian proposed by B á l d i in 1969. At present, this author extends it on the whole interval between the Priabonian and the Egerian (B á l d i, 1979). Thus, the Kiscellian becomes — without any stratigraphic reason — a hybrid stage, characterized by mesogean faunas at the lower part (equivalent to the Henisian) and Transeuropean endemic faunas at the upper part (equivalent to the Rupelian *s.s.*). In our opinion a regional stage of the early Paratethys would be useful only for the interval represented by the Bizușa Beds + the Ileanda Beds = upper part of the Tard Clay (non-laminated argillites with brackish fauna) + the Kiscell Clay, which probably corresponds to zones NP 23 and NP 24 of calcareous nannoplankton.

In the lower part of the Ileanda Beds there is an intercalation with marine faunas of normal salinity — *Nucula comta* level (R u s u, 1972) — — whose elements are found also in the Dîncu Beds in the Gilău Area. The fossil assemblage is represented by mollusca [*Nucula comta* G o l d f., *Nuculana westendorpi* (N y s t.), *Angulus nysti* (D e sh.), *Polinices catena achatensis* (R é c l)] (R u s u, 1977), ostracoda [*Cytheridea helvetica* (L n k s.), *Bairdia* aff. *oviformis* S p e y e r, *Loxoconcha* aff. *kuiperi* K e i j, *Cytheromorpha* aff. *zindorfi* (L n k s.), *Cytherella* sp., *Cadona* sp.)] (O l t e a n u, 1980), foraminifera [*Dorothia textilaroides* (H a n t k.), *Glandulina aequalis* R e u s s., *Pyrulina fusiformis* (R o e m e r), *Nonion granosum* (d'O r b.)] (R u s u, 1977), nannoplankton [*Reticulofenestra ornata* M ü l l e r, *R. clatrata* M ü l l e r, *Dictyococcites dictyodus* (D e f l. & F e r t.), *Ortozygus aureus* (S t r a d.), *Transversopontis fibula* G h e ț a, *Cyclicargolithus floridanus* (R o t h . & H a y), *Coccolithus rupeliensis* M ü l l e r, *C. pelagicus* (W a l l i c h), *Discolithina desueta* M ü l l e r)] (G h e ț a et al., 1976; M é s z á r o s et al., 1977). This fauna points to the Rupelian age and the nannoplankton is considered to belong to the standard zone NP 23 (Sphenolithus predistentus zone) in the lower part of the Rupelian *s.s.* The Rupelian *s.s.* is represented, in NW Transylvania, by the terminal part of the Mera Beds, the Dolheni Formation (entirely) and the basal part of the Vima Beds.

Rupelian/Chattian Boundary. So far this limit is poorly defined in Transylvania. The brackish mollusca faunas in the Gilău and Meseș area (M o i s e s c u, 1972; 1978; R u s u, 1977) seem to be useless for the tracing of the Rupelian/Chattian boundary. Because of the lack of pectinids typical of these stages, the marine malacofauna in the Preluca area (R u s u, 1977) cannot bring specifications in this sense.

Important biostratigraphic elements have been brought by the palaeoflora studies on the basis of which P e t r e s c u (1970) assigned the Cetate Beds to the Middle Oligocene and the Zimbor Beds to the Upper Oligocene.

The microfauna of the marine formations constitutes the starting point of the discussions on the Rupelian/Chattian boundary in Transylvania. The foraminifera assemblages studied by P o p e s c u (1972, 1975) in the Vima Beds (= Gostila Beds) — a formation overlying the Ileanda Beds — allowed him to separate some zones of planktonic foraminifera. The first zone — Globigerina tapuriensis/G. tripartita zone, whose upper limit is marked by the extinction of the index species and of forms of *Globigerina gortanii* (Borsetti), *G. pseudovenezuelana* Blow., *G. selli* (Borsetti) and *Tritaxia szaboi* (Hantken) and the appearance of the species *Globigerina ciperoensis* Bolli and *Globorotalia opima opima* Bolli — would belong entirely to the Rupelian. According to Popescu the bionomic threshold between the mentioned zone and Globigerina ciperoensis/G. anguliofficinalis zone, situated approximately half the thickness of the Vima Beds (which amounts to 300—400 m), corresponds to the Rupelian/Egerian boundary. The studies on nannoplankton do not ascertain this conclusion. Gheța (in Bombiță et al., 1979) points out the presence of the standard zone NP 24 (Sphenolithus distentus zone) with *Helicosphaera recta* Haq, *H. euphratis* (Haq), *Reticulofenestra lockeri* Müller, *Dictyococcites dictyodus* (Defl. & Fert.), *Discolithina latelliptica* Báldi-Beke, *Sphenolithus distentus* (Martini), *Cyclicargolithus abisectus* Müller up to 10—30 m above the base of the Vima Beds, followed by zone NP 25 (Sphenolithus ciperoensis zone), the basis of which is marked by the first appearance of the species *Discolithina enormis* Locker and *Triquetrorhabdulus carinatus* Martini. As known, the lower boundary of the Chattian is situated within zone NP 24 or, according to the same authors, at the base of zone NP 25, therefore in our case, towards the base of the Vima Beds.

Globigerina ciperoensis/G. anguliofficinalis zone, which includes beside the index species the taxa *Globorotalia opima opima* Bolli and *G. opima nana* Bolli (see Popescu, 1975), would correspond only to the upper part of the Chattian, its lower part being represented by the Globigerina tapuriensis/G. tripartita zone (almost entirely) (Figure).

Considering the nannoplankton, the Chattian corresponds to the standard zone NP 25, about 200 m thick, including towards the upper part the acme of the species *Cyclicargolithus abisectus* Müller.

In NW Transylvania, in the marine and brackish psamitic facies, Chattian mollusca are found in the Buzaș Beds, where it forms the *Pycnodonte callifera* level, *Turritella* and *Thracia* level and *Amusiopecten burdigalensis* level (see Rusu, 1977), and in the Zimbor Beds (*Crassostrea cyathula* level) (Moisescu, 1972, 1978; Șuraru, 1975; Rusu, 1977). The Chattian age of the Zimbor Beds — of brackish facies — is also supported by the discovery of a marine malacofauna at their upper part (Șuraru, 1969), within which Mészáros et al. (1975) point out standard zone NP 25 and the lower part of nannoplankton zone NN 1. Also at this level Rusu (1977) reported a foraminiferal assemblage from the *Amusiopecten burdigalensis* level, which attests the presence of the standard zone P 22 (Globigerina ciperoensis zone) (Blow, 1969).

The Chattian would include an important part of the Vima Beds, having the extension of zone NP 25, most of the Buzaș Beds (interval between the *Pycnodonte callifera* level and *Amusiopecten burdigalensis* level) and the Zimbor Beds with the Red Clays Horizon of the Sînmihai Beds (Figure 1).

Oligocene/Miocene Boundary(=Chattian/Aquitanian). Still debated in the world, in Transylvania this limit has been established according to Popescu's delimitation (1972) in the marine pelitic deposits and to Rusu's delimitation (1974) in psamitic deposits of marine and brackish facies.

In the Vima Beds the Oligocene/Miocene boundary is established at the base of the Globigerinoides primordius zone — approximately equivalent to the standard zone N 4 — marked by the first occurrence of the species *Globigerinoides primordius* Blow & Banner (see Popescu 1972, 1975). Approximately at the same level there occur the first typical specimens of *Helicosphaera carteri* (Wallich) and *H. ampliaperta* (Braml. & Wilcox.) (see Gheța in Bombiță et al., 1979), which would mark the base of zone NN 1 (Triquetrorhabdulus carinatus zone), considered of Miocene age.

In psamitic deposits this limit is established on the basis of the mollusca fauna at the base of the *Callista lilacinoides* level in the Buzaș Beds (marine formation) and immediately under the *Crassostrea aginensis* level in the Sînmihai Beds (brackish member) (see Rusu, 1974, 1977). Therefore, the Oligocene/Miocene boundary is situated under the Coruș Beds — Larger Pectens Horizon which marks the first Miocene transgression in Transylvania — but not at the base of the mentioned horizon as Popescu mentioned in 1972. The stratigraphic conception presented in this paper is at present adopted by most of researchers who study the Transylvania Basin.

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QUESTIONS

J. Mitrović. In the lower and upper levels with *Scutella subtrigona* there are also other genera of echinits. In what types of deposits does this assemblage occur?

Answer. In the two levels with *Scutella* there also occur echinits belonging to the genera *Diplosalenia, Coelopleurus, Echinolampas, Schizaster* and *Euspatangus.* This faunal assemblage is encompassed in bioclastic limestones, arenaceous limestones and calcareous sandstones.

M. Nosovsky: 1. Dr. A. Rusu presented very interesting data on the basis of which a correlation can be made not only within the Carpathian area but also in the Euxino-Caspic area. This is possible due to a brackish fauna found out in Transylvania, which includes *Corbula sokolovi* and *Cardium serogosicum,* first described in the south of Ukraina and then in other parts in the south of the U.S.S.R.

2. As regards G. Bizon and C. Müller's report (1977) on the correspondance of the nannoplankton zone NP 25 with the planktonic foraminifera zone N 4 don't you think that the nannoplankton reported from the Buzaş Beds prove the appurtenance of these beds to the Lower Miocene?

Answer. The correlation of the planktonic foraminifera zones with the nannoplanktonic zones shown on the table had been made according to the evidence found in Transylvania. It is possible that here *Globigerina primordius* – index species for zone N_4 at the basis of which the Oligocene-Miocene boundary has been establishd – might occur later on and consequently the real limit is beyond it. In this case a larger section of the Buzaş Beds would belong to the Miocene; however, their lower part belongs undoubtedly to the Oligocene as indicated by the other groups of organisms, the nannoplankton included.

It is to be mentioned that the correspondance of the zone N 4 with the zone NP 25 is only partial at Bizon and Müller, at least in the paper of 1979 the whole zone NP 25 is assigned to the Oligocene.

BIOZONEN VON ORGANISCHSKELETTINGEN MIKROPLANKTONS IN DEN PANNONISCHEN SCHICHTEN UNGARNS [1]

VON

MARIA SÜTŐNÉ SZENTAI [2]

In den pelitischen Ablagerungen der pannonischen Schichten kommen Überreste von organischskelettigen Mikroplanktons /Algen/ stellenweise massenhaft vor. Bedeutend sind besonders die im unteren Teil der Formationsgruppe verbreiteten Dinoflagellaten, da ihre Ausbildung, ihre sehr schnelle Verbreitung und schnelle Selektierung nachher, wichtige Veränderungen der ökologischen Verhältnisse innerhalb einer kurzen erdgeschichtlichen Zeiteinheit festhält.

Mit der Beschreibung von Dinoflagellaten befasste sich zuerst Nagy Lászlóné, dann gab in Rumänien N. Balteş Bericht über die mit der ungarischen identischen Gemeinschaft und beschrieb neue Arten. Nachher arbeiteten L. E. Stover und W. R. Evitt eine neue Methode der Bestimmung der vorpleistozänen Dinoflagellaten aus, auf Grund dessen man mit einheitlichen Gesichtspunkten diese Gemeinschaft bestimmen konnte.

Die bearbeiteten Bohrungen befinden sich am Fuss der Cserhát-Mátra- und Bükk-Gebirge, auf dem Gebiet zwischen den transdanubischen Mittelgebirgen und dem Mecsek-Gebirge, sowie südlich von dem Mecsek-Gebirge (Abb. 1).

Innerhalb der lithostratigraphisch begrenzten pannonischen Formationsgruppe (À. Jámbor, 1980) unterschied ich fünf Biozonen. Die Grenze der Biozonen ist nicht immer gleich mit der Grenze der Lithofazies-Einheiten oder der Faunengemeinschaften.

Die Untersuchung der litoralen und beckeninneren Schichtenfolgen bestätigte die weite horizontale Verbreitung dieser Mikrofossilien Ihre vertikale Verbreitung wird jedoch durch die Veränderung der Umweltsfaktoren (Salzgehalt — Temperatur — pH) begrenzt.

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[1] Vorgetragen am 12. Kongress der Karpato-Balkanischen Geologischen Assoziation, 8 – 13 September, 1981, Bukarest, Rumänian.

[2] Ungarn.

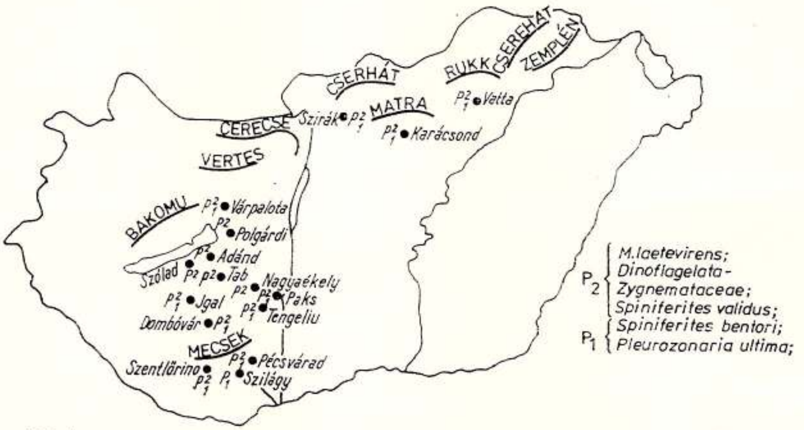

Abb.1. – Platz der untersuchten Bohrungen und das Alter der Schichten.

Biozonen von organischskelettigen Mikroplanktons

Biozone Pleurozonaria ultima

Auf Grund der Lithofazies, sowie mit Untersuchung des Makrofaunas kann man bei der Schichtenfolge von Zala die Zugehörigkeit zu den pannonischen Schichten beweisen (À. J á m b o r, 1980). Ich untersuchte diese Schichtenfolge in Transdanubien in der Umgebung von Tengelic (ihre Molluskenfauna ist das *Limnocardium praeponticum* auf Grund der Untersuchungen von Korpásné Hódi M.). Charakteristisch für ihr organischskelettiges Mikroplankton ist das massenhafte Vorkommen einer einzigen Art, der *Pleurozonaria ultima* n. sp. Ausser dieser Art fand man nur 1 — 1 schlecht bewahrte *Cymatiosphaera* und den Typ *Spirogyra* 3c. Die Faunengemeinschaft deutet auf eine brackige, euxine Fazies, ähnlich den Schichten von Tard mit Pflanzen- und Fischüberresten, wo die Art *Pleurozonaria minor* (H u t t e r) Rákosi massenhaftes Vorkommen hatte.

Oberhalb der Mergelschichtenfolge von Zala sind die Dinoflagellaten für die Tonmergelschichtenfolgen von Csákvár und Dráva charakteristisch. Die makroskopisch gleich aussehende Tonmergelschichtenfolge kann man in zwei Biozonen gliedern. Die untere kann man mit der Art *Spiniferites bentori*, die obere mit der Ausbildung der Art *Spiniferites validus* kennzeichnen.

Biozone Spiniferites bentori

In dieser Biozone (Abb. 2, 3) findet man mehrere, morphologisch wechselvolle Variationen der Art *Spiniferites bentori* (R o s s i g n o l, 1964) S a r j e a n t, 1970. Im unteren Teil der Biozone trifft man häufig die zonenanzeigende Art und ihre dimorphe Form *Gonyaulax digitale* (P o u c h e t) K o f o i d 1911; im oberen Teil zusammen mit dieser art die *Pontiadinium* Arten (*P. pécsváradensis*, *P. obesum*, *P. inequicornutum*) und Formen aus dem Prevalvat Stadium.

Im oberen Teil der Biozone wird zusammen mit der Ausbildung der Pontiadinien die mit 15–20 Mikron grössere Form der Art *Gonyaulax digitale* häufig.

Die zonenanzeigende Art entstand im unteren Teil der Schichten mit *Congeria banatica*. Ihre sehr schnelle Verbreitung mit morphologisch

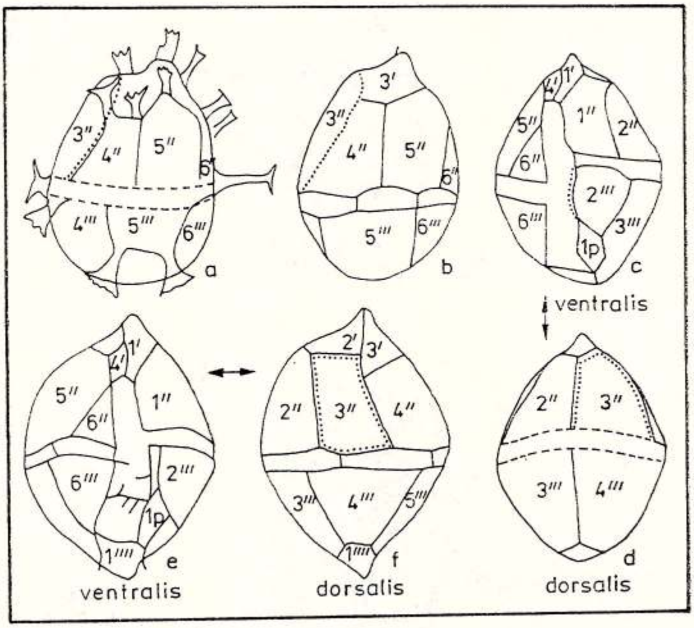

Abb.2. – a, *Spiniferites bentori* (Ross.) Sarjeant ; b – d, *Gonyaulax digitale* (Pouchet) Kofoid; e-f, *Pontiadinium pécsváradensis* n. sp., 600-fache Vergrösserung.

wechselvollen dimorphen Formen zeigt den für die Art optimalen Lebensraum in den Schichten mit *Congeria cžjžeki*.

Die auch heute lebenden und während ihrer Ontogenese zustande gebrachten planktonischen und bentonischen dimorphen Formen der zonenanzeigenden Algenhat man identifiziert (D. Wall und B. Dale, 1970). Die Identifizierung der dimorphen Formen der fossilen Art ist wichtig, weil unter Umständen das Stadium der schlafenden Spore oder das mit Theken und Planktonen vorkommt, abhängend davon ob die Ausbildung in der Nähe der Küste oder entfernt davon geschah. Die dimorphen Formen findet man in den seichtwässerigen, Küstensaum-Fazies zusammen, in den vom Ufer fernen Ausbildungen ist die Form mit Theken und Planktonen die häufigere. Mit der Identifizierung der bentonischen und planktonischen Formen kann man auch die heteropischen Fazies identifizieren, bei häufiger Probenahme kann man aber in den vertikalen Profilen auch die Oszillationserscheinungen gut verfolgen.

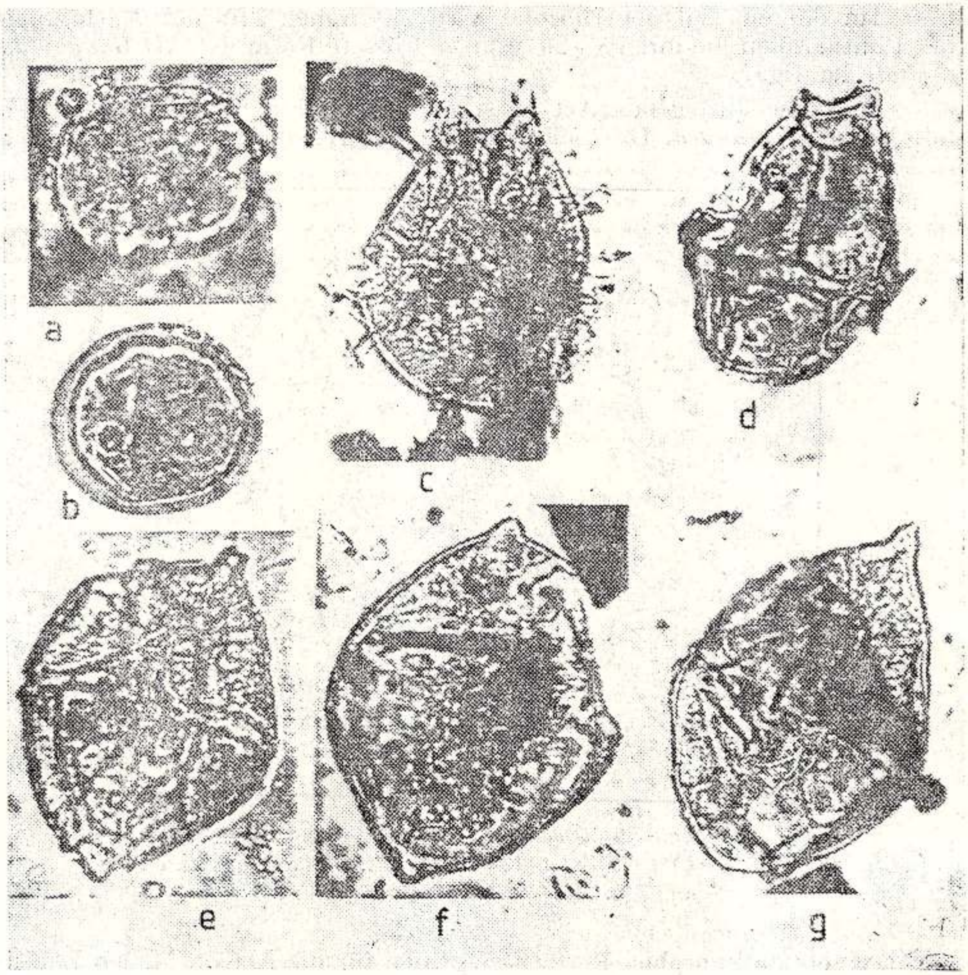

Abb.3. – a-b, *Pleurozonaria ultima* n.sp. Tengelic-2.663,9 – 665,0m. 1000 x; c, *Spiniferites bentori* (Rossignol, 1964) Sarjeant, 1970. Szirák-2. 439,2 – 439,5 m. 600x; d, *Gonyaulax digitale* (Pouchet) Kofoid, 1911. Szirák-2.439,2 – 439,5 m. 600x; e, *Pontiadinium pécsváradensis* n. sp. Pécsvárad – 15.63,0 – 64,0 m.600x; f, *Pontiadinium* sp. Prevalvát stadium. Pécsvárad – 15,65,0 – 66,0 m.600x; g, *Gonyaulax digitale* (Pouchet) Kofoid, 1911. Szirák-2 373,0 – 383,0m.600x.

Biozone Spiniferites validus

Charakteristisch für diese Biozone (Abb. 4) ist die Ausbildung der Art *Spiniferites validus* n. sp. und die Dinoflagellaten mit verdickten Theken, mit den Arten *Impagidinium globosum* n. sp., *I. spongianum* n. sp., *Tectatodinium pellitum* Wall. Ausser den genannten sind in der Biozone noch häufigere Arten *Romanodinium areolatum* Balteş, *Spiniferites paradoxus* Cookson et Eisenack, *Chytroeisphaeridia tuberosa* n. sp., seltenere *Pontiadinium inequicornutum* (Balteş). Formen 29.

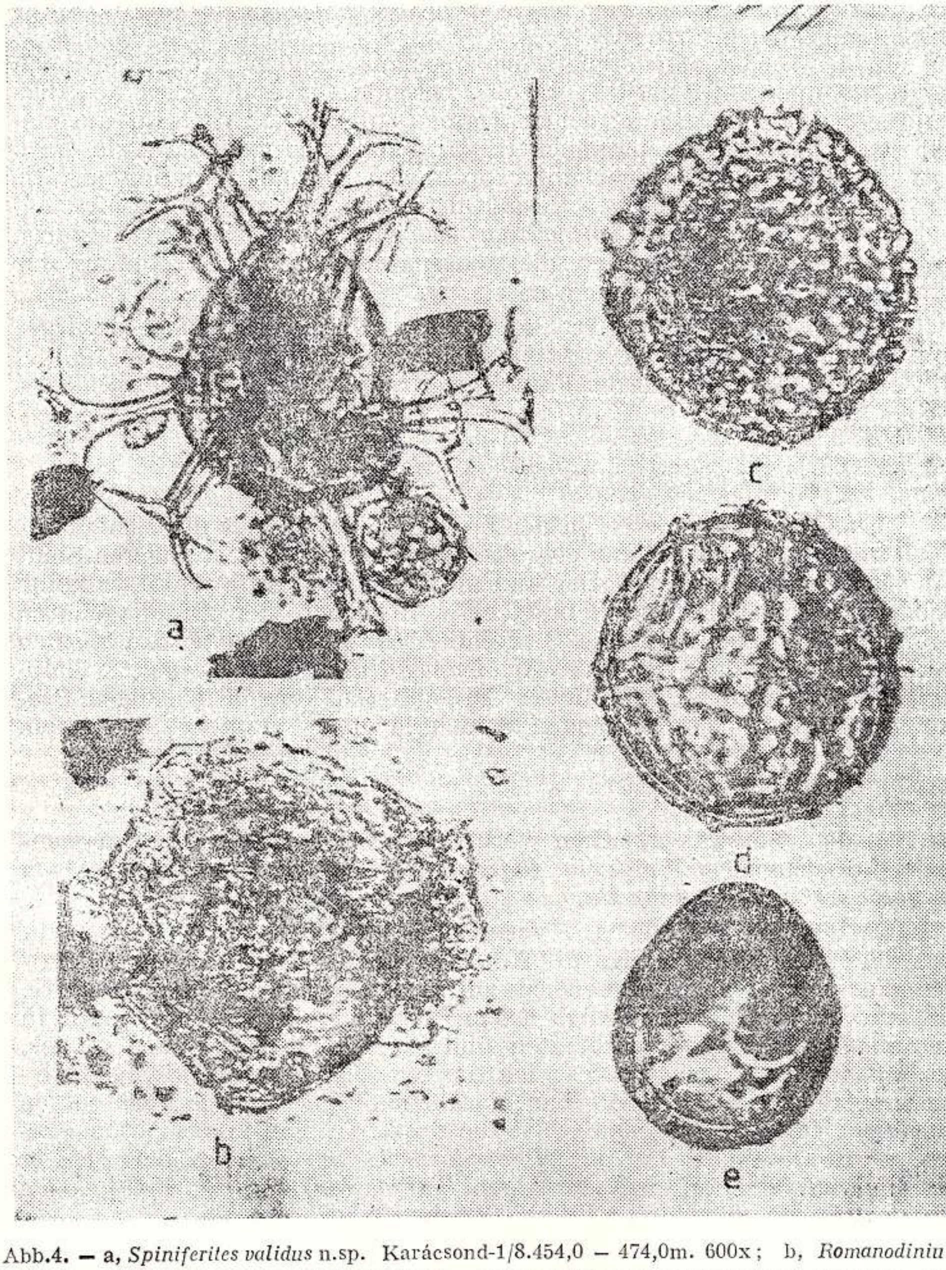

Abb.4. — a, *Spiniferites validus* n.sp. Karácsond-1/8.454,0 — 474,0m. 600x ; b, *Romanodinium areolatum* B a l t e ş, 1971. Szirák-2. 225,8 — 231,5m. 600x ; c, *Impagidinium globosum* n.sp. Szirák-2. 185,1 — 187,4 m.600x ; d, *Impagidinium spongianum* n.sp. Szentlörinc-XII.376,3 — 381,5 m. 600x ; e, *Tectatolinium pellitum* W a l l, 1967. Szirák-2. 194,7 — 195,6 m. 600x.

und 139. des *Pontiadinium*. Ich beobachtete bei der zonenanzeigenden Art mehrere Variationsformen mit verschiedener Grösse und mit verschiedener Länge der Anhängsel.

In der Begleitgemeinschaft der zonenanzeigenden Art gibt es auch einige Exemplare der Art *S. bentori*, besonders in den Fazies des Beckeninneren. Die dick gewordenen Formen mit Theken der im unteren Teil der Biozone, noch häufigen Dinoflagellaten, zeigen, dass sie trotz ihrer Dominanz schon nicht unter optimalen Verhältnissen lebten. Die Veränderung der ökologischen Verhältnisse wird auch durch das Erscheinen der Süsswasseralgen (*Mougeotia laetevirens*, *Cooksonella circularis*, *Spirogyra* Typ 1). angedeutet, die sich später zur Zeit der Bildung der lignitflössigen Schichtenfolge verbreiteten.

Die zonenanzeigende Art kommt am Fusse des Mátra-Gebirges zusammen mit *Kaladacna steindachneri*, *Limnocardium mayeri* (B a r t h a, 1971) vor, bei Tengelic mit der Fauna *Congeria zagrabiensis* (K o r p á s n é H ó d i M.) vor.

Die Dinoflagellata — Zygnemataceae Zwischenzone

Diese Zwischenzone ist in der beckeninneren Fazies der Schichtenfolgen von Somló und Tihany charakteristisch. In diesen sandigeren Schichten bilden die Dinoflagellaten und die Jochalgen der Süsswässer keine horizontal identifizierbare Schicht. Die Selektion der Dinoflagellaten begann schon in der Biozone S. validus. Ihr massenhaftes Aussterben überlebten nur einige Arten, die Art *Impagidinium spongianum*, *I. globosum* und ein Teil der Pontiadinien. Diese Arten können in einigen Proben der brackig — süsswerdenden Schichten noch vermehrt auftreten.

Biozone Mougeotia laetevirens

In den Süsswasserteichen vermehrten sich stellenweise massenhaft die Jochalgen (*Mougeotia*, *Closterium*, *Spirogyra*), sowie die Algen *Botryococcus* und *Cooksonella*.

Die zonenanzeigende Art vermehrte sich in grossen Mengen in Zentral-Transdanubien (Polgárdi, Tab, Nagyszékely), wo im sehr kalkhaltigen Wasser ein optimaler Lebensraum für die Art sich gebildet hatte. Ihre gut erhaltene, dickwändige Exemplare findet man aber überall in den lignitflössigen Pannonschichten und in ihren heteropischen Fazies. Von den Jochalgen beobachtete ich die Art *Closterium kützingii* B r é - b i s s o n in den südlich von dem Nördlichen Mittelgebirge abgeteuften Bohrungen, in Transdanubien fehlte diese Art. In den untersuchten Profilen war die Dominanz der Alge *Mougeotia laetevirens* in entgegengesetzter Beziehung zu der Häufigkeit der Algen *Botryococcus braunii* K ü t z i n g und *Cooksonella circularis* N a g y. (siehe Tabelle).

Die vertikale und horizontale Verbreitung der Dinoflagellata Biozonen

Die am meisten charakteristische Form der Dinoflagellaten Gemeinschaft der Pannonschichten ist die Art *Spiniferites bentori* (R o s s.) S a r - j e a n t. Man findet sie zusammen mit den in den Meeren typischen Hys-

TABELLE

Korrelationstabelle der vertikalen verbreitung der organischskelettigen microplanktons, gesteinfazies und molluskenfauna

Lithostratigraphische Aufteilung (À. Jámbor, 1980)			Mollusken (Bartha, Hódi, Széles Straus, Kro-lopp)	Organischskelettige Mikroplanktons: Dominanz	Organischskelettige Mikroplanktons: Biozone	Vertikale Verbreitung der organischskellettigen Mikroplanktons	
PANNONISCHE FORMATIONSGRUPPE	Oberpannonisch	Zone Torony	*Unio wetzleri*	Süsswasser – Mikroplanktons	*Mougeotia laetevirens*	*P. ultima*; *S. bentori*; *Pontiadinium*; Prevalvat Stadium; *S. validus*; Verdickte Dinofl.; *Spirogyra* Typ. I. maior; *Spirogyra* Typ I. minor; *Spirogyra* Typ 3c; *M. laetevirens*; *Closterium kützingii*; *Zygnema*; *Cooksonella circularis*; *Botryococcus braunii*	Transgression Klimaveränderung Salzgehaltverminderung Aussüssung; Verbreitung der Zygnemataceae
		Zonen Somló – Tihany	*Congeria balatonica*	schnelle Selektion der Dinoflagellaten	Zwischenzone Dinoflagellata-Zygnemataceae		Selektion der Dinoflagellaten; Oberpannon
	Unterpannonisch	Zone Dráva; Zone Csákvár	*C. ungulacaprae*; *C. zagrabienssis*; *C. cžjžeki*; *C. banatica*	massenhafte Dinoflagellaten	*S. validus*; *Spiniferites bentori*		Unterpannon; Subtropischausgeglichenes Klima, optimaler Lebensraum der Algen Typ *bentori*
		Zone Zala	Kleines *Limnocardium*	*Pleurozonaria ultima*			euxine Fazies an der Basis der pannonischen Schichten

trichosphaeriden der Salzwässer in den Tabiano Schichten Norditaliens (D. Habib); in den Quartärschichten Israels (Rossignol, 1962); in den postpleistozänen Schichten Englands (R. Harland) und in den Quartärschichten der Bohrungen im Karibischen Meer (D. Wall).

Einige Formen ihrer Begleitgemeinschaft kennt man aber schon aus der oberen Kreide. Die Art *Spiniferites paradoxus* wurde von Co-o k s o n und E i s e n a c k in den Sedimenten der oberen Kreide Australiens beschrieben, die Art *Pontiadinium inequicornutum* (B a l t e ş) S t o v e r et E v i t t fand ich auch im Sand des oberen Oligozäns (Bohrung Budafok-1.).

Die Alge *S. bentori* bildet in den Tabiano Schichten Norditaliens, in den Schichten des unteren Pontien Rumäniens und in den Schichten mit *C. banatica* Ungarns das älteste Vorkommen der Art.

Charakteristisch für ihr Vorkommen in Ungarn ist, das man sie in den Schichten mit *C. banatica* zusammen mit Nannoplanktons der Salzwässer und Schwachsalzwässer findet, die eine entgegengesetzte Dominanz haben. In den Schichten mit *C. czjzeki* ist diese morphologisch mannigfaltige Algengemeinschaft nur von wenigen Nannoplanktons begleitet, sie kommt mehr mit Kalk-und Kieselschwammnadeln vor. (Untersuchungen von J. B o n á und M. G á l).

Die Art lebt heute in den Lagunen des Karibischen Meeres bei einem tropisch -subtropischen Klima in normalem Salzwasser und bildet 4,5 % aller Planktons (D. W a l l und B. D a l e, 1968). Versuche mit der heute lebenden Art zeigten, dass sie auf die Veränderung der Temperatur empfindlich ist, die Angaben über ihre Salztoleranz sind aber nicht klar.

Die im unteren Teil der Pannon-Stufe massenhaft vermehrte Alge *S. bentori* lebte – ihre heutigen ökologischen Verhältnisse in Betracht nehmend- bei einem sich vermindernden Salzgehalt und subtropisch-ausgeglichenem Klima. Die Ausbildung der Art *S. validus*, sowie die Überverdickung einzelner Arten mit Theken geschah gleichzeitig in dem Beckenrandfazies, sowie in dem beckeninneren Fazies ober der Biozone *S. bentori*. Unter den Umweltsfaktoren konnte ausser der Verminderung des Salzgehalts die *Temperaturveränderung* entscheidend sein, die die morphologische Veränderung und die schnelle Selektion der Dinoflagellaten verursachen konnte. An der Grenze der zwei Biozonen zeigen hauptsächlich die autochtonen Algenüberreste die Klimaveränderung. *Die Art S. validus und ihre Begleitgemeinschaft bilden jenes biologische Objekt, dessen Ausbildung an der Grenze der unteren und oberen Pannonschichten auch die Zeitzonengrenzen zeigt.*

Der Veränderung der Dinoflagellaten-Gemeinschaften folgt verspätet die Veränderung der Kontinentalfauna. Auch der Umweltsanspruch der einzelnen biostratigraphischen Einheiten, Faunen und Floras ist verschieden. Ihre Korrelation ist nur dann möglich, wenn wir auch ihre unterschiedlichen ökologischen Ansprüche in Betracht nehmen.

Neu ist die Untersuchung der Dinoflagellaten, die Ausarbeitung der Gemeinschaften ist noch in Gange. In der Zukunft wird nach der Untersuchung der beckeninneren Schichtenfolgen vollkommener die Erkennung der wechselvollen Reihe der dimorphen Formen und die detailliertere Aufteilung der unterpannonischen Schichtenfolge.

DISCUSSIONS

N. B a l t e ş : There are known the difficulties in the horizontation and especially in the correlation of Pannonian beds in the Pannonian Depression with their equivalents in the extra-Carpathian regions (Carpathian Foredeep and Moesian Platform). Therefore, in my opinion this paper is very important as it brings new contributions and constitutes a practical tool in the knowledge of the Pannonian and generally of the whole evolution of the Pannonian Basin.

In this way I should like to point out the fine work of dr. Sütóne Szentai concerning the organic microplankton (Dinoflagellates included) and particularly two main aspects should be mentioned: 1. the taxonomic one – some new species have been created – and 2. the characteristic stratigraphic levels on Dynocysts and their sharp correlation.

Of course there are also other problems which are to be solved and I am sure that the author will do her best to solve them.

Once again many congratulations to dr. Mária Sütóne Szentai and new other papers like this one.

P. Stevanović: Wir müssen heute diese schöne Funde so auffassen, dass sie nicht nur dem Pannon sondern auch dem Pont angehört. Angefangen von der Molluskenzone *Congeria zagrabiensis* ist schon Pont, Schichten mit *Congeria cžjžeki* sind noch Pannon *s. str*.

NEWER KNOWLEDGE ON THE STRUCTURE AND STRATIGRAPHY OF QUATERNARY SEDIMENTS AT THE ŽITNÝ OSTROV ISLAND IN THE DANUBE LOWLAND, CZECHOSLOVAKIA [1]

BY

E. VAŠKOVSKÁ [2], I. VAŠKOVSKÝ [2]

The Žitný ostrov island on the territory of Czechoslovakia extends in the southwestern part of the Danube lowland between Bratislava and Komárno; it is bordered by the Danube and Little Danube River. The areal extension of the territory is around 1600 sq. km. (Fig.). The region of the island forms an almost ideal young riverain plain (dejection cone) formed by activity of the Danube. The total inclination of the region is from NW to SE. The highest points of the surface are near Bratislava of 134 m altitude and the lowest ones near Komárno of around 107 m altitude. The surface proper of the island is complicated by alternating flat and wide elevations and depressions and also by remnants of the hydrographic system (meanders and arms of the Danube and Little Danube). Sheets of drift sands of less areal extension are also found here. With geomorphological regional subdivision of the Žitný ostrov island are dealing more in detail Lukniš, Mazúr (1959), Vaškovský, Vaškovská (1977), Mazúr, Lukniš (1980).

In the sense of geological-tectonic subdivision of the Danube lowland (Adam, Dlabač, 1961) the region of the Žitný ostrov island takes up the essential part of the so-called central depression, the island part of which is often designated as the "Gabčìkovo Depression" with the centre of sinking near Gabčíkovo. This depression is dish-shaped, its pre-Neogene substratum, according to the newer conception and to available bases (Fusan, Biely, Plančár, 1979), is formed in direction to Bratislava by crystalline schists by the Paleozoic in the area of Komárno and by the Mesozoic of the Hungarian Midmountains immediately near Komárno.

Thickness of Neogene-Quaternary sediments, forming the filling of the Gabčíkovo Depression, is not equal. Its greatest thickness(according to interpretation of geophysical measurements) is up to 5400 m near

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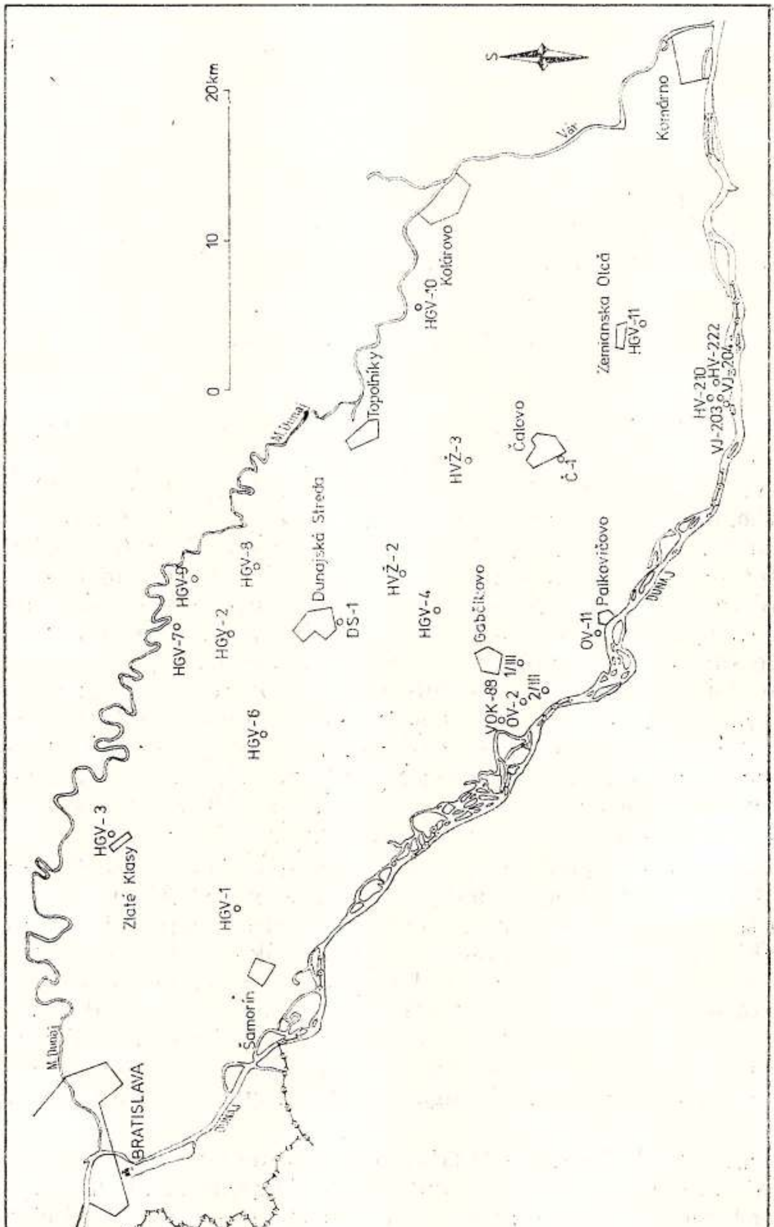

Situation of deeper boreholes in the Žitný ostrov island.

Gabčíkovo; from there in NW direction the pre-Neogene substratum is rising quite equally, filling thicknesses decrease, changing to step-like near Bratislava and practically wedging out at the borders with the Little Carpathians. Thickness of the sedimentary filling decreases also in direction east of Gabčíkovo; in boreholes near Komárno its thickness is 1224—1141 m. The sedimentary filling of the Gabčíkovo Depression, beginning with the Badenian gradually up to present, underwent a complicated geological development.

So far there is no more detailed knowledge about stratigraphy and lithofacial composition of Neogene sediments at the Žitný ostrov island, as no deep boreholes drilling through these sediments in their full thickness in the Gabčíkovo Depression were sunk here. The deepest boreholes, are only near Dunajská Streda (DS-1) and Čalovo (Č-1); these, however, were finished in the Upper Pannonian — Gaža et al. (1972), Čermák, Gaža (1973), Holeczyová, Gaža (1972). The total thickness of the Neogene at the Žitný ostrov island was drilled through by a borehole near Komárno (Pagáč, 1966). Regarding this fact our description of Neogene sediments is based especially more on several deep boreholes in close neighbourhood of the Žitný ostrov island (Kolárovo, Homola, 1966; Diakovce, Homola, 1960) and others.

The presence of Badenian and Sarmatian sediments in the Gabčíkovo Depression (thickness up to 1000 m) is supposed by Adam, Dlabač (1969). The Badenian was directly encountered in borehole Komárno-1. The Pannonian sediments in the marginal part of the depression are lying transgressively on older formations of the pre-Neogene substratum and in the central part on the Sarmatian. Most often they are formed by monotonous greenish-grey to grey, slightly fine sandy calcareous clays of quiet deep sedimentation (thickness up to 2000 m). In the Pontian the sedimentation of calcareous clays of light-greenish-grey and light-grey colour, prevailingly highly sandy, was predominating. Thickness of this sedimentation is up to 1000 m. Dacian sediments are divided here into the lower variegated series and upper coal series. Their thickness attains up to 1500 m. The Dacian underlies the Quaternary sediments in the eastern part of the Žitný ostrov island near Kližská Nemá, Kameničná, Nová Stráž. The Upper Pliocene (Romanian) is formed by the so-called "Gabčíkovo sands" at the Žitný ostrov island. Their occurrence was pointed out already earlier by Janáček, who designated them as "regressive sands" and assigned them to the uppermost Pontian in the year 1966—1966 a and 1966 b. Later (in the years 1969, 1971), this author admitted their assignment to the Upper Pliocene. These sands are spread in the deepest part of the Gabčíkovo Depression, then to the west as far as Šamorín and Rajka in Hungary, to the east or SE to Čalovo, where they are bordered by a tectonic line trending SW—NE. Thickness of the "Gabčíkovo sands", as up to present boreholes show, is 250 to 400 m, so that their base is about 650 m below the surface.

The main component of the sequence of the "Gabčíkovo sands" are sands. In general, medium- to coarse-grained sands with transition into fine pea-gravels predominate. Characteristic of these sequences are also layers of calcareous clays. Their thickness attains 10 to 40 m and they may be traced well at greater distances. The colour of calcareous clays is most often light-grey to grey and light-greenish-grey. The colour of gravels and sands is usually grey to greenish-grey.

The "Gabčíkovo sands" are poor in fossils. So far only Brestenská (1966 a, b) was successful in finding four species and nine individuals of ostracods from the deepest part in the area of the Gabčíkovo Depression near Vojka (in boreholes 1/I and 2/I) and also near Gabčíkovo (in boreholes 1/III, 2/III). In this assemblage, there are Candoniells with *Candoniella marcida*, *Candoniella schubinae*, *Candoniella albicans* and *Ilyocypris* sp. Finds of mollusc have not been known so far, only in borehole OV-11 at depths 367 — 381 m and 530 — 540 m. Brestenská (1966 b) distinguished branches and leaves of coniferous *Glyptostrobus europaeus* (Brgnt.) Here Pacltová (1971) also found an Upper Pliocene assemblage in the pollen spectrum from borehole HVŽ-3 (*Sequoia*, *Taxodium*, *Sciadopytis*, *Liquidambar*, *Cedrus*, than *Carya*, *Juglans*, *Tsuga*, *Castanea* and *Pinus haploxylon*).

It will be possible to correlate stratigraphically the "Gabčíkovo sands" with the so-called Kolárovo Beds, which are found in the wider area of the Žitný ostrov island (Kolesík 1949, Dlabač 1959, Buday 1959, Buday et al. 1962, 1968 etc.); however, still more data will be necessary for it, requiring further studies.

The Quaternary sediments unequally cover the whole surface of the Žitný ostrov island; they are characterized by unequal thickness — at the margins 8 — 12 — 20 m and in the central part up to 340 m or even more. Several authors were dealing with their study (Horusitzky, 1908, 1909; Timkó, 1904; Zoubek, Koutek, 1936; Čepek 1938; Szadecky-Kardoss, 1938; Hromádka, 1956; Lukniš, Mazúr, 1959; Dlabač, 1960; Buday et al., 1962, 1968; Janáček, 1966, 1967, 1969, 1971; Ložek, 1955; Krippel, 1962, 1963; Brestenská, 1966, 1967, 1977; Planderová, 1966, 1967, 1977: Schmidt, 1977; Horniš, 1977; Halouzka, Minaříková, 1977; Minaříková, 1966; Vašková, 1977; Vaškovský, Vašková, 1977, and others). On the basis of up to present more detailed lithogeochemical, mineralogical-petrographical investigations, several paleontological finds and also several data of paleomagnetism (willingly provided by P. Pagáč from the Slovak Academy of Sciences in Bratislava) we distinguish the following genetic types of Quaternary sediments: A) Lacustrine or lacustrine-fluvial sediments; B) Fluvial sediments; C) Sediments of the surficial cover (with specific genetic types of sediments and soils).

A) **Lacustrine or lacustrine-fluvial sediments**. Extension of these sediments in the region of the Žitný ostrov island is roughly bound to its deepest (depression) part, so that it coincides, to a considerable extent, with extension of the above-mentioned so-called "Gabčíkovo sands". They are deposited approximately in depth intervals from 95 to 160 m below the surface. According to HGV boreholes their thickness will be probably more than 200 m. The lacustrine or lacustrine-fluvial sediments in the Žitný ostrov island, as results from the already earlier descriptions by Janáček (1968, 1969, 1971) mainly, however, on the basis of our evaluation of HGV boreholes, are composed of sands and fine-grained gravels. Typical for these sediments is, however, the prevalence of clayey-loamy layers.

The sands found in lacustrine or lacustrine-fluvial sediments are fine, more often medium to coarse-grained, polymict, in places micaceous, of grey to light-grey colour. The pebbles are well worked up. In the pebble component of these sediments M i n a ř í k o v á (1966) has pointed out that quartzites and quartz are predominating 70 – 80%. In the residue are mostly present mainly silicites (20% of pebble component). The silicites are brown, ochreous; less represented are glauconite sandstones. Pebbles of limestones and dolomites are sporadical. Unlike the higher-lying fluvial sediments, no crystalline schists and granites are observed in lacustrine sediments. Their occurrences are seldom around 1%.

In the sandy component of lacustrine or lacustrine-fluvial sediments quartz is predominating in the area under study, the admixture is formed by silicites and feldspars. Working up of grains of the sandy fraction is little; the grains are mostly angular, less subangular and half-rounded.

Clayey-loamy sediments are of important position in lacustrine or lacustrine-fluvial sediments at the Žitný ostrov island. Their layers attain thickness of 5–10 m on an average, maximum 30 m. Visible are sandy to highly sandy clays. They are characterized by grey, brownish-grey, greenish-grey and light-grey etc. colours. We studied these layers more in detail in the following boreholes and depth. In borehole HGV-2 (depth 187, 192, 196, 213 and 233 m), borehole HGV-4 (162, 173, 176, 322 m), borehole HGV-6 (196 an 198 m), borehole HGV-8 (216 m), borehole HGV-9 (135, 164 and 172 m).

The complex litho-geochemical study makes it possible essentially to distinguish two types of lacustrine or lacustrine-fluvial clayey sediments: a) of so-called basal clays and b) of clay layers with more frequent alternation with layers of sands. The basal clays are characterized by greater thicknesses; we follow them mainly in the lower parts of the sequence (e.g.in borehole HGV-4 at depth of 187–233 m, then in borehole HGV-4 depth 159 – 176 and 322 m, in borehole HGV-8 at depth of 212 – 220 m etc.). The second type, characterized by lesser thicknesses of clay layers and more frequent alternation with sands, can be seen for example, in boreholes HGV-6 at depth of 194–234 m, HGV-9 at depth of 161 – 174 and borehole HGV-11 at depth of 69 – 72 m.

The distinguished two types of clayey sediments in lacustrine or lacustrine-fluvial sequences differ not only in thickness but also in lithological-geochemical properties.

The basal clays in lacustrine or lacustrine-fluvial sediments at the Žitný ostrov island represent mono-, bi- and polydispersive loams with granulometric maxima in fractions 0.05 – 0.01 mm, then 0.25 – 0.1 and < 0.002 m; they are medium-sorted to sorted, characterized by a variable content of $CaCO_3 = 1.2 - 17.5$ %, are weakly humose (humus up to 0.3 %), active pH reaction in $H_2O = 8.4 - 8.95$ (or weakly alkalic to alkalic) exchangeable pH in $KCl = 7.4 - 7.8$. An interesting particularity of these clayey sediments is that with the commonly used dispergator (6% $Na_4P_2O_7$) they induce peptization instead of dispergation.

The second type of clayey sediments distinguished in lacustrine or lacustrine-fluvial sequences at the Žitný ostrov island is characterized mainly by a low coefficient of microaggregation (K micr 5—9); they are polymodal with granulometric maxima in fractions 0.25 — 0.1; 0.05 — 0.01 and < 0.002 mm. The content of $CaCO_3$ also varies within the range of 2.3 — 17.9%. They are characterized by higher activity of pH in $H_2O = 8.6 - 9.0$.

The difference in the litho-geochemical properties of the distinguished two types of clayey sediments in lacustrine or lacustrine-fluvial sequences is explained by different fluvial conditions of their forming and also by different diagenetic processes.

The content of heavy minerals (according to J. Horniš, 1977), in lacustrine or lacustrine-fluvial sediments at the Žitný ostrov island, is not equal, considerably varying is the representation of individual minerals. So, in borehole HGV-4 the content of opaque minerals is within the range of 7.7 — 16.6%, of garnets 2.6 — 13.0%, clorite 0.5 — 2.4%, carbonates 36.6 — 75.1%. In borehole HGV-6 opaque minerals are represented 8.5 — 82.2%, garnets 1.0 — 38.3%, chlorite 1.0 — 41.6% and carbonates 4.2 — 54.5%. Particularly characteristic of this type of sediments when compared to other is a relatively high representation of carbonates and a low content of garnets. Rarely, the occurrence of hypersthene is also recorded in these sediments.

The lacustrine or lacustrine-fluvial sediments distinguished by us at the Žitný ostrov island are generally known in not being particularly rich in paleontological remnants. More important finds of ostracods in the frame of this investigation were recorded by Brestenská (1977) in borehole HGV-4 (depth 97, 99.3 and 102 m), in borehole HGV-3 (depth 106.5 — 106.6 m), in borehole HGV-7 (depth 161.3 — 161.4 m and 193.0 —195.5 m). In the ostracod fauna the following species are essentially represented: *Candonea candida* (Müller), *Candida* sp. div. juv. str., *Cyclocipros ovum* (Jurine), *Scottia browniana* (Jones), *Scottia tumida* (Jones), *Limnocythera inopinata* (Baird), *Metacypris cordata* Brady et Robertson, *Ilyocypris gibba* (Ramdohr), It is worth mentioning the authoress', find of macrospore *Azolla ficuloides* Lam. *fossilis* Flrorschütz in borehole HGV-6 at depth 223.5— — 223.6 m. The finds of spores and pollen from clayey layers from boreholes HGV-6 (depth 197 — 198 m) and HGV-7 (depth 183.5 m) by Planderová (1977) are also important. In borehole HGV-6 there are spores and pollen of the following familiers and genera: Gramineae, *Cedrus* sp., Chenopodiaceae, Liliaceae, Cyperaceae, *Quercus*, *Pinacea*, *Pinus silvestris*, *Pleurozonaria* and in borehole HGV-7 practically of similar representation: *Quercus*, *Keteleeria*, *Abies*, *Picea*, freshwater plankton of the genus *Pleurozonaria*, NAP, mainly Ranunculaceae, Rosaceae.

Within the complex study of deep hydrogeological boreholes at the Žitný ostrov island we tentatively took samples from three boreholes (HGV-5, HGV-6 and HGV-9) of the core from clayey layers for the study of paleomagnetism. From borehole HGV-5 samples from the depths of 135.5, 137.9, 140.5 and 142.8 were taken. The upper three samples (from depths of 135.5, 127.4 and 137.9 m were characterized by positive magnetization and the lower by negative. From borehole HGV-6 we took samples from depths of 197.5, 203.0, 230.8 and 238 m. All of them (except the sample from the depth of 203.4 m) displayed negative magnetization. In borehole HGV-9 samples were taken from depth 161.2 to 161.7 m (on the whole 25 samples). Eleven upper samples except one displayed direct magnetization to the depth of 161.4 and positive magnetization from the depth of 161.4 m to 161.7 m.

Setting out from paleomagnetic scale of C o x (1968), the obtained values of paleomagnetism make it possible the following interpretation in the individual boreholes. The magnetic inversion found in borehole HGV-5 is bound to the inner part of the loamy-clayey sequence, which shows traces after pedogenetic processes. We suppose that the mentioned magnetic inversion in this sequence (depth 137.9—140.5 m) corresponds to the inversion boundary Brunes-Matujama (in the chronological scale it is around 0.7 m.y. This whole sequence (at depth of 118.9—143.0 m) corresponds to the Günz/Mindel-Cromerian interglacial period. We correlate this sequence in age then also with the sequences in borehole HGV-3 (depth 95 m) and borehole HGV-4 (depth 144 m) with the already mentioned finds of ostracodes and assign them to the Cromerian.

The paleomagnetic data obtained from borehole HGV-6 show that except one sample from the depth of 203 m they have negative magnetization. The complex of sediments in this borehole (perhaps representing a fossil soil horizon) from the depth of 197.5 m corresponds in time to the Matujama magnetic epoch. The positive polarization of the sample from depth 203 m could probably indicate the Jaramillo episode (inside the Matujama epoch), providing the possibility to assign this sequence to the Donau/Günz interglacial period. In any case there is a sequence of the Early Quaternary, younger than 1.8 m.y. The assignment of this sequence to the interglacial is also supported by finds of flora (from the depth of 197 — 198 m) as well as by the occurrence of *Azolla filiculoides* L a m. in this borehole from the depth of 223.5 — 223.6 m.

Similarly as in the mentioned borehole HGV-6, also in borehole HGV-9 at depth 161.4 — 161.6 m a magnetic inversion was observed, most probably corresponding to the Jaramillo episode (of the Matujama epoch). In this borehole, as in the preceding one, the underlying beds also formed in the Early Quaternary. Their common feature, as we have already mentioned, is the low coefficient of microaggregation, the low content of humus, variable content of $CaCO_3$ and high pH values.

We range to period of the Earliest Interglacial Donau/Günz, besides the mentioned sequences in boreholes HGV-6 and HGV-9, also clayey sequences in borehole HGV-7, in which at the depth of 183 m the find of interglacial flora is recorded and at the depth of 161.3 — 161.1 m are also finds of ostracods of interglacial character.

The total of data obtained by lithological-geochemical, paleontological, mineralogical-petrographical investigation and the study of paleomagnetism makes it possible to range the distinguished lacustrine or lacustrine-fluvial sediments at the Žitný ostrov island generally to the Eopleistocene, i.e. to the stratigraphic division of the Quaternary distinguished in the sense of G r o m o v (1957), G r o m o v, K r a s n o v, N i k i f o r o v a (1958), N i k i f o r o v a, K r a s n o v et al. (1976).

In this stratigraphic division of the Quaternary the following stages are also included: Biber, Biber/Donau, Donau, Donau/Günz, Günz, Günz/Mindel. A more detailed division of the individual stages will require still further investigations.

In order to support assignment of the lacustrine and lacustrine-fluvial sediments at the Žitný ostrov island to the Eopleistocene the first finds of ostracodes established by J i r í č e k (1969) from boreholes HVŽ-3 and HVŽ-2 from the area of Dunajská Streda may be still mentioned. This author mentions in borehole HVŽ-3 8 species and 13 individuals and in the underlier of the ostracod fauna in borehole HVŽ-2 he established gastropods: *Gastocopta* cf. *noulitiana* (D u p.), *Carychium minimum* (M ü l l e r), *Segmentina* sp., *Cepaea* sp., *Bythynia* S c h w e y r. H y r o š o v a - V á v r o v á (1959) and B u d a y (1959, 1962) called attention to finds of fauna similar to those mentioned by R. J i r í č e k from the area of Dunajská Streda from Eopleistocene sediments of the eastern part of the Žitný ostrov island near Kameničná. At last, for completeness it is necessary to mention that B r e s t e n s k á (1966 a, b) determined also 15 species and 63 individuals of ostracods from the central part of the Žitný ostrov island from the area of Gabčikovo (boreholes OV-2 and VOK-88), then from the area of Palkovičovo (borehole OV-11) and Kližská Nemá (in boreholes HV-210, HV-222, VJ-203, VJ-204).

B) **Fluvial sediments build up the upper complex of Quaternary sediments at the Žitný ostrov island.** The sediments of this complex occur at surface mainly in the upper part of the core of the island. They are deposited on the above described *Eopleistocene lacustrine* and lacustrine-fluvial sediments, then on older formations (Dacian or Upper Pliocene). This upper complex was designated by J a n á č e k (1967 a, 1969) as the "Danube gravelous series". Its thickness is not equal (below Bratislava about 20 m, near Komárno 8 – 12 m and in the central part near Gabčíkovo it attains 130 – 150 m) according to J a n á č e k. In the deep hydrogeological boreholes (HGV) studied by us it attains up to 160 m.

The upper complex of fluvial sediments at the Žitný ostrov island is predominantly built up of gravels, gravelous sands, sands (sometimes also with larger pebbles); less represented are here clayey or loamy sediments, which are most often found in form of lenticles, more rarely form layers extending at greater distances.

The greater thickness of the upper complex and the prevalence of thicker loose materials in its lithological composition reflect con-strative accumulation. Though in its structure stream-bed, flood-plain and dead arm facies take part, which are superimposed in cyclic superposition on one another as is to be seen from boreholes, however, the stream-bed facies predominates and the flood-plain; the dead arm facies are usually removed or preserved at various depth in form of larger or smaller lenticles. Destruction of these facies was by the influence of lateral erosion of the migrating Danube stream-bed. In close co-activity of accumulation and synsedimentary tectonic mouvements a typically con-strative fluvial with many-stage or polycyclic development formed. Between the individual cycles lying on one another we observe wash-outs, separating the individual beds, however, usually of no stratigraphic importance.

In the area under study the gravelous material of the upper Quaternary fluvial complex is usually marked by characteristic rusty-brown to brownish yellow colouring with frequent limonitization. According to the data of M i n a r í k o v á (1966), in petrographic composition of this complex quartz and quartzites (70 – 80%) are mostly represented. In the residue there are various types of silicates (5%), mainly brown and ochreous silicites, less grey silicites and rarely reddish-brown, red and black silicites. Another type of rock are limestones and dolomites, more represented in the upper parts (around 6%), however, decreasing downward. From sandstones (content around 4 – 8%) most abundant are quartz sandstones, less polymictic sandstones, more rare red quartz sandstones, micaceous fine-grained sandstones, glauconite sandstones and calcareous sandstones. Approximatelly equally represented as sandstones are also crystalline schists, the content of which decreases downward and the amount of silicites increases.

From crystalline schists are mainly represented two-mica gneisses, less muscovite, light-coloured biotite and amphibole, rarely garnet-biotite gneisses and granulites. From granitic rocks, forming only a small admixture (1%), two-mica and muscovite granites and fragments of pegmatite character are present.

Among heavy minerals in gravelous sandy sediments of the upper fluvial complex, according to H o r n i š (1977), opaque minerals and garnet are mostly represented. The content of opaque minerals varies within the range of 9.5 – 69%, their greatest representation is observed in horeholes HGV-1 (up to 69%), HGV-2 (up to 51%) and HGV-3 (up to 47.5%). Garnets are found within the range of 1 – 45.2% the greatest content is recorded in borehole HGV-3 (up to 45.2%), in borehole HGV-8 (up to 40.8%), HGV-9 (up to 37.0%) and HGV-2 (up to 40.8%). Whereas opaque minerals are relatively equally represented in the complex, in garnets we observe a decreasing trend in direction from above downward practically in all observed boreholes. Representation of other heavy minerals is as amphiboles (up to 14%), chlorite (up to 15%), epidote (up to 18.5%). From other heavy minerals there are still represented apatite, zircon, rutile, tourmaline, zoisite, staurolite, disthene, augite, titanite, biotite, sillimanite, andalusite and dimmed minerals. Carbonates in this complex are represented with low contents. Especially interesting in this complex is the presence of hypersthene, bound to the northwestern and central part of the Žitný ostrov island and to the depth of around 80 – 109 m; its content attains up to 4%. Its presence convincingly confirms the already earlier expressed opinion (V a š k o v s k ý, 1970) of influx of the Váh River into the Danube River in the Older Quaternary in the southwestern part of the Danube lowland and its gradual migration to SE.

The upper gravelous sandy complex is mostly sterile paleontologically. The only microfauna of freshwater ostracods was found by B r e s t e n s k á (1977) in borehole CHGP-1 near Benková Potôň at depth 84 and 93 m : *Candona* div. sp. juv., *Cyclocypris* sp., *Limnocythere* sp., *Ilyocypris* sp., and in borehole HGV-6 near Michal na Ostrove at depth 98 –

— 99 m she established *Ilyocypris gibba* (R a m d o h r). The earlier finds of the mentioned authoress from the year 1966 from boreholes near Gabčikovo /OV-2, OV-12 / and Palkovičovo / OV-11 and OV-10/ where she found 9 species and 21 individuals may be also ranged to this circle.

According to the superposition of the upper gravelous sandy fluvial complex we presume that it was forming during the periglacial Pleistocene, i.e. from the Mindel to the end of the Würm. So far we have not sufficient data for its more detailed stratigraphical subdivision. We may only suppose about clayey layers in boreholes HGV-2 (83—84 m), HGV -8(93 m) and HGV-1 (80 — 85 m), in which finds of interglacial fauna are observed, that there is most probably the Mindel/Riss Interglacial. The mineralogical-petrographical analyses point out that the fluvial complex is built up of Danube material mixed with the material from Carpathian streams.

C) **Sediments of the surficial cover.** The sediments of the surficial cover represent a complex of sediments, which are covering the above described complex of fluvial gravelous sandy sediments at the Žitný ostrov island. Thickness of sediments of the surficial cover is variable, up to 5 — 8 m. In the structure of the surficial cover the following facial-genetic types of Quaternary sediments take part : a) fluvial-eolian (loess-like) loams ; b) sediments of the flood-plain facies ; c) sediments of the facies of dead arms ; d) drift and re-blown sands ; e) buried soils—as a particular product of the Quaternary. The sediments of the surficial cover represent the time of the Late Glacial, the Würm and Holocene. These facial-genetic types of sediments were described more in detail by V a š k o v s k á, V a š k o v s k ý, S c h m i d t (1979).

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НОВЫЕ ДАННЫЕ О ПЕРМСКИХ ОТЛОЖЕНИЯХ СЕВЕРО-ЗАПАДНОГО ОКОНЧАНИЯ МАРМАРОШСКОГО МАССИВА[1]

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Длительное время считалось, что некоторые пестроцветные образования северо-западного окончания Мармарошского массива и Чивчинских гор являются одновозрастными и сформировались в триасовое время (Т к а ч у к, Г р у ж и й, 1957; С л а в и н, 1963). Такой вывод явился следствием сложных тектонических условий, затрудняющих изучение последовательности накопления осадков, и внешнего сходства верхнепалеозойских и мезозойских отложений.

На основании изучения литолого-фациального состава, выяснения структурно-тектонического положения и биостратиграфических особенностей в толще пестроцветных вулканогенно-осадочных образований в настоящее время установлены каменноугольные, пермские и нижнетриасовые породы (Ж у к о в, С е р г е е в а, П а с е ч н и к, 1964; С л а в и н, Ж у к о в, 1966; С е р г е е в а, Ж у к о в, 1966; С е р г е е в а, 1974; Ж у к о в, В о з а р, Я н е в, 1976; С е р г е е в а, 1980).

Обстоятельный обзор микрофлористического изучения метаморфизованных пород верхнего палеозоя Украинских Восточных Карпат приведен в нашей публикации 1980 г. (С е р г е е в а, Ж у к о в, 1980). В частности здесь изложен палеопалинологический материал, свидетельствующий присутствии среди исследуемых осадков образований карбона (преимущественно верхнего отдела) и проведен сравнительный анализ палиноморф с таковыми из аналогичных образований польской части Карпат, южной части гор Бихор (район Арьяшени), Малых Карпат (окрестности Ламача), Восточной Чехии и Балканид.

На отложениях, содержащих микрофитофоссилии каменноугольного возраста, несогласно залегает толща пестроцветных пород, относящаяся ранее к триасу. В настоящее время в пределах Раховского массива она выделяется в качестве самостоятельной красноплесненской свиты, принадлежность которой к перми подтвердилась нашими нанодками спор и пыльцы (Ж у к о в, С е р г е е в а, П а с е ч н и к, 1964; С е р-

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геева, Жуков, 1966; Сергеева, 1974). Это оболочки *Perisaccus pumicosus* (Ibr.) Isch., *Rhytisaccus subnotatus* Naum., *Pemphygaletces auritus* Lub., *Florinites* sp., *Trilobozonotriletes inciso-trilobus* (Waltz) Naum., *T. trivialis* (Waltz) Isch., *Hymeno-zonotriletes granulatus* (Naum.) Isch. (Красный Плес, стратотипический разрез свиты).

Единичные пермские формы встречались также нами в графитизированных кварцитах участка Кобылецкая Поляна (*Vittatina striata* Lub., *Pemphygaletes auritus* Lub.) и в алевролитовых породах, обнажающихся по рр. Большой Угольке и Перкалаб.

Таким образом, в базальной части пестроцветных вулканогенно-осадочных образований Раховского массива и Чивчинских гор нами были обнаружены и описаны палиноморфы пермских преимущественно хвойных и кордаитовых растений. Однако, вследствие редкой встречаемости этих оболочек, принадлежность микрофитофоссилий к нижнему или верхнему отделам пермской системы не указывалась.

Материал, полученный нами в последнее время, позволяет дополнить палинологическую характеристику базальной, вулканогенно-осадочной и гипсоносной свит Раховского массива и уточнить возраст слагающих их пород.

Здесь впервые встречены углефицированные споры: *Raistrickia heteromorpha* (Andr.) Siverc., *Nigrisporites nigritellus* (Lub.) Oschurk., *Calamospora plicata* (Waltz) Siverc., *Verrucosisporites verrucosus* Ibr., *Lueckisporites* sp. Пыльца кордаитовых представлена видами *Cordaitina rotata* (Lub.) Medv., *C. uralensis* (Lub.) Samoil. Отмечены единичные *Vittatina vittifer* (Lub.) Samoil., *V. striata* Lub., пыльца голосеменных растений—*Ginkgocycadophytus*. Вместе с установленными ранее *Perisaccus pumicosus* (Ibr.) Isch., *Rhytisaccus subnotatus* Naum., *Pemphygaletes auritus* Lub., отдельными *Florinites* sp., и другими приведенный состав палиноморф отвечает верхней перми.

Микрофитофоссилии сходного состава, представленные *Florinites* sp., *F. schopfi* Medv., *F.* cf. *tener* Medv., *Cordaitina uralensis* (Lub.) Samoil., *Striatopodocarpites* sp., *Punctatisporites punctulatus* Ibr., *Convolutispora* sp., *Vittatina* sp. были обнаружены нами (Ачеев, Жуков, Газданов и др., 1976) в терригенных образованиях Северной Осетии (междуречье Геналдон-Фиагдон), проблематично относимых к карбону и залегающих под фаунистически охарактеризованными верхнепермскими известняками.

Сравнительный анализ микрофитофоссилий, выделенных из перми Украинских Восточных Карпат, Северной Осетии и аналогичных образований г. Сливена Восточных Балканид (образцы переданы И. Пашевым и Ф. Жуковым, показал следующее.

Для проб, отобранных у г. Сливена, также как для Восточных Карпат и Северной Осетии, характерна одномешковая пыльца *Florinites* sp. (*F. luberae* Samoil., *F. pumicosus* Sch., Will. et Ben., *F. ovalis* Bhardw.), занимающая в палинологическом спектре господствующее положение. Отмечаются здесь пыльцевые оболочки кордаитовых [*Cordaitina* cf. *spongiosa* (Lub.) Samoil]. и виттатин (*Vittatina* *striata* Lub., *V. vittifer* Lub.), имеющие также распространение в пермских отложениях Северной Осетии.

Наряду с пыльцой, в исследуемых породах Восточных Балканид развиты оболочки *Sporonites* sp., *Inderites compacta* (Lub.) Abram. et March., акритархи родов *Verihachium* sp., *Leiosphaeridia* sp., ? *Tasmanites* sp.

Пермский возраст пород г. Сливена, в свою очередь, подтверждают споры ископаемых грибов типа *Sporonites* (Potonié) Ibr., характерные для отложений цехштейна.

Оболочки турмы *Saccites* известны и в метаморфизированных отложениях перми Малых Карпат (Жуков, Возар, Янев, 1976). Вместе с ними в пестроцветных породах распространены *Sporonites* sp., *Reticulatisporites* sp., *Punctatisporites* sp., *Schulzospora* sp., *Taeniaesporites* sp., *Calamospora* sp., *Raistrickia* sp., *Densosporites* sp., *Cingulatizonates* sp., *Florinites* sp., *Protohaploxypinus* sp., *Protodiploxypinus* sp., *Pityosporites* и другие. Комплекс пермских спороморф выделен в метаморфитах Чешского массива близ г. Чески-Брод. Он состоит преимущественно из пыльцы с участием рода *Vittatina* (Konzalova, 1970). Пермская микрофлора, найденная в восточной части Низких Татр, характеризуется обилием форм родов *Vittatina*, *Striatosaccites*, *Protohaploxypinus*, *Ginkgocycadophites*, *Punctatisporites*, *Apiculatisporites*, *Aulisporites* и многих других. Смешанные комплексы, состоящие из верхнепермских *Vittatina*, *Lueckisporites virkkiae*, *L. parvus*, *L. globosus*, *Jugasporites delasaucei*, *Limitisporites moersensis*, *Potoniesporites* и некоторых триасовых форм, выделены Е. Пландеровой, 1973) из пелитопсаммитовых отложений мелафировой серии единицы Хог в северо-восточной части Низких Татр.

В Восточной Словакии в метаморфизованных графитовых сланцах, залегающих под породами неогена, установлена ассоциация спор *Karpatisporites*, *Klausipollenites*, *Taeniaesporites*, *Striatites* и другие, указывающие на принадлежность отдельных неовулканитов к верхней перми.

Однако по составу, облику, морфологическим особенностям палиноморфам пестроцветных образований Раховского массива Украинских Восточных Карпат наиболее соответствуют комплексы спор и пыльцы, установленные С. Чернявской и Ж. Лачевой (Жуков, Возар, Янев, 1976) в скв. Р—7 (Тырговищте) и скв. Р—3 (Ветрино) Мизийской плиты (Северная Болгария). Здесь характерны верхнепермские *Nuskoisporites klausi*, *Limitisporites rectus*, *L. latus*, *Lueckisporites virkkiae*, *Taeniaesporites noviaulensis*, *Platysaccus papiliosus*, *Falcisporites zatiei*, *Labisporites granulatus*, *Nuskoisporites gondvanensis*, *Vittatina vittifer*, *V. striata*, *Platysaccus* sp., *Jugasporites* sp., *Protodiploxypinus elongatus*, *Pemphygaletes striatus*, *Dacridium* sp., *Caytoniales* sp. Названные микрофитофоссилии ценны тем, что извлечены из терригенных и вулканогенных пестроцветных образований, сопоставленных на основании литолого-фациального анализа (Жуков, Сергеева, Янев, 1976) с обнажениями перми Балканских гор. При этом следует указать, что этот палеопалинологический материал получен из скважин, которые в ряде случаев вскрывают полный разрез перми (Ветрино). Учитывая сходство микроформ, извлеченных из названных образований Мизийской плиты и каменного материала, отобранного близ г. Сливена, возможно высказать предположение о позднепермском возрасте осадочных пород Сливенской Планины.

Интересно отметить, что сходные нашим ассоциации верхнепермских микрофитофоссилий изучены (Jerzykiewicz, 1979) из отложений, вскрытых скв. Болеславец-24 (Северосудетский синклинорий). Миоспоры здесь представлены группой Saccites: *Lueckisporites*, *Klausipollenites*, *Pityosporites*, *Labisporites*, *Jugasporites*, а также отдельными одномешковыми формами *Cordaitina*, *Potonieisporites*, *Nuskoisporites*. Этот комплекс соответствует комплексам верхней перми Свентокшиских гор и Западной Европы.

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ПАЛИНОМОРФЫ ДЕВОНА В МЕТАМОРФИТАХ СТРАНДЖИ[1]

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Странджанский горный массив — своеобразный геологический регион юго-востока Болгарии, сложенный породами архейско-четвертичного возраста, частично, претерпевшими метаморфизм. Из-за сложности тектонического строения, седиментологического и метаморфического развития, а также отсутствия характерных органических остатков, стратиграфическое расчленение метаморфизованных образований представляют здесь сложную проблему.

Детальный анализ различных точек зрения на геологию Странджи представлен И. К. Начевыми Й. Г. Маляковым Третьей научной сессии (1978 г.), посвященной этому вопросу, а также всесторонне изложен ими в публикации 1979 г. (Начев, Маляков, 1979). В этой работе авторы, учитывая главные различия во взглядах на возраст пород и характер строения Стоиловской структуры, выделяют три этапа в развитии познания геологии Странджи. При этом, основываясь на новых стратиграфических и седиментологических данных, они отдают предпочтение одному из них, подтверждающему на основании палеопалинологических исследований палеозойский возраст пород аспидной формации Странджанского антиклинория.

Впервые ассоциации микрофитофоссилий в метаморфитах проблематичного возраста Стоиловской структуры были получены нами в период совместного изучения образцов пород, отобранных в Срандже, Искырском ущелье и Краиште (Сергеева, Начев, Маляков, 1979). Результаты палинологического анализа показали, что в большинстве проб (в 14 из 28) присутствуют палиноморфы кембрия, девона и карбона (Стоиловская структура), ордовика и силура (Искырское ущелье) девона и карбона (Краиште). До наших исследований метаморфизованные образования Странджи, в основном, интерпретировались как мезозойские или относились к формациям неустановленного возраста. Что касается материала из Искырского ущелья и Краиште, то он имел достоверно установленный палеозойский возраст и был отобран нами для корреляции по органическим остаткам со Стоиловской структурой.

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На участке Дядо Костов дол (Странджа) в черных мелкозернистых породах нами установлены акритархи, свойственные, преимущественно, отложениям среднего и верхнего кембрия. Не исключено, что они могут присутствовать и в образованиях нижнего ордовика. Это редкие оболочки *Dictyotidium* cf. *cambriense* Slav., *Lophidiacrodium arbustum* Tim., *Acanthodiacrodium vestitum* Tim., *Dasyrytidiacrodium* sp., *Ooidium* sp., *Vendotaenia* sp. Приведенные растительные остатки по преобладанию диакродиевых сходны с комплексом микрофитофоссилий терригенно-карбонатной толщи Юго-Восточного Памира (Сиверцева, Смирнова, 1974). Отмечаются также скопления оболочек этого облика в кембрии Западной Европы, Северной Африки, Антарктиды и других территорий.

При исследовании филлитов и мраморов на участках Бръшлян (р. Чурка), Каменска бърчина, Катун (р. Велека) нами были обнаружены ткани растений, состоящие из фрагментов кутикул и трахеид, распространенных в палеозойских отложениях. Наиболее интересны трахеиды с окаймленными порами араукароидного типа, встреченные в глинистых примесях мраморов участка Каменска бърчина. Подобные ткани характерны для девонзких-каменноугольных отложений Украинских Восточных Карпат (Сергеева, 1966, 1974, 1980). Достаточно распространены здесь и примитивные ткани сосудистых растений, вместе с которыми были встречены единичные корродированные микроспоры *Leiotriletes* sp., *Retusotriletes* sp., отдельные деформированные акритархи *Verihachium* sp. и хитинозоа *Conochitina* sp. также свидетельствующие о палеозойском возрасте вмещающих пород.

Углефицированные остатки спор *Leiotriletes* sp., *Trachytriletes* sp., *Lophotriletes* sp., акритархи *Mycrhystridium* sp., *Verihachium* sp., *Leiosphaeridia* sp. неудовлетворительной сохранности, трахеиды с простой диагональной и супротивной поровостью были установлены также на участках вдоль рек Младежка и Стръвница. При предварительных исследованиях (Сергеева, Начев, Маляков, 1979) из-за малочисленности фактического материала, вмещающие эти микрофитофоссилии породы мы относили к палеозою. В настоящее время появились новые данные, утверждающие и уточняющие этот вывод. В частности, имеются в виду палиноморфы девона, встреченные в обнажениях при повторном палинологическом изучении филлитов участка Катун и по рекам Младежка и Стръвница.

Микрофлористические остатки девона характеризуются здесь следующими чертами: во всех образцах присутствуют споры с ребристой структурой видов *Emphanisporites neglectus* Vigr., *E. rotatus* Mc Greg. в сочетании с *Trachytriletes* ex gr. *minor* Naum. Эти палиноморфы нами ранее не встречались. Спорадически наблюдались *Retusotriletes simplex* Naum., *R.* sp., Однолучевые формы *Azonomonoletes costatus* Tschibr. были отмечены только на уч. Катун. Найдены акритархи подгрупп (родов) Acanthomorphitae—*Mycrhystridium* sp., Netromorphitae — *Leiofusa* ? *striata* Mach. Polygonomorphitae — *Verihachium* sp. 1 и 2, *V. polyaster* Stapl. и другие. Интересны *Psophosphaera* aff. *safes* Tschibr. также обнаруженные впервые (уч. Катун). Растительные ткани однообразны и представлены фрагментами лестничных трахеид.

Получены новые данные и о возрасте образца с. 1650 (м. 502, 80). Здесь отмечены споры *Emphanisporites rotatus* McGreg. и *Retusotriletes minor* Naum. И, хотя эти оболочки немногочисленны и обуглены, их присутствие свидетельствует в пользу девонского возраста вмещающего их филлита. В начальной стадии палинологического изучения образца С. 1650 отмечалось, что в пробе присутствует достаточное количество углефицированного материала, который на данном этапе исследований невозможно определить даже до рода. Тем не менее, облик, форма, характер скульптуры и другие морфологические признаки растительных остатков уже тогда говорили о том, что они являются палеозойскими.

Удалось определить до вида и некоторые разрушенные акритархи рода *Mychrystridium*, найденные в образце С. 1650 (м. 481). Их также, по-видимому, следует считать девонскими.

В сланцевых аргиллитах близ с. Долна Мелна и с. Стайчовцы (Краиште) присутствуют микрофитофоссилии нижнего девона в общих чертах напоминающие девонские оболочки Странджи. К ним относятся споры высших растений *Emphanisporites rotatus* Mc Greg., *Retusotriletes divulgatus* Tschibr. var. *plicatus* Tschibr., *R. simplex* Naum., *R. minor* Naum., *R. pychovii* Naum., *Reticulatisporites* cf. *emsiensis* All., *Leiotriletes* sp., *Acanthotriletes parvispinosus* Naum., акритархи *Winvaloeusia* cf. *distracta* (Deunff) Deunff, *Verihachium trispinosum* (Eis.) Cram., *V. europaeum* Stock. et Will., *V. downiei* Stock. et Will., *Leiofusa striata* Mach., *Cymatiosphaera* cf. *miloni* Deunff., *Domasia* sp., *Multiplicisphaeridium* cf. *ramusculosum* (Defl.) List.

Ранее (Сергеева, 1966, 1974, 1980) споры девона были выявлены нами при палинологическом изучении метаморфизованных разностей терригенно-карбонатных пород деловецкого комплекса, распространенных в пределах северо-западного окончания Мармарошского массива (Сергеева, 1974). Это *Leiotriletes minutissimus* Naum., *Trachytriletes minor* Naum., *Retusotriletes* cf. *devonicus* Naum., *R. subgibberosus* Naum., *R. punctatus* Tschibr., *Acanthotriletes* cf. *serratus* Naum., *Camarozonotriletes* sp., *Emphanisporites* sp., *Archaeozonotriletes* sp., *Azonomonoletes laevis* Tschibr. и др., многочисленные ткани растений. Отмечены отдельные акритархи родов *Verihachium* sp. и *Baltisphaeridium* sp. Хотя местонахождения этих микрофоссилий единичны, о некоторые споры порой имеют нечеткую скульптуру, их присутствие в терригенно-карбонатной толще пород деловецкого комплекса позволяет судить о возможном присутствии среди метаморфитов Украинских Восточных Карпат и отложений девона, скорее всего, нижнего.

У с. Кобылецкая Поляна девонские формы отмечены в графитизированных сланцах как в виде самостоятельных групп, так и совместно со спорами каменноугольного возраста (*Hymenozonotriletes* sp., *Trilobozonotriletes inciso-trilobus* Naum., *Tripartites trifoliolatus* (Horst) Dyb. et Jach.) Последние, очевидно, вмыты, так как приурочены к трещинам, заполненным кальцитовым материалом. Всречены девонские микрофитофоссилии и в графитизированных сланцах участка Красное Плесо.

Что касается широкого распространения в Страндже метаморфизованных пород карбона, то пока уверенно судить по этому поводу не приходится, так как здесь не встречены четкие ассоциации каменноугольных микрофитофоссилий.

Сохранность вышеупомянутых палиноморф неудовлетворительная — оболочки часто разрушены, подвержены коррозии и пиритизации. Вследствие углефикации изменен их цвет. Очевидно, подобные изменения микрофитофоссилий связаны с эпигенетическими процессами.

Однако приведенный состав спор растений и микрофитопланктона, отвечающий аналогичным ассоциациям Восточно-Европейской платформы, Южно Урала и Приуралья, ФРГ, Англии, Канады и других территорий, позволяет уверенно считать метаморфизованные породы отдельных участков Странджи девонскими (нижнедевонскими).

Таким образом, комплексное изучение микрофлористических, литологических и тектонических данных подтверждает концепцию палеозойского возраста (Начев, 1972, 1976; Маляков, 1976) метаморфитов Странджи.

Полученные результаты свидетельствуют о плодотворности такого рода совместных исследований и способствуют положительному решению общих проблем строения, развития и корреляции метаморфических образований Карпато-Балканской складчатой зоны, а особенно Странджи.

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СТРАТИГРАФИЯ И СТРУКТУРНОЕ ПОЛОЖЕНИЕ СУХОВСКОЙ ЗОНЫ УКРАИНСКИХ КАРПАТ [1]

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В 60-х годах в юго-восточной части южного склона Украинских Карпат были получены интересные новые данные, значительно пополнившие существовавшие представления о стратиграфии мелового флиша, а также позволившие по-новому подойти к решению некоторых вопросов структурно-фациального районирования. Так, к северу от Мармарошской зоны, в полосе развития ,,раховского и буркутского флиша'', известной в то время под названием Раховской зоны, удалось выделить (Дабагян, Кульчицкий, Лозыняк, 1967) три самостоятельных структурно-фациальных единицы — Раховскую (*sensu stricto*), Суховскую и Буркутскую, причем для каждой из них были установлены характерные типы разреза мела. В частности, Сиховской зоне[3] свойственно широкое площадное развитие серого ,,курбикортикального'' флиша белотисенской свиты, перекрывающегося кверху своеобразной и палеонтологически хорошо охарактеризованной толщей отложений суховской свиты. Эта структурно-фациальная зона является естественным продолжением ,,нижнего внутреннего покрова'' Румынских Карпат. Отложения белотисенской свиты (неоком-альб) смяты в мелкие складки и представлены тонкоритмичным алевролитово-аргиллитовым серым флишем, в разрезе которого встречаются вкладки (мощность до 100 м и больше) грубослоистых песчаников и реже мелкогалечных конгломератов и гравелитов (богданские и броньковские конгломераты). Белотисенские слои (можность 1000 м) хорошо охарактеризованы палеонтологически (Дабагян, Кульчицкий, Лозыняк, 1967) и в возрастном отношении соответствуют баррему-альбу. Обнаруженный П. Ю. Лозыняком в окрестностях хутора Пригодь (бассейн Лужанки) в нижней части разреза аммонит близкий к *Polyptychites stubendorfi* (Schmidt) дает некоторое основание предполагать, что низы свиты опускаются в готерив и возможно даже в валанжин. Если это так, то нижняя часть белотисенской свиты является фациальным аналогом вовчинской и воз-

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можно раховской свит, развитых в Раховской структурно-фациальной зоне.

Кверху белотисенская свита постепенно переходит в отложения суховской свиты, нижняя часть которой (нижнесуховская подсвита) представлена почти черными известковистыми песчаными аргиллитами и мергелями с прослоями песчаников и редко конгломератобрекчий.

Приоритет выделения суховской свиты принадлежит Я. О. Кульчицкому и П. Ю. Лозыняку, которые уже в первых публикациях (Дабагян, Кульчицкий, Лозыняк, 1967; Кульчицкий, Лозыняк, Дабагян, Мархель, 1965; Кульчицкий, Лозыняк, Пастернак, 1966) установили ее объем и возраст. Согласно первоначальному определению, в разрезе суховской свиты (название дано от с. Суха, расположенного у слияния рек Красной и Бороньки) выделяются две основные части: нижняя, сложенная темно-серыми, до черных, мергелями и аргиллитами с фауной головоногих моллюсков, и верхняя, представленная пестроцветными (красными, зеленоватыми, зеленовато-серыми) и серыми песчано-глинистыми породами. Свита занимает промежуточное положение между тонкоритмичным флишем белотисенской и грубослоистыми песчаниками терешовской свит. Обнаруженная в суховских слоях фауна моллюсков и фораминифер позволила отнести вмещающие породы к верхам альба-турону. В качестве эталонных рекомендованы разрезы в с. Суха и по руч. Поркулецу в бассейне р. Терешовы.

При последующих исследованиях, проводимых геологами производственных и научных организаций, выходы суховской свиты были закакартированы в ряде других районов (по р. Белый Черемош в бассейнах рек Тисы, Теребли, Рики и Боржавы). Вместе с тем в последние годы появились отдельные работы (Беэр, Бызова, Маслакова, 1968; Круглов, Смирнов, 1979), в которых без всякого для этого основания пересматриваются объемы свиты и даже, вопреки существующим правилам, предпринимаются попытки введения новых наименований. Так С. С. Круглов и С. Е. Смирнов (1979) предлагают выделить пестроцветную часть разреза в самостоятельную ,,поркулецкую свиту'', причем рекомендуют распространить этот термин на пестроцветные части разрезов яловецкой, лолинской и других свит более северных тектонических зон, занимающих несколько иное стратиграфическое положение. Это вызывает необходимость приведения дополнительного более подробного описания стратотипа (окрестности с. Суха) и лектостратотипа (руч. Поркулец) суховской свиты с попутным рассмотрением новых данных о фауне из самой верхней части свиты, ранее палеонтологически недостаточно охарактеризованной.

Стратотип с. Суха. Низы суховской свиты и ее контакт с подстилающей белотисенской свитой хорошо обнажаются на южной окраине села по руч. Плыняку, правому притоку р. Броньки. Вверх по течению ручья от точки его пересечения с дорогой, соединяющей села Бронька и Суха, на протяжении 100 м прослеживаются выходы тонкопереслаивающихся темно-серых аргиллитов и скорлуповатых алевролитов белотисенской свиты. В конце данного интервала обнаружен характерный для верхнего альба *Inoceramus tenuis* Mant. (определение С. П. Коцюбинского). Выходы таких же пород наблюдаются по р. Красной в с. Суха, откуда известна находка обломка верхнеальбского *Neohibolites* cf. *minimus* (Lister) (II).

Выше точки с иноцерамом до развилки руч. Плыняка среди характерных для белотисенской свиты тонкоритмичных серых алевритово-аргиллитовых пород появляются темно-серые, до черных, мергели и глинистые известняки, а также единичные прослои аргиллитов с зеленоватым оттенком. Эти породы являются переходной пачкой (мощность около 20 м) между белотисенской и суховской свитами и на общем фоне доминирующего юго-западного падения образуют мелкие складки. Данная пачка начинает разрез суховской свиты.

Выше развилки как по левому, так и по правому разветвлениям, слагая небольшую синклиналь, обнажаются более высокие звенья разреза суховской свиты, представленные снизу вверх:

а) 18—20 метровой пачкой сильно перемятых темно-серых, зеленовато-серых и единичных красных мергелистых пород;

б) 20 метровой пачкой темно-серых, до черных мергелей и толстоплитчатых аргиллитов с редкими прослоями мелкозернистых песчаников; вверху наблюдаются единичные прослои зеленоватых аргиллитов и песчаников с налетами окислов марганца;

в) 30—40 метровой пачкой сизо-зеленоватых мергелей и аргиллитов с включениями линзовидных прослоев и булообразных стяжений серых плотных известняков и алевролитов. Подробное изучение фораминифер из этой части разреза суховской свиты и переходных слоев показало, что они содержат враконские планктонные и бентосные фораминиферы. В частности в пачках, перечисленных кроме планктонных [*Globigerinelloides bentonensis* (Mor.), *Hedbergella delrioensis* (Carsey), *H.* (*Asterohedbergella*) *asterospinosa* Hamaou, *Thalmanninella praebalernensis* (Sigal), *Th. balernensis* (Gand.)], выявлены и агглютинирующие виды: *Hormosina crassa* Geroch, *Glomospirella gaultina* (Berth.), *Ammodiscus tenuissimus* (Gümb.), *Plectorecurvoides alternans* Noth., *Haplophragmoides gigas minor* Nause.

Красные аргиллиты, как правило, содержат агглютинирующие *Hormosina crassa* Geroch, *Glomospirella gaultina* (Berth.), *Ammodiscus tenuissimus* (Gümb.), *Bathysiphon tautinensis* Sacco, *Spiroplectammina laevis* (Roemer) var. *cretacea* Cushm., *Tritaxia gaultina* (Moroz.), *Plectorecurvoides alternans* Noth, *Haplophragmoides gigas minor* Nauss, *Thalmannammina neocomiensis* Geroch., бентосные *Parella cretacea* Carb., *Gyroidina infracretacea* Moroz. и в массовых количествах такие планктонные формы, как *Planogyrina globigerinellinoides* (Subb) и *Globigerinelloides ultramicrus* (Subb.).

Следует отметить отсутствие характерных видов для сеномана, что ставит под сомнение достоверность приводимых в более ранних работах (Дабагян, Смирнов, 1963) определений сеноманских тальманнинел и роталипор, якобы найденных в переходных слоях. В 1963 г. Н. В. Дабагян за *Rotalipora appenninica* ошибочно были приняты *Thalmanninella balernensis* (Gand.) и *T. praebalernensis* Sigal.

Верхнюю часть разреза суховской свиты можно проследить по р. Кушнице в с. Кушница, где наблюдается следующий восходящий разрез:

г) 30 метровая пачка красных и зеленых мергелей и аргиллитов с единичными прослоями песчаников:

д) 30—40 метровая пачка темно-серых и серых с зеленоватым или голубоватым оттенком карбонатных и некарбонатных мягких аргиллитов и глин (мощность прослоев 0,5—1,5 м), вмещающих редкие прослои (5—15 см) темно-серых и серых глинистых алевролитов и известняков. К этой части разреза иногда (р. Боржава, села Бронька, Липецкая Поляна) бывают приурочены небольшой мощности эндоолистостромовые горизонты. Содержащаяся в породах фауна фораминифер *Uvigerinammina jankoi* Majcn., *Plectina taylleuri* (Tapp.), *Haplophragmoides herbichi* Neagu, *Praeglobotruncana imbricata* (Mornod), *Globotruncana lapparenti* (Brotz.), *G. sigali* Reich. указывает на их принадлежность к турону;

е) 20 метровая почка голубовато-серых плотных мергелей, красных аргиллитов, алевролитов и глин с характерным для турона комплексом *Trochammina globigeriniformis* (P. et J.), *Dorothia filiformis* (Bert.), *Plectina grzybowski* Neagu.

ж) Самая верхняя часть суховской свиты представлена 40—50 метровой пачкой тонкоритмично переслаивающихся серых и темносерых аргиллитов, алевролитов, редко песчаников и мергелей. Выше согласно залегают толстослоистые песчаники терешовской свиты, выходы которых можно наблюдать в истоках правого разветвления руч. Плыняка и на водоразделе рек Кушницы и Красной.

Заканчивая описание стратотипа в районе с. Суха, уместно подчеркнуть, что общая мощность суховской свиты достигает 210—230 м. Учитывая данные по фауне фораминифер, ее возраст обоснованно можно отнести к вракону-турону. Отсутствие палеонтологических остатков в верхах свиты, позволяет предполагать их возможную принадлежность к низам сенона.

Лектостратотип по руч. Поркулецу. Поркулец — левый приток реки Терешовы (его устье находится в 7 км севернее с. Тарасовки). Изучение разреза суховской свиты по руч. Поркулецу и его притокам, а также по р. Терешове позволяет выделить следующие горизонты:

а) непосредственно на белотисенской свите залегает пачка (около 60 м) черных и темно-серых карбонатных песчанистых аргиллитов с прослоями мергелей, глинистых песчаников, а также характерными для этой части разреза линзами и крупными булообразными стяжениями серых и темно-серых плотных алевритистых известняков. В породах наблюдаются явления подводного оползания осадков, сильная дислоцированность и густая сеть прожилков (1—2 см) кальцита. По всему разрезу отмечено массовое содержание вараконских планктонных фораминифер *Planogyrina gaultina* (Moroz.), *P. globigerinellinoides* (Subb.), *Globigerinelloides bentonensis* (Morrow.), *Hedbergella deerioensis* (Carsey), *H. (Asterohedbergella) asterospinosa* Hamaoui, *Clavihedbergella simplex* (Morrow.), *Thalmanninella ticinensis* (Gand.) Примерно в средней части пачки по р. Терешове в 600 м выше устья руч. Поркулеца, в черных аргиллитах обнаружены головоногие моллюски [*Puzosia planulata* (Sow.), *Sciponoceras baculoides* (Mantel), *Parahibolites tourtiae* (Weigner) и пелециподы *Aucellina gryphaeoides*

Sow.)], позволяющие отнести вмещающие породы к низам сеномана. Однако этот выводне подтверждают фораминиферы, тяготеющие еще к вракону.

б) 30 метровая пачка тонкоритмичных переслаивающихся темно-серых, нередко песчанистых аргиллитов, алевролитов и мергелей с единичными прослоями красных аргиллитов в нижней части. В песчанистых аргиллитах обнаружено несколько экземпляров сеноманских *Puzosia* cf. *subplanulata* (Schültzer) (определение Р. Й. Лещух). К верхам пачки приурочены небольшие микроолистолиты серых толстослоистых песчаников.

в) 6—8 метровая пачка фукоидных песчанистых мергелей зеленовато-серого и вверху слабо розоватого цвета с редкими прослоями алевролитов. Характерная для нижней части ассоциация верхнесеноманских фораминифер *Hedbergella delrioensis* (Carsey), *H. praehelvetica* (Trujillo), *H. portsdownensis* (Will.-Mitch.), *H. porculecensis* Dab., *Thalmanninella deeckei* (Fran.), *Rotalipora cushmani* (Morrow.) кверху сменяется нижнетуронским комплексом *Rotalipora cushmani* (Morrow.), *Praeglobotruncana stephani* Gand.), *P. delrionensis* (Plumm.), *P. oraviensis* Scheibn. *Helvetoglobotruncana helvetica* (Bolli), *Globotruncana sigali* Reich.

По левым притокам руч. Поркулеца вскрывается верхняя часть разреза, в составе которой выделяются:

г) 50 метровая пачка толстоплитчатых зеленовато-серых мергелей с прослоями темно-серых, до черных, песчанистых аргиллитов в нижней части и красными разностями аргиллитов и мягкими глинами, мергелями и тонкослоистыми песчаниками в верхней. В породах встречаются крупные (диаметром до 10 см) конкреции марказита и кристаллы пирита.

д) 10—15 метровая пачка зеленовато-серых грубослоистых мергелей с характерными для коньяка *Globotruncana conica* White, *G. renzi* Gand., *G. angusticarinata* Gand.;

е) мелкогалечные конгломераты и гравелиты, переходящие вверх по разрезу и по простиранию в грубозернистые песчаники. Кластический материал хорошо окатан и состоит из обломков гранитов, амфиболитов, биотитовых гнейсов, полосчатых доломитов и известняков, напоминающих породы Мармарошского кристаллического массива. В конгломератах местами наблюдаются включения неокатанных глыб размером до 0,5 м. Общая мощность грубообломочных пород 10 м;

ж) 30—40 метровая пачка тонкоритмично чередующихся темно-серых и серых иногда зеленоватых песчанистых аргиллитов, алевролитов, редко песчаников. В аргиллитах кровельной части пачки выявлены сенонские агглютинирующие *Proteonina complanata* (Fr.), *Trochammina irregularis* White, *Ammodiscus angusta* Fried.

Общая мощность суховской свиты в бассейне руч. Поркулеца не превышает 210 м. Занимаемое ею стратиграфическое положение соответствует вракону-сантону.

Как следует из приведенных описаний стратотипа и лектостратотипа, общий литологический состав слагающих свиту пород меняется незначительно. Почти повсеместно можно выделить три части: а) нижнюю, в основном представленную черными и темно-серыми глинистыми породами с фауной моллюсков; б) среднюю — пестроцветную, состоящую из пород окрашенных в красные, зеленоватые и темноцветные цвета; в) верхнюю, сложенную серыми алевролитами аргиллитами. Перечисленные горизонты не являются стратиграфически выдержанными и могут менять свои объемы в результате литологофациальных замещений. Это еще раз доказывает неправомерность выделения С. С. Кругловым и С. Е. Смирновым (1979) на южном склоне Ураинских Карпат ,,поркулецкой свиты'', включающей пестроцветные породы, так как в разных структурно-фациальных зонах она будет иметь неодинаковые стратиграфические объемы. Отложения суховской свиты кверху переходят в толщу (500 м) грубослоистых серых песчаников терешовской свиты (сенон-низы палеоцена (?), в верхней части разреза которой появляются отдельные пачки мелкогалечных конгломератов, серых мергелей, а также известен маломощный (2 м) горизонт пестроцветных (красных и зеленовато-серых) аргиллитов. Обломочный материал конгломератов (общая мощность грубообломочных пород в бассейне Терешовы достигает 20—25 м) сложен галькой гранитоидов пегматитовой структуры, амфиболитов, диабазов, очковых гнейсов, кварц-серицитовых и хлоритовых сланцев. Интересно, что сходный по составу, обломочный материал наблюдается также в конгломератобрекчиях суховской свиты, в которых кроме упомянутых разностей встречается еще галька и валуны биотитовых гнейсов, полосчатых доломитов, пелитоморфных известняков, конгломератов и кварцитов.

Из средней части терешовской свиты известны находки *Hormosina gigantea* Geroch; в верхней части разреза установлено присутствие кампанских глоботрункан [*G. arca* (Cushm.), *G. stuartiformis* (Dalb.) и др.). Встреченная фауна и стратиграфическое положение терешовской свиты дают основание отнести ее к сенону-низам палеоцена (?). В последнее время некоторые исследователи (Круглов С. С., 1972) допускают возможную принадлежность верхов терешовской свиты к эоцену. Подобное предположение необосновано палеонтологически и кажется мало вероятным.

Суховская структурно-фациальная зона — крупный скальпированный покров, амплитуда горизонтального перемещения которого превышает 30 км. Внутри покрова закартировано несколько косо расположенных чешуй, сложенных в основном отложениями белотисенской свиты (баррем-альб), а также породами суховской (вракон-сантон) и терешовской (кампан-низы палеоцена) свит. До последнего времени многие геологи полагали, что в основании белотисенской свиты залегает раховская свита (неоком), известная на смежной территории Румынии под названием слоев Синая. В действительности в пределах Суховского покрова выходов раховской свиты нет (в прошлом к ним иногда ошибочно причислялась литологически сходная враконская часть разреза суховской свиты). Неудивительно, что сейчас предположение о наличии и раховских слоев в основании белотисенской свиты вызывает сомнение. Как известно, отложения раховской свиты являются характерными для Раховской структурно-фациальной зоны (*sensu stricto*), но в пределах последней они перекрываются не белотисенскими слоями, а породами вовчинской свиты (баррем-апт), представленной серым песчаным флишем (мощность 280—350 м) с прослоями и пачками гравелитов и мелкогалечных конгломератов. В качестве стратотипа этой свиты рекомендован разрез по руч. Вовчому, левому притоку р. Белой Тисы (район г. Рахова). Таким образом, существенные отличия между меловыми образованиями Раховской и Суховской единиц подтверждают правомерность присвоения им ранга самостоятельных структурно-фациальных зон.

Особого внимания заслуживает вопрос о структурной принадлежности Буркутской единицы, которая представляет собой крупную надвинутую скибу, сложенную буркутскими песчанистыми породами (верхний мел), залегающими на шипотских слоях (баррем-альб) и фациально замещающими пестроцветные отложения яловецкой свиты (сеноман-турон). Некоторые геологи ошибочно причисляют ее к Суховскому покрову и даже иногда на всю Суховскую зону распространяют название ,,Буркутская зона'' (Бызова, Беэр, 1974). В действительности Буркутская единица (ее можно выделить в самостоятельную подзону) является составной частью более северной структурно-фациальной зоны — Петросской (Габинет, Кульчицкий, Матковский, 1976) или Климовской (Кульчицкий, Лозыняк, 1977), особенностью которой является развитие нижнемеловых отложений в фации шипотских слоев.

[3] В 1971 г. С. С. Круглов без всякого для этого основания переименовал ее в ,,Поркулецкую зону''. Термин ,,Поркулецкая зона'' является излишним и, учитывая право приоритета, им не следует пользоваться.

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ВОПРОС

Ион Жана: Существуют ли в последовательности пластов, напластования уровней типа черных сланцев валанжин-барремского возраста?

Ответ: Не существуют такие напластования; в последовательности отложения зон Сухово существуют отложения типа черных сланцев с микрофауной враконского возраста (?!?)

К КОРРЕЛЯЦИИ ПАЛЕОГЕНОВОГО ФЛИША УКРАИНСКИХ И РУМЫНСКИХ КАРПАТ[1]

О. С. ВЯЛОВ [2], Н. В. МАСЛУН [2], Я. В. СОВЧИК [2]

Одной из наиболее важных задач стратиграфии Карпатского флиша является корреляция различных типов палеогеновых отложений. В настоящей статье делается попытка сопоставления основных палеогеновых разрезов северного склона Украинских Карпат и соответствующих им в Румынии разрезов Маргинального покрова (Вранча) и Таркэу. Более внутренние покровы в пограничных районах сложены породами мела, а отложения палеогена имеют локальное развитие и не прослеживаются на смежных территориях, в связи с чем в данной работе не рассматриваются.

Анализ комплексов мелких планктонных и бентосных фораминифер, а также нуммулитов, содержащихся в палеогеновом флише Украинских и Румынских Карпат показал их большое сходство и позволил произвести некоторые сопоставления и развить существовавшие в литературе представления о возрасте и корреляции разрезов.

В Украинских Карпатах к палеоцену относятся верхнестрыйская подсвита, яремчанские слои и песчаники ямненскои свиты (Вялов, 1961; Маслун, 1976; Мятлюк, 1970). Отложения верхнестрыйской подсвиты, представленные тонкоритмичным флишем, содержат, главным образом, агглютинирующие фораминиферы: *Carpathiella ovulum* var. *gigantea* (Grzyb.), *Rzehakina fissistomata* Grzyb., *Asanospira grzybowskii* (Mjatl.), *Trochamminoides ammonoides* Grzyb., *Glomospira serpens* Grzyb., *Dendrophrya maxima* (Fried.) и редкие *Globoconusa daubjergensis* Bronn., *Globigerina trivialis* Subb., *Subbotina triloculinoides* (Plum.), *Planorotalia pseudobulloides* (Plum.), *Subbotina varianta* (Subb.) и др. (Влялов, 1961; Грузман, Дабагян, Круглов идр., 1966; Маслун, 1976; Мятлюк, 1970).

В залегающих выше по разрезу яремчанских слоях содержится комплекс фораминифер значительно более богатый по видовому составу, чем во флише верхнестрыйской подсвиты. Здесь, кроме появившихся и

[1] Доклад была представлена на XII-ом Конгрессе Карпато-Балканской Геологической Ассоциации, 8—13 сентября 1981 г., Бухарест, Румыния.

[2] СССР.

преобладающих в комплексе особей *Carpathiella ovulum ovulum* (Grzyb.), *Hyperammina exilis* Mjatl., *H. intermedia* Mjatl. встречаются секреционные бентосные и планктонные фораминиферы: *Anomalina danica* (Brotz.), *Cibicides variantus* Dab., *C. proprius* Brotz., *Subbotina triloculinoides* (Plum.), *Globigerina nana* Chal.

Несмотря на то, что комплексы агглютинирующих фораминифер яремчанских слоев во многом сходны с комплексами из нижележащих отложений, между ними отмечаются определенные отличия. Различить эти комплексы можно по преобладанию криптокристаллических кремнистых раковин родов *Asanospira*, *Grzybowskiella*, *Carpathiella*. Исчезают характерные для верхнестрыйской подсвиты *Carpathiella ovulum gigantea* (Geroch.) и частыми становятся *Carpathiella ovulum ovulum*. *Dendrophrya maxima* (Fried.), *Hyperammina primitiva* Mjatl. сменяются более тонкостенными, меньшими по размерам *H. jamnensis* Maslun, часты в разрезе *Rzehakina fissistomata* Grzyb. Из крупных фораминифер в яремчанских слоях обнаружены *Nummulites deserti* и *Discocyclina* sp.

Для залегающих на яремчанских слоях ямненских песчаников характерна частая встречаемость *Hyperammina cylindrica cylindrica* (Glaessn.), *Cystamminella pseudopauciloculata* Mjatl., *Recurvoides varius* Mjatl., *Grzybowskiella angusta* (Friedb.), *Rzehakina inclusa* Grzyb. и др. Отличительной чертой комплекса фораминифер ямненской свиты является развитие большого числа мелких форм агглютинирующих фораминифер, появление видов, которые массовое развитие получили в более позднее время: *Recurvoides varius* Mjatl., *Karreriella horrida* Dab., *Glomospira saturniformis* Majzon и др. Кроме того в ямненской свите присутствуют *Globorotalia angulata* White, *Globigerina nana* Chal., *Globorotalia velascoensis* Cushm.

По видовому составу содержащихся в верхнестрыйской подсвите фораминифер она датируется дат-монсом, а ямненская свита — тенетом.

В Румынских Карпатах аналогичные комплексы фораминифер содержаться в верхней части слоев Путна, Извор. Здесь обильны агглютинирующие фораминиферы: *Carpathiella ovulum ovulum* (Grzyb.), *Hyperammina cylindrica* (Glaess.), *H. grzybowskii* Dylaz: *Dendrophrya maxima* (Friedb.), *Nodellum velascoense* (Cuschm.), *Rzehakina fissistomata* Grzyb., *Glomospira serpens* Grzyb. и др.

Планктонные фораминиферы более многочисленны и разнообразны в Румынских Карпатах, что позволило Брату (1975) выделить по этой группе зоны: Subbotina triloculinoides [S. pseudobulloides (даний), Acarinina inconstans, Globorotalia angulata (монс-тенет), Acarinina acarinata, Globorotalia aequa (G. marginodentata) илердский ярус]. В одновозрастных отложениях Украинских Карпат выделение аналогичных зон невозможно из-за спорадической встречаемости планктонных фораминифер. Условно верхнестрыйская подсвита и ямненская свита сопоставляются с зонами Subbotina triloculinoides, Globorotalia angulata.

Нижний эоцен на большей части северного склона Украинских Карпат представлен серо-зеленым, тонкоритмичным флищем манявской свиты, которому на территории Румынских Карпат по возрасту соответствуют нижние горизонты песчаников Таркэу и их базальный горизонт в Маргинальной зоне, слои Стража, нижняя часть слоев Тазлэу в покрове

Тазлэу и слои Сучевицы в фации Лешунт-Путна, покров Таркэу. Указанные стратиграфические подразделения содержат около 50 видов агглютинирующих фораминифер, большая часть которых имеет узкий стратиграфический диапазон (Bratu, Alexandrescu, 1970; Tocorjescu, 1960).

Изучение вертикального распространения агглютинирующих фораминифер в манявской свите позволяет выделить ряд слоев. В основании свиты (надъямненский пестроцветный горизонт по О. С. Вялову) выделяются слои с *Hyperammina cylindrica crassa* Mjatl., *Dendrophrya manjavica* Maslun. Затем следуют слои с *Glomospira*, *Radiolaria* и спикулами губок. В кровле свиты выделяется пачка с пестроцветными аргиллитами, где доминируют *Karreriella horrida* Dab., *Cyclammina intermedia* Mjatl., *Dendrophrya robusta* (Grzyb.) (Маслун, 1976).

В монофациальных отложениях нижнего эоцена комплексы агглютинирующих фораминифер аналогичные манявским развиты во всей Скибовой зоне Украинских и Польских Карпат, в покровах Таркэу и Маргинальном Румынских Карпат. В разнофациальных отложениях отмечаются общие элементы, по которым возможно выделение слоев с микрофауной определенного типа, имеющих региональное значение — это слои с *Glomospira*, *Radiolaria* и с *Karreriella*, *Dendrophrya* и др.

В слоях Стража, в пестроцветных породах нижней части песчаников Таркэу и в базальном горизонте румынскими исследователями (Bratu, 1975; Bratu, Alexandrescu, 1970; Bratu, Gheţa, 1972) указывается обилие тех же видов *Glomospira*, *Hyperammina*, *Dendrophrya*, *Karreriella*, *Radiolaria* и спикул губок.

По планктонным фораминиферам нижнеэоценовые отложения в Румынских Карпатах выделяются в объеме зон Globorotalia marginodentata и G. aragonensis. Аналогами этих зон в Украинских Карпатах являются слои с *Globorotalia aequa*, *G. marginodentata*, охватывающие манявскую свиту.

Нижнеэоценовый возраст коррелируемых толщ подтверждается крупными фораминиферами идентичного видового состава: *Nummulites planulatus*, *N. burdigalensis*, *N. partschi* и др. Нуммулиты не обнаружены только в слоях Стража, но именно слои Стража содержат наибольшее количество агглютинирующих фораминифер сходного видового состава с комплексами из манявской свиты.

Наиболее характерным литофациальным типом среднеэоценовых отложений пограничных районов Украинских Карпат являются песчаники выгодской свиты, которые соответствуют верхним песчаникам Таркэу, верхней части слоев Тазлэу и Кольци (Ionesi, 1966). О среднеэоценовом возрасте коррелируемых отложений свидетельствует богатый комплекс нуммулитид: *N. gallensis*, *N. distans*, *N. murchisoni*, *N. globulus*, *N. laevigatus* и др. О сопоставимости указанных комплексов говорит также общность видового состава бентосных секреционных фораминифер родов *Cibicides* агглютинирующих фораминифер родов *Dendrophrya*, *Grzybowskiella*, *Cyclammina*, *Trochamminoides*, *Cystamminnella* и планктонных видов *Globigerapsis subconglobatus*, *Subbotina frontosa*, *Acarinina bullbrooki*. Как в Румынских, так и в Украинских Карпатах в среднем эоцене выделяются слои с *Cibicides*, *Nummulites*. В тех разрезах, когда в выгодских песчаниках появляются многочисленные некарбонатные глинистые прослои в них доминируют агглютинирующие фораминиферы: *Dendrophrya*, *Cystamminella*, *Psammosphaera*.

По планктонным фораминиферам эти отложения соответствуют зонам Acarinina bullbrcoki, Hantkenina liebusi (Румынские Карпаты) и A. bullbrooki—Hantkenina alabamensis (Украинские Карпаты).

Что касается пасечнянских известняков и мергелей буковинских слоев, то на территории Румынии они имеют литофациальные аналоги, что позволяет проводить прямую корреляцию: пасечнянской свите отвечают известняки Доамна, а буковинским слоям — слои Вицеу. Те и другие содержат типичные среднеэоценовые *N. laevigatus*, *N. gallensis*, *N. perforatus*, *N. atacicus* и др. (Tătărâm, 1968).

В пограничной южной части Скибового покрова развит груборитмичный песчаный флиш, который находится на продолжении фации типа разреза Таркэу, столь распространенного в покрове Таркэу Румынии, что также не вызывает сомнений при корреляции. В кернах скважины Плоскоя — I в верхней части разреза песчаников были обнаружены *N. millecaput*, что позволило датировать вмещающие породы верхами среднего эоцена. В аналогичных условиях комплекс нуммулитов с *Nummulites millecaput* был найден и в покрове Таркэу.

Верхняя часть эоценовых отложений на северном склоне Украинских Карпат представлена разнофациальными породами, которые выделены в быстрицкую, попельскую и довжинскую свиты.

Довжинская свита развита в крайних южных скибах, представлена толстослоистыми массивными слюдистыми песчаниками с маломещными прослоями алевролитов и аргиллитов. В верхней части разреза этой свиты обнаружены типичные верхнеэоценовые нуммулиты: *Nummulites fabiani*, *N. chavannesi*, *N. retiatus*.

Попельская свита имеет ограниченное распространение в северных скибах и повсеместно развита во внутренней зоне Предкарпатского прогиба. Она замещает в большей или меньшей степени нижнюю часть быстрицкой свиты. Представлена попельская свита известковистыми породами — аргиллитами и даже мергелями, всегда с примесью алевритового материала. Она содержит известковистые бентосные фораминиферы: *Asterigerina rogalai* Mjatl., *Cibicides popeliensis* Mjatl., *C. tallahattensis* Mjatl., *C. coniferus* Mjatl., *Uvigerina jacksonensis* Cusch., *Nonion halkyardi* Cuschm., *Grammostommum nobilis* (Hantken), *G. elongatum* (Hantken), *Bulimina sculptilis* Cuschm., *Planulina costata* (Hantken), *Globigerapsis index* (Finlay), *Turborotalia centralis* (Cushm. et Berm.), *Subbotina eocaenica* (Terq.) и др.

Быстрицкая свита — основная фация этого стратиграфического уровня. Она сложена некарбонатными зеленовато-серыми аргиллитами с прослоями алевролитов и песчаников. Повсеместно в основании свиты прослеживается пестроцветный горизонт. Комплекс фораминифер состоит уз агглютинирующих видов: *Silicobathysiphon pseudoloculus* Mjatl., *Hyperammina lineariformis* Mjatl., *H. karpatica* Maslac., *Reophax planus* Halk., *Ammodiscus latus* Grzyb., *Grzybowskiella subangusta* Mjatl., *Gr. macilenta* Mjatl., *Asanospira walteri* Grzyb., *Cyclammina amplectens* Grzyb., *C. rotundidorsata* (Hankten), *Tro-*

chamminoides subtrullisatus (Rz. et G.), *Spiroplectammina spectabilis* Grzyb. Наиболее характерными видами являются: *Cyclammina amplectens*, *C. rotundidorsata*, *Ammodiscus latus*, *Asanospira walteri*.

В Румынии этим отложениям соответствуют в элементе Таркэу слои Поду-Секу, в элементе Тазлэу — слои Плопу и в Маргинальном покрове — слои Биссерикани. Во всех перечисленных слоях содержится богатый комплекс агглютинирующих фораминифер: *Dendrophrya robusta* Grzyb., *Cyclammina amplectens* Grzyb., *Ammodiscus latus* (Grzyb.), *Haplophragmoides* (= *Asanospira*) *walteri* (Grzyb.), *Reophax planus* Halkyard, *Saccammina placenta* (Reuss).

Удельный вес планктонных фораминифер в комплексе верхнеэоценовых отложений Украинских Карпат сравнительно низок. Условно можно выделить только слои с *Globigeropsis index*, которые соответствуют верхней части быстрицкой свиты.

В Румынских Карпатах по массовому развитию планктонных фораминифер выделены зоны Globigerapsis tropicalis и G. index. Несмотря на некоторые различия видового состава агглютинирующих и плактонных фораминифер мы можем говорить о прямой корреляции слоев Поду-Секу, Плопу, Биссерикани и с быстрицкой и попельской свитами. Верхняя часть указанных литостратиграфических подразделений всеми исследователями датируется поздним эоценом. Возраст же нижней части понимается не однозначно.

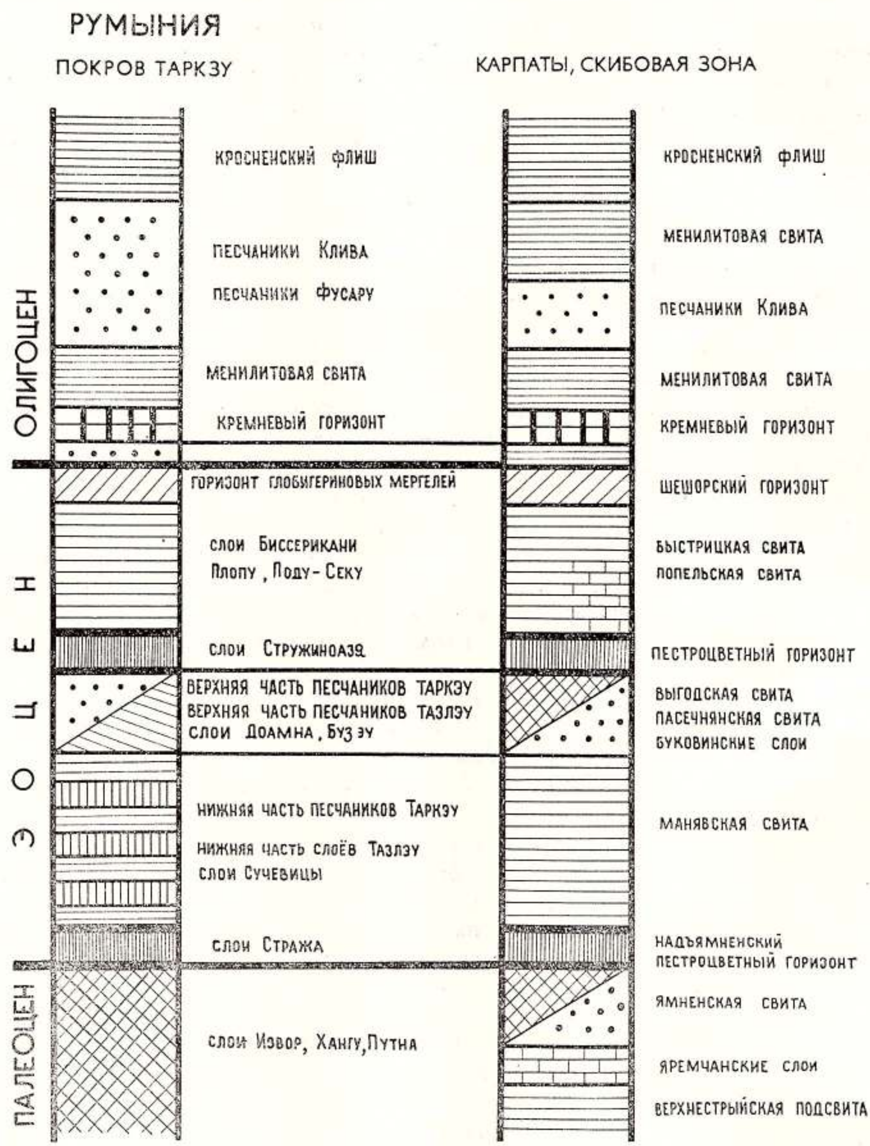
Нами принимается позднеэоценовый возраст быстрицкой, свиты. Однако в низах ее фациального аналога — попельской свите в разрезе на р. Опур указываются (Мятлюк, 1970) находки *Acarinina bullbrooki*, *A. rotundimarginata*, что делает возможным предположение о том, что самая нижняя часть свиты может еще озносится к среднему эоцену. В прочем, с другой стороны, в богатом комплексе фораминифер в пестроцветном горизонте в основании быстрицкой свиты наравне с другими имеются и те виды агглютинирующих фораминифер, которые численно преобладают в верхней части быстрицкой свиты: *Cyclammina rotundidorsata* (Hantken), *Ammodiscus latus* Grzyb., *Asanospira wal-* Grzyb.), *Hyperammina leariformis* и др.

Во всяком случае большая верхняя часть — несомненный верхний эоцен. Это доказывается наличием в попельской свите таких характерных форм как *Globigerapsis index*, *Subbotina eocaena*, *Turborotalia centralis* и др., а в довжинской фации — типичный верхнеэоценовых нуммулитов — *Nummulites fabiani*. Списки моллюсков верхеэоценового возраста из попельской свиты приводит В. Рогаля. Все же вопрос о возрасте пестроцветного горизонта остается для нас не ясным.

Румынские геологи неоднозначно датируют нижнюю часть разрезов слоев Биссерикани, Стружиноаза, Поду-Секу и др. В одной из последних сводок (Săndulescu, Ştefănescu, 1981) слои Биссерикани отнесены к верхнему эоцену, а пестроцветный горизонт в их основании — к верхам среднего эоцена. Отнесены к верхам среднего эоцена и нижняя часть слоев Плопу и Поду-Секу, где пестроцветы отсутствуют. В то же время в другой сводке (II) слои Биссерикани, Плопу и Поду-Секу — это только приабон. Е. Брату указывает на массовое развитие *Cyclammina amplectens* в пестроцветных отложениях и датирует их средним эоценом.

Несколько слов следует сказать о вертикальном распространении *Cyclammina amplectens* (*s.l.*), характерной для быстрицкой фации и цитируемой из Польских и Румынских Карпат.

В Украинских Карпатах виды рода *Cyclammina* в эоценовом флише распределены по всему разрезу. Первые представители этой группы поя-



вляются в нижнем эоцене (манявская свита) — это *Cyclammina amplectens* var. *ornamenta* Dabagjan in lit. и *I. intermedia* Mjatl. Последний вид встречается и в выгодской свите. Вид *Cyclammina ample-*

РАСПРЕДЕЛЕНИЕ ФОРАМИНИФЕР В ПАЛЕОГЕНОВЫХ ОТЛОЖЕНИЯХ СЕВЕРНОГО СКЛОНА УКРАИНСКИХ КАРПАТ

Возраст			Украина	Румыния
ПАЛЕОГЕН	Эоцен	Верхний	**Шешорский горизонт:** *Subbotina eocaenica eocaenica* (Terq.), *S. eocaenica irregularis* (Subb.), *Catapsydrax dissimilis* (Cush. et Berm.), *Turborotalia tripartita* (Koch.), *Globigerapsis index* (Finlay), *Subbotina corpulenta* Subb.	**Горизонт глобигериновых мергелей:** *Subbotina corpulenta* Subb., *S. eocaenica eocaenica* (Terq.), *S. eocaenica irregularis* (Subb.), *S. tripartita* (Koch.), *S. praebulloides leroyi* (Blow. et Bann.), *Catapsydrax dissimilis* (Cushm. et Berm.), *Globigerapsis index* (Finlay).
			Быстрицкая свита: *Hyperammina neariformis* Mjatl., *Asanospira walteri* Grz., *Cyclammina amplectens* (Grz.), *Reophax duplex* (Grz.), *Bogdanovicziella complanata* (Franke), *Dendrophrya excelsa* Grz., *Nodellum velascoense* (Cushm.), *Saccammina placenta* (Franke), *Reophax planus* Halk., *Grzybowskiella subangusta* Mjatl. **Довжинская свита:** *Nummulites fabianii*, *N. retiatus* Grig., *N. chavannesi*.	**Слои Поду-Секу, Плопу, Биссерикани:** *Cyclammina amplectens* Grz., *Haplophragmoides walteri* (Grz.), *Ammodiscus latus* Grz., *Dendrophrya robusta* Grz., *D. excelsa* Grz., *Hyperammina neariformis* Mjatl., *Grzybowskiella subangusta* Mjatl., *Reophax planus* Halk., *Cystamminella pseudopaucilocula* Mjatl.
		Средний	**Выгодская свита:** *Cibicides westi* Howe, *C. ventratumidus* Mjatl., *C. agrestis* Mjatl., *C. grossoconulus* Mjatl., *Psammosphaera pseudofusca* Masl., *Hyperammina intermedia* Mjatl., *H. neariformis* Mjatl., *Dendrophrya vygodensis* Masl., *Karreriella indigena* Mjatl. *Grzybowskiella subangusta* Mjatl., *Acarinina bullbrooki* (Bolli), *Hankenina alabamensis*. *Nummulites murchisoni*, *N. atacicus*, *N. partschi*, *N. gallensis*, *N. burdigalensis*, *Assilina placentula*, *A. exponensis*. **Буковинские слои:** *Nummulites murchisoni*, *N. globulus*, *N. atacicus*, *N. rotolarius*, *N. gallensis*, *N. partschi*, *N. perforatus*, *N. praelucasi*, *N. laevigatus*, *N. burdigalensis*, *N. inkermanensis*. **Пасечнянская свита:** *Nummulites globulus*, *N. partschi*, *N. atacicus*.	**Песчаники Таркэу, Мойнешти, пасечнянские слои, слои Капин:** *Cibicides westi* Halk., *C. conoideus* (Czjek.), *C. refulgens* Monf., *Cystamminella subgaleata* Vasicek., *Hyperammina neariformis* Mjatl., *Psammosphaera* sp., *Trochamminoides* sp., *Reophax grandis* Grz. *Karreriella indigena* Mjatl., *Acarinina bullbrooki*, *Hantkenina liebusi*. **Слои Вицеу:** *Nummulites partschi*, *N.* ex gr. *perforatus*, *N.* cf. *globulus*, *N.* cf. *irregularis*. **Слои Колци:** *Nummulites distans*, *N. pratti*, *N. anomalus*, *N. millecaput*, *N. gallensis*, *N. atacicus*, *N. striatus*, *Assilina exponensis*. **Пасечнянские слои:** *Nummulites distans*, *N. murchisoni*, *N. globulus*, *N. atacicus*, *N. partschi*. **Слои Тазлэу (верхи):** *Nummulites distans*, *N. millecaput*, *N. pratti*, *N. gallensis*, *N. perforatus*, *N. partschi*, *N. murchisoni*, *N. globulus*, *N. rotularis*, *N.* cf. *variolarius*. **Верхняя часть песчаников Таркэу:** *Nummulites* cf. *distans*, *N. globulus*, *N. gallensis*, *N. millecaput*, *N. striatus*.
		Нижний	**Манявская свита:** *Hyperammina intermedia* Mjatl., *H. karpatica* Masl., *Silicobathysiphon dubia longoloculus* Mjatl., *Reophax splendidus* Grzyb., *Trochamminoides irregularis* White, *Karreriella horrida* Dab., *K. aegra* Finlay, *Glomospira corona* C. et J., *G. saturniformis* Majzon, *G. gordialis* (J. et P.), *Recurvoides smugerensis* Mjatl., *R. varius* Mjatl., *Cyclammina intermedia* Mjatl., *Globorotalia subbotinae*, *G. marginodentata*, *Radiolaria*, спикулы губок. *Nummulites planulatus*, *N. globulus*, *N. partschi*, *N. bolcensis*, *N. irregularis*, *N. praelucasi*.	**Слои Стража, нижняя часть песчаников Таркэу, слои Извор (верхи), слои Сучевицы:** *Hyperammina cylindrica* (Brady), *H. intermedia* Mjatl., *Glomospira charoides* (J. et P.), *G. saturniformis* Majzon, *G. favilla* Emiliani, *Reophax splendidus* Grz., *Cyclammina elegans* Grz., *Karreriella horrida* Dab., *K. indigena* Mjatl., *Saccammina placenta* Grz., *Recurvoides varius* Mjatl., *Radiolaria*, спикулы губок. **Слои Извор (верхи):** *Nummulites planulatus*, *N. globulus*, *N. solitarius*, *N.* ex gr. *distans*. **Нижняя часть песчаников Таркэу:** *Nummulites planulatus*, *N. aquitanicus*, *N. exilis*, *N. vounderschmitti*.
	Палеоцен	Верхний	**Ямненская свита с яремчанскими слоями в основании:** *Carpathiella ovulum ovulum* (Grzyb.), *Saccammina scabrosa* Mjatl., *Hyperammina cylindrica cylindrica* (Glaessn.), *Dendrophrya maxima* (Frieb.) *D. jamnensis* (Maslun), *Grybowskiella glabrata* (C. et J.), *Cystamminella pseudopauciloculata* Mjatl., *Anomalina danica* Brotzen, *Cibicides variantus* Dab., *Globorotalia angulata* White, *Subbotina triloculinoides* Plum., *Nummulites deserti*, *Discocyclina* sp.	**Слои Хангу, Извор:** *Carpathiella ovulum ovulum* (Grz.), *Hyperammina cylindrica* Glaessner, *H. grzybowskii* Dylaz, *Dendrophrya maxima* (Friedb.), *Rzehakina fissistomata* Grz., *Glomospira serpens* Grzyb., *Subbotina triloculinoides* (Plum.), *Globorotalia angulata* (White), *G. conicotruncata* Subb., *G. crassata* Cushm., *Acarinina inconstans* Subb., *A. acarinata* Subb., *Discocyclina seunesi* Subb.
		Нижний	**Верхнестрийская подсвита:** *Silicobathysiphon dubia dubia* (White), *Spiroplectammina clotho* Grzyb., *Rzehakina epigona epigona* (Rzehak), *Carpathiella ovulum* var. *gigantea* (Grzyb.), *Globoconusa daubjergensis* (Bronnim.), *Subbotina varianta* (Subb.), *Planorotalia pseudobulloides* (Plum.).	

ctens Grzyb. в единичных экземплярах отмечается в выгодской свите, но массовое развитие его наблюдается в быстрицкой свите. *Cyclammina placenta* (Reuss) встречается по всему разрезу быстрицкой свиты. Вид *Cyclammina rotundidorsata* (Hantken) приурочен к верхней части быстрицкой свиты. Наибольшего расцвета виды рода *Cyclammina* достигают в позднебыстрицкое время и основной вид этой группы — *Cyclammina amplectens* является у нас весьма характерным для быстрицкой свиты. В литературе по Румынским Карпатам он указывается начиная с сенона (слои Присака) до верхнего эоцена слои Плопу, Поду-Секу, Биссерикани, Струженоаза). Быть может, там вид *Cyclammina amplectens* понимается в очень широком объеме без выделения форм, мало известных в литературе, появившихся на более низком стратиграфическом уровне. Корреляция отложений, из которых он приводится, должна поэтому проводится с учетом всего комплекса мелких и крупных фораминифер.

Под пестроцветными слоями Стружиноаза, в самых верхах литофации Колци и Тазлэу встречен комплекс нуммулитид, характерный для самых верхов среднего эоцена (биаррица) — *Nummulites jonesi* и *N. millecaput* и др. Эти же виды обнаружены в аналогах верхов песчаников Таркэу в скважине Плоское-I в юго-восточной части Скибовой зоны, недалеко от границы с Румынией. Таким образом, по данным нуммулитид, породы залегающие под пестроцветным горизонтом быстрицкой свиты и Стружиноаза относятся к верхам среднего эоцена, пестроцветный горизонт (см. рис.) находится на границе среднего и верхнего эоцена, но возраст его остается неясным.

Повсеместно в Украинских Карпатах и в Румынских Карпатах в кровле верхнего зоцена выделяется мергельный шешорский горизонт или зона крупных глобигерин. Эти отложения характеризуются изменением систематического состава фораминифер по сравнению с нижележащими. В разрезах всех структурно-фациальных зон Карпат агглютинирующие фораминиферы весьма редки, а в комплексе доминируют планктонные: *Catapsydrax dissimilis*, *Subbotina corpulenta*, *S. pseudoeocaena*, *S. yeguensis*, *S. praebulloides leroyi*, *S. tripartita*.

Корреляция олигоценовых отложений из-за недостаточности палеонтологических данных проводится, главным образом, по литофациальным признакам (Совчик, 1976).

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ЗОНАЛЬНАЯ КОРРЕЛЯЦИЯ ПАЛЕОГЕНОВЫХ ОТЛОЖЕНИЙ УКРАИНСКИХ КАРПАТ ПО МЕЛКИМ ФОРАМИНИФЕРАМ [1]

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В основу большей части стратиграфических подразделений палеогеновых отложений Украинских Карпат положены исследования микрофауны. Терригенный флиш содержит преимущественно агглютинирующий бентос, в карбонатно-терригенных образованиях встречается так же и секреционный бентос. К карбонатным породам приурочены находки планктонных фораминифер, иногда в массовых количествах.

Зональная стратиграфия палеогена, как по планктону, так и по бентосу, нашла свое отражение в работах Маслаковой (1955, 1957), Мятлюк (1950, 1970), Маслун (Иваник, Маслун, 1977; Маслун, 1979), Грузман и Дабагян (1979). Однако схемы, разработанные ими для отдельных районов развития палеогеновых отложений Украинский Карпат, до настоящего времени разобщены, а ряд выделенных биостратиграфических единиц нуждается в пересмотре с точки зрения современного понимания их ранга и возрастного объема. В связи с этим давно назрела необходимость обобщения имеющихся данных по биозональному расчленению палеогена в виде единой сводки, включающей сопоставление планктонных зон с бентосными ассоциациями, что безусловно, увеличит стратиграфическую ценность последних.

В качестве эталонной принята зональная шкала Средиземноморья. При сопоставлениях с зарубежными Карпатами имеются определенные трудности, заключающиеся прежде всего в том, что подобные зональные схемы или сводки разработаны для отдельных структурно-фациальных единиц и подчас имеют существенные различия. Кроме того существуют разногласия в понимании зональных видов, стратиграфического объема зон и принципов их выделения. Поэтому на данном этапе исследований возможны лишь сравнения отдельных зональных схем на уровне подотделов палеогена.

Для Румынских Карпат использована схема фации Шотриле (Восточные Карпаты), как наиболее полная по планктону (Bratu, Ghe-

[1] Доклад была представлена на XII-ом Конгрессе Карпато-Балканской Геологической Ассоциации, 8—13 сентября 1981 г., Бухарест, Румыния.

[2] СССР.

ţa, 1972) и схема для внешнего флиша Краевой единицы, где выделяются ассоциации планктонных и бентосных фораминифер (Bratu, 1975). В Словацких Карпатах детально разработана сводная зональная схема для палеогенового флиша по планктону и бентосу (Samuel, Borza, Köhler, 1972; Samuel, 1973). Польские исследователи в целом ве склонны выделять в карпатском флише какие-либо биостратиграфические единицы, отдавая предпочтение характерным комплексам. Однако в последней работе Б. Ольшевской (Olszewska, 1980) приеводится биозональная схема для палеогена Дукельской единицы по планктону. Бентосные ассоциации фораминифер подробно изучены в Скольской единице (Morgiel, Szymakowska, 1978). Эти данные и были приняты за основу при сравнении с Польскими Карпатами.

В Украинских Карпатах относительно полное биозональное расчленение палеоцена и эоцена имеется для Мармарошской структурно-фациальной единицы (Грузман, Дабагян, 1979), хотя сами авторы считают эту схему несколько условной в силу сложного геологического строения района, прерывистости разрезов и других причин. В качестве основного разреза, в котором выделены биостратиграфические единицы по бентосу, может быть взят сводный разрез Скибовой зоны. Основная суть работы заключена в таблице. В тексте остановимся лишь на некоторых комментариях и дискуссионных вопросах.

В последнее время широкое распространение как в СССР, так и в других странах получила точка зрения о необходимости причисления датского яруса к палеогену. Анализ фауны из пограничных слоев мела и палеогена в непрерывных разрезах на северном склоне Украинских Карпат позволяет присоединиться к такому мнению. Комплекс планктона, найденный в верхней части стрыйской свиты и ее аналогах в других зонах, содержит набор видов, типичный для дания Средиземноморья, юга СССР и карпатского региона в целом. Однако подразделить его на зоны у нас пока не представляется возможным. В составе комплекса установлены: *Globigerina trivialis* Subb., *G. pseudobulloides* Plumm., *G. microcellulosa* Moroz., *G. varianta* Subb., *G. quadrata* White, *Subbotina triloculinoides* (Plumm.), *Acarinina inconstans* Subb., *Globorotalia compressa* (Plumm.), *Globoconusa daubjergensis* (Bronn).

Отличительную особенность нижнепалеоценовой зоны Globorotalia angulata, выделенной только в метовской свите Мармарошской единицы, составляет появление большого числа представителей зонального вида на фоне комплекса, в котором преобладают еще типично датские фораминиферы. Кроме *Globorotalia angulata* здесь встречены *Globigerina varianta* (Subb.), *G. trivialis* Subb., *Subbotina triloculinoides* (Plumm.), *S. nana* (Chalil.), *Globorotalia compressa* (Plumm.), *Acarinina inconstans* Subb. и др. Немногочисленные экземпляры *Globorotalia angulata*, найденные в ямненской свите, относящейся к верхнему палеоцену (Скибовая единица), и в ее возрастном аналоге лютской свите (Дуклянская единица) не могут служить основанием для выделения здесь биостратиграфической зоны с упомянутым зональным видом, как это было сделано в свое время Н. И. Маслаковой, а вслед за ней и другими авторами.

Верхний палеоцен метовской свиты А. Д. Грузман и Н. В. Дабагян охарактеризовали зоной Globorotalia pseudomenardii (по появлению многочисленных экземпляров этого вида), считая ее аналогом зоны G. velascoensis средиземноморской шкалы. Сама *G. velascoensis* здесь не обнаружена. Сопутствующими видами являются: *Globorotalia elongata* Boli. *G. aequa* Cushm. et Renz., *Acarinina acarinata* Subb., *A. Subsphaerica* Subb., *A. triplex* Subb., *Subbotina triloculinoides* (Plumm.), *S. quadritriloculinoides* (Chalil.), *S. nana* (Chalil.). Многочисленный верхнепалеоценовый планктон был найден в Скибовой зоне (Грузман, Портнягина, 1976) в разрезах рек Рушора, Брустурки, Сереteля и др. Авторы считают его аналогом нижней части зоны Acarinina subsphaerica Крымо-Кавказской области. По современным представлениям это скорее всего аналоги зоны *G. velascoensis*. Подобные комплексы планктона с *Globorotalia pseudomenardii* Bolli, *G. marginodentata* (Subb.), *Subbotina nana* (Chalil.) и др. известны в Дуклянской зоне в басс. р. Дусинки (Лозыняк, 1971) и в олистолитах по р. Абранке (материалы авторов).

По бентосным фораминиферам разделить палеоцен на нижний и верхний и отграничить его от датского яруса пока не удается. Большая часть разрезов датско-палеоценовых отложений представлена терригенными слабо карбонатными разностями пород, содержащими ассоциацию фораминифер с *Hormosina ovulum ovulum*, *Rzehakina epigona*, *Rzehakina fissistomata*, *Modellum velascoense*. В более карбонатных породах иногда развит комплекс секреционного бентоса с характерными видами *Stensioina caucasica* и *Anomalinoides danicus*.

В метовской свите Мармарошской единицы выделяются две нижнеэоценовых зоны. Первая из них — *Globorotalia subbotinae*, кроме зонального вида представлена многочисленными *Globorotalia marginodentata* Subb., *Acarinina acarinata* Subb., *A. triplex* Subb., *Subbotina triloculinoides* (Plumm.), *S. quadritriloculinoides* (Chalil.). Вторая зона Globorotalia aragonensis содержит *Globorotalia lensiformis* Subb., *Globigerina pseudoeocaena* Subb., *G. boweri* Bolli, *Acarinina pentacamerata* Subb., *A. acarinata* Subb., *A. interposita* Subb., *A. triplex* Subb. Здесь появляются единичные *Acarininna pseudotopilensis* Subb., *A. bullbrooki* Bolli и др.

Значительно раньше рассматриваемые зоны были выделены Н. И. Маслаковой (1955) на южном склоне Карпат и без достаточных оснований распространены на всю их территорию. Пользоваться этими материалами трудно, так как они не привязаны к конкретным разрезам. Вслед за Н. И. Маслаковой эти исследования были продолжены и конкретизированы Н. В. Дабагян (Кульчицкий, Лозыняк, Дабагян ... 1965), установившей в эоцене Дуклянского покрова ряд планктонных зон. Правда, следует оговориться, что по современным представлениям большая часть пород, вмещающих обильный планктон, принадлежит к олистолитам.

В нижнем эоцене повсеместно во всех структурно-фациальных зонах (в манявской свите и ее аналогах) присутствует характерная ассоциация бентосных фораминифер с *Glomospira charoides* (Park et Jones), *Recurvoides smugarensis* Mjatl., *Karreriella horrida* Dab. и др.

Зона Acarinina bullbrooki прослеживается в Мармарошской единице, в той же метовской свите, где вместе с зональным видом встречено множество *Acarinina pentacamerata* Subb., *A. acarinata* Subb., *A. pseudotopilensis* Subb., *Globigerina pseudoeocaena* Subb., *G. posttriloculinoides* Chalil., *G. frontosa* Subb. и единичные *Globorotalia aragonensis* Nutt., *Globigerapsis kugleri* Bolli, Loebl. et Tappan. В Дуклянской единице известны находки многочисленного планктона зоны *A. bullbrooki* в пестроцветах (скорее всего это олистолиты) бассейна Латорицы (материалы Н. В. Дабагян и авторов). Кроме того она установленна и в коренном разрезе в одном из правых притоков реки Дусинки (Габинет, Кульчицкий,... 1976). Наконец Е. В. Мятлюк (1970) указывает на наличие массовых количеств *Acarinina bullbrooki* и *Acarinina rotundimarginata* в пестроцветном горизонте в основании быстрицкой свиты и низах ее попельской фации (Скибовая единица) и считает возможным выделить здесь зону *Acarinina rotundimarginata*. Названные виды характеризуют две самостоятельные зоны, относящиеся к низам среднего эоцена, что в целом не увязывается с положением быстрицкой свиты в сводном разрезе. К среднему эоцену здесь относятся выгодская и пасечнянская свиты, лежащие ниже. Поэтому вопрос об аналогах зон A. bullbrooki и A. rotundimarginata в Скибовой единице приходится оставить открытым.

В среднем эоцене Мармарошской единицы прослеживается еще одна зона — Hantkenina albamensis. Кроме зонального вида здесь встречены многочисленные *Globigerapsis kugleri* Bolli, Loebl. et Tappan, *Acarinina bullbrooki* Bolli, *Globigerina pseudoeocaena* Subb. и единичные *Truncorotaloides rohri* Bronn. et *Truncorotaloides rohri* Bronn. et Berm., *Globanomalina micra* (Cole).

Бентосные фораминиферы среднеэоценовых отложений Украинских Карпат (большей частью это массивные песчаники) не отличаются особым разнообразием, позволяющим выделять здесь какие-либо характерные комплексы. Лишь на северном склоне в средней и верхней частях пасечнянской и выгодской свит отмечен своеобразный секреционный бентос с коническими цибицидоидесами: *Cibicidoides westi* (Howe), *C. ventratumidus* (Mjatl.), *C. grossoconulus* (Mjatl.).

Планктон, найденный в верхней части метовской свиты, позволил А. Д. Грузман и Н. В. Дабагян выделить здесь две верхнеэоценовых зоны — Globigerapsis index и Globigerina corpulenta. Первый из названных видов встречен в массовых количествах совместно с единичными *Globigerapsis rubriformis* (Subb).), *Globigerina corpulenta* (Subb.). Е. В. Мятлюк (1970) указывает на присутствие большого числа экземпляров этого вида [*Globigerinoida index* (Finl.)], в попельской свите Скибовой зоны.

Вторая верхнеэоценовая зона Globigerina corpulenta охватывает стратиграфический горизонт, известный под названием шешорский или горизонт крупных глобигеринид, имеющий широкое распространение в пределах всей карпатской дуги. Мергелистые породы этого горизонта буквально переполнены планктоном. Среди него преобладают *Catapsidrax dissimilis* (Cushm. et Berm.), *Globigerapsis index* (Finl.), *Globigerina corpulenta* Subb., *G. galavisi* Bermud., *Subbotina eocaenica eocaenica* (Terq.), *S. eocaenica irregularis* (Subb.).

Основным показателем верхнеэоценового возраста пород по бентосу в Украинских Карпатах обычно считалась ассоциация агглютинирующих фораминифер с *Cyclammina amplectens*, приуроченная к быстрицкой свите Скибовой зоны. Отложения с подобной ассоциацией в других тектонических зонах рассматривались как аналоги быстрицкой свиты и относились к верхнему эоцену. Однако со временем практика исследований показала, что комплекс с *C. amplectens* встречается и ниже по разрезу, подчас совместно со среднеэоценовым планктоном и нуммулитами. В Польских и Словацких Карпатах (Morgiel, Szymakowska, 1978; Samuel, 1973), *Cyclammina amplectens* характеризует средний эонец, а в верхнем развита *Cyclammina rotundidorsata*. Маслун (1979) считает, что объем вида *Cyclammina amplectens* понимается слишком широко. Онхарактерен только для верхнего эоцена, а ниже встречаются *Cyclammina amplectens* var. *ornamenta* Dabag. и *Cyclammina intermedia* Mjatl. А. Д. Грузман и Н. В. Дабагян (1979) отмечают присутствие *Cyclammina amplectens* с сопутствующими *Asanospira walteri* и *Hyperammina lineariformis* по всему разрезу среднего и верхнего эоцена (метовская свита, Мармарошская зона). Из сказанного следует, что ассоциация фораминифер с *Cyclammina amplectens* сама по себе вопроса о возрасте однозначно не решает, а разногласия по поводу ее вертикального распространения требуют специальных дополнительных исследований.

В олигоценовых отложениях Украинских Карпат микрофауна встречается большей частью спорадически, однако в процессе многолетних исследований здесь выявлен достаточно многочисленный секреционный бентос и планктон. Зональное подразделение олигоцена затруднено, так как планктонные фораминиферы не особенно богаты в систематическом отношении и незначительно меняются по вертикали.

В основании менилитовой толщи Скибовой, Кросненской и Дуклянской зон в подроговиковой части разреза прослеживается горизонт с мелкими глобигеринами, резко отличающимися по своему составу от комплекса крупных глобигерин подстилающего шешорского горизонта. Эта часть разреза выделяется в местную зону Globigerina officinalis, Subbotina vialovi и Planorbella (Вялов, Дабагян, Мятлюк ... 1965). Именно по смене этих комплексов и по границе шешорского горизонта и подроговиковой части менилитовых слоев проводится граница между эоценом и олигоценом в Украинских Карпатах.

В нижних частях менилитовой и кросненской олигоценовых толщ встречаются *Globigerina officinalis* Subb., *G. praebulloides* Blow, *Globorotalia pseudoscitula* Glaessn., *G. brevispira* Subb., *Cibicidoides lopjanicus* (Mjatl.), *Elphidium karpaticum* Mjatl., *Uvigerinella majkopika* Kraeva. Выше по разрезу известны *Globigerina ampliapertura* Bolli, *G. angustiumbilicata* Bolli, *G. ciperoensis* Bolli, *G. pseudoedita* Subb., *Cibicidoides amphysiliensis* (Andr.), *Baggatella altiuscula* Subb. и др. (Грузман, 1972, Андреева-Григорович, Грузман, 1978).

Неоходимо отметить, что по последним представлениям А. Д. Грузман (1981) большую часть упомянутых толщ следует отчленить от олигоцена, так как в среднекросненской и среднеменилитовой (лопянецкой) свитах установлена миоценовая микрофауна. На наш взгляд такое решение имеет свои неясные стороны и не может считаться окончательным.

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СРАВНЕНИЕ ЗОНАЛЬНЫХ СХЕМ ПАЛЕОГЕНА ВОСТОЧНЫХ И ЗАПАДНЫХ КАРПАТ

С.О. ВЯЛОВ, Д.Л. ПОНОМАРЕВА. Зональная корреляция Палеогеновых отложений Украинских Карпат по Мелким Фораминиферам.

ОТДЕЛ	ПОДОТДЕЛ	СТРАТИГРАФИЧЕСКАЯ ШКАЛА ПАЛЕОГЕНА СРЕДИЗЕМНОМОРЬЯ И ОКЕАНОВ. Крашенинников, Немков, 1975	УКРАИНСКИЕ КАРПАТЫ. МАРМАРОШСКАЯ ЗОНА. Трузман, Дабагян, 1979. ПЛАНКТОН	МАРМАРОШСКАЯ ЗОНА. БЕНТОС	ДУКЛЯНСКАЯ ЗОНА. Кульчицкий, Лозыняк, Дабагян, 1965; Кульчицкий, Дабагян, 1980; Маслакова, 1955; материалы авторов. ПЛАНКТОН	ДУКЛЯНСКАЯ ЗОНА. БЕНТОС	СКИБОВАЯ ЗОНА. Андреева-Григорович, Гризман, 1978; Гризман, Портнягина, 1976; Вялов, Дабагян, Мятлюк...1965; Мятлюк, 1970; материалы авторов. ПЛАНКТОН	СКИБОВАЯ ЗОНА. БЕНТОС	СКИБОВАЯ ЗОНА. СВИТЫ	РУМЫНСКИЕ КАРПАТЫ. ФАЦИЯ ШОТРИЛЕ. Bratu, Gheţa, 1972. ПЛАНКТОН	РУМЫНСКИЕ КАРПАТЫ. КРАЕВАЯ ЗОНА. Bratu, 1975. ПЛАНКТОН, БЕНТОС	СЛОВАЦКИЕ КАРПАТЫ. Samuel, 1973. ПЛАНКТОН	СЛОВАЦКИЕ КАРПАТЫ. Samuel, 1973. БЕНТОС	ПОЛЬСКИЕ КАРПАТЫ. ДУКЕЛЬСКАЯ ЕДИНИЦА. Olszewska, 1980. ПЛАНКТОН	ПОЛЬСКИЕ КАРПАТЫ. СКОЛЬСКАЯ ЕДИНИЦА. Morgiel, Szymakowska, 1978. БЕНТОС
ОЛИГОЦЕН		*Globigerina ciperoensis* *Globorotalia opima* *Globigerina ampliapertura* *Globigerina sellii – Pseudohastigerina barbadoensis* *Globigerina tapuriensis*			*Subbotina vialovi*		*Subbotina vialovi*	*Cibicidoides amphysiliensis* *Elphidium Karpaticum Cibicidoides lopjanicus* *Grammostomum danvillensis Uvigerinella majcopica*	ВЕРХНЕМЕНИЛИТОВАЯ ЛОПЯНЕЦКАЯ НИЖНЕМЕНИЛИТОВАЯ			*Globigerina postcretacea*		*Turborotalia increbescens*	
ЭОЦЕН	ВЕРХНИЙ	*Globigerina corpulenta*: *G. gortanii – G. centralis*; *Globorotalia cocoaensis*; *Globigerapsis semiinvoluta*	*Globigerina corpulenta* *Globigerapsis index*		*Globogerina corpulenta*		*Globigerina corpulenta* *Globigerapsis index*	*Cyclammina placenta Ammodiscus latus Cycl. amplectens, Reophax planus* *Asterigerina rogalai, Brizalina pseudoaenariensis* *Cibicidoides tallahatensis*	БЫСТРИЦКАЯ ПОПЕЛЬСКАЯ	*Globigerina corpulenta* *Globigerapsis tropicalis* *Globigerapsis index*	*Globigerina (Subbotina) corpulenta* *Globigerapsis index Spiroplectammina*	*Globigerina officinalis* *Globigerapsis index*	? – – – – – – – – ? *Cyclammina rotundidorsata*	*Globigerina ampliapertura* *Globorotalia cocoaensis* *G. cerroazulensis*	*Cyclammina rotundorsata Ammodiscus latus* *Rotalia lithotamnica, Asterigerina rogalai,* *Almaena taurica*
ЭОЦЕН	СРЕДНИЙ	*Truncorotaloides rohri* *Orbulinoides beckmanni (Hantkenina alabamensis)* *Globorotalia lehneri (Acarinina rotundimarginata)* *Acarinina bullbrooki*: *Globigerapsis kugleri*; *Hantkenina aragonensis*	*Hantkenina alabamensis* *Acarinina bullbrooki*	*Cyclammina amplectens Hyperammina leariformis Asanospira walteri*	? *Acarinina bullbrooki*	*Cyclammina amplectens Hyperammina lineeariformis Asanospira walteri*	?	конических *Cibicidoides* *C. westi* *C. ventratumidus* *C. grossoconulus*	ВЫГОДСКАЯ	*Hantkenina liebusi* *Acarinina bullbrooki*	*Cyclammina amplectens Cystamminella elongata Asanospira walteri. Recurvoides walteri* *Cyclammina amplectens Subbotina frontosa Acarinina Globigerapsis subconglobatus*	*Truncorotaloides rohri* *Turborotalia (A) crassata densa*	*Cyclammina amplectens* *Ammodiscus latus* ? – – – – ?	*Globorotalia densa – Turborotalia rotundimarginata*	*Cyclammina amplectens* *Reophax pilulifer* *Ammodiscus latus*
ЭОЦЕН	НИЖНИЙ	*Globorotalia aragonensis*: *Acarinina pentacamerata (G. palmerae)*; *Globorotalia aragonensis* *Globorotalia marginodentata (Globorotalia formosa)* *Globorotalia subbotinae*	*Globorotalia aragonensis* *Globorotalia subbotinae*	*Glomospira charoides Recurvoides smugarensis*	*Globorotalia aragonensis* *Globorotalia subbotinae*	*Glomospira charoides Recurvoides smugarensis*		*Glomospira charoides Recurvoides smugarensis Karreriella horrida*	МАНЯВСКАЯ	*Globorotalia aragonensis* *Globorotalia lensiformis*	*Dendrophrya Trochamminoides Recurvoides*	*Globorotalia aragonensis crater* *Globorotalia subbotinae marginodentata*	*Glomospira Ammodiscus*	*Globanomalina wilcoxensis* *Globorotalia subbotinae*	*Saccaminoides carpathicus Glomospira*
ПАЛЕОЦЕН	ВЕРХНИЙ	*Globorotalia velascoensis*: *Globorotalia velascoensis*; *Globorotalia pseudomenardii*	*Globorotalia pseudomenardii*	*Hormosina ovulum Rzehakina fissistomata Nodellum velascoense* *Stensioina caucasica Anomalinoides danicus*	*Globorotalia velascoensis*	*Hormosina ovulum Rzehakina fissistomata Nodellum velascoense* *Stensioina caucasica Anomalino-ides danicus*	*Globorotalia velascoensis*	*Hormosina ovulum Rzehakina fissistomata Nodellum velascoense* *Stensioina caucasica Anomalinoides danicus*	ЯМНЕНСКАЯ	*Globorotalia aequa Globorotalia marginodentata*	*Rhabdammina discreta, Rzehakina, Carpathiella, Ovulum, Karreriella* *Cystamminella pseudopauciloculata, Subbotina triloculinoides, Morozovella angulata*	*Globorotalia aequa*	*Rzehakina inclusa – Rzehakina complanata* *Hormosina ovulum ovulum*		*Recurvoides Trochamminoides* *Glomospira diffundens Rzehakina fissistomata*
ПАЛЕОЦЕН	НИЖНИЙ	*Globorotalia angulata (Globorotalia pusilla)*: *Globorotalia conicotruncata*; *Globorotalia angulata*	*Glorotalia angulata*		?		?				*Rhabdammina cylindrica cylindrica Carpathiella ovulum Bogdanovicziella complanata, Cystamminella gezybowskii, Anomalina danica, Subbotina triloculinoides, Subbotina pseudobulloides, Subbotina varianta*	*Globorotalia pusilla pussilla*		*Globigerina triloculinoides – Globigerina varianta*	
ДАНИЙ		*Acarinina uncinata* *Globorotalia pseudobulloides (G. trinidadensis)* *Globigerina taurica*			*Globigerina trivialis Globoconusa daubjergensis Globorotalia compressa*		*Globigerina trivialis Globoconusa daubjergensis Globorotalia compressa*		СТРЫЙСКАЯ			*Turborotalia (A.) inconstans praecursoria* *Globigerina compressa*	*Rzehakina epigona*		*Hormosina ovulum Rzehakina fissistomata Nodellum velascoense*

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STRATIGRAPHIE ET PALÉONTOLOGIE