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Xth Congress RCMNS
Bucharest, September 4-9, 1995

TOPICS:

NEW LIGHTS ON MEDITERRANEAN CHRONOLOGY
TETHYS-PARATETHYS CONNECTIONS

AND

SYMPOSIA:

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AVANT-PROPOS

Ce numéro du Romanian Journal of Stratigraphy publie une sélection des quelques 150 travaux présentés au 10ème Congrès du "Regional Committee on Mediterranean Neogene Stratigraphy" (Bucarest, 4-9 septembre 1995).

A côté des deux thèmes principaux du congrès "New lights on Mediterranean chronology" et "Tethys-Paratethys connections", les communications ont concerné cinq autres sujets, objets des symposia "Neogene volcanism", "Quantifying global climate on the Pliocene", "Definition and chronology of the major Neogene tectonic events", "Biosedimentology" et "Economic potential of the Neogene formations".

Les articles qui composent ce volume représentent les différents thèmes du congrès. Celui-ci a été ouvert par les conférences de M. Săndulescu ("L'histoire géologique paléogéographique de la Roumanie") suit par Fl. Marinescu avec des détails (sur "Le Néogène de Roumanie"), par F. J. Hilgen ("Towards an astronomical time scale for the Mediterranean Neogene, a major advance in resolution and accuracy") et par M. L. Bordenave ("Influence of the Neogene tectonic on the entrapment of the oil reserve in the Zagres orogenibelt of Iran"). Les conférences ont été suivies par un exposé de J. E. Meulenkamp, ancien président du RCMNS qui a retracé l'histoire de cet organisme, dont on trouvera le texte ci-après.

Le 10ème Congrès du RCMNS a permis de nombreuses discussions notamment sur les questions qui sont toujours d'actualité comme la crise de salinité messinienne ou les échanges entre la Méditerranée et la Parathéthys. Il a été parfaitement organisé par Fl. Marinescu, entouré d'une équipe très efficace et extrêmement accueillante. Il convient de les féliciter très chaleureusement d'autant plus que cette organisation n'a pas toujours été facilitée par un contexte favorable, surtout du point de vue économique.

J'adresse mes félicitations et remerciements aux organisateurs ainsi qu'aux participants et je donne rendez-vous à tous pour les prochains interim-coloquia de Catania (novembre 1997) et de Patras (mai 1998), en attendant le 11ème Congrès, qui se tiendra au Maroc (Rabat, 2000).

Jean-Pierre SUC
Président du RCMNS



RCMNS - PAST, PRESENT, FUTURE

J. E. MEULENKAMP

It is most interesting to look back on one and half century of research on the Neogene of the Mediterranean and the Paratethys in particular here in Bucharest, at the 10th international congress of the Regional Committee on Mediterranean Neogene Stratigraphy. In fact, the activities of what is now the RCMNS started in the late fifties after about a century of research on the Neogene, which started with the introduction of Neogene stages and ended with the introduction of marine microfossil biostratigraphy as a major tool in relative datings and correlations. During this first century age assignments and correlation were primarily based on the combination of lithological and macropaleontological criteria. As we know now, this led to erroneous correlations, but in retrospect such correlations made sense and were based on a sound way of reasoning at the time.

A nice example is given by the sequence of the late Middle and early Late Miocene stages. The sandy Helvetien with *Cardita jovanetti* was followed by the clayey Tortonian with *Ancilla glandiformis*, a succession recognized both in the Paratethys and the Mediterranean proper. In both areas, the top parts of the Miocene sequence yielded ample evidence of environmental conditions with salinities strongly deviating from normal. So, it seemed quite logical to correlate the brackish Sarmatian of the Paratethys with the sediments deposited during the Messinian salinity crisis in the Mediterranean. Moreover, both the Sarmatian and Messinian were considered to be overlain by the Pliocene (Pannonian and Tabianian stages, respectively).

In the late fifties and early sixties it became evident that this hitherto generally accepted frame was no longer tenable. This awareness was, amongst other things, based on the rapidly increasing importance of marine micropaleontology in dating and correlating Neogene rocks, also in response to the needs of industry. In continental biostratigraphy micromammalia started to play a pronounced part at the time. Throughout the sixties and the early seventies zonal schemes were established for both the Paratethys and the Mediterranean. These efforts went along with inventories of stages and renewed studies of stratotype sections and they resulted in new stratigraphic frames presented in Bratislava in 1975. These, basically, are the schemes still in use today.

The 1975 "spike" in the history of the RCMNS is also of interest because the discussions at the Bratislava Congress marked and in fact further accentuated three different lines of approach to Neogene stratigraphic studies in the later seventies and the early eighties. The first line of approach, followed in particular by those involved in planktonic studies, was based on the definition and application of first and last occurrences of planktonic species as "datum planes", Mediterranean data and interpretations became elaborated in the context of global stratigraphic studies related to ocean drilling, which provided an overwhelming set of data pertinent to the evolution of the biorecord. There was great confidence in the possibility to establish refined bio- and chronostratigraphic interregional correlations (chromozones). As a logical consequence, boundary rather than "body" stratotypes were considered the best references for the chronostratigraphic frame.

In contrast, studies pertinent to the second line of approach, followed in particular by those involved in quantitative studies of marine microfossil assemblages, focussed on the degree of accuracy and precision that could be reached in stratigraphic correlations. Bodies rather than boundaries were considered the appropriate tools to characterize the chronostratigraphic framework. By principle, there was no great confidence in the concept of "datum planes" and its application in time-stratigraphic correlations. It is not surprising that specialists working along the second line of approach were amongst the initiators and strong defenders of the concept of superstages, which resulted in the establishment of four such superstages for the Mediterranean Neogene at the Bratislava congress in 1975.

In retrospect, it is quite understandable that specialists working along the two lines of approach drifted apart and that some kind of scientific "barrier" developed in the late seventies - early eighties: "ocean drilling-based" further refinement of bio- and chronostratigraphic scales with the incorporation of paleoceanographical and paleoclimatological concepts versus "land-based" critical analyses of the accuracy of stratigraphic correlation tools through high-resolution quantitative studies of selected sections with the incorporation of modern mathematical/statistical concepts. No doubt this is a gross sim-



plification of the developments as they really were, but "pushing the extremes" may be of help in order to better understand the history of the evolution of (Mediterranean) Neogene stratigraphic concepts.

The third line of approach, followed in particular by those working with problems of the Paratethyan realm, dealt with the integration of various types of stratigraphic correlation tools into comprehensive biostratigraphic (multiple zonation) and chronostratigraphic schemes, including regional stages and substages. In fact, the recurrent and profound environmental changes which had occurred in the Paratethys in the Neogene impeded the establishment of "simple" marine or continental zonations based on selected groups of marine or terrestrial organisms. Instead, the integrated, chronostratigraphic framework established in the course of the seventies and further refined in the eighties, mirrors the application of "ecostratigraphy *avant-le-lettre*", the construction of time planes through environmentally-defined time stratigraphic units.

Parallel to these developments increasing attention was paid to the impact environmental control had on the composition of faunal/floral associations and on the temporal and spatial distribution of index species. Awareness of this environmental control and its potential use in time-stratigraphic correlations during the RCMNS congress in Athens, 1979, led to the conclusion that ecostratigraphy should be the unifying concept in Mediterranean/Paratethyan stratigraphic research for the eighties. In addition, the role to be played by the application of the hitherto achieved stratigraphic results in palaeogeographic/geodynamic studies and in studies pertinent to the origin and distribution of mineral resources was emphasized. Paleoenvironments, paleoclimates and paleoecology rather than bio- and chronostratigraphy were in the centre of interest at the one side, basin analysis and paleogeography at the other during the Budapest (1985) and, in particular, the Barcelona (1990) congresses.

The outlines of RCMNS research between Athens, 1979, and Bucharest, 1995, clearly demonstrate the

large impact, the explosive (further) development and ensuing application of chemical and physical tools in global (marine) stratigraphic and environmental/climatic studies had on the knowledge and understanding of the Mediterranean Neogene. In return, the application of high-resolution correlations in paleoenvironmental/paleoclimatic reconstructions in the Mediterranean had an indisputable impact on the development of global studies. Also in the eighties and early nineties paleoecological models were established to account for the evolution of Mediterranean biocommunities and first attempts were made to intercalibrate in greater detail environmental and climatic signals inferred from the marine and terrestrial records, respectively in basin analysis and paleogeography important results were obtained with respect to the construction of fairly accurate paleogeographic map series, both for the Mediterranean and Paratethyan Neogene. New and most promising, recent developments included the establishment and application of astronomical polarity time scales for part of the Neogene, which allowed to date, correlate and interpret Neogene sequences in unprecedented detail, i.e., with a degree of resolution of less than 20.000 years. This is also of great importance for basin analysis and geodynamics, in order to constrain e.g. the timing and duration of tectonic events and paleogeographic reorganizations.

The establishment and intercalibration of such high-resolution astronomical time scales for both the marine and terrestrial records of the Mediterranean and the Paratethys should be one of the major and challenging goals of future RCMNS activities. The application of these scales will allow to perform detailed, integrated paleoenvironmental/paleoclimatic studies along marine-continental environmental gradients. These, in turn, may be expected to result in a "total system view" on the evolution of circum-Mediterranean climates, ecosystems and environments. The RCMNS should face the challenge to initiate, stimulate and coordinate this type of pioneer research.



THE OLIGOCENE-MIOCENE SEDIMENTARY AND VOLCANIC SUCCESIONS OF CENTRAL SARDINIA, ITALY

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Key words: Oligocene. Miocene. Sardinian calc-alkaline volcanism. Ignimbrites. Biostratigraphy. Tectonics.

Abstract: The Upper Oligocene-Middle Miocene sedimentary and volcanic successions, cropping out with a thickness of more than 1000 m in an extensive area in Central Sardinia, are described in detail from stratigraphical, volcanological and tectonical points of view. The following lithostratigraphic successions are evidenced: A) Upper Oligocene- Burdigalian continental volcano-sedimentary sequence; B) Upper Oligocene- Middle Burdigalian marine sedimentary cycle; C) Late Burdigalian-Upper Langhian marine sedimentary cycle. The continental sequence (A), around 300 m thick, lies unconformably on the Paleozoic metamorphic basement and begins with a continental formation (CPC) related to the top of the Ussana Formation from Central-Southern Sardinia; it consists of conglomerates, silts and reddish clays belonging to an alluvial fan depositional system fed by a tectonically uplifted Paleozoic relief. A calc-alkaline sequence follows, in which several eruptive units were distinguished: a) Luzzana Unit (LU); b) Allai Unit (AL); c) Dacitic Lava Unit (DL); d) Ruinas Unit (RU); e) Monte Ironi Unit (MI). They are dacitic and rhyolitic ash-pumiceous pyroclastic flows (ignimbrites) and lava-like welded ignimbrites, followed by dacitic lava-flows and lava-domes, raised along fractures related to volcano-tectonic collapse phenomena. In particular, the Allai Unit ignimbrite, about 50 m thick, represents the most important Cenozoic explosive event occurring in the area. Furthermore, inside the volcanic sequence some sedimentary episodes (TT, AS) are pointed out and related to alluvial fan and lacustrine environments, which seem to mark relative pauses in the volcanic activity. The first marine sedimentary cycle (B), about 500 m thick, is constituted by silici-clastic, mixed-silici-clastic-carbonatic and partly carbonatic sediments of littoral to epibathial environment, with interbedded subaqueous basaltic and andesitic-basaltic volcanics. Near the Omodeo Lake (Sa Manenzia Unit) and near Paulilatino, overlying pyroclastic eruptive products, sedimentary rocks belonging to a second transgressive cycle (C) occur. Through an erosional unconformity siliciclastic, arenaceous and arenaceous-marly sediments, carbonatic lithofacies and pumiceous pyroclastic intercalations, belonging to infra-circalittoral and epibathial environments, crop out. This succession is about 500 m thick. In the studied area, the calc-alkaline volcanic activity and the most important Cenozoic structural modifications are genetically related to the main geodynamic events of the Western Mediterranean: particularly, to the compressional phase of the Northern Apennines during Oligocene-Aquitaniian and to the extensional tectonics as a consequence of the gravitational collapse of Northern-Appenninic orogen in the Burdigalian time. Then, the separation and the drift of Sardinia-Corsica microplate from the South-European margin can be connected to this extensional post-collisional phase.





Fig. 1 - Simplified geological sketch map of Central Sardinia. 1, Metamorphic and magmatic Paleozoic basement; 2, Mesozoic carbonatic cover; 3, Late Oligocene-Early Aquitanian Continental Polygenic Conglomerates (CPC); 4, Rhyodacitic Large Volume ignimbrites of Luzzana and Allai Units (LU, AL); 5, Dacitic lava flows (DL) and rhyodacitic lava-like ignimbrites (RU, MI) with interbedded Terrigenous-Tuffaceous Unit (TT) and Asuni Unit (AS); 6, Undifferentiated marine sediments of Late Oligocene-Burdigalian and Late Burdigalian-Upper Langhian age; 7, Submarine basalts and basaltic andesites of Upper Basic Intermediate Series (UBLS); 8, Basaltic lava flows of Pliocene-Pleistocene alkaline volcanic cycle; 9, Quaternary continental deposits; 10, Tertiary faults.

1. Geological setting

The Oligocene-Miocene sedimentary and volcanic successions cropping out in Central Sardinia are located in the eastern side of the Lower Miocene Sardinian Rift (Figs. 1 and 3). They lie unconformably on a Paleozoic basement constituted by low- and medium-grade Hercynian metamorphic rocks belonging to different allochthonous tectonic units, affected by several deformation phases (Carmignani et al., 1982) and intruded by Late-Hercynian granitoids.

To the west, the Paleozoic rocks form the Monte Grighini massif (673 m), while, to the east, they extend from Samugheo to Nureci.

Unconformably overlying the Paleozoic basement, Mesozoic carbonatic rocks represent the Mesotriassic and Mesojurassic marine ingressions (Damiani & Gandin, 1973). They display a tabular morphology, and are displaced to different altitudes by Cenozoic tectonics.

In the Oligocene-Miocene successions it is possible to distinguish: A) a volcano-sedimentary succession, ca. 300 m thick, of Upper Oligocene-Burdigalian age, built by reddish, continental, conglomeratic-siltitic-argillaceous deposits, followed by rhyolitic-rhyodacitic calc-alkaline volcanic rocks of prevalently ignimbritic facies; volcanoclastic and sedimentary continental products as well as dacitic lava-flows are also interbedded; B) a first marine sedimentary cycle, heteropic with the above volcano-sedimentary succession, constituted by littoral and bathial sediments Upper Oligocene to Middle Burdigalian in age, represented by conglomerates and sandstones, with associated (Marmilla area: Maccioni, 1974) subaqueous calc-alkaline volcanic products (pillow-lavas, hyaloclastites) varying from basalts to andesites; C) with a hiatus of Middle-Upper Burdigalian age, follows a second marine sedimentary cycle, constituted by Late Burdigalian-Upper Langhian marly-arenaceous sediments belonging to infra-circalittoral to epibathial environment, where pumiceous pyroclastic levels (Pecorini, 1975), representing the last calc-alkaline volcanic event in Central-Southern Sardinia, are present. Finally, above the described Oligocene-Miocene successions, through a sharp erosional limit, unconformably lie basaltic lava-flows belonging to the Pliocene-Pleistocene alkaline cycle (Beccaluva et al., 1987). They are linked to tectonic movements related to the opening of the Southern Tyrrhenian Sea.

2. The Oligocene-Miocene successions

2.1. Continental volcano-sedimentary successions: Upper Oligocene-Aquitania-Burdigalian

2.1.1. Continental Polygenic Conglomerates (CPC): Upper Oligocene-Early Aquitanian

Along the eastern slopes of the Monte Grighini Massif, both interbedded in the first two Cenozoic ignimbrites (LU, AL) and lying directly on the Paleozoic basement, crops out a thick (up to 100 m) continental terrigenous formation (CPC) constituted by polygenic conglomerates with a reddish argillaceous-arenaceous matrix, and associated lens-shaped layers of reddish siltstone and mudstone, deposited in an alluvial fan system under semiarid climatic conditions.

Their age, firstly ascribed to Permian (Atzeni, 1962), has been later correctly assigned to the Oligocene-Miocene by Coccozza & Valera (1968). Analogous continental deposits, underlying "Oligocene-Miocene" volcanics are known also in other areas of the Island (cfr., Negretti, 1957; Dieni & Massari, 1970; Barca, 1973; Porcu, 1983; Assorgia et al., 1988; Assorgia et al., 1991).

Field relations between this continental terrigenous formation (CPC) and the volcanic units of Luzzana (LU) and Allai (AL), the latter dated (K/Ar, see Tab.) 20.3 ± 1.0 Ma by Assorgia et al. (1995), suggest for CPC a more precise Upper Oligocene-Early Aquitanian age. Furthermore, the CPC were originated by erosion of morpho-structural paleoreliefs, essentially constituted by Paleozoic basement, subjected to abrupt rising probably as a consequence of transpressional movements linked to the collisional tectonics of the Northern Apennines (Carmignani et al., 1994).

The sedimentological characteristics of these syn-tectonic clastic deposits have allowed to define the following sedimentary processes:

a) massive gravitative flows and channelled flows in the proximal areas, localized on the eastern slopes of the Monte Grighini, and characterized as lenticular layers bounded by erosional surfaces and as reverse and/or normal graded-bedding attributable to high density turbulent flows; the transport of the clastic material occurred both through gravitacional flows along channels, and by tractive actions;

b) reworking flows in distal areas, whose deposits consist of planar or lenticular beds, with planar-parallel lamination; they are also characterized by both higher percentage of the argillaceous-arenaceous matrix and cementation degree. The morphometric characteristics of the conglomeratic fraction, and, in particular, the considerable frequency of the coarse clasts ($\phi > 50$ cm), seem to point out the proximity of their source area that coincides with a rising relief affected by intense erosion due to the rapid incision of the hydrographic system.



Table - K/Ar radiometric data of some large volume ignimbrites from Sardinia

| Sample locality | Material | % K | $^{40}\text{Ar}_{\text{rad}}(\text{nl/g})$ | $^{40}\text{Ar}_{\text{rad}}/^{40}\text{Ar}_{\text{tot}}$ | Conv. Age $\text{Ma} \pm 1\sigma$ |
|-----------------|----------|-------|--|---|--------------------------------------|
| Funtanazza* | Pl | 0.454 | 4.216 | 0.535 | 23.8 ± 1.2 |
| Ottana* | Bt | 5.750 | 4.721 | 0.245 | 21.1 ± 1.4 |
| Allai* | Pl | 0.550 | 4.364 | 0.360 | 20.3 ± 1.0 |
| Villanovaforru* | Pl | 0.567 | 4.548 | 0.676 | 20.5 ± 0.8 |
| Bosa** | Bt | 6.50 | 5.200 | 0.712 | 20.5 ± 0.8 |
| Bosa** | Bt | 6.16 | 4.732 | 0.809 | 19.7 ± 0.8 |
| Bosa** | Bt | 6.36 | 4.819 | 0.743 | 19.4 ± 0.8 |

Conventional ages are calculated using $^{40}\text{K}/\text{K}_{\text{total}} = 1.167 \cdot 10^{-4}$ as atomic ratio and the following decay constants: $\lambda_{\text{total}} = 5.543 \cdot 10^{-10}/\text{y}$, $\lambda_{\beta} = 0.581 \cdot 10^{-10}/\text{y}$, and $\lambda_{\beta} = 4.962 \cdot 10^{-10}/\text{y}$. Errors at 1σ level are calculated according to Dalrymple and Lamphere's (1969) formula. * = Data from Assorgia *et al.* (1995); ** = New data.

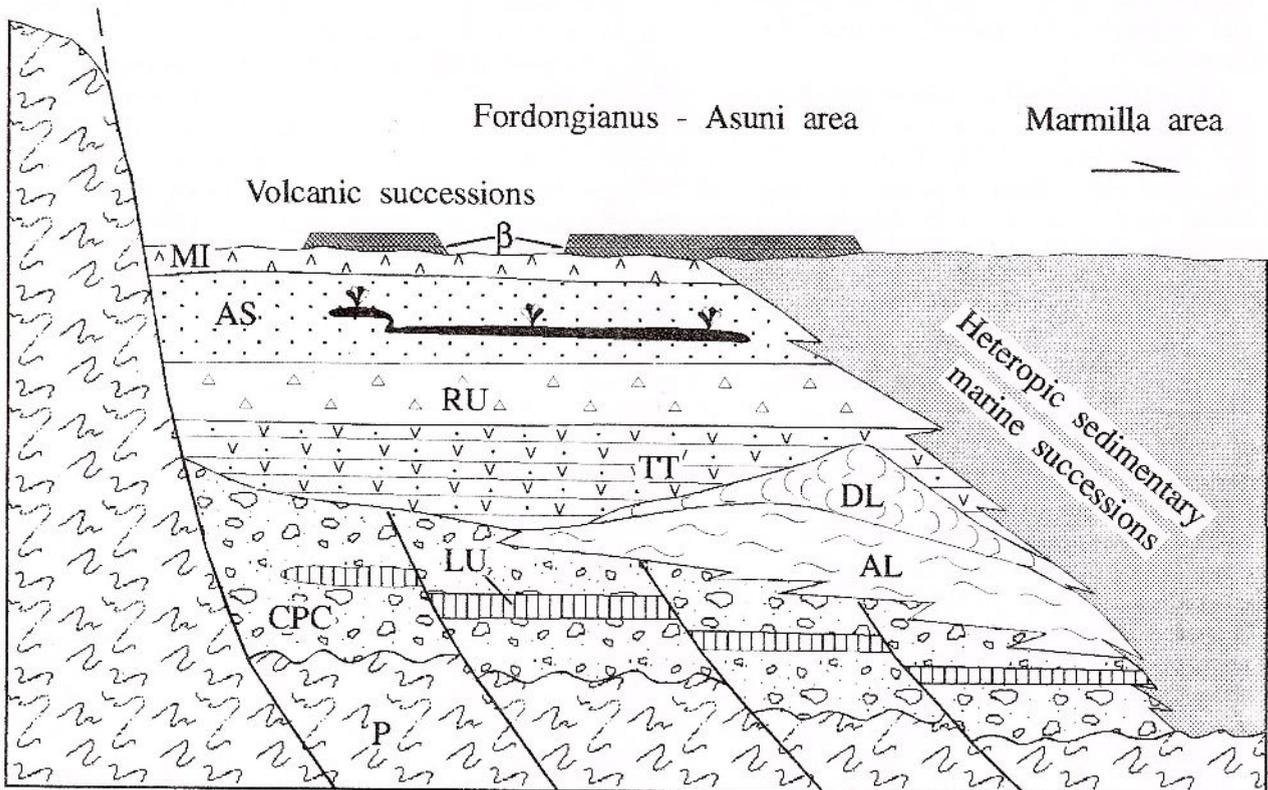


Fig. 2 - Simplified stratigraphical sketch of the Oligo-Miocene volcanic and sedimentary successions in Central Sardinia (not in scale). P: Paleozoic basement; CPC: Continental Polygenic Conglomerates; LU: Dacitic ignimbrite of Luzzana Unit; AL: Rhyodacitic ignimbrite of Allai Unit; DL: Dacitic lava-flows; TT: Terrigenous-Tuffaceous Unit; RU: Dacitic ignimbrite of Ruinas Unit; AS: Epiclastic and volcanoclastic deposits with fluidal globular peperites of Asuni Unit; MI: Dacitic ignimbrite of Monte Ironi Unit; β: Pliocene alkaline basalts.

2.1.2. Volcanic Units: Aquitanian-Burdigalian

The lower volcano-sedimentary succession is articulated in different pyroclastic, effusive and volcanoclastic-sedimentary units; from the oldest to the youngest, they are (Fig. 2):

- a) Luzzana Unit (LU);
- b) Allai Unit (AL);
- c) Dacitic Lava Unit (DL);
- d) Terrigenous-Tuffaceous Unit (TT);
- e) Ruinas Unit (RU);
- f) Asuni Unit (AS);
- g) Monte Ironi Unit (MI);

a) Luzzana Unit (LU)

The first volcanic event is represented by the Luzzana Unit (LU), a grey-greenish dacitic (Pl+Cpx) ignimbrite, characterized by lithic fragments belonging to both Cenozoic volcanics and Paleozoic basement xenoliths, and by small pumices embedded in a cineritic matrix of glass shards, sometimes with a globular aspect. The juvenile pumiceous fragments are essentially whitish, centimetric pumices, generally collapsed, and subordinate greyish scoriaceous fragments with a porphyritic texture due to abundant crystals of plagioclase. Among the lithic fragments, the cognates are brownish in color, centimetric in sizes, with aphyric and porphyritic texture (abundant plagioclase phenocrysts); they prevail over the metamorphic xenoliths. Furthermore, the latter, up to 15 cm in sizes, tend to concentrate in the central-basal part of the ignimbrite body, which crops out as an individual pyroclastic flow unit; the bottom contains abundance of carbonised wood and plant remnants.

The LU Unit, with an average thickness of 10 m, crops out almost always as erosional windows due to post-depositional tectonic processes.

b) Allai Unit (AL)

The second eruptive event is represented by the large volume ignimbrite of Allai (AL). This ignimbrite is well marked in the landscape due to its particular characteristics, e.g., considerable thickness (40-80 m) and widespread cavernous erosional structures.

As a whole, the AL ignimbrite is massive, with rare laminar textures due to internal motions of the pyroclastic flow. The centimetric pumices are well visible in the matrix constituted by cusped glass shards and free crystals. In the matrix are evident the lithic fragments, both angular accidental clasts (belonging to Paleozoic metamorphic rocks and, more commonly, to Cenozoic andesites), and juvenile clasts with fringed margins.

The ignimbrite texture is vitroclastic with often corroded quartz, plagioclase, always characterized by polysynthetic twinning (according to the Albite law), zoned and with more calcic cores. The mafic mineral phases are constituted by biotite, as large flakes sometimes twisted and often altered, and less frequent and almost always altered brown hornblende.

Among all the volcanic units the AL Ignimbrite shows the largest area of emplacement; in fact, it crops out with continuity ca. 20/7 km. It also crops out to the north, more than 5 km NW of Samugheo.

The provenance of this pyroclastic flow was likely from the northern sectors as they seem to point out the decrease from north towards south of the amount of clastic and crystalline component, as well as the thickness that decreases in the same direction (Asorgia et al., 1995).

In the AL ignimbrite the 2b layer (Sparks et al., 1973) is massive at the bottom, while in its central part there is a grading probably due to flow rate changes during progressive aggradation (Branney & Kokelaar, 1992). Furthermore, in the central-basal part, the 2b layer is characterized locally by degassing pipes, evidenced by magmatic and metamorphic lithic fragments concentration arranged in pseudocylindric chimneys.

In the upper part of the ignimbrite body (layer 3 of Sparks et al., 1973), owing to the erosion occurred after its emplacement, ash-cloud surge deposit characterized by repeated centimetric laminations can be observed and contains pockets of accretionary lapilli.

Overlying the AL Unit a prevalently brown-reddish arenaceous-argillaceous deposit, attributable to the terrigenous continental formation (CPC) crops out. These sediments indicate similar morphological, tectonic and paleoclimatic conditions in time during the alluvial material accumulation. Nevertheless, this sedimentary episode differs from the precedent one with regard to argillaceous-arenaceous component predominance and better internal organization in more regular layers.

Such characteristics suggest that the source areas of these terrigenous deposits were more distal than those related to the precedent sediments, or that the new erosive event has been less intense.

c) Dacitic Lava Unit (DL)

Viscous dacitic magmas, (Pl+Cpx) building up lava-domes and small plateaus constituted by repeated lava flows, like those observable east of Samugheo rised from central vents, are aligned along fractures bordering the volcano-tectonic depression.

These lavas are massive and compact; sometimes they show conchoidal fracture due to post-depositional silicification processes. Even macroscop-



ically it is possible to observe a laminated texture in the ground-mass, due to flow bands and alignment of whitish, almost always zoned, plagioclases. The predominant mafic mineral phase is clinopyroxene, which sometimes forms aggregates with plagioclase.

d) *Terrigenous-Tuffaceous Unit (TT)*

The huge amount of erupted magma (LU, AL, DL), together with the extensional movements connected with the Miocene Sardinian Rift, probably accelerated the volcano-tectonic collapse; in these depressions, continental basins formed, wherein subsequently both terrigenous sediments and partly re-worked pyroclastic products accumulated.

In fact, overlying the Dacitic Lava Unit (DL) it is possible to observe a terrigenous-volcaniclastic sequence 30 m thick, constituted by mass-flow deposits, derived by resedimentation of the pyroclastic material in a fluvial-lacustrine environment. It is formed by siltitic-arenaceous-conglomeratic material organized in metric layers characterized by parallel and/or cross laminations and normal grading. The layers are laterally continuous for some hundred metres, but with frequent thickness and grain size variation, especially in the lower part where it is also possible to observe rare large clasts of 30-50 cm in diameter.

e) *Ruinias Unit (RU)*

The Ruinas Unit (RU) is a strongly welded dacitic (Pl+Cpx+Bt) lava-like ignimbrite up to 20 m thick; it is reddish-brown in its upper part due to thermal oxidation phenomena. Instead, at the bottom, the ignimbrite is grey-whitish, sometimes greenish, with lamination and unwelded consistency.

In this pyroclastic unit the juvenile fragments are pinky or greyish centimetric pumices, with generally aphyric textures and reaction rims; the lithics are cognate and brownish in color and centimetric in size, generally glassy, with aphyric and porphyritic textures due to the presence of plagioclase phenocrysts.

The rock shows a porphyroclastic texture and at the top of the unit the groundmass often exhibits a strongly devitrification and vapour phase crystallization.

The plagioclases phenocrysts are altered or generally zoned with a turbid core; the mafic mineral phases are generally broken clinopyroxenes and scarce small lamellae of biotite sometimes altered and twisted.

The lithic fragments are frequent and constituted by different lithologies among which volcanic rocks with a porphyritic texture, due to generally zoned and turbid plagioclases crystals, and with a vitreous blackish tachylitic groundmass, are abundant.

From the observed volcanological characteristics, it is possible to suppose that the emplacement mechanism of this volcanic unit is an overflow of foamy material fluidized due to the presence of magmatic gases responsible for the oxidation phenomena.

f) *Asuni Unit (AS)*

This unit is constituted by a sequence approximately 40 m thick of centimetric and decimetric layers of epiclastics and volcanoclastics deposited in fluvial-lacustrine environment. These deposits contain carbonised woods and plant remnants, sometimes enriched in particular layers.

The volcanoclastics show a matrix constituted by cusped glass shards, small collapsed pumices and broken plagioclase phenocrysts; within the sediment, sometimes fragmented in globules (fluidal-globular peperite), are embedded andesitic scoriaceous clasts constituted by dark tachylitic glass, in which plagioclase and clinopyroxene phenocrysts, sometimes with a glomeroporphyritic texture, are immersed. These peperitization structures have been referred by some authors (e.g., Mac Donald, 1939; Brooks *et al.*, 1982; Kokelaar, 1982; 1986) to interaction processes between magma and wet sediments.

This basic to intermediate volcanism, coeval to upper basic-intermediate lavic series (UBLS, see later), is represented in this area by subaerial small explosion pipes, whereas in Marmilla the same volcanic materials were erupted in submarine environment.

g) *Monte Ironi Unit (MI)*

Overlying the preceding unit, another pyroclastic flow unit, the Monte Ironi Unit (MI) crops out. It is about 10 m thick and in some sectors of the studied area it constitutes the uppermost part of the Aquitanian-Burdigalian volcano-sedimentary succession. This ignimbrite unit is orange in color and welded; in addition it shows an evident fluidal texture and fragmentation with frequent autobrecciation at its base.

It is characterized by a low porphyritic index underlined by the presence of euhedral plagioclase and scarce clinopyroxene phenocrysts. In the upper part of the unit the groundmass exhibits isotropic textures, and it is often characterized by devitrification processes of variable intensity, with spotted textures (patchy devitrification and lithophysae). Moreover, cognate fragments of maximum 2 mm in size are present.

2.2. *First marine sedimentary cycle: Upper Oligocene-Middle Burdigalian*

In the studied area the first marine sedimentary cycle probably started in the Upper Oligocene (Upper



Chattian) with the deposition of the Gesturi Sandstones, the heteropic Isili Limestones and the Ales Marls Formation (Cherchi, 1974) and it ended between the Middle Burdigalian and the Upper Burdigalian, according to the N6 and, maybe, N7 zone of planktonic foraminifera of Blow's Scale (1969) (Smit, 1974).

At the base of the Gesturi Sandstones, the sedimentary environment was fluvial-deltaic and fluvial-marine, and it evolves towards the top into a marine one. The fluvial-marine levels from Riu Su Rettori, south of the studied area, are older, Upper Oligocene in age, and they are interbedded with the continental sediments of the top of the Ussana Formation (Upper Oligocene-Early Aquitanian).

These deposits locally contain an abundant littoral-infralittoral fauna with *Tympanotonos margaritaceus* BROCCHI, *Turritella vermicularis* (BROCCHI), *Astrea carinata* (SACCO), *Mytilus haidingeri* HÖRN, *Gigantostrea gigantea oligoplana* SACCO, *Lucina multilamella* DESHAYES. Thus they correspond to the first inception of the first cycle marine ingression.

At the same time, in the eastern part of the studied area, along the Ottana-Allai-Nureci-Genoni alignment, continental conditions predominated (continental volcano-sedimentary succession), owing to the presence of extensive morpho-structural rises. As a consequence, the sediments of the first marine sedimentary cycle crop out beginning from Northeastern Marmilla.

On the basis of currently available data the oldest deposits lying at Nureci-Genoni are Lower-Middle Aquitanian in age, in correspondence to the medium part of the N4 zone of the Blow's Scale (1969), while the youngest deposits do not seem to be more recent than the Lower Burdigalian (Leone et al., 1984 and references).

Upon the morpho-structural rises, heteropic with the sediments of the Marmilla Formation, the volcanic products of the AL Unit would have been deposited, cropping out in the homonymous locality, at Villanovaforru and Ottana, as well as the LU, DL, RU and MI volcanic units (from the most ancient to recent).

Regarding the age of AL Unit, it is to note that the available radiometric data, as shown in Table, conflict in part with the biostratigraphic ones. In fact, the oldest marine sediments cropping out in the Genoni-Nureci area, and lying on the AL Unit, document a microfaunal association attributable to the N4 zone of Blow's Scale (1969) (Leone et al., 1984); while the radiometric data show a maximum age of 22.5 Ma and a minimum age of 19.3 Ma (Assorgia et al., 1995). On the other hand, as the age of $21.1 \pm$

1.4 Ma has been calculated on separated biotite, it is more probable that this datum is actually the more reliable among those available for the AL Unit, and then it approaches the biostratigraphic data.

Furthermore, in the studied area, the first marine cycle is represented by the sediments of the Ales Formation and chiefly by those of the Marmilla Formation. They are silici-clastic, mixed silici-clastic-carbonatic and tuffitic. The macrofauna, sometimes very abundant, belongs to shelf and epibathial environments and are characterized by benthic infralittoral-circalittoral and pelagic species (Pteropoda). Among the former predominate the Turritellidae: *Protoma cathedralis* (BRONGNIART), *Turritella turris* (BASTEROT), *T. terebralis* LAMARCK, the Pectinidae: *Aequipecten northamptoni* (MICHELOTTI), *Gigantopecten tournali* (DE SERRES), *Amussiopecten expansus* (SOWERBY), the Arcidae: *Anadara turonica* (DUJARDIN), *A. umbonaria* (MAYER), the Lucinidae: *Lucina multilamella* DESHAYES; while among the latter, the Pteropoda: *Vaginella austriaca* KITTL, *V. depressa* DAUDIN, *V. rotundata* BLANCKENHORN, *Clio triplicata* AUDENINO.

2.3. Second marine sedimentary cycle: Late Burdigalian-Upper Langhian

In the whole Island a stratigraphic gap marks the limit between the first and the second cycle of marine sedimentation. The pause in the sedimentation can be localized within the Middle-Upper Burdigalian since the first deposits of the second cycle overlying explosive volcanic units, as well as Chiaramonti in Northern Sardinia, Paulilatino (Odin et al., 1994) and Omodeo Lake area (Sa Manenzia Unit) in Central Sardinia, are Late Burdigalian (Cherchi, 1974). The volcanics from the subsurface of Porto Torres area (Melis, 1988) and the basal volcanics of the Calancui (Sassari) sequence (Pani, 1982), that constitute the base of the transgressive second cycle in most parts of Northwestern Sardinia, are also correlated.

In particular, Sa Manenzia Unit, evidenced for the first time, is an ash and pumice pyroclastic flow constituted by centimetric to decimetric pumiceous fragments within a groundmass of glass shards and free crystals of quartz, plagioclase and biotite. The cylindrical degassing pipe structures are frequent; the medium thickness is around 20 m and its provenance is from the northern sectors as evidenced by the degassing pipe orientation and the increase of the lithic amount.

In the Omodeo Lake area, concomitantly with the above-mentioned lack of marine sedimentation, this volcanic unit follows the fluvial sediments of the Sedilo Unit (Porcu, 1983); these latter contain a rich



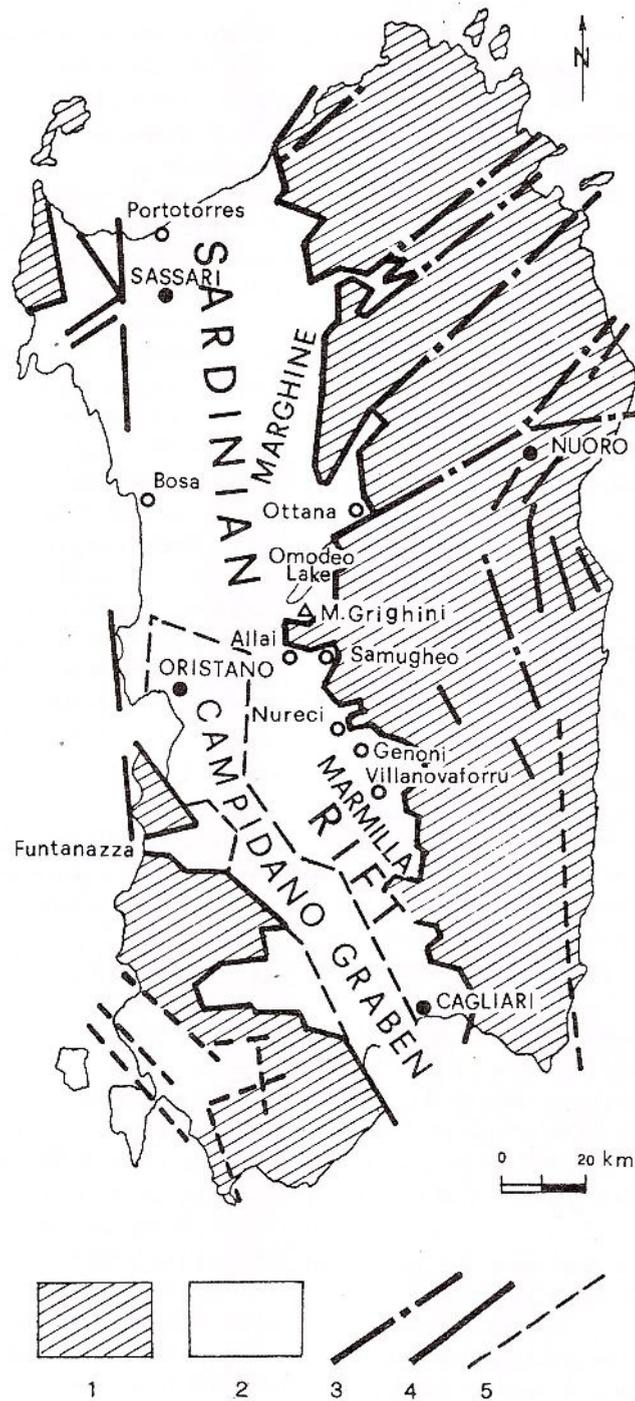


Fig. 3 – Schematic map of the Lower Miocene Sardinian Rift. 1, Paleozoic basement with Mesozoic and Paleogene covers; 2, Oligocene-Miocene and Plio-Quaternary sediments and volcanics; 3, Oligocene-Aquitania transcurrent faults; 4, Aquitanian-Burdigalian faults; 5, Pliocene-Pleistocene faults.

macroflora (i.e., "*Foresta di Zuri, Auct.*") represented by: *Arganioxylon sardum* BIONDI, *Sclorocaryoxylon chiarugii* BIONDI, *Palmoxylon cavallottii* LOVISATO & STERZEL, *Palmoxylon sardum* (UNGER) STERZEL, *Robinioxylon zuriensis* FALQUI, *Bombacoxylon owenii* (CARR.) GOTTWALD. (Charrier & Maxia, 1970; Biondi, 1981).

The rather widespread Dualchi sedimentary Formation overlies the Sa Manenzia Unit chiefly north of the Omodeo Lake. Essentially fluvial-lacustrine and fluvial-deltaic, at the top it shows transitional features to the overlying marine deposits. It probably contains the first marine pulsations of the second cycle in the studied area, maybe the whole of Sardinia. Thus, it can be correlated with the Lower Sands Formation from Sassarese (Mazzei & Oggiano, 1990).

According to Cherchi (1974) and Leone et al. (1984), in the studied area the beginning of the above-mentioned transgressive event is related to a span of time between Middle and Late Burdigalian. These authors assign the top of the Marls of the Gesturi Formation to this transgressive event. This formation is characterized by a monotonous alternation of marls and arenaceous-siltitic marls, with interbedded volcanoclastic deposits, containing marine microfauna. Within these sediments, in particular cropping out along the eastern part of the investigated area, two lithological sequences are recognizable, one marly-arenaceous and another tuffitic-marly-calcareous (Sanna, 1982; Casu, 1994; Serra, 1994).

These sequences are typical of deep circalittoral and epibathial environments owing to the constant and abundant presence of benthonic associations with: *Gigantopecten ziziniæ* (BLANCKENHORN), *Ostrea edulis lamellosa* BROCCHI, *Ostreinella neglecta* (MICHELOTTI), *Aequipekten submalvinæ* (BLANCKENHORN), *A. scabrellus* (LAMARCK), *Amusiopecten spinulosus* (MÜNSTER), *Lentipekten denutatum* (REUSS), *Propeamussium duodecimlamellatum* (BRONN), *Lucina orbicularis* DESHAYES, *Lucinoma borealis* LINNÉ, *Abra longicallus* (SCACCHI), *Turritella turris* (BASTEROT), *Xenophora infundibulum* (BROCCHI), *Schizaster desori* WRIGHT.

3. Tectonics and volcanism

3.1. Geodynamic events

The most important structural and tectonic modifications and the calc-alkaline volcanic activity displayed in this area during the Lower-Middle Miocene are commonly connected with the drift and rotation movements of the Sardinia-Corsica microplate (Alvarez, 1972). In particular, this volcanism is connected to subduction processes of oceanic crust along a Benioff plane dipping N or NW and localized east

of Sardinia (Beccaluva et al., 1987 and references). The geodynamic context of the magmatic cycle would have been dominated by the progressive collision between the African Plate and the European Plate which would have produced a rift system (Cherchi & Montadert, 1982) that evolved in a mixed lateral expulsion of the Sardinia-Corsica microplate (Tapponier, 1977). Within this microplate and upon the subduction plane, the Sardinian Rift (Fig. 3) would have built as an inter-arc extensional structure, constituted by different transtensional sub-basins.

According to Carmignani et al. (1994), the transcurrent tectonics of Central-Northern Sardinia (and of the Hercynian Corsica) would be considered as the effect of the deformation of the Sardinian-Corsican hinterland during the Oligocene-Aquitainian in the scenario of the Northern Appenninic collisional dynamics between the Apulian-Dinaric Plate and the South-European margin, of which Sardinia and Corsica were part before the drift and Burdigalian anticlockwise rotation (Montigny et al., 1981). In this hypothesis the extensional dynamics is linked to the Northern Appennine collapse during the Burdigalian that would have connected the opening of the Balearic Basin, the Sardinian Rift and the drift and rotation of the Sardinia-Corsica microplate.

In particular, in the eastern sector of the examined area, the normal faults related to the rifting phase, which displace the Mesozoic carbonatic formations about 200 m towards SW, are in part syneruptive because they displace only some calc-alkaline volcanic units. Furthermore, west of these displacements a thick and heterogeneous volcanic succession (5 volcanic units for a total compressive thickness of 150 m), crops out whose basement could not be observed, while in the east, only some thin volcanic units (AL, RU) occur on pre-Cenozoic basement; it could be supposed that the sector localized west of the Laconi-Samugheo alignment coincided with a morpho-structural depression during the Lower Miocene volcanic cycle.

Regarding the Late Burdigalian-Upper Langhian marine sedimentary formations, cropping out south of Ruinas and Asuni, these show displacements generally towards west and south due to normal fault systems linked to the extensional Pliocene-Quaternary tectonics of the Campidano Graben (Central-Southern Sardinia: Fig. 3), regarded in the framework of the opening of Southern Tyrrhenian basin, with associated widespread, mainly basaltic-alkaline volcanism.

3.2. Volcano-tectonic movements

Related to both compressional and extensional tectonic processes, which played a fundamental role



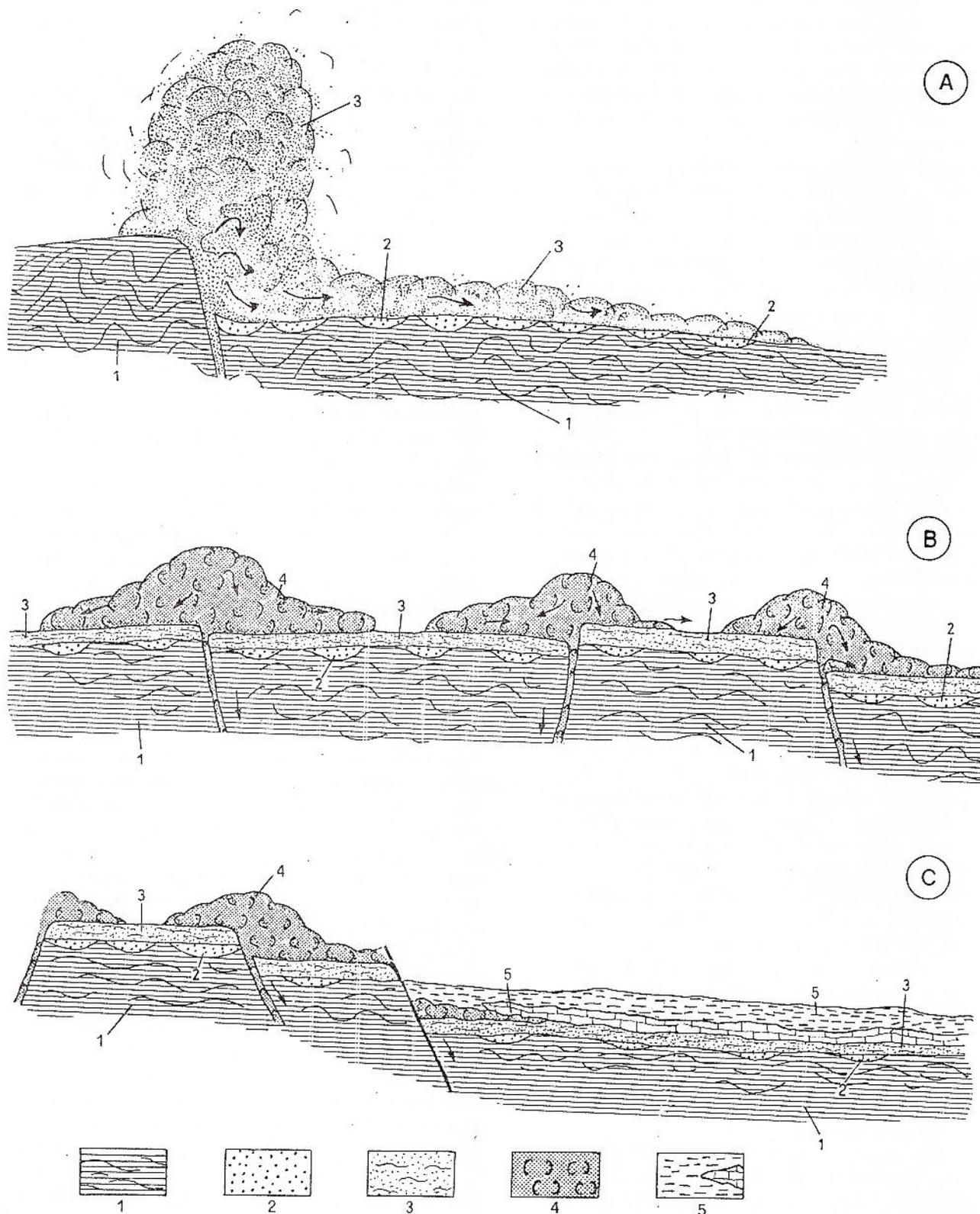


Fig. 4 - Sketch of the volcanic evolution of Central Sardinia during the Upper Oligocene-Lower Miocene time (see in the text). 1, Paleozoic basement; 2, Late Oligocene-Early Aquitanian Continental Polygenetic Conglomerates (CPC); 3, Ash and pumice pyroclastic flows (LU, AL); 4, Dacitic lava flow (DL) and rhyodacitic lava-like ignimbrites (RU, MI); 5, Miocenic marine sediments.

to promote volcanic activity, important volcano-tectonic movements are present in the studied area. They are connected with collapse phenomena as a consequence of large-volume ignimbrite formation of Allai (AL), the dacitic lava production (DL) and lava-like ignimbrite formation of Ruinas (RU) emplacement.

The existence of caldera structures is suggested by the presence of volcano-sedimentary successions, some hundred metres thick, which are deposited in generally lacustrine environments, and by the emplacement of viscous dacitic (DL) lavas as domes and small lava-flows along bordering fractures.

The Miocene volcanic activity evolution in this area could be synthesized as follows:

- As a consequence of the Luzzana and Allai Units emplacement, caldera collapse phenomena took place. They could be put in relation with an isostatic disequilibrium caused by the rapid escape of a large volume of magmatic material (about 300 km³) as ignimbrites (Fig. 4A);

- after the volcano-tectonic collapse rhyodacitic-dacitic lavas were erupted (Fig. 4B) along the volcano-tectonic depression fractures-bordering or along a ring-fracture system;

- a new collapse stage occurred after the Ruinas Unit (RU) emplacement. In the volcano-tectonic depression lacustrine clastic sediments about 40 m thick (Asuni Unit) accumulated;

- by the progression of the Sardinian Rift, these depressions have been displaced and submerged during the first marine transgression. At the same time, in the Marmilla area subaqueous basic to intermediate volcanic activity (basalts, basaltic andesites, andesites) occurred in facies of pillow-lavas, hyaloclastic breccias and hyaloclastites interbedded with marine sediments (Fig. 4C);

- finally, the last calc-alkaline event occurred during the second marine transgression, as the interbedded Langhian pumiceous tuffs testify.

4.2. Comparison with other Sardinian volcanic districts

Notoriously, only in the Logudoro-Bosano area (NW Sardinia) is present the complete sequence of the Sardinian "Oligocene-Miocene" volcanic cycle, whose total thickness is of about 1,500 m. In that area four main volcanic series are distinguishable (Deriu, 1962; 1964; Coulon, 1977); from bottom to top they have been redefined as follows (Guerrera et al., 1995):

- *Lower basic-intermediate lavic series (LBLS)*

Essentially constituted by basalt and andesite lava-flows, with associated subordinate explosion breccias. The beginning of this volcanic activity took place

33 Ma (cfr., Savelli et al., 1979; Beccaluva et al., 1985); however, some recent radiometric investigations tend to rejuvenate it (28-23 Ma, K. Balogh, personal comm.).

- *Lower acid-intermediate explosive series (LAES)*

Constituted both by highly explosive products (rhyolitic and rhyodacitic ash and pumice pyroclastic flows), and by extremely welded products (rhyolitic and rhyodacitic lava-like ignimbrites). To this series belong some large volume ignimbrites erupted in a span of time approximately of 3 Ma (e.g., Funtanazza Unit: 23 Ma; Allai Unit: 21 Ma; Bosa Unit: 20 Ma) (Tab.).

- *Upper basic-intermediate lavic series (UBLS)*

Essentially constituted by basalt and andesite lava-flows with associated subordinate explosion breccias emplaced partly in a submarine environment. The acme of this volcanic activity is placed around 18-20 Ma (cfr., Savelli et al., 1979; Beccaluva et al., 1985);

- *Upper acid-intermediate explosive series (UAES)*

Constituted both by highly explosive products (ash and pumice pyroclastic flows), and by products characterized by low eruptive mechanic energy, erupted as high temperature more or less foamy and overflow lavas (boiling over), and strongly welded products (lava-like ignimbrites). They are from dacitic to rhyolitic in chemism. To this series belong ash and pumice pyroclastic flows with an age comprised between 18 Ma, e.g., Chiaramonti, Paulilatino (Odin et al., 1994) and Sa Manenzia Unit and 15-16 Ma e.g., Acqua Sa Canna Unit, comenditic pyroclastic flows (Morra et al., 1994) and marine pumiceous pyroclastic levels (Pecorini, 1975).

As it occurred for other Sardinian districts (i.e., Sulcis area in the Southwestern Sardinia), only incomplete volcanic successions crop out in the studied area which can be correlated with the complete Logudoro-Bosano sequences chiefly using stratigraphic and chemical criteria.

In the studied Central Sardinia area no outcrops attributable to the LBLS series have been discovered; its existence even though is supposed in depth because of the presence of frequent magmatic lithic fragments within the AL Unit probably incorporated from underlying volcanics (Assorgia et al., 1995). The products (LU, AL, DL, RU, MI) belonging to the LAES series are predominant; however, according to Deriu (1962), Coulon (1977) and Savelli et al. (1979), they would be more recent (Aquitanian-Burdigalian) on the basis of the new radiometric ages (K/Ar) obtained on mineral phases (Bt and Pl) belonging to the Allai Unit (21±1.4 Ma and 20.3±1.0 Ma, Assorgia et al., 1995). Instead, to the UBLS series would belong the andesitic breccia pipes observed near Asuni and Mogorella, cropping out into Aquitanian-Burdigalian



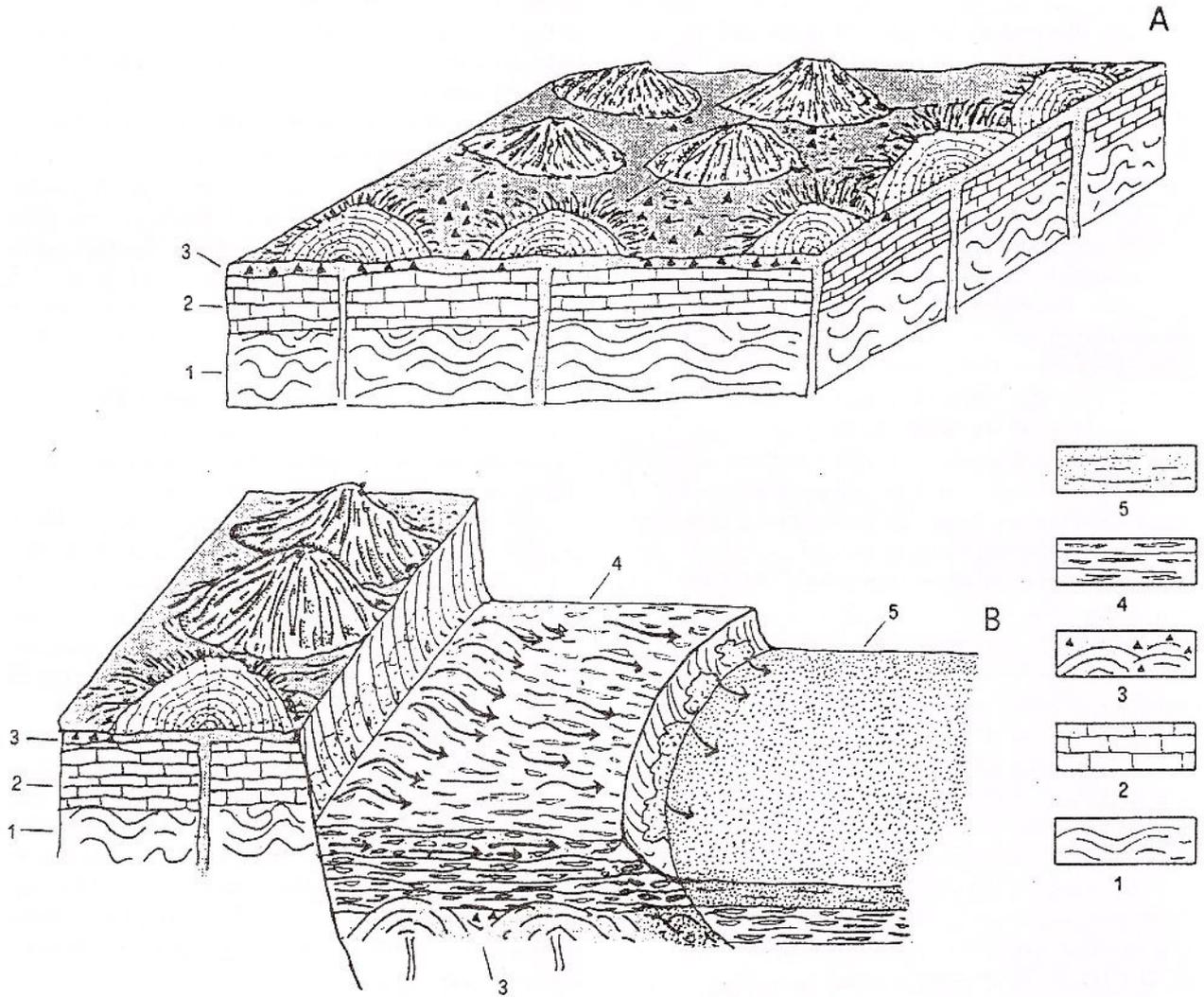


Fig. 5 – Evolutive schematic model of UBLs (A) and UAES (B) emplacement in SW Sardinia. 1, Paleozoic basement; 2, Paleogene locally also Mesozoic marine and continental sediments; 3, UBLs: basic to intermediate volcanics (breccias and lava-domes); 4, UAES: lava-like ignimbrites and pyroclastic flows; 5, UAES: small-volume pyroclastic flows and pyroclastic surges.

marine sediments. Finally, in the studied area, effusive products belonging to the UAES series are lacking.

The genetic relations between the basic-intermediate (UBLs-LBLS, generally effusive in character) and acid-intermediate (LAES-UAES, generally explosive in character) volcanic series from Sardinia are still not well defined, despite some contributions furnished by Coulon (1977), Coulon *et al.* (1978); Beccaluva *et al.* (1987). For the genesis of the acid-intermediate products an origin due to partial melting of lower granitic crust, generated by the heat-flow determined by the persistence in depth of ascending mantle-derived magmas (Coulon *et al.*, 1978; Becca-

luva *et al.*, 1987) has been supposed. However, the acid rocks may also be derived by differentiation processes and in particular by crystal/liquid fractionation of primary dacitic magmas (Morra *et al.*, 1994). Nevertheless, the crystal fractionation processes do not account for the large volumes of acid rock, and thus of magmas, involved during the eruptions.

The question about the observed alternation of the acid and basic volcanic activity, and in particular why the large volume ignimbrites (Allai Unit: Farris *et al.*, 1992; Assorgia *et al.*, 1995; Bosa Unit: Rizzo *et al.*, 1992; Assorgia *et al.*, 1994) are always the first magmatic events of a volcanic series linked to syneruptive tectonics leading to the development of a volcano-tectonic depression, still has not an ex-

haustive answer. It is probable, as demonstrated the experimental and theoretical studies of Huppert & Sparks (1988), that the produced heat-flow by a basic magma stored at the base of the crust can determine the partial melting of the surrounding crust according to the "melting roof theory" yielding a certain amount of silicic magma around the magmatic chamber.

A volcano-tectonic structural model regarding UBS and UAES relationships (Fig. 5) is given by the volcanic sequence from Sulcis (SW Sardinia). In this area, to a widespread basic-intermediate effusive volcanic activity follows an essentially explosive one, characterized earlier by welded pyroclastic flows and later by phreato-magmatic ash and pumice pyroclastic flows and surges emplaced in volcano-tectonic depressions.

5. Concluding remarks

The tectonic and volcanic evolution in the studied area from Central Sardinia during the Upper Oligocene-Middle Miocene time could be summarized as follows:

- During the Upper Oligocene-Aquitanean time thick conglomeratic continental deposits (CPC) originated due to erosion of raising Paleozoic reliefs as a consequence of transpressive movements likely connected to the Northern-Appenninic collision;

- during and after the continental sedimentation, in the Aquitanian-Burdigalian, a calc-alkaline volcanic activity occurred, essentially explosive and responsible for the emplacement of some rhyolitic-rhyodacitic ignimbrite units (LU-AL). In particular, the "Large Volume Ignimbrite of Allai", characterized by a large extension (certain surface of about 300 km², presumed surface of about 500 km²) and by a considerable thickness (40-80 m), constitutes an eruptive event that for its extension and volumes can be considered among the biggest ones occurred during the Cenozoic in the Mediterranean area;

- as a consequence of the emplacement of this *large-volume ignimbrite*, the studied area underwent important morphological-structural changes, due to the evolution of volcano-tectonic depressions;

- along fractures bordering these volcano-tectonic depressions acid-intermediate magmas raised yielding lava domes and lava-flows, while at the bottom of these depressions lacustrine sediments were deposited together with pyroclastic-epiclastic materials, 100 m thick;

- at about the same time, in the neighbouring areas probably submitted to transpressive movements, a first marine transgression occurred, which continued

in the Aquitanian up to Middle Burdigalian, extending further from littoral to epibathial facies;

- with the progress of the Sardinian Rift, probably in relation with the extensional dynamics connected to the gravitational collapse of the Northern Appennine belt, this area was subjected to a second marine transgression which developed at the end of the Burdigalian and continued in the Langhian giving rise to infra-circalittoral to epibathial sediments, with pumiceous pyroclastic intercalations, last evidence of the Miocene calc-alkaline volcanism;

- finally, a sharp discordance, marked by a deep erosional surface, separates the described Miocene volcano-sedimentary successions by the overlying basaltic lava-flows of the Plio-Pleistocene alkaline volcanism; the latter manifested as a consequence of extensional tectonics connected with the opening of the Southern Tyrrhenian Basin.

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BOOK REVIEW

László Trunkö. *Geology of Hungary*. Gebrüder Borntraeger, Berlin - Stuttgart, 1996. Hardcover, 464 pp. Price: 15 DM, ISBN 3-443-11023-1.

This synthesis on the geology of Hungary represents a concise and complete presentation of the stratigraphy and structure of the main structural units of Hungary. This new edition of a book first published in German in 1969 required an entirely new conception not only for the general ideas, but also for the details, in order to keep up with the tremendous amount of papers published during the last three decades which shed a new light on the tectonics, sedimentological and stratigraphic issues.

The geology of Hungary is treated in thirteen chapters. The book also includes a list of geographical names and one of stratigraphical names and major tectonic units and structures. The book is based on a large volume (47 pages) of references.

The almost generally accepted structural model of Hungary is a puzzle of plates of different origin, whose recent configuration was achieved in the Lower or Middle Miocene, whilst the beginning of the movements, though still debated, probably started in the Middle or Upper Cretaceous. There are three main units, separated by the Mid-Hungarian Lineament: Transdanubicum and Bükkium (Pelso Unit) to the North and Tisza Unit to the South. A few smaller units are scattered north of Transdanubicum and Bükkium: the Alpine Unit (regarded as part of the Alps, consists of a Penninic window near Koszeg, Lower Austroalpine series near Sopron and Upper Austroalpine - Graz Paleozoic - in the basement of the Little Plain), Vepor Unit, Zemplen Subunit. They are largely covered by Tertiary deposits.

The first eight chapters of the book present the stratigraphy and tectonics of the major structural units of the country: the Transdanubian Mid-Mountains, Igal Unit, Vepor Unit, the NE-Hungarian Mountains (Bükk and Rudabnya) and Tisia (Tisza) Unit.

A substantial chapter of the book is dedicated to the Cenozoic in Hungary, including detailed stratigraphy of the Paleogene, Neogene and Quaternary deposits, with their flora and fauna. A special chapter treats the tectonics of the Paleogene and Neogene basins.

Magmatism is grouped in pre-Variscan events, Variscan plutonism and volcanism, Triassic basic-intermediate volcanism, Cretaceous basic-alkaline volcanism in Tisia and Cretaceous subvolcanic rocks, the andesitic volcanism of the Eocene, the acidic-intermediate volcanism of the Miocene and the basaltic volcanism of the Pannonian.

Chapter 12 represents a general outline of the pre-Variscan, Variscan and Alpine tectonic events and of geotectonical conceptions.

A special chapter is dedicated to natural resources: ores (non-ferrous metals, iron-manganese-titanium ores, uranium and bauxite), sources of energy (crude oil and gas, coal, geothermal energy and thermal water) and nonmetallic raw materials.

This volume provides the reader new data and good, up-to-date reviews. It is therefore a good source and will be useful for all people interested in the region, which makes this volume worth having in personal or institutional libraries.

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CALCAREOUS NANNOPLANKTON AND MOLLUSCS FROM A FEW BADENIAN-SARMATIAN OCCURRENCES OF THE WESTERN TRANSYLVANIAN BASIN AND VIENNA BASIN

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Key words: Nannoplankton (NN5-NN9). Molluscs ("Neopycnodonte navicularis Zone", Rissoa Beds, Ervilia Beds). Badenian-Sarmatian *s.str.* Western Transylvanian Basin. Vienna Basin.

Abstract: A correlation of calcareous nannoplankton and molluscs of 6 sections from the Western Transylvanian Basin with 4 sections from the Vienna Basin is given. Two of the sections of the Vienna Basin are referential sections for the stratotypes: Baden-Sooss and Nexing, and the two other sections are now for the first time analysed from the nannoplankton viewpoint. Furthermore, the assemblage from Nexing is richer than previously known. From among 30 species of nannoplankton (the most frequent and representative ones), 20 species occur in both basins and 12 taxa have a large distribution throughout the entire Middle Miocene. There occur both guide-fossils and the assemblages specific to the zones: NN5 and NN6. The absence of guide-fossils for the NN7, NN8 and NN9 zones in both basins raises difficulties for a specification during the Sarmatian *s. str.* The NN5 is correlated with "Neopycnodonte navicularis Zone" (Livada, Borzești) (Lower Badenian). From among the 60 species of Sarmatian molluscs found in three sections: Aiton, Siebenhirten, Nexing, 12 species are common for both basins. The mollusc assemblages from Aiton present affinities with those from Siebenhirten (Rissoa Beds) and also Nexing (Ervilia Beds).

A correlation of calcareous nannoplankton and molluscs from several occurrences of the Badenian and Sarmatian deposits from the Western Transylvanian Basin and Vienna Basin is given.

From the Western Transylvanian Basin, the sections studied are from Livada (Valea Mare), Borzești (Porcărei Brook-Costița Hill, Gruicul and Săliște Hill) (Mészáros et al., 1992), Cheia quarry (Ghergari et al., 1991), Aiton (Valea Cioltului), Ceanu Mic (Valea Cheița) and Turda-quarry (Fig. 2).

And as far as the Vienna Basin is concerned there were analysed the sections from Siebenhirten and St. Margarethen Kiesgrube, and reanalysed the sections from the holostatotypes of the Badenian: Baden-Sooss and Sarmatian: Nexing (Fig. 2).

In the Western Transylvanian Basin, similarly to the Vienna Basin, the Badenian deposits of the littoral and shallow-water facies are developed as conglomerate, Leitha Limestone (lithothamnium limestones - typically developed and described in the Leithagebirge) and calcareous marls. There are also sediments of the deep-water facies with clay and intercalated sands (in the surrounding of Baden, in the

surroundings of Baden, in the Borzești Basin in the three mentioned sites and in the Livada Basin).

At Baden-Sooss, as known, the main part of the large clay pit consists of Baden Tegel. It represents a grey-blue plastic clay with intercalated sandy layers and lenses (recent description in W. Piller & N. Vavra, 1991).

The calcareous nannoplankton belongs to NN5 - Sphenolithus heteromorphus Zone (Fuchs, 1978).

The mollusc fauna (gastropods, bivalves and scaphopods) is characterized by infaunal elements. The most frequent ones are the species of *Turritella*, *Cassidaria*, *Semicassis*, *Fusus*, *Mitra*, *Conus*, *Ringicula*, *Corbula*, *Chlamys*, *Glycymeris*, *Amussium*, *Pinna* (O. Schultz, F. Steininger and F. Stojaspal, 1978); *Neopycnodonte navicularis* is also present.

At Livada (Valea Mare) the basal conglomerate is overlain by the Dej Tuff and by marls with *Neopycnodonte navicularis*, which bear intercalations of sands and tuffites. The nannoplankton assemblage corresponds to NN5 - Sphenolithus heteromorphus Zone. The assemblage contains: *Coccolithus miopelagicus*,



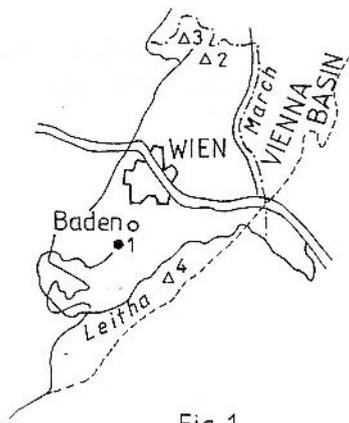


Fig. 1



Fig. 2

Location map of the studied occurrences from the Transylvanian Basin (Fig. 2) and Vienna Basin (Fig. 1)

- Badenian occurrences: 1, Baden Sooss; 1-Livada; 2-Borzești; 3-Cheia
- ▲ Sarmatian occurrences: 2-Nexing; 3-Siebenhirten; 4-St.Margarethen Kiesgrube; 4-Turda, 5-Aiton; 6-Ceanu Mic

C. pelagicus, *S. neoabiens*; *Sphenolithus moriformis*, *Discoaster variabilis*, *D. musicus*, *D. exilis*, *Cyclococcolithus rotula*, *Calcidiscus leptoporus*, *C. macintyreii*, *Helicosphaera carteri*, *Reticulofenestra pseudumbilicus*, *Braarudosphaera bigelowii* etc. (Pl. I, Figs. 1-7).

The Badenian sediments occur in the Borzești Basin in three sites.

On the Porcărei Brook near the Costița Hill the Badenian deposits begin with a polymictic conglomerate with calcareous matrix. The overlying formation contains calcareous clay and marls. The marls are rich in *Neopycnodonte navicularis* (BROCCHI) (Pl. II, Figs. 1-3) and subspecies: *Neopycnodonte navicularis henei* (NYST). A level of vitroclastic bentonized tuff is interlayered in these sediments.

The second site (south-west of the Borzești-Gruiu village) has a basal level of marls, followed by marly tuffite and marls bearing *Neopycnodonte navicularis* and covered again by marly tuffite.

In the third Miocene site, north-east of the Borzești village (Săliște Hill), the deposits begin with marly tuffite, marls and calcareous deposits represented by micritic pelletal limestone and biomicritic limestone bearing diagenised spicules of sponges. The spongy limestone has preserved juvenile specimens of burrowing bivalves. A few meters above the limestone lies an alternation of marls and sands bearing scarce specimens of *Limacina* sp. (= *Spiratella*). Gábos (1974) has described from Borzești five species of *Spiratella*: *S. andrussowi andrussowi* KITTL, *S. hospes* (ROLLE), *S. cf. stenogyra* (PHILIPPI), *S. subtarchanensis* ZHIZCENKO, *S. koeneni* (KITTL).

The nannoplankton assemblages correspond to the first two sites to NN5 - *Sphenolithus heteromorphus* Zone.

In the upper part of the last Badenian site, due to the absence of the guide-fossils, the age cannot be specified on the basis of the nannoplankton, yet, there occur the mentioned forms of *Spiratella*.

The *Neopycnodonte navicularis* Zone is recorded in many outcrops in Romania at the Lower Badenian level. Nearby our recorded area, Lubenescu et al. (1978) have recorded this zone in the Copăceni-Tureni area.

Neopycnodonte navicularis (BROCCHI) is considered (Freneix, 1975) to be a paleontological indicator of temperate and euhaline environment for the circalittoral and bathial levels.

The Badenian sediments begin in the Cheia-quarry area from a conglomeratic level overlain by a gypsiferous facies (15-20 m thick), then a marly and clayey level which bears a few calcareous intercalations and a pyroclastic sequence. The nannoplankton assemblages from the upper part of the gypsum complex belong to NN5 Zone. The whole assemblage was considered to have a Middle Badenian character (Wielićian) (Ghergari et al., 1991).

The upper part of the section belongs to NN6 - *Discoaster exilis* Zone (Upper Badenian-Kossovian). They are also reworked forms from Lower Badenian, even from Priabonian. During the Middle and Upper Badenian, the basin lowered even more, and the post-gypsiferous formation bearing pyroclastics was formed.

The statistical approach to the basic character of



the seven levels of pyroclastics in this formation does not allow their correlation with the other tuffs from the Transylvanian Basin (Ghergari et al., 1991) and, it has been estimated that they belong to local volcanic sources, some with underwater activities.

The following species are also mentioned in the calcareous nannoplankton assemblage from the upper part of the quarry: *Calcidiscus pataecus*; the abundant development of the species *Calcidiscus leptoporus* and *C. macintyreii* and the lack of discoasterids. Although the assemblage does not contain calcareous nannoplankton typical of a certain zone, it is supposed to belong to NN6 Zone (Mărunțeanu in Popescu, Mărunțeanu et al., 1995).

The Sarmatian holostatotype from Nexing contains lumachelle sands and marls with molluscs described by Papp (1954) (Upper Ervilia Beds).

The calcareous nannoplankton was studied by Stradner & Fuchs (1979). They mentioned only rare forms of *Coccolithus pelagicus* and reworked forms from the Cretaceous.

In the samples studied now there are also: *Coccolithus miopelagicus*, *Calcidiscus leptoporus*, *Reticulofenestra pseudoumbilicus*, *Braarudosphaera bigelowii*, *Helicosphaera carteri*.

Within the facies stratotype from Siebenhirten there have been described deposits specific to for the sublittoral area-sands with very frequent *Pirenella*. The Lower Sarmatian deposits occur overlying gravels which belong to the Badenian. These Sarmatian deposits are represented by marls which include in the upper part sand lenses with *Pirenella*. Papp (1954) mentioned the deposits from Siebenhirten and also presented a list of the molluscs retrieved there. Within marls the prevalence of the *Mohrenstermia*, *Pseudammicola* and *Cerastoderma* forms is obvious, and for the sandy sequences that of *Pirenella*, *Cerithium* and *Calliostoma*.

The list of calcareous nannoplankton studied by me contains: *Helicosphaera carteri*, *H. minuta*, *H. walbersdorfensis*, *Discoaster variabilis*, *Reticulofenestra pseudoumbilicus*, *Umbilicosphaera jafari*, *Coccolithus pelagicus*, *C. miopelagicus*, *Calcidiscus leptoporus*, *Rhabdosphaera procera*, *R. sicca*, *Sphenolithus compactus* etc. and reworked forms from Cretaceous and Badenian.

At St. Margarethen Kiesgrube in the sand and marls with intercalated gravels the following taxa are very frequent: *Mactra*, *Pirenella*, *Bittium hartbergense* (Upper Sarmatian-Mactra Beds).

Calcareous nannoplankton is totally absent in some samples, in others there rarely occurs *Coccolithus pelagicus*.

The studied section from Aiton (Valea Cioltului) contains marls and calcareous clays with intercalated

sands and tuffs and microconglomerates with the richest Sarmatian mollusc fauna assemblages from the western border of the Transylvanian Basin (north of the Arieș Valley and Turda, nearby Cluj). These assemblages present affinities with those from Siebenhirten (Rissoa Beds) and also from Nexing (Ervilia Beds) (Pl. II, Figs. 4-17; Pl. III, Figs. 1-4).

The calcareous nannoplankton assemblages are represented by: *Coccolithus miopelagicus*, *C. pelagicus*, *Calcidiscus leptoporus*, *C. macintyreii*, *Cyclargothus floridanus*, *Discoaster variabilis*, *Braarudosphaera bigelowii*, *Reticulofenestra pseudoumbilicus*, *Pontosphaera multipora*, *Helicosphaera carteri*, *Sphenolithus neoabies* etc. (Pl. I, Figs. 8-13) and reworked forms from Cretaceous, Paleogene and Badenian.

The deposits from Valea Cheița-Ceanu Mic are represented by marls, clays and sands with thin tuff intercalations. There also occur microconglomerates with mollusc casts.

The nannoplankton assemblages contain: *Coccolithus miopelagicus*, *C. pelagicus*, *Calcidiscus macintyreii*, *Discoaster musicus*, *D. variabilis*, *D. exilis*, *Syracosphaera hystrica*, *Umbilicosphaera jafari*, *Rhabdosphaera procera*, *Sphenolithus abies*, *Braarudosphaera bigelowii*.

In the Turda-quarry marls occur, and in the upper part of the succession a tuff level, followed again by marls. The molluscs are represented by rare forms of *Ervilia*. The nannoplankton is represented by: *Coccolithus pelagicus*, *C. miopelagicus*, *Calcidiscus macintyreii*, *C. leptoporus*, *Helicosphaera carteri*, *H. minuta*, *Discoaster variabilis*, *Braarudosphaera bigelowii*, *Pontosphaera multipora*, *Umbilicosphaera jafari*, *Rhabdosphaera procera*, *R. sicca* etc. and reworked forms from Cretaceous and Paleogene.

Lack of useful guide-fossils for the Sarmatian *s.str.* precludes the correlation with the Standard Nannoplankton Zonation, as it was also remarked for Nexing and for other occurrences from Austria, then those mentioned here, by Stradner & Fuchs, 1979. The nannoplankton from the Sarmatian *s.str.* gives evidence of reduced salinity and near-coastal environment.

The deposits south of Borzești as well as those from Livada and Cheia are lithostratigraphically assigned to the Ciccu-Giurgești Formation (Lower Badenian-Langhian) (Popescu, 1970) (conglomerates and clays) and to the Dej Formation (Complex of the Dej Tuffs, Popescu, 1970) which includes tuff, tuffites, marls and clays. The chemical precipitation rocks have been set into the frame of the Ocna Dejului Member included by Popescu (1972) in the Mireș Formation just as the Radiolarian Shales and Spirialis Marls, found north from Borzești (Săliște Hill) and Cheia



CORRELATION OF CALCAREOUS NANNOPLANKTON AND MOLLUSCS AND STRATIGRAPHICAL DISTRIBUTION OF SECTIONS

| MARINE BIOCHRONOLOGY | | MARINE CHRONOSTRATIGRAPHY | | | SECTIONS STUDIED | |
|--------------------------|---|---------------------------|---------------|----------------------|---|--|
| CALCAREOUS NANNOPLANKTON | MOLLUSCS | SERIES | STAGES | | Western Transylvanian Basin | Vienna Basin |
| | | | Mediterranean | Central Paratethys | | |
| Martini & Müller 1986 | Papp 1954 Lubenescu & al. 1978 Moiescu & Popescu 1980 | UPPER | SERRAVALLIAN | SARMATIAN s. str. | ALTON Cioabui Valley CEANU MIC Cheia Valley TURDA Quarry | BADEN-SOOSS Quarry SIEBENHIRTEN NEXING ST. MARGARETHEN-KIESGRUBE |
| NN9 | | | | | | |
| NN8 | <i>Mactra</i> Beds <i>Ervilla</i> Beds | MIDDLE | LANGHIAN | BADENIAN | ALTON Cioabui Valley CEANU MIC Cheia Valley TURDA Quarry | BADEN-SOOSS Quarry SIEBENHIRTEN NEXING ST. MARGARETHEN-KIESGRUBE |
| NN7 | <i>Rissoa</i> Beds | | | | | |
| NN6 | | LOWER | LANGHIAN | BADENIAN | ALTON Cioabui Valley CEANU MIC Cheia Valley TURDA Quarry | BADEN-SOOSS Quarry SIEBENHIRTEN NEXING ST. MARGARETHEN-KIESGRUBE |
| NN5 | <i>Neopycnodonte navicularis</i> Zone | | | | | |

partly after the correlation chart for marine / continental Neogene chronology of Europe (F.F. Steininger, R.L. Bernor et V. Fahibusch, 1990)

Fig. 3

| TABLE 1. CALCAREOUS NANNOPLANKTON | | ZONES | |
|--|--|-------|-----------------|
| DISTRIBUTION (— common species for both basins) (- - - only in western Transylvanian Basin) (- - - - only in Vienna Basin) | | CN4 | CN5 |
| | | CN6 | CN7 |
| | | NN5 | NN6 NN7 NN8 NN9 |
| 1. <i>Sphenolithus heteromorphus</i> DEFLANDRE | | | |
| 2. <i>Discoaster formosus</i> MARTINI & WORSLEY | | | |
| 3. <i>Discoaster musicus</i> STRADNER | | | |
| 4. <i>Cyclicargolithus abisectus</i> MULLER | | | |
| 5. <i>Cyclicargolithus floridanus</i> (ROTH & HAY) BUKRY | | | |
| 6. <i>Discoaster trinidadensis</i> HAY | | | |
| 7. <i>Coccolithus miopelagicus</i> BUKRY | | | |
| 8. <i>Coronocylus nifescens</i> (KAMPTNER) BRAMLETTE & WILCOXON | | | |
| 9. <i>Pontosphaera multipora</i> KAMPTNER | | | |
| 10. <i>Discoaster exilis</i> MARTINI & BRAMLETTE | | | |
| 11. <i>Discoaster variabilis</i> MARTINI & BRAMLETTE | | | |
| 12. <i>Braarudosphaera bigelowii</i> DEFLANDRE | | | |
| 13. <i>Sphenolithus abies</i> DEFLANDRE | | | |
| 14. <i>Sphenolithus neobabes</i> BUKRY & BRAMLETTE | | | |
| 15. <i>Sphenolithus compactus</i> BACKMAN | | | |
| 16. <i>Helicosphaera carteri</i> (WALLICH) KAMPTNER | | | |
| 17. <i>Helicosphaera minuta</i> (MULLER) | | | |
| 18. <i>Reticulofenestra pseudoumbilicus</i> (GARTNER) | | | |
| 19. <i>Coccolithus pelagicus</i> (WALLICH) SCHILLER | | | |
| 20. <i>Calcidiscus leptoporus</i> (MURRAY & BLACKMAN) | | | |
| 21. <i>Cyclococcolithus rotula</i> (KAMPTNER) KAMPTNER | | | |
| 22. <i>Calcidiscus macintyreii</i> BUKRY & BRAMLETTE | | | |
| 23. <i>Syracosphaera histrica</i> KAMPTNER | | | |
| 24. <i>Discoaster aulakos</i> GARTNER | | | |
| 25. <i>Discoaster brouweri</i> TAN EMENDED BRAMLETTE & RIEDEL | | | |
| 26. <i>Helicosphaera philippinensis</i> MULLER | | | |
| 27. <i>Helicosphaera walbersdorffensis</i> MULLER | | | |
| 28. <i>Rhabdosphaera sicca</i> (STRADNER) FUCHS & STRADNER | | | |
| 29. <i>Rhabdosphaera procera</i> MARTINI | | | |
| 30. <i>Umbilicosphaera jafari</i> MULLER | | | |



TABLE 2. STRATIGRAPHICAL DISTRIBUTION OF SARMATIAN MOLLUSCS FROM THE WESTERN TRANSYLVANIAN BASIN: AITON (N. MÉSZÁROS, E. NICORICI 1962; C. CHIRA 1995) AND VIENNA BASIN: SIEBENHIRTEN (A. PAPP 1954) AND NEXING (A. PAPP 1954)

| SARMATIAN MOLLUSCS | Mohrensternia Beds | Ervilia Beds | | Mactra Beds |
|---|--------------------|--------------|-------|-------------|
| | | Lower | Upper | |
| 1. <i>Mohrensternia inflata inflata</i> (M. HOERNES) | ----- | | | |
| 2. <i>Mohrensternia angulata</i> (EICHWALD) | ----- | | | |
| 3. <i>Mohrensternia hydroboides</i> (HILBER) | ----- | | | |
| 4. <i>Mohrensternia benatica</i> JEKELIUS | ----- | | | |
| 5. <i>Hydrobia stagnalis andrussowi</i> HILBER | ----- | | | |
| 6. <i>Lunatia catena sarmatica</i> (PAPP) | ----- | | | |
| 7. <i>Calliostoma (Feneoniana) angulatum angulatum</i> (EICHWALD) | ----- | | | |
| 8. <i>Calliostoma (Feneoniana) guttenbergi</i> (HILBER) | ----- | | | |
| 9. <i>Calliostoma (Feneoniana) styriacum</i> (HILBER) | ----- | | | |
| 10. <i>Stenothyrella schwartzi</i> (FRAUENFELD) | ----- | | | |
| 11. <i>Cerastoderma pseudopicatum</i> (FRIEDBERG) | ----- | | | |
| 12. <i>Cerastoderma janoschecki</i> (PAPP) | ----- | | | |
| 13. <i>Abra reflexa</i> (EICHWALD) | ----- | | | |
| 14. <i>Cerithium (Theridium) rubiginosum rubiginosum</i> (EICHWALD) | ----- | | | |
| 15. <i>Cerithium (Theridium) rubiginosum prahovensis</i> SIMIONESCU | ----- | | | |
| 16. <i>Acteocina pupa</i> EICHWALD | ----- | | | |
| 17. <i>Clithon (Vitellithon) pictus pictus</i> (FERUSSAC) | ----- | | | |
| 18. <i>Modiolus (Modiolus) incrassatus</i> (D'ORBIGNY) | ----- | | | |
| 19. <i>Musculus (Musculus) sarmaticus</i> (GATUJEV) | ----- | | | |
| 20. <i>Mactra eichwaldi</i> LASKAREV | ----- | | | |
| 21. <i>Mactra cf. naviculata</i> BAILY | ----- | | | |
| 22. <i>Mactra cf. venjukovi</i> ANDRUSOV | ----- | | | |
| 23. <i>Pseudamnicola (Staja) inflata</i> (JEKELIUS) | ----- | | | |
| 24. <i>Pseudamnicola (Staja) immutata</i> (M. HOERNES) | ----- | | | |
| 25. <i>Pseudamnicola tholosa</i> JEKELIUS | ----- | | | |
| 26. <i>Pirenella picta picta</i> DEFRANCE | ----- | | | |
| 27. <i>Pirenella picta mitralis</i> (EICHWALD) | ----- | | | |
| 28. <i>Pirenella picta nympha</i> (EICHWALD) | ----- | | | |
| 29. <i>Dorsanum duplicatum duplicatum</i> (SOWERBY) | ----- | | | |
| 30. <i>Acteocina lejnonkaireana lejnonkaireana</i> (BASTEROT) | ----- | | | |

(----- common species for both basins)

(----- only in western Transylvanian Basin)

(----- only in Vienna Basin)

TABLE 2. (continued)

| SARMATIAN MOLLUSCS | Mohrensternia Beds | Ervilia Beds | | Maetra Beds |
|---|--------------------|--------------|-------|-------------|
| | | Lower | Upper | |
| 31. <i>Ocenebrina sublavata striata</i> (EICHWALD) | ----- | | | ----- |
| 32. <i>Melanopsis impressa impressa</i> KRAUSS | ----- | | | ----- |
| 33. <i>Ervilia dissita dissita</i> (EICHWALD) | ----- | | | ----- |
| 34. <i>Hydrobia elongata elongata</i> (EICHWALD) | ----- | | | ----- |
| 35. <i>Pirenella disjuncta disjuncta</i> (SOWERBY) | ----- | | | ----- |
| 36. <i>Dorsanum opinabile trabale</i> KOLESNIKOV | ----- | | | ----- |
| 37. <i>Donax lucidus</i> EICHWALD | ----- | | | ----- |
| 38. <i>Gibbula balatro</i> EICHWALD | ----- | | | ----- |
| 39. <i>Gibbula subbalatro</i> KOLESNIKOV | ----- | | | ----- |
| 40. <i>Gibbula bessarabica</i> SINZOW | ----- | | | ----- |
| 41. <i>Gibbula papilla</i> EICHWALD | ----- | | | ----- |
| 42. <i>Gyraulus</i> sp. | ----- | | | ----- |
| 43. <i>Modiolus subpapilio</i> GRATELOUP | ----- | | | ----- |
| 44. <i>Cerastoderma obsoletum</i> (EICHWALD) | ----- | | | ----- |
| 45. <i>Irus gregarius gregarius</i> (PARTSCH-GOLDFUSS) | ----- | | | ----- |
| 46. <i>Ervilia dissita macrodon</i> ANDRUSOV | ----- | | | ----- |
| 47. <i>Donax dentiger eichwaldi</i> (EICHWALD) | ----- | | | ----- |
| 48. <i>Acteocina lajonkairaana sinzowi</i> KOLESNIKOV | ----- | | | ----- |
| 49. <i>Calliostoma (Sarmates) podolicum enodis</i> (TOULA) | ----- | | | ----- |
| 50. <i>Gibbula (Rollandia) picta</i> (EICHWALD) | ----- | | | ----- |
| 51. <i>Theodoxus (Theodoxus) crenulatus crenulatus</i> (KLEIN) | ----- | | | ----- |
| 52. <i>Cepaea gottschicki</i> (WENZ) | ----- | | | ----- |
| 53. <i>Cerastoderma vindobonense vindobonense</i> (PARTSCH-LASKAREV) | ----- | | | ----- |
| 54. <i>Cerastoderma latisulcum latisulcum</i> (MUNSTER) | ----- | | | ----- |
| 55. <i>Cerastoderma latisulcum jammense</i> (HILBER) | ----- | | | ----- |
| 56. <i>Cerastoderma latisulcum nexingense</i> (PAPP) | ----- | | | ----- |
| 57. <i>Cerastoderma ghergutai</i> (JEKELIUS) | ----- | | | ----- |
| 58. <i>Ervilia dissita podolica</i> (EICHWALD) | ----- | | | ----- |
| 59. <i>Calliostoma (Feneoniana) orbignyanum orbignyanum</i> (HOERNES) | ----- | | | ----- |
| 60. <i>Blittium hartbergense hartbergense</i> (HILBER) | ----- | | | ----- |

(—) common species for both basins
 (---) only in western Transylvanian Basin
 (- - -) only in Vienna Basin

quarry (NN6-Zone). The Sarmatian deposits are included in the Feleac Formation and are represented by sands, clays, sandstones, conglomerates (Aiton, Ceanu Mic, Turda-quarry). The sands contain convolute concretions named: "concrețiuni de Feleac"; present at Aiton.

In conclusion, this study presents a correlation of 6 sections from the western part of the Transylvanian Basin with 4 sections from the Vienna Basin, based on nannoplankton and molluscs.

Figure 3 indicates the stratigraphical distribution for the 10 sections. Two of the sections from the Vienna Basin are referential sections, for the stratotypes Badenian-Baden-Sooss and Sarmatian-Nexing, and the two other sections from Siebenhirten and St. Margarethen Kiesgrube are now for the first time analysed from the nannoplankton viewpoint. Furthermore, as we have already mentioned, the assemblage from Nexing is richer than previously known.

In Table 1, from among the 30 species of nannoplankton presented (the most frequent and representative ones) out of the studied sections in the western part of the Transylvanian Basin and Vienna Basin, 20 species occur in both basins, 6 species are mentioned only from the Transylvanian Basin and 4 species only from the Vienna Basin. At the same time, 11 taxa have a large distribution through the entire Middle Miocene.

There occur both guide-fossils and the assemblages specific to the zones: NN5 - *Sphenolithus heteromorphus* and NN6 - *Discoaster exilis*. The absence of guide-fossils for the NN7, NN8 and NN9, in both basins raises difficulties for a specification during the Sarmatian *s. str.*

In Table 2, from among the 60 species of Sarmatian molluscs found in three sections: Aiton (Western Transylvanian Basin), Siebenhirten and Nexing (Vienna Basin), 12 species are common for both basins, 20 species are mentioned only from Aiton, and 28 species only from the Vienna Basin. Three of them are marine relics: *Lunatia catena sarmatica* (PAPP), *Stenothyrella schwartzi* (FRAUENFELD), *Clithon (Vitolithon) pictus pictus* (FERUSSAC).

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Plate I

Figs. 1, 2 — *Sphenolithus heteromorphus* - Livada.

Figs. 3, 4 — *Sphenolithus neoabies* - Livada.

Fig. 5 — *Discoaster exilis* - Livada.

Figs. 6, 7 — *Discoaster variabilis*, *Coccolithus pelagicus* - Livada.

Figs. 8, 9 — *Cyclicargolithus floridanus*, *Helicosphaera carteri*, *Coccolithus pelagicus* - Aiton.

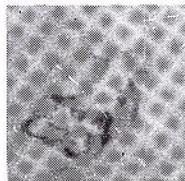
Figs. 10, 11 — *Calcidiscus macintyreii* - Aiton.

Figs. 12, 13 — *Braarudosphaera bigelowii* - Aiton.

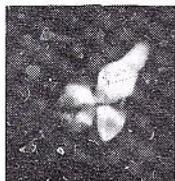




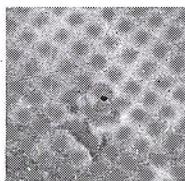
C. CHIRA CALCAREOUS NANNOPLANKTON AND MOLLUSCS



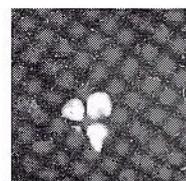
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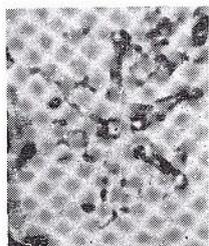
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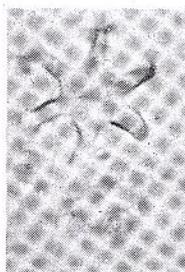
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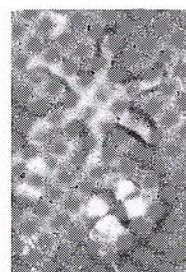
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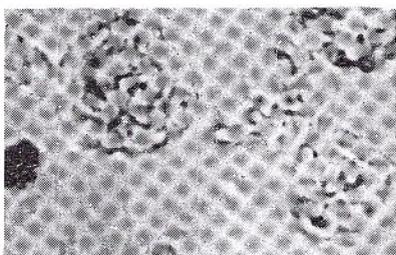
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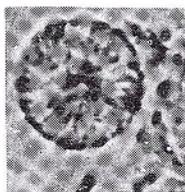
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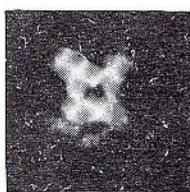
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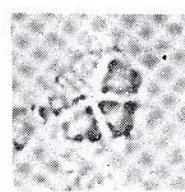
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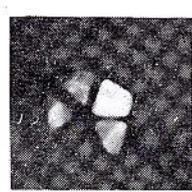
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Plate II

Figs. 1, 2 — *Neopycnodonte navicularis* BROCCHI Borzești (L.V.) (x 1)

Fig. 3 — *Neopycnodonte navicularis* BROCCHI Borzești (L.V.) (x 1)

Figs. 4, 6, 7, 9, 10, 13, 14 — *Ervilia dissita dissita* AITON

Figs. 5, 8, 12 — *Ervilia dissita podolica* AITON

Fig. 11 — *Ervilia dissita crassa* AITON

Figs. 15, 16 — *Cerastoderma vindobonense vindobonense* AITON

Fig. 17 — *Mohrensternia inflata inflata* AITON



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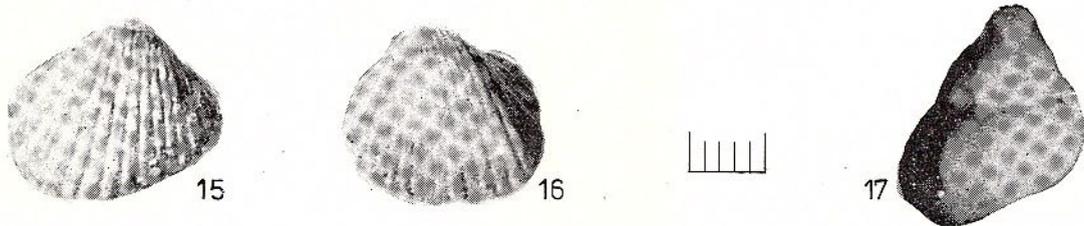
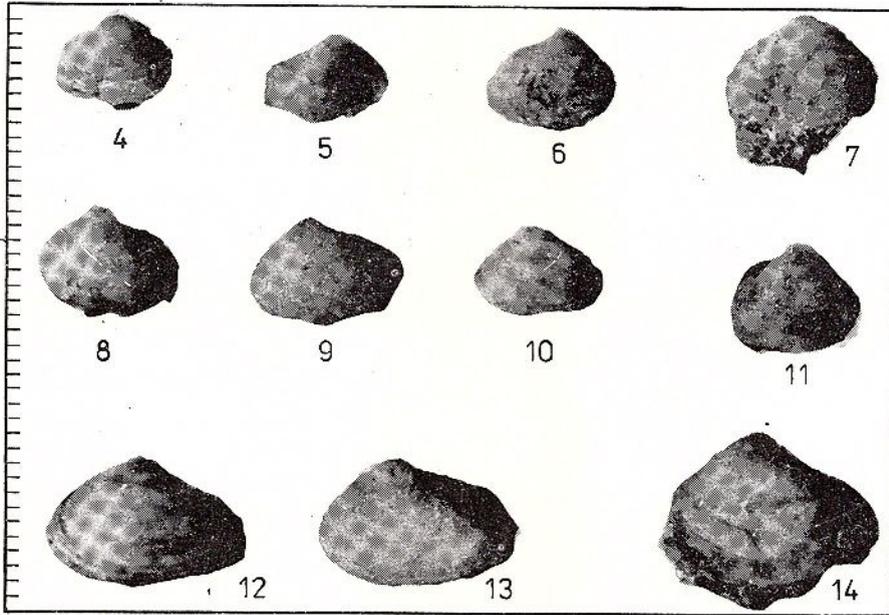
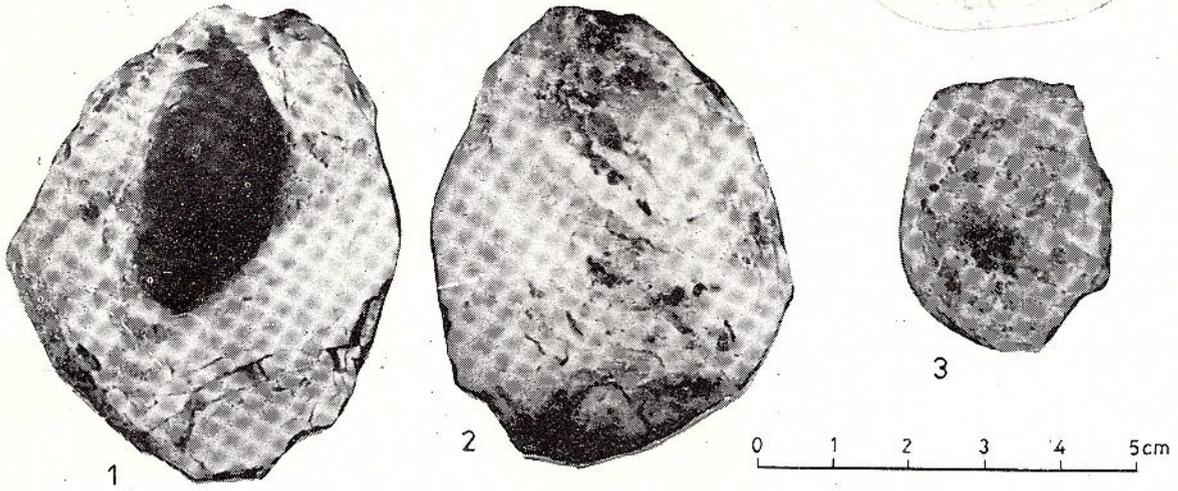


Plate III

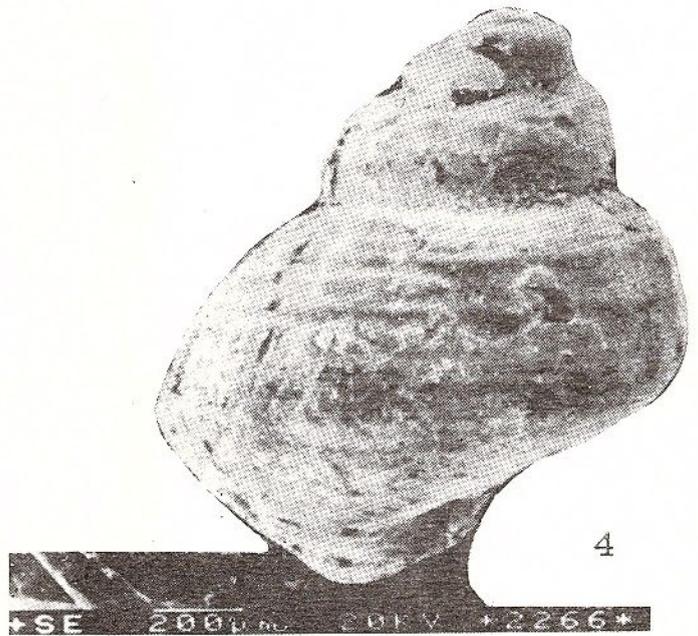
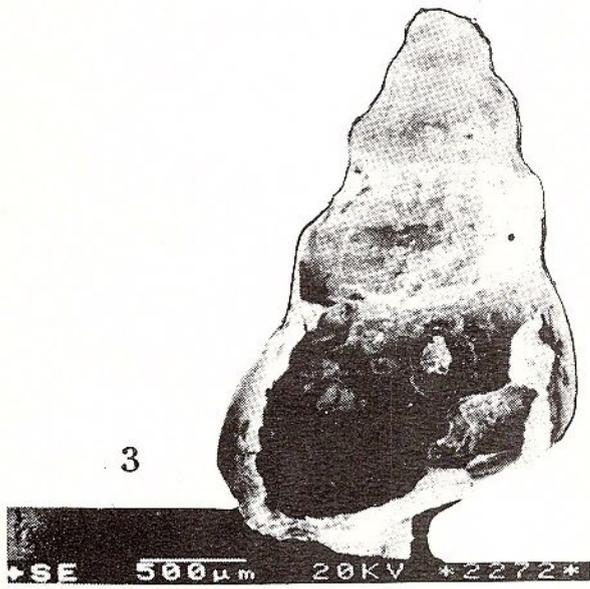
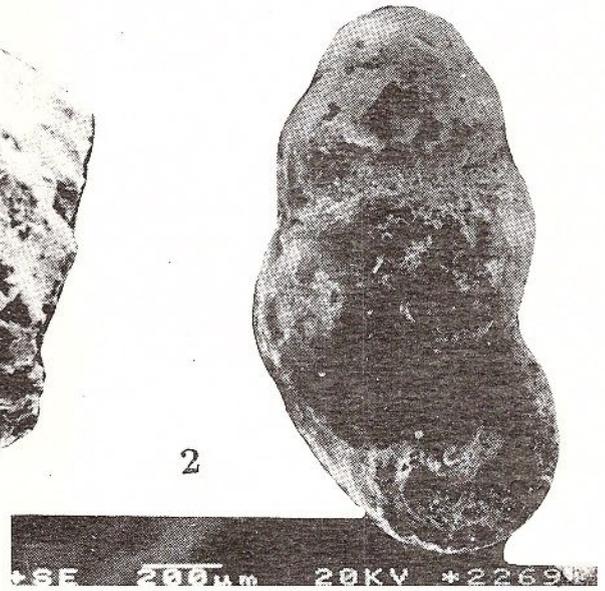
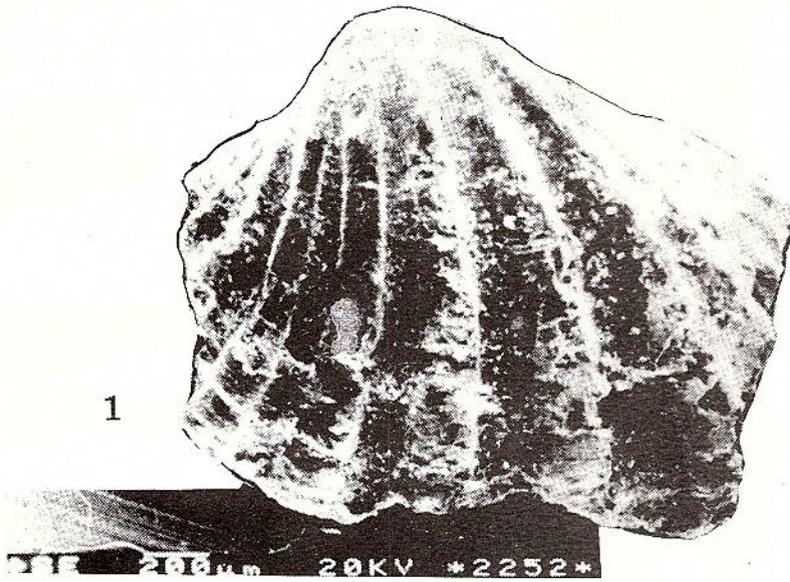
Fig. 1 — *Cerastoderma lalisulcum* AITON

Fig. 2 — *Stenothyrella schwartzi* AITON.

Fig. 3 — *Calliostoma angulatum* AITON.

Fig. 4 — *Calliostoma orbygnianum* AITON.





LA BIOSTRATIGRAPHIE DES DÉPÔTS DU MIOCÈNE EN ALBANIE, BASÉE SUR LES NANNOFOSSILES CALCAIRES

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Resumé: Les dépôts de l'Aquitaniens au Messinien, de la zone Ionienne et la Dépression Préadriatique (D.P.) d'Albanie, ont été l'objet de ce travail. La richesse en nannofossiles calcaires et en foraminifères planctoniques permet d'établir une échelle biostratigraphique détaillée. La corrélation des zones de nannofossiles calcaires appliquée ici, avec les zonations de Okada & Bukry (1980), de Martini (1971) et des foraminifères planctoniques est donnée dans les tableaux 1 et 2. La pauvreté des quelques marqueurs (*T. carinatus*, *D. druggii*, *D. kugleri* etc.) rend difficile l'application des zonations de Martini ou de Okada & Bukry dans certains intervalles du Miocène inférieur et moyen. Deux nouvelles zones (à *Discoaster deflandrei* et à *Dictyociccites hesslandii*) sont employées (Çobaj, 1992) pour caractériser le Miocène inférieur (partiellement).

Introduction

De nombreux travaux biostratigraphiques ont été effectués sur les dépôts du Miocène de la zone Ionienne et la D.P. (Albanie Sud, Sud-Ouest). Ces dépôts sont bien exposés, présentant des épaisseurs considérables. Ils sont riches en foraminifères planctoniques, nannofossiles calcaires et spores et pollens. Dans le cadre de ce travail, j'ai observé plusieurs échantillons des coupes prélevées, des échantillons ponctuels et des forages, dispersés dans presque toute la région étudiée (fig. 1). Parallèlement à cette étude, un travail sur les foraminifères planctoniques des mêmes échantillons a été effectué (Prillo et al., 1993), afin de réaliser de bonnes corrélations entre les deux biozonations.

Brièvement sur le cadre géologique

La zone Ionienne est limitée à l'Est par la zone de Kruja (analogue de la zone Gavvora en Grèce et de la zone Dalmate en Yougoslavie). Le passage entre ces zones est continu. Au Nord elle plonge sous une couverture Miocène moyen-Pliocène et Quaternaire. Au Sud, elle continue en Grèce sous le même nom. À l'Ouest cette zone contacte tectoniquement la zone de Sazani (zone Paksos en Grèce et Préapulien en Italie)

(Bakia et al., 1987). Les dépôts qui la composent sont du Trias supérieur au Miocène moyen.

Les dépôts de l'Aquitaniens successifs sur l'Oligocène supérieur (coupe de Ndroq, Lapardha, Levan, Ballaj etc.), se caractérisent par une alternance de couches d'argiles, de silts et de grès. Dans quelques coupes, des turbidites et des grès massifs sont présents. La partie inférieure du Burdigalien est présente par des argiles, alors que la partie supérieure et le Langhien inférieur soient composés de marnes. Des argiles remplacent graduellement ces derniers en dessus (Ndroq, Ballaj, Levan, Lapardha etc.).

La D.P. représente un avant-pays qui se pose transgressivement sur l'orogène. Les formations qui la construisent sont d'âge Serravallien-Pliocène.

Les dépôts du Serravallien, formés par des argiles avec des intercalations de couches de grès et de calcaires lithothamniques, se déposent en discordance sur les dépôts de la zone Ionienne. La succession mollassique du Tortonien contient des argiles où de rares couches de grès et des silts s'intercallent. Ces dépôts sont identifiés dans les coupes de Kavaje, Rrogozhine, Zvervec, Bisht Pallc, Patos etc. Des grès massifs et de minces couches de silts sont rencontrés au Messinien. De l'Est à l'Ouest, la lithologie change graduellement passant de grès avec de rares intercalations d'argiles, aux argiles. Parmi les argiles, on y trouve des niveaux de gypses (Kavaje, Vlore etc.).



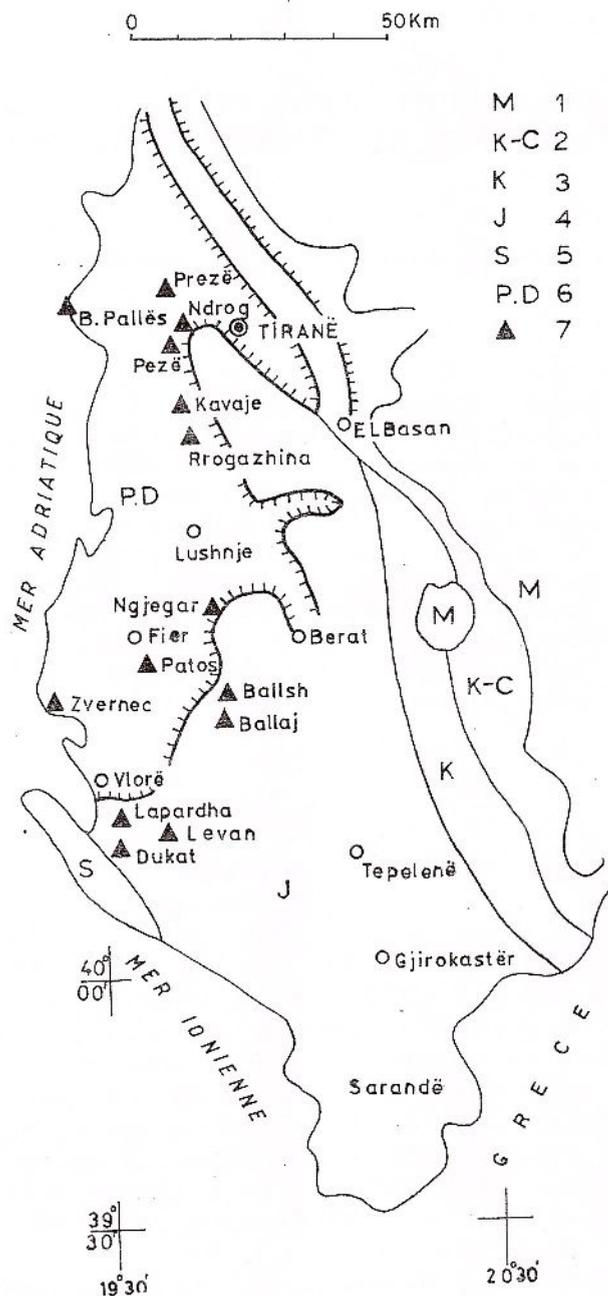


Fig. 1 - La position géographique des échantillons étudiés

1, Zone de Mirdita; 2, Zone de Krast-Cukali; 3, Zone de Kruja; 4, Zone Ioni-
enne; 5, Zone de Sazan; 6, Dépression Préadriatique; 7, échantillons étudiés.

La biostratigraphie

Les associations de nanfossiles calcaires de l'Aquitaniens-Burdigalien inférieur sont très riches, de petite taille, bien conservées, mais pas diversifiées. Dans tous les échantillons observés *T. carinatus* et *D. druggii* sont extrêmement rares. Cette pauvreté rend difficile l'application des zonations de Martini (1971) et ou d'Okada & Bukry (1980) dans ces niveaux. En

conséquence deux nouvelles zones, à *D. deflandrei* et *D. heslandii* sont proposées (Çobaj, 1992). L'intervalle entre la réapparition de *H. carteri*, près de la limite Oligocène supérieur-Aquitaniens et les disparitions de *S. conicus* et de *S. dissimilis* représente la zone à *D. deflandrei* (NN 1-NN 2 à part ou CN 1 à part). L'association des nanfossiles est riche, bien préservée, de petite taille, mais pas variée. Les espèces plus communes sont: *H. carteri*, *H. intermedia*, *H.*



Tableau 1
Corrélation des zones de nanfossiles calcaires appliquée ici dans les zones de Okada & Bukry (1980) et de Martini

| AGE | Okada & Bukry (1980) | Martini (1971) | L. Çobaj (1992) | Marqueurs |
|--------------|----------------------|----------------|---------------------------|--|
| MESSINIEN | CN 10a+b | NN 12 | <i>C. tricorniculatus</i> | † <i>D. quinqueramus</i> |
| | CN 9b | NN 11 | <i>D. quinqueramus</i> | † <i>D. quinqueramus</i> |
| TORTONIEN | CN 9a | | | |
| | CN 8 | NN 10 | <i>D. pentaradiatus</i> | † <i>D. pentaradiatus</i> |
| SERRAVALLIEN | CN 7 | NN 9 | <i>D. variabilis</i> | † <i>D. hamatus</i> |
| | CN 6 | NN 8 | | † <i>C. coalitus</i> |
| | CN 5 | NN 7 | | † <i>D. kugleri</i> |
| | | NN 6 | | † <i>D. exilis</i> |
| LANGHIEN | CN 4 | NN 5 | <i>S. heteromorphus</i> | † <i>S. heteromorphus</i> |
| | CN 3 | NN 4 | <i>H. ampliaperta</i> | † <i>D. exilis</i> |
| BURDIGALIEN | CN 2 | NN 3 | <i>H. scissura</i> | † <i>H. ampliaperta</i> |
| | CN 1 | NN 2 | <i>D. hesslandii</i> | †† <i>S. conicus</i> <i>S. dissimilis</i> |
| AQUITANIEN | | NN 1 | <i>D. deflandrei</i> | † <i>H. carteri</i> |

Tableau 2
Corrélation des zones de nanfossiles calcaires et de foraminifères planctoniques appliquée en Albanie

| AGE | L. Çobaj 1992 | Marqueurs | Marqueurs | Prillo et al. 1993 |
|--------------|---------------------------|---------------------------|--|--|
| MESSINIEN | <i>C. tricorniculatus</i> | † <i>D. quinqueramus</i> | <i>G. multiloba</i> | Indéterminable |
| | | | <i>G. multiloba</i> | † <i>G. multiloba</i> |
| TORTONIEN | <i>D. quinqueramus</i> | † <i>D. quinqueramus</i> | <i>G. conomiozea</i> | † <i>G. conomiozea</i> |
| | | | <i>D. pentaradiatus</i> | † <i>G. o. extremus</i> |
| | | | † <i>D. pentaradiatus</i> | <i>G. o. extremus</i> |
| SERRAVALLIEN | <i>D. variabilis</i> | † <i>S. heteromorphus</i> | <i>D. hamatus</i> | † <i>G. acostaensis</i> |
| | | | <i>C. coalitus</i> | † <i>G. menardii s.l.</i> |
| | | | <i>D. kugleri</i> | † <i>G. mayeri</i> |
| | | | <i>D. exilis</i> | † <i>G. peripheroranda</i> |
| LANGHIEN | <i>S. heteromorphus</i> | † <i>D. exilis</i> | <i>O. universa</i> | † <i>O. universa - G. peripheroranda</i> |
| | | | <i>H. ampliaperta</i> | † <i>O. suturalis</i> |
| BURDIGALIEN | <i>H. scissura</i> | † <i>H. ampliaperta</i> | <i>P. glomerata</i> | † <i>P. glomerata s.l.</i> |
| | | | <i>D. hesslandii</i> | † <i>G. bisphaericus</i> |
| | | | †† <i>S. conicus</i> <i>S. dissimilis</i> | † <i>G. bisphaericus</i> |
| AQUITANIEN | <i>D. deflandrei</i> | <i>H. carteri</i> | <i>G. dehiscens</i> | † <i>G. dehiscens</i> |

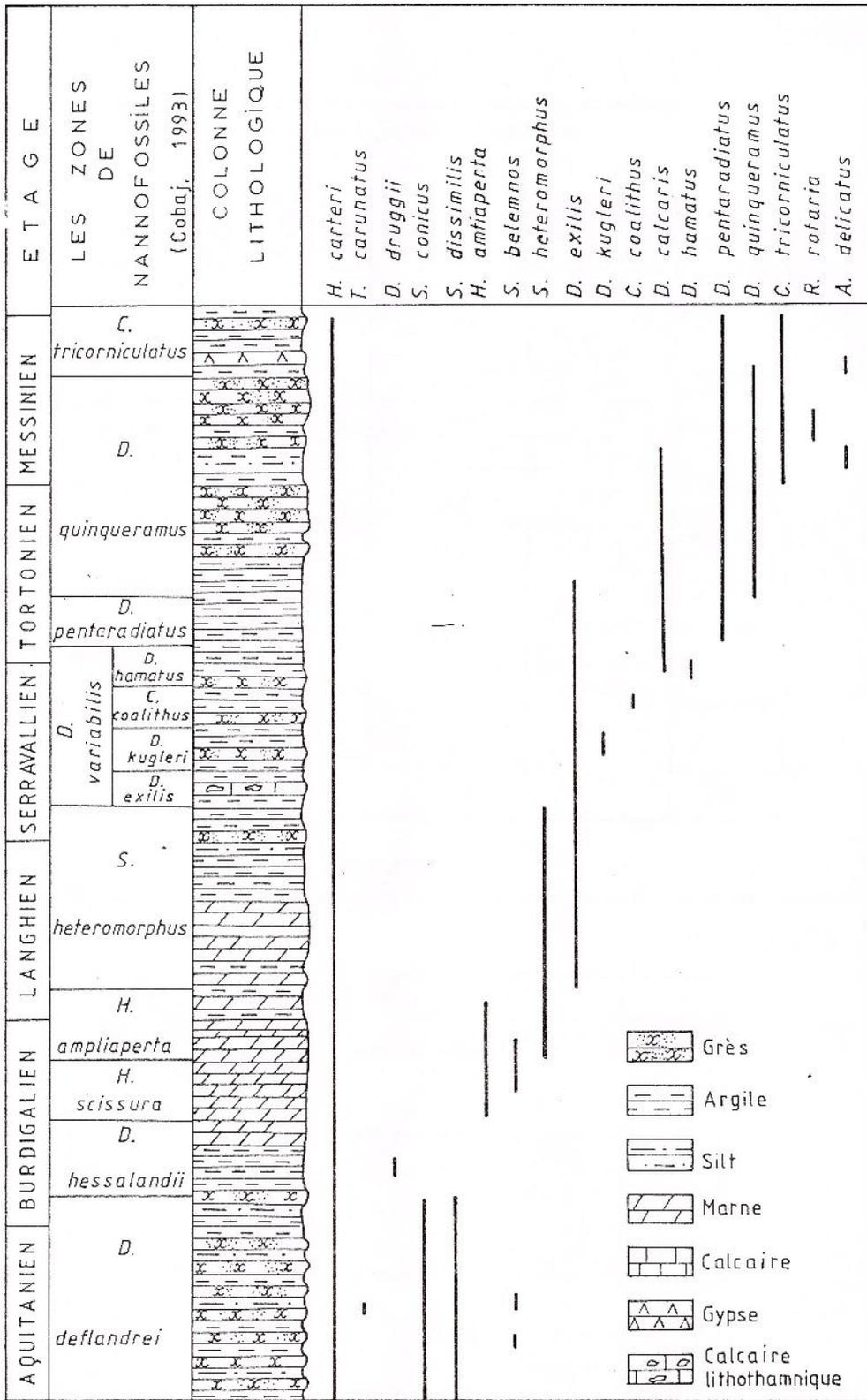


Fig. 2 - Lithologie, zones de nannofossiles et répartitions stratigraphiques de quelques marqueurs.

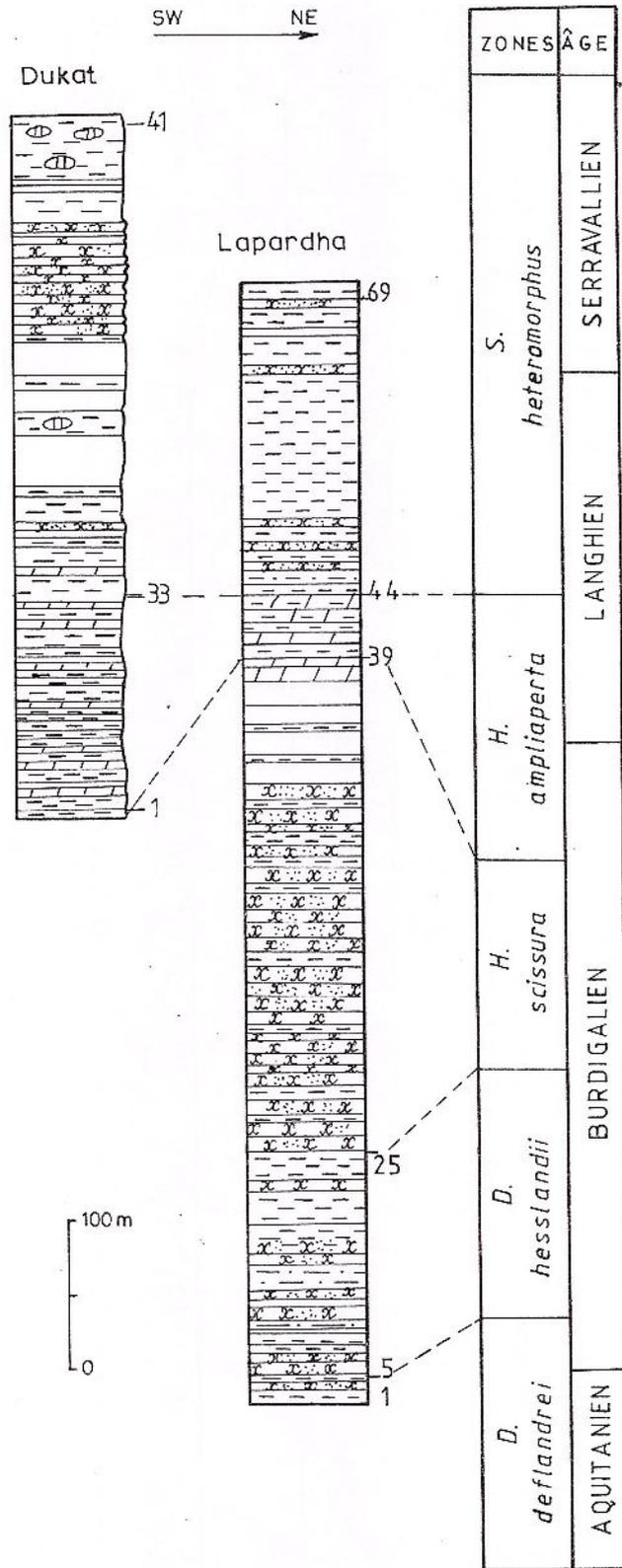


Fig. 3 - La corrélation entre les coupes de Dukat et Lapardha basée sur les zones de nannofossiles.

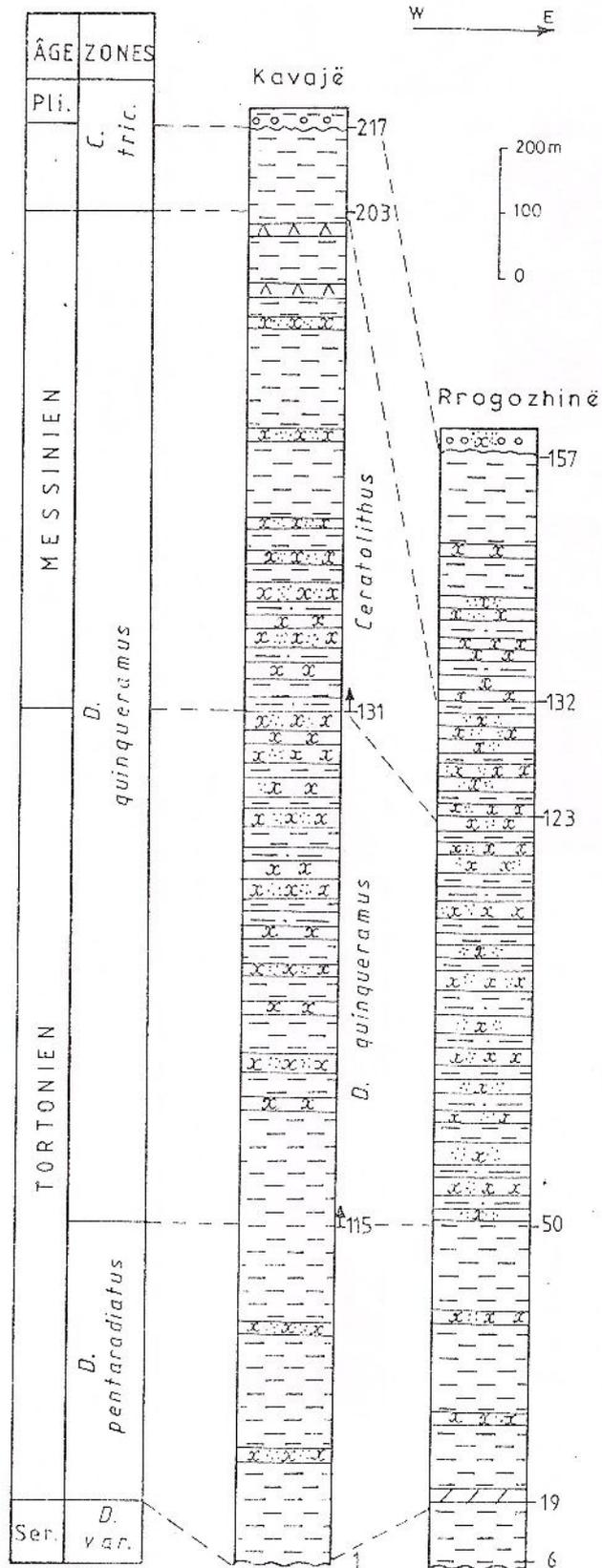


Fig. 4 - La corrélation entre les coupes de Kavaja et de Rrogozhinë basée sur les zones de nanofossiles.

euphratis, *S. conicus*, *S. dissimilis*, *S. belemnus*, *S. moriformis*, *C. pelagicus*, *D. deflandrei*, *D. adamanteus*, *C. floridanus*, *C. abisectus* etc. Elle est reconnue dans un grand nombre de coupes, forages et échantillons ponctuels. Les disparitions de *S. conicus* et de *S. dissimilis* se produisent à la base du Burdigalien (zone à *G. trilobus*-*G. dissimilis*, de foraminifères planctoniques).

La zone à *D. hesslandii* est définie comme intervalle entre les disparitions de *S. conicus* et *S. dissimilis* et l'apparition de *H. ampliapertura* (NN 2 à part; CN 1 à part). Les nannofossiles y sont nombreux, représentés presque par les mêmes espèces de la zone à *D. deflandrei*. L'absence de *S. conicus*, *S. dissimilis*, l'apparition de *H. mediterranea* et *C. macintyreii* au sommet de la zone, aide à la distinguer facilement. Il faut noter que la disparition de *S. conicus* est un peu plus tardive que celle de *S. dissimilis*, toutefois cela n'empêche pas de l'appliquer comme marqueur.

La présence de *H. ampliapertura*, *H. scissura*, *H. mediterranea*, *S. heteromorphus*, *C. macintyreii*, *C. premacintyreii*, *G. rotula*, *R. pseudoumbilica*, l'apogée de *S. belemnus* etc., permet d'identifier les dépôts plus élevés du Burdigalien et du Langhien inférieur et de reconnaître les zones à *H. scissura* (analogue à NN 3 à part; CN 2) et à *H. ampliapertura* (NN 4; CN 3). À partir de la base de la zone à *H. ampliapertura*, on observe un renouvellement des espèces du genre *Discoaster*, telles que: *D. variabilis*, *D. pseudovariabilis*, *D. subsurculus*, *D. intercalaris*, *D. exilis*, *D. signus*, etc. La richesse et la diversité des nannofossiles calcaires qui commencent au Burdigalien supérieur, deviennent évidentes au Miocène moyen.

L'association composée de *D. exilis*, *D. singus*, *D. protoexilis*, *D. variabilis*, *S. heteromorphus*, *H. mediterranea*, *C. macintyreii*, *C. premacintyreii*, *R. pseudoumbilica*, *C. miopelagicus* etc. sert à distinguer les dépôts du Langhien et à déterminer la zone à *S. heteromorphus* (NN 5 ou CN 4). Ces dépôts sont rencontrés dans les coupes de Ndroq, Ballaj, Lapidha, Ngjeqar, Levan etc. La disparition de *S. heteromorphus* se fait à la partie inférieure du Serravallien. Plus haut, dès niveaux élevés du Serravallien jusqu'à la base du Tortonien, les associations de nannofossiles s'appauvrissent. *D. kugleri*, *D. bollii*, *D. exilis*, *D. variabilis*, *D. calcaris*, *D. hamatus*, *D. brouweri*, *C. coalitus*, *H. stalis*, *H. walberdorfenensis*, *R. pseudoumbilica*, *S. miopelagicus*, etc., mettent en évidence la présence des autres zones à *D. exilis*, *D. kugleri*, *C. coalitus*, *D. hamatus* et à *D. calcaris* (NN 6-NN 8; CN 5-CN 7). Il est nécessaire de noter que dans tous les échantillons observés *D. kugleri* est très rare et atypique (Ndroq, Zvernec etc.). *C. coalitus* et *D. hamatus* sont aussi très rares. J'y

ai trouvé seulement quelques individus à Zvernec. Par rapport aux espèces citées au dessus *D. calcaris* est plus commun. La présence sporadique de ces marqueurs permet de distinguer seulement les parties isolées des zones sous les même noms. C'est pour cette raison que la zone à *D. variabilis*, intervalle compris entre la disparition de *S. heteromorphus* et l'apparition de *D. pentaradiatus*, ait été appliquée en Albanie pour caractériser ces dépôts.

L'association assez riche composée de: *D. pentaradiatus*, *D. quinquaramus*, *D. calcaris*, *D. variabilis*, *D. brouweri*, *D. exilis*, *D. neorectus*, *S. pulchra*, *M. convalis*, *H. pacifica*, *S. abies*, *S. verensis*, *C. leptopora* etc., est livrée par la succession du Tortonien. Elle sert à reconnaître les zones à *D. pentaradiatus* (Schmidt, 1973) (NN 9 à part, NN 10; CN 7 à part, CN 8) et à *D. quinquaramus* (NN 11; CN 9a). L'apparition des *Discoaster* à cinq bras a été employée pour tracer approximativement la limite Serravallien-Tortonien.

En passant aux dépôts du Messinien, les associations s'appauvrissent de nouveau. Les discoasters sont très rares et les espèces mal conservées. Le remaniement qui existe dans presque tout le Miocène, au Miocène moyen et surtout à celui supérieur est très évident. Les espèces remaniées du Crétacé au Miocène moyen, telles que *W. biporta*, *W. barnasae*, *M. staurophora*, *P. cretaceae*, *C. ehrenbergi*, *Z. bijugatus*, *R. umbilica*, *S. predistentus*, *S. distentus*, *S. ciproensis*, *S. belemnus*, *S. heteromorphus*, *H. ampliapertura* etc., sont communes, alors que sur place elle soient très rares. *R. pseudoumbilica*, *H. carteri*, *S. moriformis*, *C. pelagicus* etc., plus tolérantes aux conditions défavorables (la crise de la salinité), y sont présentes, particulièrement dans l'Est de la D.P. Leurs répartitions stratigraphiques sont très larges et ne permettent pas de préciser l'âge (les coupes de Ndroq, Preze, Patos, Lushnje cct.). Les échantillons prélevés à l'Ouest contiennent des associations assez riches en nannofossiles par rapport à celles de l'Est. On y observe *S. abies*, *S. verensis*, *R. pseudoumbilica*, *C. primus*, *C. tricorniculatus*, *A. delicatus*, *D. quinquaramus*, *D. pentaradiatus*, *D. calcaris*, *D. surculus*, *D. brouweri*, *D. variabilis*, *H. carteri*, *C. pelagicus* etc. C'est le cas de la coupe de Kavajc, des échantillons de Seman, Zvernec. L'apparition du genre *Ceratolithus* a servi à tracer la limite Tortonien-Messinien. *R. rotaria* et *R. pseudoumbilica* 7 microns, ont aidé aussi à identifier le Messinien en absence de *Ceratolithus*. Il faut dire que les représentants de ce dernier et de *D. quinquaramus* ne sont pas fréquents. Les derniers exemplaires de *D. quinquaramus* sont trouvés dans les dépôts du Messinien supérieur, bien avant la limite Messinien-Pliocène. À cause de cette pauvreté il est presque impossible de dater cette li-



mite, en base de sa disparition. Dans ces conditions, la zone à *C. tricorniculatus* (NN 12; CN 10 a+b), dans la région étudiée comprend les dépôts du Messinien supérieur et du Pliocène inférieur.

Conclusions

Les dépôts du Miocène sont riches en nannofossiles calcaires. A cause de la présence sporadique de *T. carinatus* et de *D. druggii*, la réapparition de *H. carteri*, les disparitions de *S. conicus*, *S. dissimilis* et l'apparition de *H. ampliapertura* sont utilisées pour diviser l'Aquitainien-Burdigalien inférieur en deux zones: à *D. deflandrei* et à *D. hesslandii*.

A partir du Burdigalien supérieur jusqu'au Serravallien inférieur, les associations de nannofossiles sont très riches et diversifiées, ce qui a permis d'établir une biozonation détaillée.

La pauvreté en *D. kugleri*, *C. coalitus*, *D. hamatus* rend difficile l'application des zones sous les mêmes noms de Martini (1971) et ou d'Okada & Bukry (1980). Toutefois, leurs présence sert à reconnaître en partie ces zones.

Le remaniement est présent partout, mais il est plus puissant au Miocène supérieur.

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AN OVERVIEW OF THE UPPER OLIGOCENE-LOWER MIOCENE VOLCANOGENIC SEDIMENTS IN THE WESTERN MEDITERRANEAN AND THEIR POSSIBLE SOURCE AREAS

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Key words: Volcanogenic sediments. Volcanic clast populations. Volcaniclastic depositional processes. Late Oligocene-Early Miocene. Palaeotectonics. Palaeogeography. Western Mediterranean.

Abstract: Volcanogenic sediments are widespread in the Western Mediterranean, interbedded within several Upper Oligocene-Lower Miocene marine formations along the Apennines, Maghrebides and Betic Cordillera. These volcanogenic sediments are representatives of periods of intense volcanically-induced sedimentation and their compositions have roughly been referred, by several authors, to calcalkaline magmas erupted in volcanic arc systems. No comprehensive hypotheses have been put forward on the palaeogeographic locations of the volcanic arc systems related to the volcanogenic sediments. Tuffaceous sandstones, volcanolithic sandstones and crystal- to vitric-rich volcanoclastic sandstones are very widespread, although, fine-grained ash layers, generally altered to clay minerals, are also recognized. Depositional mechanisms inferred for most of the studied volcanogenic sediments are related to erosion and/or reworking of pyroclastic material and lavas shortly after their eruptions and slumping into the basins (mass-flow processes). Very often, regarding the Apennines volcanoclastic deposits, Sardinia had been indicated as the volcanic source area but its location during Late Oligocene-Early Miocene cannot fit with the inferred epiclastic mass-flow processes, also for the complex palaeogeography. Nevertheless, considering a fallout process, the Sardinia volcanic source area can plausibly be invoked for some thin and very fine-grained pelites, but not for thick and coarse volcanoclastic deposits. Available data seem to support the existence of volcanic arc systems adjacent to the sedimentary basins, and located in internal position with respect to the basins themselves. The preliminary models of Dewey et al. (1989) and Guerrera et al. (1993) about kinematic evolution of the Western Mediterranean oceanic basins provides, during Aquitanian-Burdigalian time, the existence of a further primary volcanism (matched to the Sardinia eruptions) linked to an extensional tectonics and in agreement with the distribution of the volcanogenic sediments. In addition, whole-rocks geochemical data for some of the Western Mediterranean volcanogenic sediments seem to support a link with oceanic island-arc tectonic settings. Remnants of volcanic edifices (after quick erosional and/or subsidence processes) could have been buried and concealed at depth by underthrusting processes during the post-Early Miocene tectonic evolution.



1. Introduction

Sedimentary rocks piles with significant volcanoclastic intercalations are widespread in the Western Mediterranean. Attention in this study is focused on the relationships between Upper Oligocene-Lower Miocene volcanogenic sediments and the inferred volcanic source areas (Guerrera *et al.*, 1995).

The Western Mediterranean is bordered by three alpine systems: the Betic Cordillera, the Maghreb Chain (Rif, Tell and Sicily), and the Apennines. The Mesozoic and Tertiary successions involved in these chains were deposited in three major palaeogeographic realms (Guerrera *et al.*, 1993; Pl. I): Internal Domain (I.D.), "Flysch" Basin Domain (F.D.) and External Domain (E.D.).

The units representative of the I.D. are made up of ophiolitic nappes and several basement nappes, in places with Mesozoic and Tertiary cover. They appear strongly deformed and generally affected by alpine metamorphism. The sediments of the F.D. (clayey and terrigenous formations of Cretaceous to Miocene) were deposited in deep, oceanic to oceanoid basins ("Maghreb Ocean" of Guerrera *et al.*, 1993), and structured in several nappes and melanges during the Early Miocene. The E.D. is represented by cover nappes which derive from the divergent continental margins that, during the Mesozoic (platform area), bordered the Iberia, Africa and Adria Plates. During the Miocene these formerly divergent margins evolved to foreland basins with prevalently terrigenous sedimentation (Guerrera *et al.*, 1993).

Western Mediterranean volcanogenic sediments are interbedded with marine formations and frequently recorded in the Betic Cordillera, Maghrebides and Apennines; they are mainly related to calcalkaline volcanic arc systems developed during the Upper Oligocene-Lower Miocene (Pl. I).

These deposits are consistent with periods of coeval volcanically induced sedimentation and the record of volcanic activity preserved in sedimentary rocks has been used to develop palaeotectonic models (Dickinson, 1985, 1988; Valloni, 1985; Zuffa, 1987). General criteria to discriminate between neovolcaniclastics generated by active volcanism, and palaeovolcanic detritus derived from erosion of primary volcanic rocks, are discussed by several authors (e.g. Pettijohn *et al.*, 1972; Zuffa, 1985, 1987; Guerrera, Veneri, 1989 a, 1989 b; Critelli, Ingersoll, 1995). Having in mind that all the investigated deposits referring to neovolcaniclastics (i.e. volcanogenic sediments formed within a basin during contemporaneous-penecontemporaneous volcanic activity) the term *epiclastics* is here generally used for (*sensu* Cas, Wright, 1987): (a) pyroclastics and lavas which are reworked shortly after the eruption (in sub-

aerial and submarine environments), (b) remobilization of primary subaerial and subaqueous volcanic deposits through mass-flow processes. All the air-fallout deposits from active volcanoes within adjacent or not directly related to basin-system are here grouped with the term *pyroclastics*.

Generally, in deep to shallow marine environments the volcanic clast populations are largely dependent on eruption styles (effusive versus explosive). Fragmentation of lava flows, in intrabasinal systems or in the fringes of island volcanoes, will give rise dominantly to volcanolithic sandstones whereas subaerial to shallow marine pyroclastic eruptions will involve syn- post-eruptive tuffaceous sandstones (ash turbidites) to vitric- crystal-rich volcanic sandstones (Cas, 1979; Wright, Mutti, 1981; Fisher, 1984; Cas, Wright, 1987; Critelli, Ingersoll, 1995).

An overview of the areal distributions, thicknesses, and petrological aspects of Upper Oligocene-Lower Miocene volcanogenic sediments in the Western Mediterranean will be presented in order to constrain possible volcanic source areas and therefore to establish a more realistic geodynamic reconstruction during that time.

2. Distribution of volcanogenic sediments

The main characters of the volcanogenic sediments recorded in various units along Betic Cordillera, Maghreb Chain and Apennines Domains are listed below. Only deposits which are the result of a volcanically induced sedimentation will be described on the basis of published and unpublished data. These volcanogenic sediments are grouped in Plate II where (a) the stratigraphic column, (b) sectors and domains, (c) inferred composition and (d) depositional processes of the volcanic detritus are shown.

2.1. Betic Cordillera Sector (Spain: "1" in Plate I). Volcanoclastic deposits are dispersed within the three recognized domains (Pl. II): Internal (Viñuela, Burdigalian p.p.), "Flysch" Basin (Algeciras, Late Chattian-Aquitania p.p.; Guerrera *et al.*, 1993; M. Martin Martin, pers. comm.) and External (Almidar, Burdigalian p.p.).

In the Viñuela Group (Martin-Algarra, 1987; Guerrera *et al.*, 1993), sandstones and siltstones are related to volcanically induced sedimentation during eruptions of andesitic and/or rhyolitic magmas (Rivière, Courtois, 1975; Rivière, 1988). Volcanic imprint is mainly given by crystals and glass-shards. Depositional processes include both pyroclastic fallouts and epiclastic mass-flow processes. The stratigraphic interval with volcanogenic sediments (hereafter SIV) is about 45 m, although the total thickness of volcanoclastic beds (hereafter VTB) is 3 m,



and individual strata of volcanogenic sediments have a maximum observed thickness (hereafter MOT) of 0.5 m (Chauve et al., 1973; Rivière, Courtois, 1975; Boulin et al., 1973).

In the Algeciras Flysch, arkoses and lithic arkoses were emplaced by turbiditic processes; they are characterized by lithic clasts of felsitic lavas up to 7 % vol. and subordinate crystals (Puglisi, Carmisciano, 1992). The presence of these volcanic clasts could be due to a possible dacitic to rhyolitic volcanism.

In the Almidar Fm, sandstones and siltstones show, in places, components with hyaloclastic features. The main volcanic components comprise glass-shards and crystals; strata, having a MOT of 0.25 m, are the result of mass-flow processes during rhyolitic-rhyodacitic eruptions (Soria et al., 1992).

2.2. Rifian Sector (Morocco: "2" in Plate I). Volcaniclastic deposits (Pl. II) are recorded in the Internal (Sidi Abdeslam, Aquitanian p.p.-Early Burdigalian) and "Flysch" Basin (Beni Ider and Mixed Successions, Late Aquitanian-Early Burdigalian) Domains (Feinberg et al., 1990; Guerrero et al., 1993).

The volcanic imprint within the sandstones of the Sidi Abdeslam Fm is mainly given by crystals. They are included in a SIV of about 10 m and probably emplaced by epiclastic processes; volcaniclastic strata have a MOT of 0.15 m whereas the VTT is 0.5 m. The inferred composition of volcanic material is rhyolitic.

In the Beni Ider Flysch and Mixed Successions, sandstones containing calcalkaline lithic clast of andesitic and/or basaltic lavas were emplaced by syn- and/or post-eruptive turbiditic processes (Chiocchini et al., 1980).

2.3. Tunisian-Algerian Sector (Tell: "3" in Plate I). The volcanogenic material occurring in the late-orogenic flysch deposits of the Oligo-Miocene Kabyle (Burdigalian p.p.; 19.1 ± 1.0 Ma) belongs to the Internal Domain (Pl. II); here sandstones and siltstones with potassic calcalkaline volcanic detritus of possible rhyolitic composition (crystals and glass-shards) were emplaced by epiclastic mass-flow processes (Bellon, 1976; Rivière et al., 1977; Rivière, 1988; Guerrero et al., 1993).

2.4. Southern Calabria-Sicily Sector (Italy: "4" in Plate I). Volcanogenic sediments are recognized within the Internal (Stilo-Capo d'Orlando, Aquitanian p.p.-Early Langhian) and "Flysch" Basin (Troina-Tusa, Mixed Successions and Numidian) Domains (Zuppetta et al., 1984; Guerrero et al., 1990; Guerrero et al., 1993).

In the Stilo-Capo d'Orlando Fm the sandstones mostly contain lithic clasts of andesitic and/or

basaltic lavas (Critelli, 1991) and were probably emplaced by turbiditic mass flow processes.

Within the Troina-Tusa Flysch (Aquitanian p.p.-Early Burdigalian; *sensu* Guerrero, Wezel, 1974) arenites are characterized by 80-85 % vol. of volcanic detritus, mainly constituted of andesitic and/or basaltic lava clasts. These deposits, emplaced by turbiditic processes, have a grain size up to 2.5 mm and they are comprised within a SIV of 600 m. The MOT is 1.5-2 m in a VTT of about 200 m (Wezel, Guerrero, 1973; Guerrero, Wezel, 1974; Puglisi, 1992; Loiacono, Puglisi, 1983; Ardito et al., 1985; Balogh et al., 1993; Puglisi, 1994).

In the Mixed Successions (Late Aquitanian-Early Burdigalian) sandstones have a grain size >2 mm within a SIV of 43 m. They comprise calcalkaline andesitic and/or basaltic lava clasts and strata, mainly emplaced by mass-flow processes, having a MOT of 1 m (Hoyez, Andreieff, 1975; Grasso et al., 1987; Carmisciano et al., 1989).

Within the Numidian Fm (Late Aquitanian-Early Burdigalian) the volcanogenic material is represented by calcalkaline volcanic material of possible rhyolitic composition (glass-shards, crystals, pumices and lava fragments; Patacca et al., 1992). Lava clasts show andesitic and/or basaltic textures. These deposits, within a SIV of 20 m, were generated by epiclastic mass-flow processes (Faugères et al., 1992).

2.5. Southern Apennine-Northern Calabria Sector (Italy: "5" in Plate I). Volcaniclastic sediments (Pl. II) occur in the Internal (Paludi), "Flysch" Basin (Calabro-Lucano, Saraceno, Tusa, Pollica, S Mauro and Numidian) and External (Macchialupo and Roccadaspide) Domains (Amore et al., 1992; Guerrero et al., 1993).

In the Paludi Fm (Late Chattian-Early Aquitanian), sandstones were emplaced by turbiditic processes and volcanic imprint is given by the presence of andesitic and/or basaltic lava fragments and crystals (Critelli, 1991).

Within the Calabro-Lucano Flysch (Late Oligocene-Early Aquitanian) sandstones with a MOT of 1.5 m were generated by syn- and/or post-eruptive turbiditic processes. They contain andesitic and/or basaltic and felsitic lava clasts, crystals and glass-shards related to a calcalkaline volcanism (Critelli, Monaco, 1993).

In the Saraceno Flysch (Late Oligocene-Early Aquitanian) sandstones were also emplaced by syn-eruptive turbiditic processes, being characterized by andesitic to dacitic lava clasts and crystals (Critelli, 1991).

Within the Tusa Tuffites (Aquitanian p.p.-Burdigalian p.p.) arenites have 80 % vol. of vol-



canic detritus represented by lithic clasts of andesitic and/or basaltic lavas and minor amounts of crystals. Pelitic strata contain crystals, pumices and glass-shards (ash turbidites). The VTT of 46 m is included within a SIV of 54 m and their strata, emplaced by syn- and/or post-eruptive mass-flow processes, have a MOT of 5 m (Critelli *et al.*, 1990; Balogh *et al.*, 1993).

In the Pollica Flysch/upper member (Burdigalian p.p.) conglomerates (with pebbles up to 1 m across) and sandstones are characterized by levels having a MOT of 3 m and they seem to be consistent with mass-flow processes (syn- and/or post-eruptive turbidites). The volcanic clast populations are mainly constituted of rhyolitic-rhyodacitic lavas and fragments of tuffs (Critelli, Le Pera, 1990).

In the S. Mauro Flysch (Late Burdigalian-Early Langhian) conglomerates (with pebbles up to 15 cm) and arenites are characterized by > 90 % vol. of volcanoclastic material within a SIV of 15-20 m. The deposits, mostly characterized by rhyolitic-rhyodacitic lava fragments and subordinate crystals, are probably the result of a short-distance mass-flow transport (Crisci *et al.*, 1988).

Within the Numidian Fm (Late Burdigalian-Early Langhian) the volcanogenic material, having a calcalkaline composition (probably rhyolitic), was emplaced by pyroclastic fallout processes (Patacca *et al.*, 1992).

In the Macchialupo Unit (Daunia Complex, Burdigalian p.p.) conglomerates and sandstones (1-2 up to 4 m thick) contain about 70-89 % vol. of calcalkaline andesitic and/or basaltic lava fragments within a SIV of 80 m; volcanogenic strata, emplaced by epiclastic processes, have a MOT of 4 m (Pieri, Rapisardi, 1973).

The calcarenites within the Rocchaspide Fm (Late Burdigalian-Langhian p.p.) are characterized by 3-15 % vol. of volcanic detritus made up of calcalkaline andesitic and/or basaltic lava clasts, crystals and glass-shards; strata have a MOT of 5 m and the deposition of the volcanic material seems to be related to pyroclastic fallout and/or mass-flow processes (Perrone, 1987).

2.6. Northern Apennines-Internal Sector, Epi-Ligurian Domain (Italy: "6" in Plate I).

In the Northern Apennines three main structural-palaeogeographic domains (Chain, Foredeep and Foreland) and related sedimentary basins can be pointed out (Boccaletti *et al.*, 1990; Delle Rose *et al.*, 1994; Amorosi *et al.*, 1995).

In the internal palaeogeographic sector (Chain), piggy-back basins have been developed over the Ligurian thrust sheet (Epi-Ligurian) translating towards

the external domains. The Piemonte Tertiary Basin was generated along the suture between Alpine and Apennine Chains, representing the extension towards NW (Monferrato) of this internal sector.

In the internal sector of the Northern Apennines (Epi-Ligurian Domain) the volcanogenic materials occur in different units (Antognola, Val d'Aveto-Petrignacola, Tripoli di Contignaco and Bisciario-like).

Within the Antognola Fm (Late Chattian-Early Aquitanian, Amorosi *et al.*, 1995) arenites and tuffaceous siltstones are emplaced by epiclastic processes within a SIV of 15 m; strata of volcanogenic sediments have a MOT of 5 m (i.e. the whole VTT) and are characterized by: (i) calcalkaline andesitic and/or dacitic lava clasts and crystals, and (ii) crystals and glass-shards of rhyolitic-rhyodacitic composition (Tateo, 1993).

In the Val d'Aveto and Petrignacola Fms (Late Chattian-Early Aquitanian) conglomerates (with pebbles up to 1 m) and volcanolithic sandstones were emplaced by epiclastic mass-flow processes within a SIV of 35 m for Val d'Aveto and a SIV of 20 m for Petrignacola. Volcanic clast populations are characterized by andesitic and/or basaltic lava clasts and felsitic (rhyolites to dacites) fragments (Aiello, 1976; Vannucci, Wezel, 1978).

In the Tripoli di Contignaco Fm (Aquitanian-Burdigalian p.p.) tuffaceous siltstones, with a grain size < 0.5 mm were probably emplaced by both fallout and/or epiclastic processes within a SIV of 15 m; strata with volcanogenic sediments have a MOT of 10 m in a VTT of about 15 m. Volcanoclastic material is calcalkaline rhyodacitic-dacitic in composition and is mainly represented by glass-shards (< 90 % vol.) and subordinate crystals. Imprints from hyaloclastic processes seem to be also present (Giammetti *et al.*, 1968; Borsetti *et al.*, 1984).

The siltstones of the Bisciario-like Fm (Late Chattian-Burdigalian p.p.) were emplaced by epiclastic processes, within a SIV of 80 m, and they have a grain size < 0.5 mm; the VTT of these deposits is about 10 m with strata having a MOT of 3 m. The calcalkaline volcanic detritus derives from explosive eruptions and is rhyodacitic-dacitic in composition (Mezzetti, Olivieri, 1964; Mezzetti, 1969; Borsetti *et al.*, 1979, 1983, 1984).

In the Epi-Ligurian Domain sandstones containing volcanogenic detritus and probably emplaced by epiclastic processes within a SIV of 40 m (Aiello, 1976) are also present in the M. Senario Fm (Oligocene *s.l.*). Nevertheless, due to their uncertain stratigraphic attribution they are not reported in Plate II.



2.7. Piemonte Tertiary Basin-Monferrato Sector (Italy: "7" in Plate I). The Piemonte Tertiary Basin was generated along the tectonic suture between Alpine and Apennine Chains. Within this basin, active since the end of Eocene to Miocene, it is possible to recognize several stratigraphic successions corresponding to different palaeogeographic structural domains (basins on "Alpine" crust and basins of Apenninic relevance) generally developed in episutural conditions (Biella et al., 1992). Particularly, in the Monferrato, which represents the NW extension of the Northern Apennines, the Neogene succession is made up of a quickly evolving basin, developed on the Apenninic arc front which overthrust Padano-Adriatic Foredeep; it is characterized by a thin thickness and is constituted of different sedimentary units divided by unconformities.

Volcaniclastic products (D'Atri, Tateo, 1994; Ruffini, 1995) are recognized within formations belonging to two different groups characterized by trachytic (Tonengo Calcarenites, Antognola, M. Zucaro and Rigoroso Marls) and rhyolitic composition (Pietra da Cantoni, Pteropodi Marls and Tonengo Calcarenites).

Within the first group (from Chattian p.p. to Aquitanian), tuffaceous siltstones have a volcanic imprint mainly given by the presence of glass-shards and crystals. They were generated by pyroclastic fallout and/or epiclastic processes; they have a homogeneous trachytic (alkaline) composition within a SIV of 40 m; strata of volcanogenic sediments have a MOT of 2 m, in a VTT of 28 m. Alterations of glass into Bazeolites and smectites are also present (Clari et al., 1988; Ruffini et al., 1991; Tateo, 1993; Bonci et al., 1994; D'Atri, Tateo, 1994; Ruffini, Cadoppi, 1994; Ruffini et al., 1994).

Within the second group (from Aquitanian p.p. to Langhian p.p.; 18.7 ± 0.1 Ma; 16.4 ± 0.2 Ma) the tuffaceous sandstones and siltstones having a calcalkaline rhyolitic composition contain 90 % vol. glass-shards and subordinate crystals, and were emplaced by fallout processes. They are present within a SIV of 50 m and volcanogenic strata have a MOT of 10 m. Glass is partially altered to zeolites (Clari et al., 1988; Ruffini et al., 1991; Ruffini et al., 1994).

2.8. Northern Apennines-External Sector, Adria Margin (Italy: "8" in Plate I). The external palaeogeographic sector of the Northern Apennines is characterized by continental thrusts and it is subdivided in Foredeep and Foreland (Boccaletti et al., 1990; Amorosi et al., 1995) of the Adria Margin. In this sector volcaniclastic deposits are found both in the Foredeep (Cervarola and Vicchio) and in the Foreland ramp (Bisciaro Group).

Sandstones and siltstones in the Cervarola Fm (Late Chattian-Early Aquitanian) were emplaced by syn- and/or post-eruptive turbiditic processes. They are characterized by a great amount of glass-shards (altered very often to analcime and zeolite) and subordinate crystals with a MOT of 6 m, VTT of 10 m and SIV of 80 m. Volcanic detritus seems to derive from differentiated magmas (Papini, Vannucci, 1993), most probably of rhyolitic composition.

In the Vicchio Marls the deposits, within a SIV of 450 m, can be subdivided into: (a) crystal- and lithic-rich sandstones characterized by calcalkaline andesitic and/or basaltic lava fragments (Late Aquitanian-Early Burdigalian) and (b) tuffaceous sandstones/siltstones whose volcanic detritus is mainly represented by calcalkaline rhyolitic glass-shards (Burdigalian p.p.). Strata of volcanogenic sediments were mainly emplaced by mass-flow processes (e.g. slumping of unstable pyroclastic deposits) and have a MOT of 0.3 m. Alterations of glass-shards into analcime is widespread as the result of burial diagenesis (Delle Rose et al., 1994 a, b).

In the Bisciaro Group (Late Aquitanian-Burdigalian p.p.) siltstones, sandstones and marly limestones have a SIV > 100 m, and a VTT of 18 m. They are characterized by calcalkaline rhyolitic to dacitic volcanic detritus (glass-shards, crystals and holocrystalline to felsitic-vitric lava clasts) which seem to be emplaced by both intra-basinal mass-flow and pyroclastic fallout processes (Guerrera, 1977; Guerrera et al., 1986; Amorosi et al., 1994; Balogh et al., 1993; Delle Rose et al., 1994 a). Particular volcanogenic markers (Coccioni et al., 1988; Montanari et al., 1988; Coccioni, Montanari, 1992; Deino, Montanari, 1992; Coccioni et al., 1994) comprise "Raffaello" Bed with a MOT of 0.3 m (21.2 Ma), "Mega P" Bed with a MOT of 6.9 m (18.5 Ma) and "Piero della Francesca" Bed with a MOT of 0.15 m (17 Ma). Volcanogenic sediments partially or totally replaced by siliceous deposits, bentonitic layers deriving from alterations of glass into clay minerals, zeolite-rich and glauconite-rich deposits are also present.

Within the S.Polo Marls (Macigno Fm, Oligocene-Early Miocene?) volcanogenic strata have a MOT of 0.06 m; the volcanic detritus consists of crystals and glass partially altered to clay minerals (Centamore et al., 1980). However, due to their uncertain stratigraphic position S.Polo Marls are not reported in Plate II.

2.9. Venetian Foreland Basin (Italy: "9" in Plate I). It was a site of active sedimentation from Upper Oligocene to Recent and is filled by a clastic wedge of more than 4000 m in thickness (Stefani,



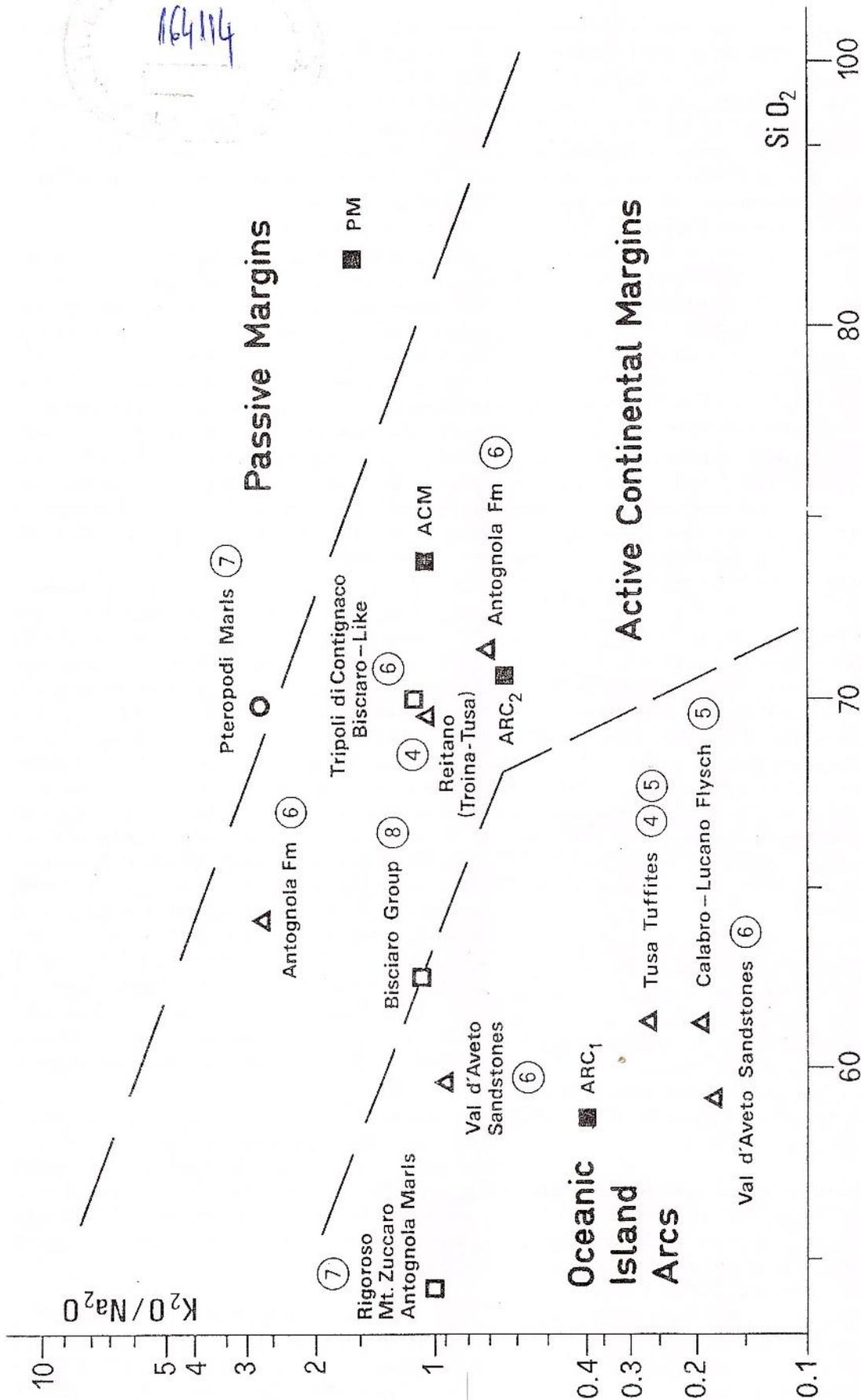


Fig. 1 - K₂O/Na₂O-SiO₂ diagram with the average whole-rock analyses of volcanogenic sediments from Table 1. Different fields related to main geotectonic settings are also reported (from Roser, Korsch, 1986). Symbols: open triangles=epiclastics; open circles=pyroclastics; open squares=epiclastics and/or pyroclastics; full squares=averages of sandstone suites related to different geotectonic settings: PM, Passive Margin; ACM, Active Continental Margin; ARC₁, Oceanic Island Arcs; ARC₂, Continental Margin Arc (from Bhatia, 1983). Numbers are referred to the various domains according to Plates 1 and 2. All the analyses have been recalculated on volatile-free basis.

Table 1
Main features of the volcanogenic sediments whose geochemical whole-rock analyses are used to discriminate between different geotectonic settings

| LOCATION IN PL. I | Basin / Domain | FORMATION | TYPE OF DEPOSIT | ANALYSED SAMPLES | REFERENCES |
|-------------------|--|---|---|------------------|---|
| 4 | Southern Calabria-Sicily Sector, "Flysch" Basin Dom. | Tusa Tuffites Sicilide Units | Lithic arenites (i.e. volcanarenites) | 3 | Ogniben, 1964; Guerrero, Wezel, 1974 |
| 4 | Southern Calabria-Sicily Sector, "Flysch" Basin Dom. | Reitano Flysch (p.p.) or Troina-Tusa | Lithic arenites (i.e. volcanarenites) | 21 | Puglisi, 1994 |
| 5 | Southern Apennines-Northern Calabria, "Flysch" Basin Dom. | Calabro-Lucano Flysch | Pheno-andesitic tuffites Pheno-quartz andesitic tuffites, Lithic arenites (i.e. volcanarenites) | 10 | Lanzafame et al. 1977; Critelli, Monaco, 1993 |
| 5 | Southern Apennines-Northern Calabria, "Flysch", Basin Dom. | Tusa Tuffites | Lithic arenites (i.e. volcanarenites) | 7 | Ardito et al., 1985 |
| 6 | Epi-Ligurian Domain Northern Apennines | Antognola Fm | Zeolite-rich volcaniclastic beds | 6 | Tateo, 1993 |
| 6 | Epi-Ligurian Domain Northern Apennines | Antognola Fm | Ash layers | 2 | Tateo, 1993 |
| 6 | Epi-Ligurian Domain Northern Apennines | Val d'Aveto Sandstones | Lithic arenites (i.e. volcanarenites) | 2 | Vannucci, Wezel, 1978 |
| 6 | Epi-Ligurian Domain Northern Apennines | Val d'Aveto Sandstones | Lithic arenites (i.e. volcanarenites) | 1 | Mattioli, unpubl. data |
| 6 | Epi-Ligurian Domain Northern Apennines | Tripoli di Contign. Bisciario-like | Ash layers | 4 | Mezzetti, Olivieri, 1964 |
| 7 | Piemonte Tertiary Basin, Monferrato | Pteropodi Marls | Ash layers | 2 | Clari et al., 1988 |
| 7 | Piemonte Tertiary Basin, SE Margin | Rigorous Marls, M. Zuccaro Marls, Antognola Marls | Volcano-sedimentary beds | 9 | D'Atri, Tateo, 1994 |
| 8 | Adria Margin Northern Apennines | Bisciario Group | Crystal-vitric tuffs | 1 | Balogh et al., 1993 |

1987). From Oligocene to Serravallian it behaved as a foreland basin of Dinaric domain. Volcaniclastic products occur within Montefumo Marls, M. Baldo Fm and S. Gregorio Sandstones (Late Burdigalian-Early Langhian). Sandstones, siltstones and marls are comprised within a SIV of 50 m, with a VTT of 2 m; volcanogenic strata whose MOT is 1 m, have a volcanic-component grain size ranging between 0.036 and 0.45 mm. Pyroclastic fallout processes seem to have supplied the deposits, which contain a small amount (3-4 % vol.) of rhyolitic glass and crystals (Stefani, 1987; Grandesso, Stefani, 1990).

3. Petrochemical features of volcanogenic sediments

The Upper Oligocene-Lower Miocene volcanogenic

sediments in the Western Mediterranean arc still not enough supported by petrochemical data (e.g. whole-rocks also including the non-volcanic components). In spite of geochemical analyses made on volcanogenic detritus (mainly lava clasts and glass-shards, very important to establish the magmatic affinity), this methodological approach disregards the non-volcanic clastic components, whose study could provide useful information about palaeogeography, geological settings and locations of volcanic source areas.

Whole-rock geochemical data made on modern sands (Maynard et al., 1982) as well as on ancient sandstone suites (Bhatia, 1983; Bhatia and Crook, 1983) have shown a great variability of major and trace elements, which reflects distinct source and tectonic setting. Furthermore, Roser, Korsch (1986), ac-



ording to the criteria suggested by Crook (1974) and Reading (1982), have recently detected three main geologic settings (active continental margin, passive margin and oceanic island arc) on a simple binary K_2O/Na_2O-SiO_2 plot.

In the diagram of Roser and Korsch (Fig. 1) we have plotted the Western Mediterranean volcanoclastic deposits, whose petrochemical data made on the whole-rock samples are known from the literature. We took into consideration the following successions: (a) Southern Calabria-Sicily Sector, (b) Southern Apennines-Northern Calabria, (c) Northern Apennines-Internal Sector (Epi-Ligurian Domain) and (d) Piemonte Tertiary Basin (Tab. 1).

The K_2O/Na_2O-SiO_2 diagram (Fig. 1) points out that not all the examined volcanogenic deposits seem to be related to sedimentary basins connected with active continental margins. In fact, some of these deposits, belonging to the Southern Calabria-Sicily Sector (Tusa Tuffites), the Southern Apennines-Northern Calabria Sector (Calabro-Lucano Flysch) or to the Epi-Ligurian Domain (Val d'Aveto Sandstones), show strong geochemical similarities with oceanic island arc settings.

The implication which stands out from these results is that the palaeogeographic context where the volcanically induced sedimentation of these deposits occurred, must be reevaluated. In fact, the subduction-related collisional nature of these volcanic events, supported by a great number of geochemical data from the literature, does not imply that the volcanic sources should be necessarily located on continental margins. Furthermore, the available data from these deposits (thickness, texture and, mainly, the ratio between volcanic and non-volcanic components) suggest that the distances between eruptive centers and source of non-volcanic sediment should be largely variable. In fact, the volcanogenic detritus can also be mixed with crystalline and/or sedimentary detritus, indicating that volcanic sources were close to continental crust. The Reitano Flysch (or Troina-Tusa; *sensu* Guerrera, Wezel, 1974) of the Sicilian Maghrebian Chain represents one example: the sandstones contain volcanic clasts only in the lower and middle part of the succession and the volcanic detritus never exceeds 25 % vol. (Puglisi, 1979; Casola *et al.*, 1992). In other cases the predominance of volcanic detritus (e.g. Val d'Aveto Sandstones, with about 80 % vol. of volcanic component) or its exclusive presence (e.g. Tusa Tuffites; Ogniben, 1964 and Calabro-Lucano Flysch; Critelli, Monaco, 1993, respectively with about 90 % and 70 % vol.) suggests a very short distance between eruptive centers and depositional areas, and a significant distance from the sediment source areas of non-volcanic clasts.

4. Distribution of primary volcanics

In the Western Mediterranean calcalkaline magmatism is widespread during the Oligocene-Miocene time, testified by many outcrops of primary volcanics (ranging from 33 to 13 Ma, Lower Oligocene-Middle Miocene). A climax of eruptions occurred between 24 and 17 Ma (Upper Oligocene-Lower Miocene) and the Western Mediterranean sectors affected by this volcanism comprise: Sardinia, Alboran Sea, Valencia Through, Gibraltar Arc, Kabyldes, Po Plain, Central-Northern Apennines and the adjoining Carpatho-Pannonian Arc. Eruptive styles range from weak to moderate lava effusions to violent pyroclastic eruptions. A summary of activity type, age and compositions of the erupted calcalkaline products in these areas is reported in Table 2. An alkaline volcanism, mainly concentrated in the Massif Central (France) also occurred since about 20 Ma.

Volcanism in the Balearic Area (Valencia Through, Mallorca island and Gata Region; "A" in Plate I) could be related to the formation of marginal basins behind the African subduction zone. The Valencia Through represents an aborted rift formed before the Early Miocene opening of the Western Mediterranean Sea (Maillard, Mauffret, 1993). Site 123 of the Deep Sea Drilling Project (Ryan *et al.*, 1973; Rivière *et al.*, 1981) drilled a rhyodacitic hyaloclastite with a thickness > 100 m, which is the result of a potassic calcalkaline magmatism; the K/Ar age of this deposit is 21.4 ± 1.2 Ma (Rivière, 1988). Within the Gata Region basaltic to andesitic lavas erupted during Burdigalian-Langhian (Bellon *et al.*, 1983), whereas at Mallorca rhyolitic lavas (18.6-19 Ma) crop out (Marti *et al.*, 1992).

In the Algerian Area ("B" in Plate I) a rhyolitic explosive phase (15-16 Ma) in the Edough Massif (Kabyldes Region) is recorded by Monié *et al.* (1992) probably connected with a northward subduction of the African Plate beneath the Betic, Rif and Kabylic zones (Rehault, 1981).

Eastwards, Late Oligocene-Early Miocene volcanism occurred in Sardinia ("C" in Plate I), between 28 and 13 Ma. The volcanic activity of Sardinia during this time interval can be summarized in four principal phases (Fig. 2):

(a) acid-intermediate upper volcanics characterized by prevailing rhyolitic-dacitic pyroclastic flows whereas effusive products are scarcely represented. The age of these deposits varies from 17 to 13 Ma and fine pyroclasts could have been transported by fallout processes to great distance during the emplacements of the larger volume ignimbrites;

(b) basic-intermediate upper volcanics spread out as subaerial and submarine basaltic to andesitic lava



Table 2
Main features of the primary volcanics related to the
Late Oligocene-Lower Miocene magmatism in the Western Mediterranean

| LOCATION IN PL. I | AREA | TYPE OF ACTIVITY | COMPOSITION AND AGE |
|----------------------|---|------------------------------|---|
| A | BALEARIC Gata Region Mallorca Island Site 123 (Valencia Trough) | explosive and effusive | -basaltic to andesitic lavas-Gata (Burdigalian-Langhian) -rhyolitic lavas - Mallorca (18.6-19 Ma) -dacitic and rhyodacitic hyaloclastite Site 123 (21.4 Ma) |
| B | ALGERIAN Kabylies (Edough Massif) | explosive | -rhyolitic "magmatism"(15-16 Ma) |
| C | SARDINIA | explosive and effusive | (a) mainly rhyolitic to dacitic pyroclastics (17 -13 Ma) (b) mainly basaltic to andesitic lavas (19 -16 Ma) (c) mainly rhyolitic to dacitic pyroclastics (22 -20 Ma) (d) mainly basaltic to dacitic lavas (28 - 23 Ma) |
| D | CORSICA | explosive | -vitroclastic ignimbrite (17.8 Ma) -rhyodacitic ignimbrite |
| E | LIGURIAN SEA | effusive | -potassic trachyandesite-trachyte (18 Ma) |
| F | PO PLAIN Mortara 1drilling | effusive | -andesite (16.1 - 20.9 Ma) -andesite/dacite Lower Oligocene-Middle Miocene) |
| G | NORTHERN APENNINES Pieve S. Stefano 1 drilling | effusive | -andesitic lavas (Oligocene-Miocene) |
| H | PANNONIAN BASIN Styrian Basin Carpathian Region | explosive and effusive | -acid to intermediate volcanics (17-15 Ma) -rhyolitic to andesitic pyroclastics /lavas (20-18 Ma) -rhyolitic to dacitic pyroclastics (17-16 Ma) -andesitic pyroclastics (16-14 Ma) |

flows with minor phreatomagmatic products. Timing of these eruptions is 19-16 Ma with no possibilities for pyroclasts to be transported away by fallout processes (Assorgia et al., 1995); (c) acid-intermediate lower volcanics made up of large-volume prevailing rhyolitic-dacitic pyroclastic flows, associated with largely dispersed pyroclastic fallout deposits; lavas are subordinate. The age of these deposits is between 22 and 20 Ma (Savelli et al., 1979; Beccaluva et al., 1985); (d) basic-intermediate lower volcanics mainly constituted by basaltic to dacitic lava flows and subordinate pyroclastic pro-

ducts. These deposits having a calcalkaline to tholeiitic affinity range from 28 to 23 Ma (Coulon, 1977; Savelli et al., 1979). The explosive activity did not give rise to widely dispersed pyroclastic fallouts from the source vents (Assorgia et al., 1995).

Probably linked to Sardinia volcanism, vitroclastic ignimbrites were identified in the southern and southwestern parts of Corsica ("D" in Plate I). The age of this volcanism is 17.8 ± 1.5 Ma (Bellon, Letouzey, 1976) for a biotite-rich ignimbrite cropping out near Bonifacio (Corsica), and to a rhyodacitic formation in the western Corsica sectors.

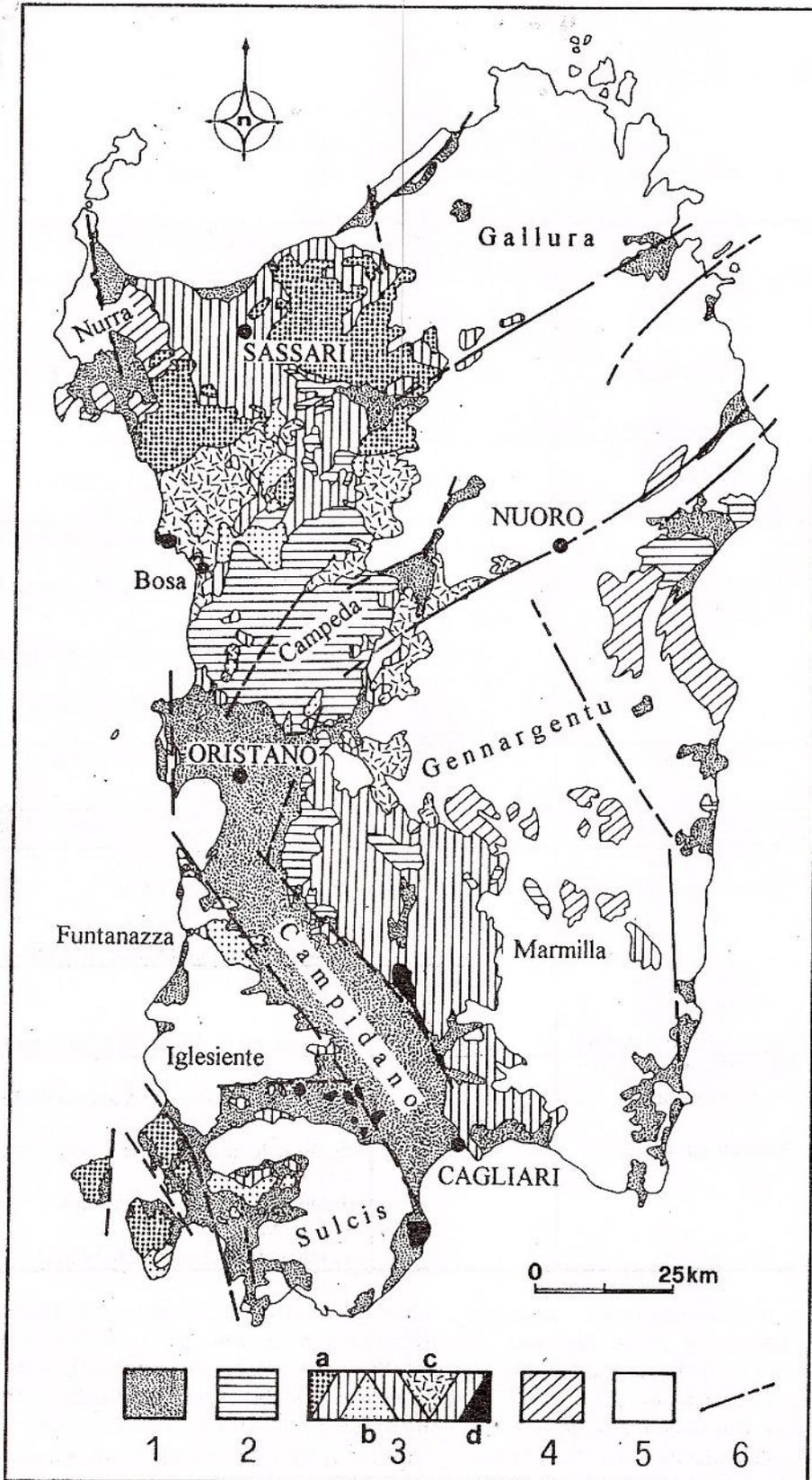


Fig. 2 - Geological scheme of Sardinia pointing out the volcanic activity occurred during the Late Oligocene-Middle Miocene. 1, Plio-Quaternary deposits; 2, Plio-Pleistocene alkali, transitional and subalkali basalts with related evolved products (trachytes, dacites, rhyolites, phonolites); 3, Miocene continental and marine polygenic sediments including (a) acid-intermediate upper volcanics (17-13 Ma), (b) basic intermediate upper volcanics (19-16 Ma), (c) acid-intermediate lower volcanics (22-20 Ma), (d) basic-intermediate lower volcanics (28-23 Ma); 4, Mesozoic carbonatic sequence; 5, Paleozoic metamorphic basement; 6, Tertiary faults.

In the Ligurian Sea, Rehaut et al. (1985) reported a "tristanite" lava (trachyandesite-trachyte) dredged *in situ* ("E" in Plate I), with a K/Ar age of 18 Ma. This fact seems to suggest that volcanoes were active during the Burdigalian spreading phase of the Western Mediterranean.

In the Italian peninsula two important deep wells drilled Oligocene-Miocene volcanics: "Mortara 1" well on the Po Plain ("F" in Plate I; Cassano et al., 1986) and "Pieve S. Stefano 1" well in the Central Apennines ("G" in Plate I; Anelli et al., 1992). As regards "Mortara 1", available data pointed out andesitic lavas of 16.1-20.9 Ma (Bonci et al., 1994) and andesite to dacitic lavas of Lower Oligocene-Middle Miocene time (Ruffini, 1995). Andesitic lavas of rather uncertain age (Oligocene-Miocene) represent the products drilled at "Pieve S. Stefano 1" well.

Calcalkaline magmatism through the Pannonian Basin ("H" in Plate I; 20-14 Ma; Szabò et al., 1992; Harangi, 1993) and Styrian Basins (17-15 Ma; Ebner, Sachsenhofer, 1995) is mainly characterized by both pyroclastics and lavas comprising rhyolitic, dacitic and andesitic compositions.

5. Discussion and final remarks

It is not easy to detect a direct relationship between primary volcanic material of Western Mediterranean and volcanogenic sediments dispersed throughout the Apennines-Maghrebides-Betic Cordillera. By contrast, in some Western Mediterranean sectors primary volcanic activity and related volcanogenic sediments are well exposed in Upper Oligocene-Lower Miocene volcano-sedimentary successions (e.g. Sardinia) or pointed out by deep-sea drillings (e.g. Valencia Trough).

Tuffaceous sandstones/siltstones (*sensu* IAVCEI, 1993) volcanolithic sandstones and crystal- to vitric-rich volcanoclastic sandstones/siltstones are the most common lithotypes among the volcanogenic sediments of Western Mediterranean. Epiclastic processes can be obviously invoked for the generation of volcanolithic sandstones but also for most of the tuffaceous sandstones/siltstones and crystal- to vitric-rich volcanoclastic sandstones/siltstones; the process implies the erosion and/or reworking of pyroclastics and lavas shortly after their fragmentation and slumping into the basins. In fact, thicknesses, grain size variations, textural features, wide but non-homogeneous areal distributions and crystals-glass shards-lavas relationships in many of these deposits are consistent with *epiclastics* rather than *pyroclastics*. Although tuffaceous sandstones and siltstones can be also commonly considered as volcanogenic sediments produced by direct fallout of volcanic ma-

terial within the basin, a comparison with modern volcanoclastics of active island-arc volcanoes indicate that the abundance of fragile glass-shards in several Western Mediterranean volcanogenic sediments could also imply mass-flow processes. In fact, delicate pyroclastic grains of fine sand and silt crossing the shorelines with practically no residence there can preferentially move away, then deposited as over-bank turbidites and/or likely turbidite fan in deep water (Kokelaar, Romagnoli, 1995). Therefore, sub-aerially produced fragile pyroclasts can be also found where modes of volcanically induced sedimentation were mainly through turbiditic processes (Cas, 1979).

Very often, regarding the Apennines and Maghreb Chain volcanoclastics, Sardinia had often been indicated as the source area; however, its location during Late Oligocene-Early Miocene could not fit with the inferred epiclastic mass-flow processes because it was located too far from the sedimentation basins. In fact, the Corsica-Sardinia block started to drift from the Provençal Margin only during the Burdigalian (Carmignani et al., 1994; Vigliotti, Langenheim, 1995). In addition, possible epiclastic processes from Sardinia would be prevented by the complex palaeogeography (i.e., Dewey et al., 1989; Doglioni, 1992; Guerrero et al., 1993; Maillard, Mauffret, 1993). Due to distances of about 1.000 km between Sardinia and the Upper Oligocene-Lower Miocene Apennine sedimentary basins (Montanari et al., 1994), a relationship with Sardinia volcanic activity through fallout processes, can be plausibly invoked only for some thin and very fine-grained volcanogenic pelites between 22-20 and 17-13 Ma, i.e. the same time intervals corresponding to the catastrophic Sardinia eruptions associated with largely dispersed fallout deposits.

Resedimentation of primary volcanic material by turbidity currents may have represented the most common style of transport of the Western Mediterranean volcanogenic sediments (*epiclastics*) into Upper Oligocene-Lower Miocene sedimentary basins. Nevertheless, fine-grained ash layers, generally altered to clay minerals, can be also recognized as possible distal *pyroclastics*.

Most of the Upper Oligocene-Lower Miocene volcanogenic material dispersed throughout the Apennines, Maghrebides and Betides was erupted by volcanic source vents closer than Sardinia: available data seem to support the existence of volcanic arc systems adjacent to the sedimentary basins, and located in internal position with respect to the basins themselves.

In Plate III palaeotectonic and palaeogeographic reconstructions along the Western Mediterranean during Late Oligocene-Aquitania and Burdigalian-



Langhian times can be seen. Volcanic activity was inferred from distribution of volcanogenic sediments. In particular, cross sections of the southern chains (Betic Cordillera, Maghrebides and Southern Apennines) and palaeogeographic reconstructions of the Northern Apennines and Southern Alps are traced out.

A microplate named "Mesomediterranean Terrane" (Guerrera et al., 1993) is located between the European and African Plates (see also Doglioni, 1992), and it is bordered northward by the Piemontese Ocean and southward by another Ocean (North Africa "Flysch" Basin or Maghrebian). The southern margin of this microplate was deformed in the Aquitanian when the North Africa "Flysch" Basin was characterized by flysch deposits and volcanoclastic material. This sedimentation continued in the Burdigalian-Langhian time and became younger toward the external zones. Most probably, an intense volcanism was located along external margins of the "Mesomediterranean Terrane". Anyway, besides the eruptions occurred in Sardinia, the preliminary models from Dewey et al. (1989) and Guerrera et al. (1993) about kinematic evolution of the Western Mediterranean oceanic basins suggest, during Aquitanian-Burdigalian time, the existence of an additional source area of volcanism, as a result of an extensional tectonics. In addition, as already suggested by Ruffini et al. (1994) and Delle Rose et al. (1994 b), respectively for volcanogenic sediments of the Piemonte Tertiary Basin and the Northern Apennine (External Sector), Upper Oligocene-Lower Miocene primary volcanics might have undergone quick erosional processes and/or subsidence and remnants of volcanic edifices could also have been buried and concealed at depth by underthrusting processes during the post-Early Miocene tectonic evolution (e.g. Central Apennines, "Pieve S. Stefano 1" well). Large amounts of buried Tertiary primary volcanics underneath the Po Plain (e.g. "Mortara 1" well) cannot be excluded as one of the main source areas.

The results obtained by analyses of whole-rock samples of volcanogenic sediments and the affinity of some of them with oceanic island-arc tectonic settings could account for the abundance of the volcanic detritus in the sediments. The palaeogeographic implication is that eruptive centers should have been not necessarily located on continental margins but rather distant from them, to give rise to a volcanic arc system adjacent to the sedimentary basins. Thus, volcanic activity could have occurred along major strike-slip faults controlling plate margin convergence (Sylvester, 1988).

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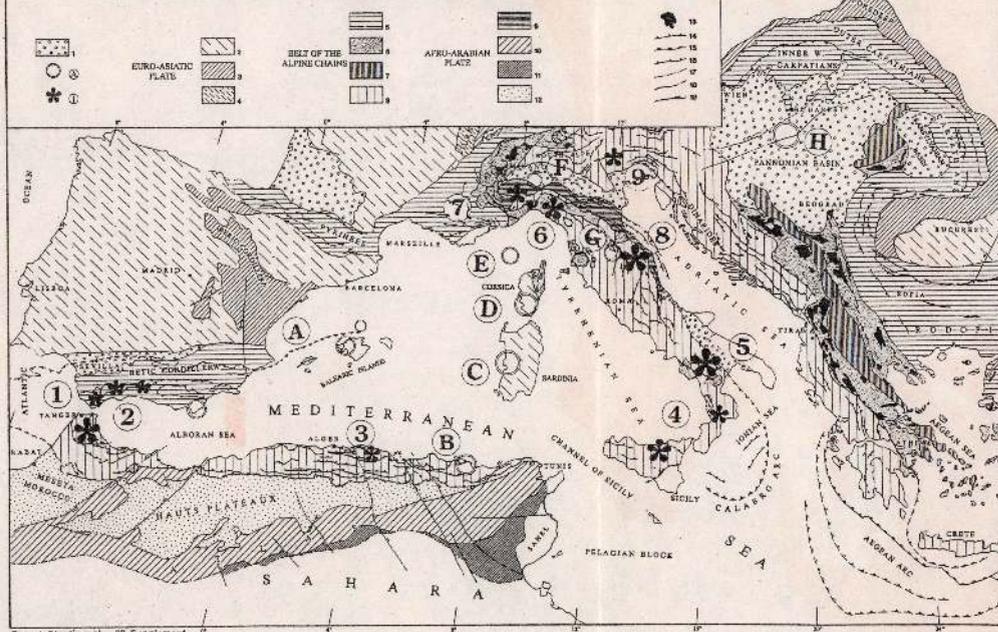
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TECTONIC SKETCH OF THE MEDITERRANEAN AREA

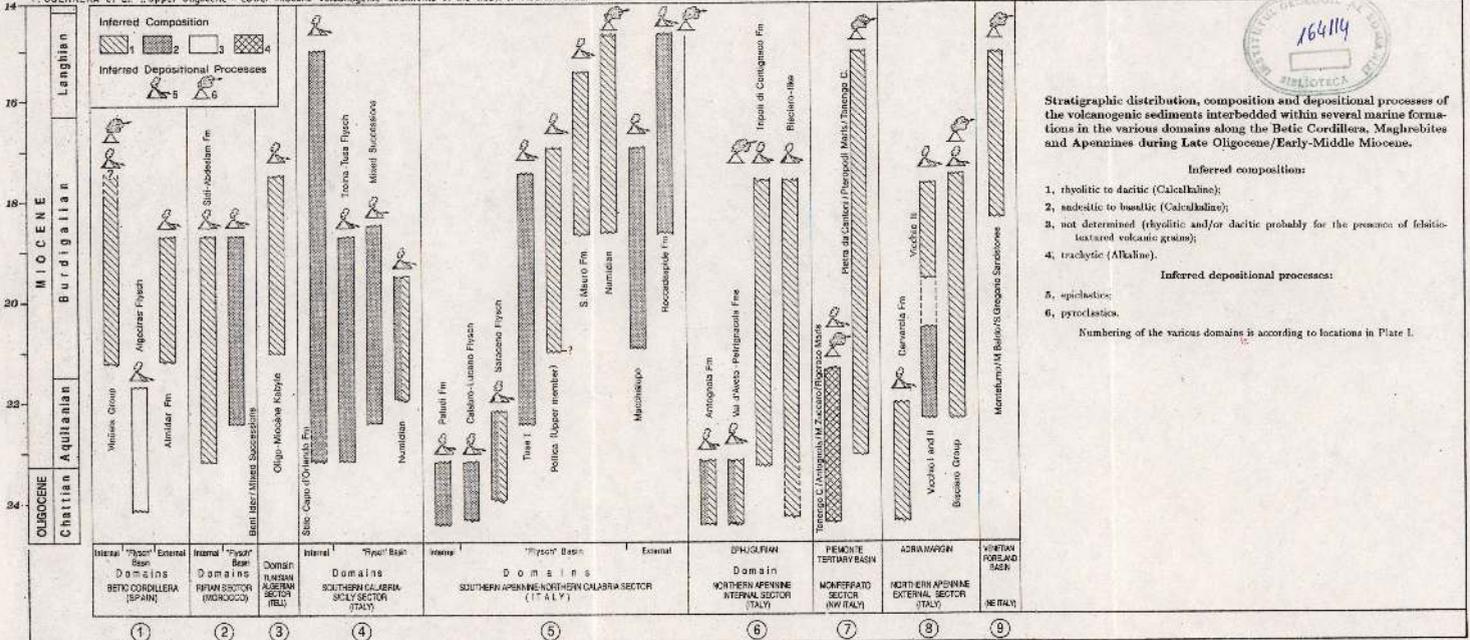


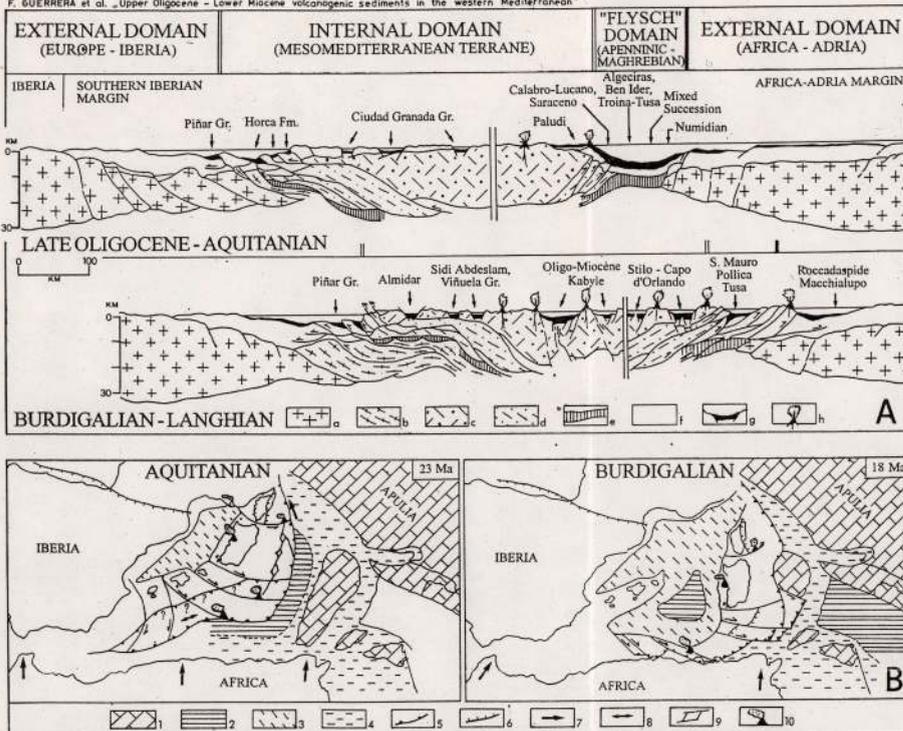
Tectonic Sketch of the Mediterranean (from Boccaletti, Dainelli, 1984; modified). Circles labelled by a capital letter = locations of Late Oligocene-Early Miocene primary volcanism; asterisks labelled by a number = locations of Late Oligocene-Early Miocene volcanogenic sediments; sizes of circles and asterisks vary as abundance and/or distribution of the deposits increase.

1. Areas of the main Neogene-Quaternary fill.
- Euro-Asiatic Plate:
2. Foreland, undeformed during the Alpine orogenesis.
3. External platform covers, more or less deformed in the Neogene-Quaternary phases and Alpine-Carpathian molasses.
4. Jura Chain.
- Belt of the Alpine Chains:
5. Elements of the Euro-Asiatic margin chains (Betics, Alps, Carpathians, Balkans), including Pyrenees and Provence.
6. Belts of outcropping orogenic remnants (flysch, ophiolites, mélange, etc.) including continental elements.
7. Intermediate massifs (Pelagonia zone).
8. Elements of the Afro-Arabian margin chains (Kil. Tell, Apennines, Southern and Eastern Alps, external Hellenides, external Helonides).
- Afro-Arabian Plate:
9. Po Plain-Adriatic buried folded zones.
10. Foreland chain (Atlas).
11. Platform covers, more or less deformed in the Neogene Quaternary phases (external Atlantic zone).
12. Hauts-Plateau and Moroccan Meseta.
13. Major ophiolitic masses of Alpine age.
14. Major thrust front.
15. Hellenic trench system.
16. Thrust systems related to the arc.
17. South-Atlantic Arcs.
18. Some major lineaments.
19. Direction of horizontal displacement along lineaments.



F. GUERRERA et al., Upper Oligocene - Lower Miocene volcanogenic sediments in the western Mediterranean*





Palaeotectonic (A) and Paleogeographic (B) reconstructions along the western Mediterranean basins in the Aquitanian-Langhian time. Schematic locations of volcanic activity within the two cross sections are mainly the results of the inferred processes of volcanically induced sedimentation, areal distribution of volcanogenic sediments and their widespread occurrence along the stratigraphic columns.

A: Paleotectonic reconstruction through the Betic Cordillera, Maghrebides and Southern Apennines cross sections (From Guerrero et al., 1993, modified).

- a. European (Iberia) and Africa (Adriatic) pre-Triassic basements;
- b. Idem, metamorphosed during the Alpine orogenesis;
- c. Pre-Triassic basement of the Mesomediterranean Terrane (for more details about the "M.T." see Guerrero et al., 1993);
- d. Idem, metamorphosed during the Alpine orogenesis;
- e. Oceanic crust;
- f. Mesozoic-Tertiary cover;
- g. Upper Oligocene/Lower-Middle Miocene marine sedimentation with associated volcanogenic deposits;
- h. Inferred volcanoes.

B: Paleogeographic reconstructions (from Dewey et al., 1989, modified).

- 1. Mesozoic Carbonate Platform;
- 2. Oceanic Crust;
- 3. Extended Continental Crust;
- 4. Basin on Continental Crust;
- 5. Active Thrust;
- 6. Old Thrust;
- 7. Africa-Europe Motion;
- 8. Extension;
- 9. Pull-Apart Basin;
- 10. Volcanism.



CONNECTIONS OF EASTERN PARATETHYAN PALEOBASINS WITH TETHYAN SEAS IN THE MIDDLE AND LATE MIOCENE

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Key words: Eastern Paratethys. Eastern Mediterranean. Early Tarkhanian. Early Konkian. Middle Karaganian. Late Veselyankian. Late Meotian. Connections. Polyhaline assemblages.

Abstract: The widest connections between Eastern Paratethyan paleobasins with the Tethyan seas were in the Early Tarkhanian and Early Konkian. In the Varnian (Middle Karaganian), Late Veselyankian (end of the Konkian), and the more so in the Late Meotian time, on the contrary, short term connections existed. Most probably the seas of Eastern Paratethys communicated with Eastern Mediterranean through a strait which might have existed in the south-east of Paratethys, in the areas of Predtalyshky and Akero-Araxinsky depressions (East Turkey and Iran). The existence of South-Eastern connections is proved by the presence of diverse polyhaline assemblages in Transcaucasia and Transcaspien. The presence of Indo-Pacific genus *Obortia* (Gastropoda) is recorded in Lower Tschokrakian deposits.

The succession of basins of different type (those with normal salinity to considerably freshened, brackish-water) which replaced each other in Eastern Paratethys depended mainly on the character of their connection with the World Ocean waters. Changes of the hydrologic regime in these seas influenced the community structure of biota. The freer the communications of Eastern Paratethys with the World Ocean were, the more polyhaline and diverse fauna inhabited it. Impediments in communication with open sea waters led to the decrease in salinity, to the extinction of polyhaline forms and flourishing of marine euryhaline species and to the origin of endemics, the numerical increase of the latter depending much on the duration of such a basin with abnormal salinity. After almost a complete closure of the basin, marine elements became extinct and replaced by brackish-water species which could adapt themselves to the existence in brackish waters with ion composition, different from that characteristic of the seas with normal salinity (Caspian type of the basin, in which ions SO_4^- , Ca^{++} etc. prevail).

Recent publications dealing with Middle Miocene fauna of Eastern Paratethys show that practically all researchers of the basins of that time are unanimous in believing that the freest communications with the World Ocean were in Tarkhanian and Konkian times. Indeed, there existed in Early Tarkhanian sea diverse mollusc species, pteropods including, benthic

and planktonic foraminifers, diverse nannoplankton; sea urchins, brachiopods and corals were also encountered. The salinity of the sea was normal at that time (more than 32⁰/₀₀) but later on the fauna became a little more impoverished due to fact that communications of the Tarkhanian sea with open sea waters weakened, though the salinity in the basin seems to have never been lower than 30⁰/₀₀ (Gontcharova, 1989).

The succeeding Chokrakian basin enlarged somewhat its limits, and though its communications with open sea waters seem to have been rather wide in Early Chokrakian time, nevertheless, the salinity in this paleobasin deviated somewhat from the normal, but almost never decreased lower than 25⁰/₀₀. The number of polyhaline species reduced, nannoplankton composition became considerably less diverse. In the second half of the Chokrakian the sea began closing and freshening, which resulted in the extinction of the overwhelming majority of marine species.

The Karaganian basin was still more isolated, its salinity having been far from normal. Its benthic fauna was sharply impoverished and endemic, nannoplankton was absent (Muzylev, Golovina, 1987). Only in the Middle Karaganian (Varnian time) a short term communication with Tethyan waters re-occurred, the result of which was the prochoresis of a number of marine mollusc species, and also of rare nannoplankton representatives. Abnormal salinity of



the Karaganian sea did not change considerably during this episode.

At the beginning of the Konkian (Sartaganian), Eastern Paratethys was rather widely connected to the World Ocean, and its salinity became close to normal (not less than 30⁰/00). The Sartaganian basin was populated by marine molluscs (among which not a few polyhaline species existed), by bryozoans, diverse benthic and planktonic foraminifers, ostracods, comparatively diverse nannoplankton, and also echinoderms could be encountered in it. The Sartaganian sea existed for a short time. Progressing impediments in the communication with open sea waters led to the formation of the semimarine Veselyanka basin which was invaded by euryhaline, often endemic species that had arisen already in most freshened marginal parts of the Sartaganian sea. However, at the very end of Konkian time a short term ingression of marine waters into the Transcaspien region (Western Turkmenia etc.), Georgia, Ciscaucasia, lower reaches of the Don River, Cisazovia and other areas reoccurred. Rather polyhaline mollusc species (*Turritella*, large *Natica*, *Murex* and other forms), and, according to data of Ts.D.Minashvili, abundant *Helicopontosphaera kamptneri* (in the sections of Western Georgia) invaded the basin.

After this ingression, due to the noticeable reduction of the links with open sea waters, the sea again began to freshen. All groups of marine organisms impoverished considerably (their genera and families reduced in number). Among the molluscs, euryhaline species, survived from Late Badenian-Konkian time, and quite a number of new endemics settled in the basin (Paramonova, 1994). Only in the west of the basin, in its Vienno-Pannonian, Dacic and partially Euxinian parts, so-called hemisteno haline (term introduced by Kojumdjieva, 1969) species continued to exist in the Early and in the beginning of the Middle Sarmatian. Impoverished planktonic foraminiferal assemblages (Bobrinskaja, Kurenkova, 1986) and most diverse nannoplankton associations (Muzylev, Golovina, 1987) are known to occur in western parts of the Paratethys at that time as well. Among benthic foraminifers, new Mediterranean immigrants appeared at the beginning of the Middle Sarmatian (Venglinsky, 1975). Isotopic content of oxygen in Sarmatian bivalves (¹⁸O) is indicative of an influx of oceanic waters in the Early and at the beginning of the Middle Sarmatian (Kijashko, Paramonova, 1987). All this suggests that during the Early and partly Middle Sarmatian the Paratethys maintained the links with open sea waters, though communications with the Ocean were rather hampered. If there had been no communications at all, then the hydrology of the sea would have changed

(at least, its salinity), marine elements would have died out, and a group of brackish-water species would have prevailed (such as Dreissenidae among Bivalvia, Neritidae and Pyrgulidae among Gastropoda, and other groups), that is, the Sarmatian sea would have become very similar to the Pannonian lake-sea (after its closure in the Middle Sarmatian), or to some other brackish-water basins of Caspian type. High endemism and peculiarity of Early and Middle Sarmatian fauna can be explained by its existence during a long time (not less than 2 m.y.) under conditions of hampered communications with open sea waters, rather than by its complete isolation from the World Ocean. Very monotonous and poor mollusc assemblage and those of other groups, the absence of the nannoplankton in Late Sarmatian sea allow to suggest that communications of the Paratethys with the World Ocean were considerably disturbed or, may-be, even interrupted completely.

In the Early Meotian, communications with the World Ocean were reestablished. Evidently this communication occurred through a number of intermediate basins, because the endemism of Early Meotian sea was high from the very beginning (Iljina et al., 1976). Unchanged marine elements penetrated the Euxino-Caspian area (Western Georgia) during the maximum transgression in the second half of the Early Meotian (Iljina, 1980). At that time the basin was inhabited also by pteropod molluscs (*Spiratella*) and by a rather diverse nannoplankton (Semenenko et al., 1995). A diverse marine mollusc assemblage was reported also from the Dafni Formation in the Strimon River basin (NE Greece). It is true, though, that alongside with some polyhaline forms, not a few number of brackish-water elements were also encountered (Stevanović, Iljina, 1982).

In the Late Meotian the basin closed and lost its links with open sea waters. Brackish-water forms (*Congerina*, *Theodoxus*, *Turricaspia*, *Pseudamnicola* etc.) became predominant in the basin. In the Late Meotian there was a short term ingression of sea waters, the traces of which can be found in Western Georgia and in the Kerch Peninsula as rare findings of some euryhaline marine forms, such as *Sphaeronassa*, *Mactra* and some others (Iljina et al., 1976; Iljina, 1980).

Such is the general pattern of the change of the hydrology (Fig.) and faunal characteristics in the paleobasins of Eastern Paratethys in Middle and Late Miocene times.

While there is no doubt about the existence and the nature of water exchange between the basins of Eastern Paratethys and World Ocean at certain times, the location of straits, through which this water exchange and prochoresis of marine fauna occurred, has not



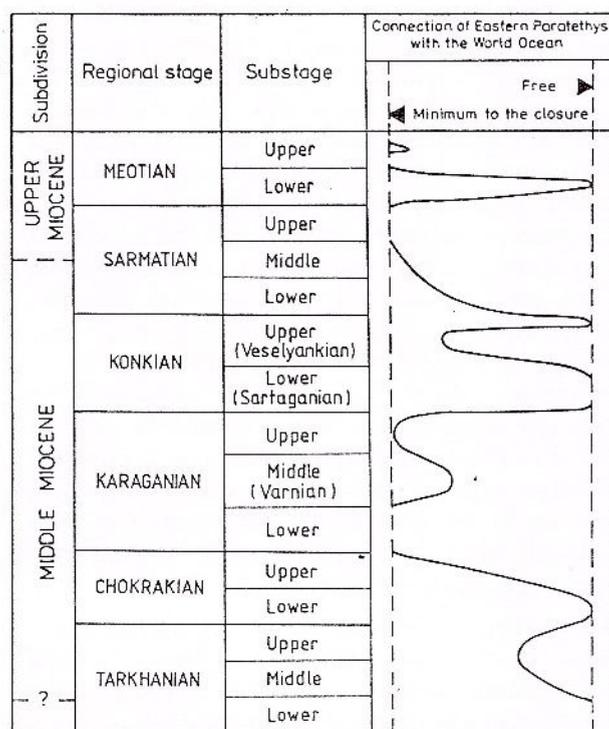


Fig. - Character of communication of Eastern Paratethyan basins with the World Ocean in the Middle and Late Miocene

been known definitely and it is still a subject of discussion.

Before proceeding with the analysis of Eastern Paratethys basins, the relevant data on the Recent Black Sea should be discussed.

The Black sea with the salinity of 17-18⁰/₀₀ is inhabited by an impoverished euryhaline marine fauna of the Mediterranean subtropical biogeographic realm (Golikov, Starobogatov, 1968). Among the molluscs (about 80 bivalve species and more than 85 species of shelled gastropods), only 20 bivalve species and 16 gastropod species are encountered near Bosphorus, that is in the area of the connection of the Black Sea with Mediterranean waters, where the salinity is, naturally, higher (20⁰/₀₀) than in the entire Black Sea aquatorium. Only from this area echinoderms are known - the starfish *Mathasterias glacialis* (L.), the sea urchin *Echinocyamus pusillus* (O. MÜLL.) and a number of marine ophiuroid species [Identification guide to...1972]. These animals are absent from other regions of the Black Sea.

It is to such areas, close to the straits through which more salted waters penetrated and prochoreses of marine forms occurred, that the most polyhaline species of the inland seas with abnormal salinity are confined. Hence, the location of the straits,

through which marine waters penetrated into Eastern Paratethys from the adjacent basins, may be determined on the basis of the distributional pattern of the most diverse polyhaline fauna.

Let us begin with the Tarkhanian. Most likely, the Tarkhanian sea was connected to the open sea waters in the north-west of the basin. One can suggest the existence of a strait and the links with Western Paratethys in the northern Black Sea region (Gontcharova, 1989). In the Dnieper River basin (village Tomakovka) 40 bivalve species and 20 gastropod species are present, the considerable percent of the molluscs being represented by polyhaline species common with those from the Carpathian and Badenian basins (among gastropods, these are the members of *Turritella*, *Nassarius*, *Aporrhais*, *Clanculus* and others). However, according to the data of Gontcharova (1989), suggestions exist that some polyhaline mollusc species, widespread in Ustjurt, Georgia and Western Ciscaucasia, penetrated into it through the south-eastern strait, stretching from Eastern Mediterranean sea to Iran, Akero- Nizhnearaxinsky and Predtalysky troughs (Gontcharova, 1989). To state it on the basis of gastropod data is prematurely for the present.

Analysis of the distribution of Chokrakian molluscs (particularly gastropods) indicates that the connection of Early Chokrakian sea with open sea waters might have been in the south-east of the basin. Gastropod assemblages are the most diverse in Ciscaucasia, in the Kerch Peninsula, in Georgia and Transcaspien region (northern Kara Bogas area, Krasnovodsk Peninsula, West Kopetdag - Geokoba Ridge). Farther westwards gastropod diversity sharply decreases.

It was in Early Chokrakian time that representatives of Indo-Pacific genus *Obortio* appeared for the first time in Eastern Paratethys and spread widely in Western and Central Ciscaucasia, in the Kerch Peninsula, in Georgia, Ustjurt, in northern Kara-Bogas area, in Krasnovodsk Peninsula, and in Western Kopetdag (Iljina, 1993). Species of this genus have not been yet reported from the region west of the Kerch Peninsula. This suggests that the Chokrakian sea had no connections with open sea waters in the west. This sea was connected with open sea waters rather in the south-east of the basin, probably by way of Akero- Nizhnearaxinsky and Predtalysky troughs, as it was pointed out by Gontcharova (1989). Link of the Chokrakian sea with the Indo-Pacific were not direct but might have occurred only through Eastern Mediterranean, from where also many other Mediterranean gastropods might have immigrated, which inhabited the eastern part of the Chokrakian sea.

A new prochoresis of the fauna occurred in the Middle Karaganian, when marine molluscs, close in their taxonomic composition to the greatly impoverished Konkian assemblages, reappeared in eastern regions of the Varnian sea (Mangyshlak, Ustjurt, Georgia, Ciscaucasia). And again, the fact that they were distributed mainly in the eastern part of the basin allows to suggest the existence of a strait somewhere in the south-east, probably in the same region where the strait had existed in Chokrakian time.

Diversity and widespreading of polyhaline and relatively polyhaline molluscs in the Sartaganian basin indicate definitely that the latter was connected with open sea waters in the south-east. In Eastern Georgia (station Agara) from among 42 mollusc species, 30 were the most polyhaline, in the Western Ciscaucasian region (settlement Tulsy) 26 from 59, in Mangyshlak 26 from 57, and in Ustjurt 53 species were polyhaline out of 102 bivalve and gastropod species. At the same time, in more western parts of the basin, the diversity of the molluscs decreased noticeably, and only single species are polyhaline. Besides, it is difficult to determine to what developmental stage, Sartaganian or Veselyanian, in the history of the Konkian sea one or another mollusc assemblage belongs.

In the Transcaspiian area (the region near Karabogaz-gol), in Georgia (village Naspere and others), Ciscaucasia (Dubrovaya Mountain), in the lower reaches of the Don River (Novocherkassk), in Cisazovia and in other areas, at the very end of Veselyanian time there occurred a new prochoresis of marine molluscs, representatives of *Turritella* inclusive, which were almost unknown in the Sartaganian, but they contain fewer polyhaline forms than Sartaganian assemblages. In Ciscaucasia, for instance, there are 15 polyhaline species out of the total 46 species, and in the lower reaches of the Don River only 6 out of 29 species are known. The confinement of these marine assemblages to the eastern and south-eastern parts of the basin again suggests its connections with the World Ocean somewhere in the south-east. At the same time, the presence of polyhaline representatives of microfauna and of nannoplankton (Barg, Nosovsky, 1993; Musylev, Golovina, 1987) in Konkian deposits, directly underlying Lower Sarmatian beds in the western part of Eastern Paratethys, may indicate communications of the Konkian sea with Late Badenian Western Paratethys, probably, through the Preddobrudzhinsky and Byrdsky troughs at the very end of Veselyanka time (Didkovsky, Nosovsky, 1975).

Thus, the confinement of the most diverse and rather polyhaline fauna to eastern parts of Early Chokrakian, Varnian, Sartaganian, and, to a con-

siderable extent, to Late Veselyanian basins allows a suggestion that the regions of Transcaucasia, Transcaspiia and Ciscaucasia were under a greater influence of normal marine waters. On the basis of these data a conclusion can be drawn that it was in the south-east where the communications of these Eastern Paratethys basins occurred. A possibility of communications through Iran and Turkey during the Konkian (Sartaganian) time was indicated long ago by Zhizhchenko (1947), Danov (1957) and Sudo (1961). These investigators based their views about the paleogeography of that time upon the fact of considerably high diversity of polyhaline molluscs in the Transcaspiian region. It seems to us that settling down of relatively polyhaline elements predominantly in south-eastern and eastern areas of Middle Miocene basins in Eastern Paratethys might have been also due to the fact that the main current in inland seas was cyclonic (Knipovich, 1938).

On the basis of available data, it is difficult to determine the location of a strait (or strait), through which a communication (though hampered) of the Sarmatian sea with open waters could have been realized. It can be only stated that an influence of the marine regime was greater in the west of Paratethys than in other regions of the aquatorium. Here, more polyhaline molluscs lived, such as *Crassostrea gryphoides*, *Gastrana fragilis*, *Brachidontes marginatus*, *Clavatula doederleini*, *Ocenebrina sublavata striata*, *Mitrella scripta* and others (Roshka, 1987; Paramonova, 1994), assemblages of planktonic foraminifers have been encountered, and the nannoplankton is more diverse than in Eastern Paratethys. In Early Meotian time, the greatest diversity of marine mollusc species, including Mediterranean polyhaline forms (*Gibberula philippi*, *Rissoa ventricosa*, *Alvania montagu* etc.), and also a richer nannoplankton, is observed in the Ryonian Gulf (Western Georgia). It is most likely that the links of the Meotian basin with the Mediterranean might have been realized through the territory of recent Eastern Turkey, or through Iran (Iljina, 1980; Neveeskaja et al., 1986). On the other hand, a rather diverse assemblage of marine Meotian molluscs discovered in the Strimon River basin suggests the possibility of the connection of Paratethys to the Mediterranean somewhere in the region of Grecian Macedonia (Stevanović, Iljina, 1982). A short term Late Meotian ingress and a prochoresis of marine mollusc species are connected with south-eastern links of Paratethys with the Mediterranean.

If not only the time but also the paths of prochoresis of marine fauna are determined, then more precision into correlations and more details into paleogeographic reconstructions could be introduced. How-



ever, a more complete solution of the above problems requires further combined investigations of these animal groups, which inhabited Paratethyan basins in the Middle and Late Miocene, though it will be difficult to carry them out because the geology and stratigraphy of the south and south-eastern margins of Paratethys (boundary territories of Azerbaijan, Iran, Armenia and Turkey) have been insufficiently studied. There is still no answer to the question, why no noticeable changes during almost 3.5 m.y. (during the Serravalian which is correlatable to the Karaganian, Konkian and to a considerable part of the Sarmatian) were observed in the Mediterranean, whereas in the Paratethys, particularly in its eastern part, a lot of important events (Varnian, Sartaganian and Late Veselyanian trasgression, a considerable reduction of the links with open waters in Early and Late Karaganian and Sarmatian) occurred.

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GEOLOGICAL AND PALAEOGEOGRAPHIC EVOLUTION OF WESTERN GREECE, DURING THE NEOGENE-QUATERNARY PERIOD IN THE GEODYNAMIC SETTING OF THE HELLENIC ARC

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Abstract: After studying more recent data concerning the tectonic and palaeogeographic evolution of Western Greece, we suggest a model for the evolution of the Neogene basins in this region. Three distinct sedimentary sequences of post-Aquitania age have been recognized. The first sedimentary sequence was deposited during the Lower Miocene (Burdigalian), the second one from Middle Miocene up to early Lower Pliocene and the third one from Lower Pliocene and continues to the present day. The tectonic evolution strongly controlled the development of Neogene sedimentary sequences and the formation of foredeeps and piggy back basins according to their position in the orogenic belt. A series of palaeogeographic maps covering the Neogene to Quaternary period are presented in this study. They are based on the correlation of surface geological data, drilling data, litho-biostratigraphic analysis and interpretation of seismic profiles.

1. Introduction

The area of study is located in the Hellenic arc, which belongs to the south branch of Tethys orogenic system. The closure of Tethys ocean since Malm and the consequent collision between the passive continental margin of Adria microplate and the Eurasian continent gave rise to the development of successive thrusts and the formation of the Hellenides. The thrust sheets movement was progressively towards the Apulian platform foreland.

Since Late Neogene, an active subduction of the Ionian oceanic crust beneath the outer Hellenic arc is taking place.

The external Hellenides, which were affected by the final stage of Alpine orogenesis during Tertiary, are subdivided in isopic zones (or geotectonic units) namely Preapulian (Paxos), Ionian, Gavrovo and Pindos zones (Fig. 1).

From Neogene to Quaternary, sediments accumulated in basins situated in the above mentioned zones. The development of these basins was strongly controlled by the progressively westward migration of the orogenic front. Based on major compressional events, we distinguish three successive time periods of Neogene sedimentation.

A selection of data from seismic profiles, surface geology and exploratory wells presented in this study, shows the palaeogeographic and structural evolution of these basins.

2.1. Lower Miocene (Burdigalian)

A deep foreland basin was developed in the Ionian and Gavrovo zones during the Oligocene as a result of Pindos orogenesis.

Flysch sedimentation took place in this asymmetric basin with progressive increase of its thickness to the east. At the end of the Aquitanian, thrusts propagation gave rise to emergence of Gavrovo and internal Ionian zone and the formation of an elongated foredeep basin which was developed parallel to the newly formed thrust front. In this basin, sedimentation took place during the Burdigalian.

A sketch map showing the extension of this sequence is presented (Fig. 2). This map is a representation of the relative palaeogeographic extension of this Burdigalian sedimentary unit. The true palaeogeographic extension cannot be accurately mapped, because postdepositional folding and overthrusting as well as a significant clockwise rotational movement (Kissel and Laj, 1988) have affected the Burdigalian sedimentary basin.

This sedimentation conformably overlies flysch de-



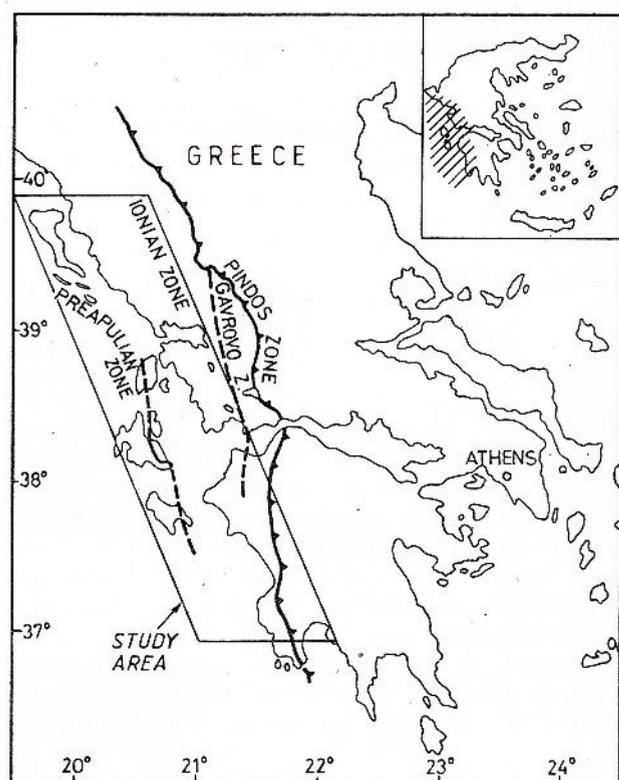


Fig. 1 – Location map of the studied area

posits and consists mainly of marls with banks of sandy organogenic limestones. It is characterized by microfauna of *Miogyssina grobulina*, *G. dissimilis*, *G. trilobus*, *Globoquadrina dehiscens*, *Melobesites* (IGME-IFP, 1966). The depositional environment is pure marine. Littoral facies and local unconformities have been noticed only at the eastern flank of the Botsara basin (IGME-IFP, 1966), close to the orogenic front in that period. Significant thickness of Burdigalian sediments have been located only in central Ionian zone (syncline of Botsara) while in external Ionian zone only thin outcrops are observed. Such a case has recently been reported from NW Corfu island near Afona village where the microfauna includes *G. trilobus*, *G. primordius*, *Ammonia becarii* and *Gl. ciperoensis angustumbilicata* (Fig. 3).

The presence of Burdigalian sediments in NW Peloponnesus remains uncertain. Only the lower drilled sediments in Sosti-1 well probably correspond to this period (Esso, 1962). Moreover, such sedimentation cannot be excluded in the deeper parts of certain basins in this area as it is recognised in some seismic profiles. Generally, we lead to assume that the greatest part of NW Peloponnesus remained uplifted during this period.

In the Early Langhian, the Ionian zone was affected by the main compressional phase (Sorel *et al.*, 1992), which caused a significant deformation of the Burdigalian sequence. On the contrary further to the west, Preapulian zone was not affected by this tectonic event and is related to a stable foreland area. Sedimentation was rather continuous with marly carbonate deposits. Nevertheless, tectonic instability gave rise to the development of relatively high arcas either without sedimentation or with deposition of condensed series (Fig. 2).

2.2. Middle Miocene-early Lower Pliocene

The post-Burdigalian compressional phase resulted to the uplift of Ionian zone. After a subsidence which took place in the frontal part of Ionian zone (Sorel *et al.*, 1992), postorogenic transgressive sediments accumulated on the eroded preexisting formations (Fig. 4). Preapulian zone was not affected by this phase. Only a change from pure carbonate to marly sedimentation is noted during this period.

Therefore, two groups of individual basins could be defined according to their position in the orogenic belt.

The first group includes the foredeep basins situated in the eastern part of Preapulian zone, in front of the newly formed Ionian thrust belt. The sedimentation was continuous and it is characterized by marls and clays deposition up to Early Pliocene. The sedimentation in this basin is characterized locally by high rate of subsidence as indicated by seismic profile interpretation. This subsidence is related to Flexural depression of the eastern area of the Preapulian zone due to superimposed load of the thrust units (Zakynthos channel).

However, many authors have recognised features indicating locally tectonic instability in Preapulian zone until the end of the Miocene. Such features include the presence of paraunconformity with basal breccia in Middle Miocene formations and horizons with pebbles of reworked Messinian gypsum in the Zakynthos island (Nicolau, 1986), as well as the occurrence of olisthostromes and thickness variations in Middle-Upper Miocene formations (Underhill, 1989) and oxydation phenomena in the Kefallinia island. Furthermore, the absence of Early-Middle Miocene formations have been observed in the northeastern part of the Zakynthos island (Alykes-1 well) and in the offshore area south of the Kefallinia island (S. Kefallinia-1 well) confirms the assumption of tectonic instability in the Preapulian zone. This is also supported by the seismic exploration in the area (Fig. 5). In S. Kefallinia-1 well, the following sequences were drilled: (1) Upper Pliocene with fauna of *Gl. inflata* and *Gl. crassaformis*; (2) Lower Pliocene with

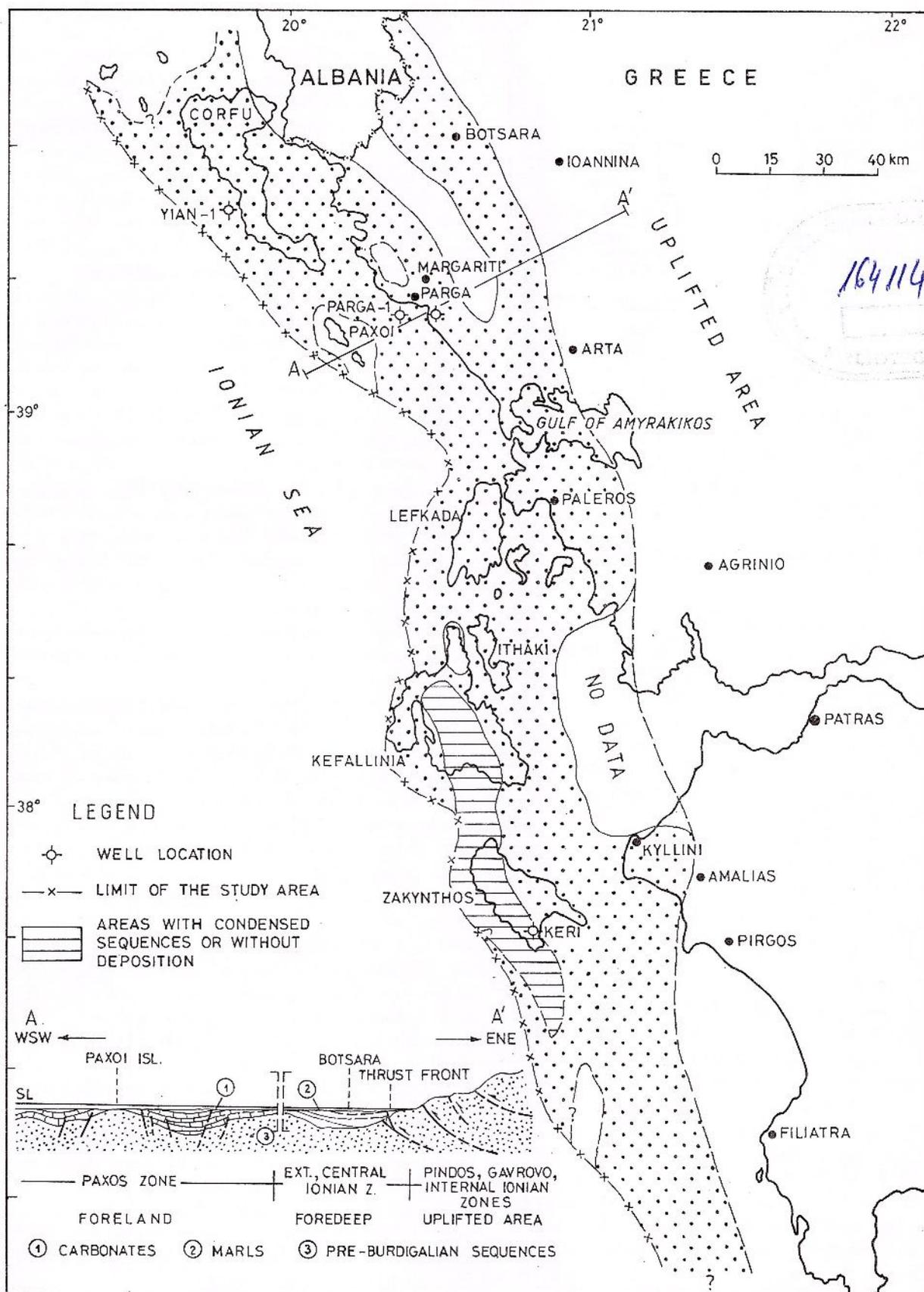


Fig. 2 - Map of Lower Miocene (Burdigalian) extent in Western Greece and restored section.

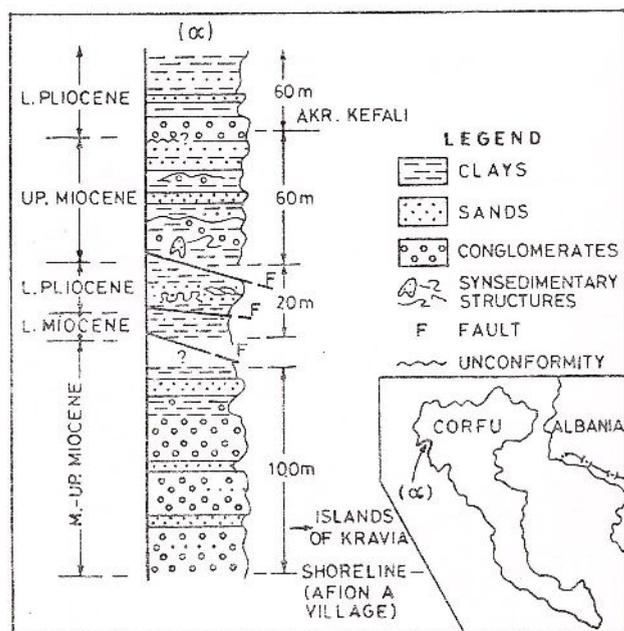


Fig. 3 – Lithostratigraphic column "Afiona village" (NW Corfu).

fauna of *Gl. puncticulata* and *Gl. margaritae*; (3) Upper Miocene with fauna of *Gl. conomiozea* and finally (4) Upper Cretaceous carbonates. These sequences are separated by the unconformities a, b, c (Fig. 5).

The second group includes basins developed on the eroded external Ionian zone. These basins were filled by transgressive sequences which overlaid unconformably older formations (Lefkas, Corfu, etc.). This marine sedimentation consists mainly of marls as well as sandstones, marly limestones and basal conglomerates and breccias. These basins could be considered as piggy back basins.

In the Late Miocene and Lower Pliocene formations of certain basins of the external Ionian zone, syndepositional features have been recognised, suggesting an increase of palaeoslope surface. Such syndepositional features include olistholites, slumps as well as wavy erosional surfaces, found in many places, as for example near Akır. Kefali in NW Corfu (Fig. 3) and Paleros in W Aetoloakarnania (Doutsos et al., 1987).

The transition from Late Miocene to Early Pliocene is generally characterized by the presence of three lithostratigraphic horizons (Sorel et al., 1992). The first horizon is Upper Messinian evaporites (gypsum) as a result of significant regression in this period. The second one is composed of sandy clays corresponding to "lago-mare facies". The last horizon consists of marly limestones with of *Shaeroidinellopsis* related to the return of marine sedimentation at the base of Pliocene.

2.3. Pliocene-Pleistocene

In the Lower Pliocene an important tectonic event took place in the Hellenic arc: this is the emplacement of the Ionian trust sheet on the Preapulian zone due to thrust propagation to the west. A subsequent subsidence of the area behind the orogenic front resulted in the deposition of transgressive sequences, unconformably on the older Neogene or Mesozoic formations (Fig. 6). However, in certain basins the sedimentation was continuous without interruption (Zakynthos channel, northern Corfu, Katakolo).

This sedimentation consists mainly of marls and clays with sandy intercalations and basal conglomerates or breccias. The depositional environment changes progressively from pure marine conditions that predominate to the west to transitional ones to the east including lagoonal, lacustrine or terrestrial environments. Thick accumulation of conglomerates took place in certain basins in the vicinity of areas with strong uplift movements (NW Peloponnesus). Further east, intermountainous basins were formed with lacustrine deposits. In several areas, unconformities have been recognised in Quaternary formations which are related to transgression or regression periods (Dermitzakis et al., 1979).

Concerning the tectonic regime in Pliocene to Quaternary period, we can distinguish two areas with different structural features.

(i) The more external area (western part) of Hellenic arc, characterized by compressional structures generally trending NNW-SSE. These are related to the active subduction that is taking place further west in the Hellenic Trench. Folding, thrusting, reactivation of earlier normal faults as reverse, as well as new diapiric movements (Ionian sea) confirm a compressional stress regime which continued up to present.

(ii) The eastern part of the Hellenic arc is affected by an extensional tectonic regime, which is associated to a N-S extension of thin continental crust in the Aegean domain. As a result, main extensional basins were formed generally trending WNW-ESE, for example gulf of Corinth, gulf of Patras, gulf of Amvrakikos etc. (Doutsos et al., 1987; Brooks et al., 1988).

The present seismicity confirms the previous different structural styles where the western part is characterized by earthquakes with compressional focal mechanisms while the eastern part by earthquakes associated with normal and strike slip faulting (McKenzie et al., 1978; Papazachos et al., 1984). Especially in NW Peloponnesus, several strike slip faults have been recognised from surface geology in Upper Pliocene-Pleistocene sediments.

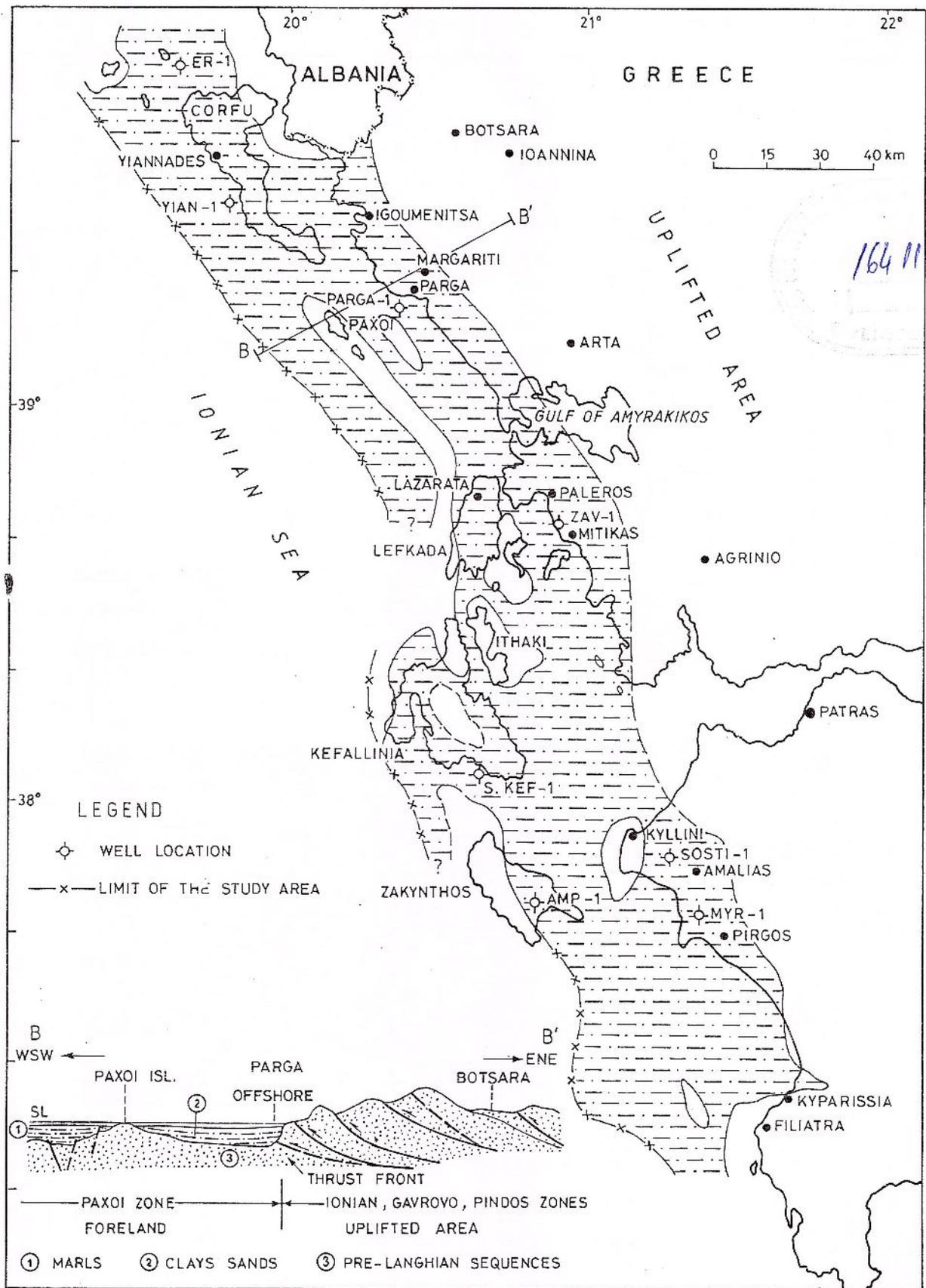


Fig. 4 – Map of Middle Miocene to early Lower Pliocene extent in Western Greece and restored section.

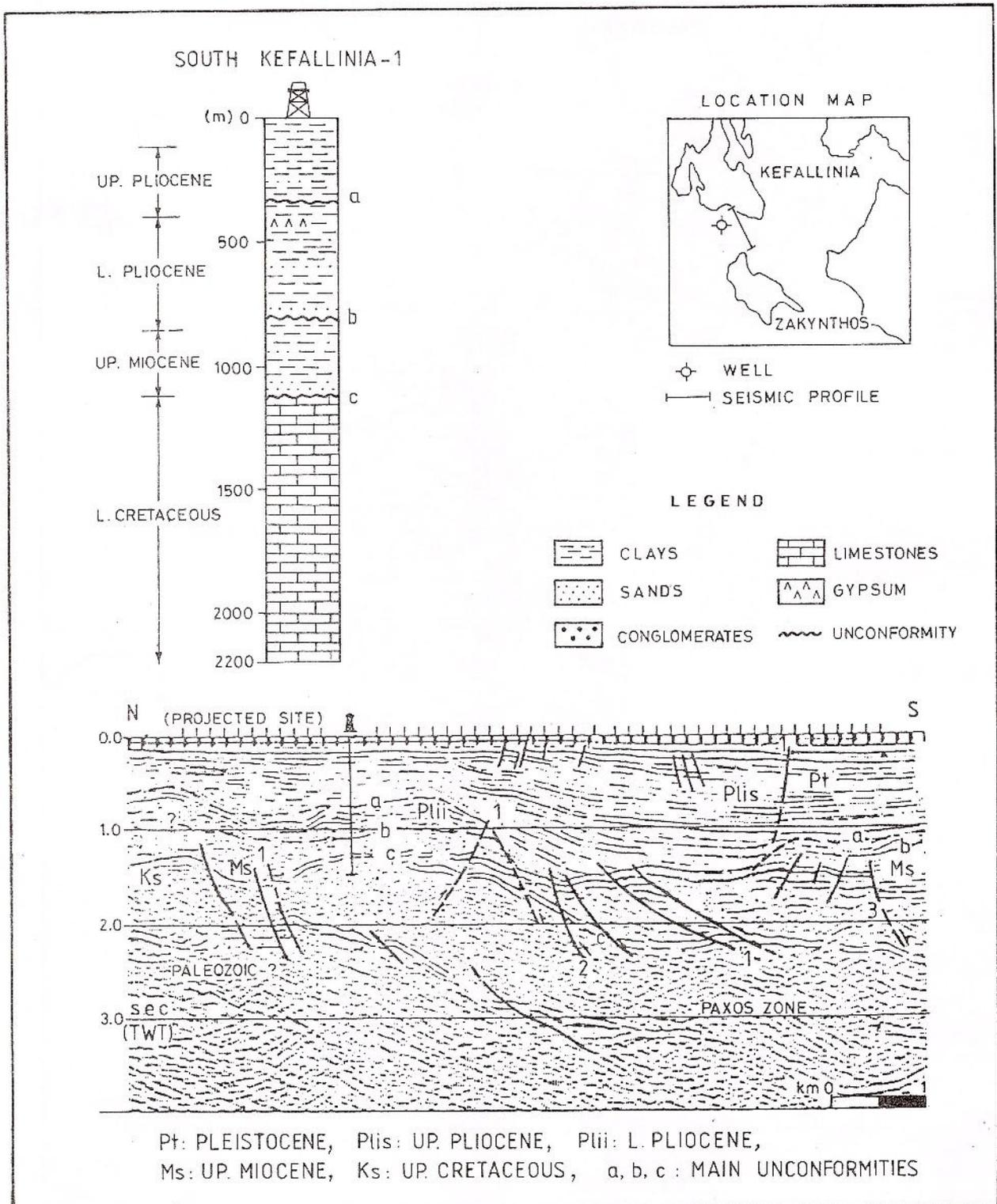


Fig. 5 - Lithostratigraphic column of S Kefallina-1 well and interpreted seismic profile.

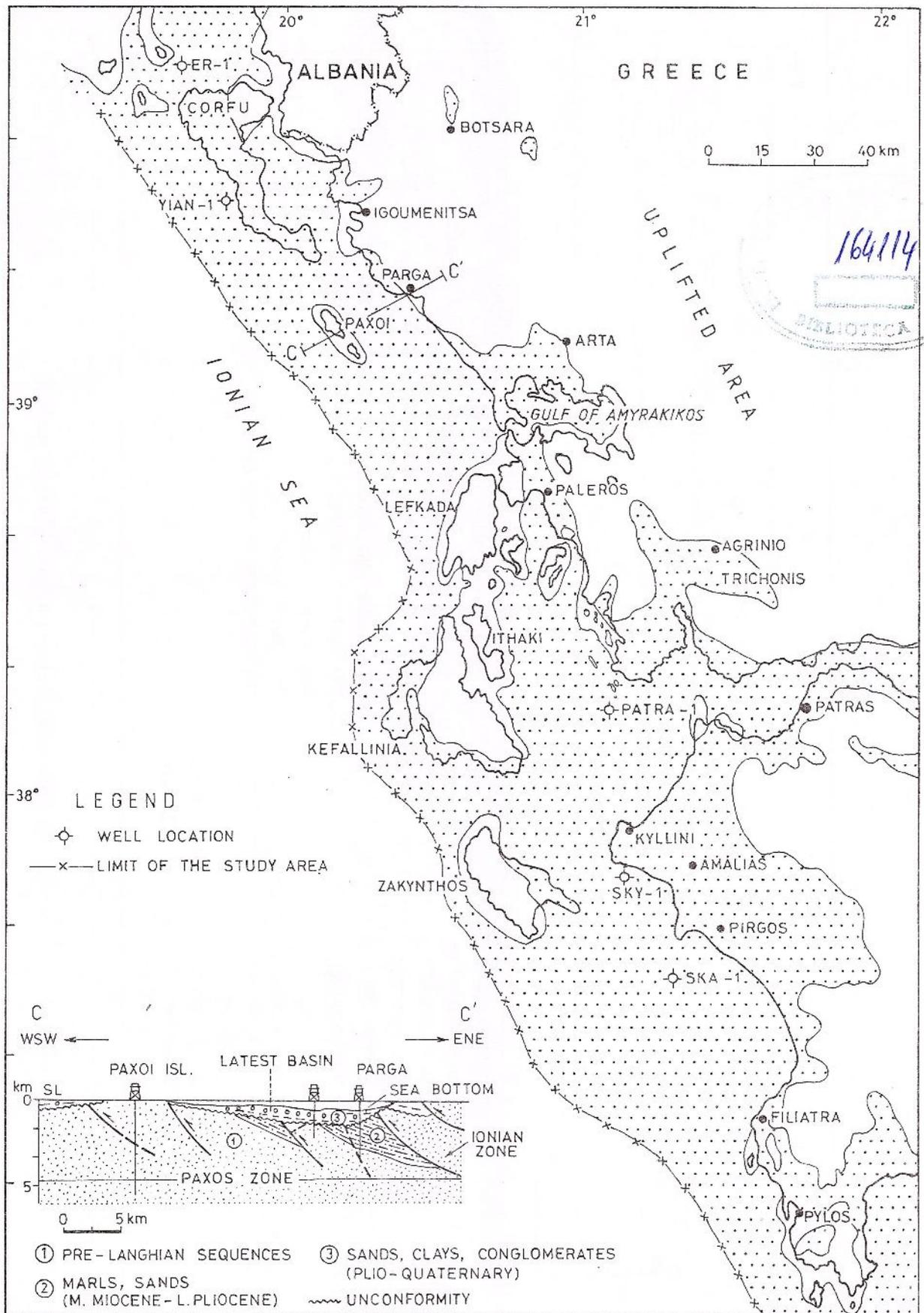


Fig. 6 - Map of Pliocene to Pleistocene extent in Western Greece and cross section.

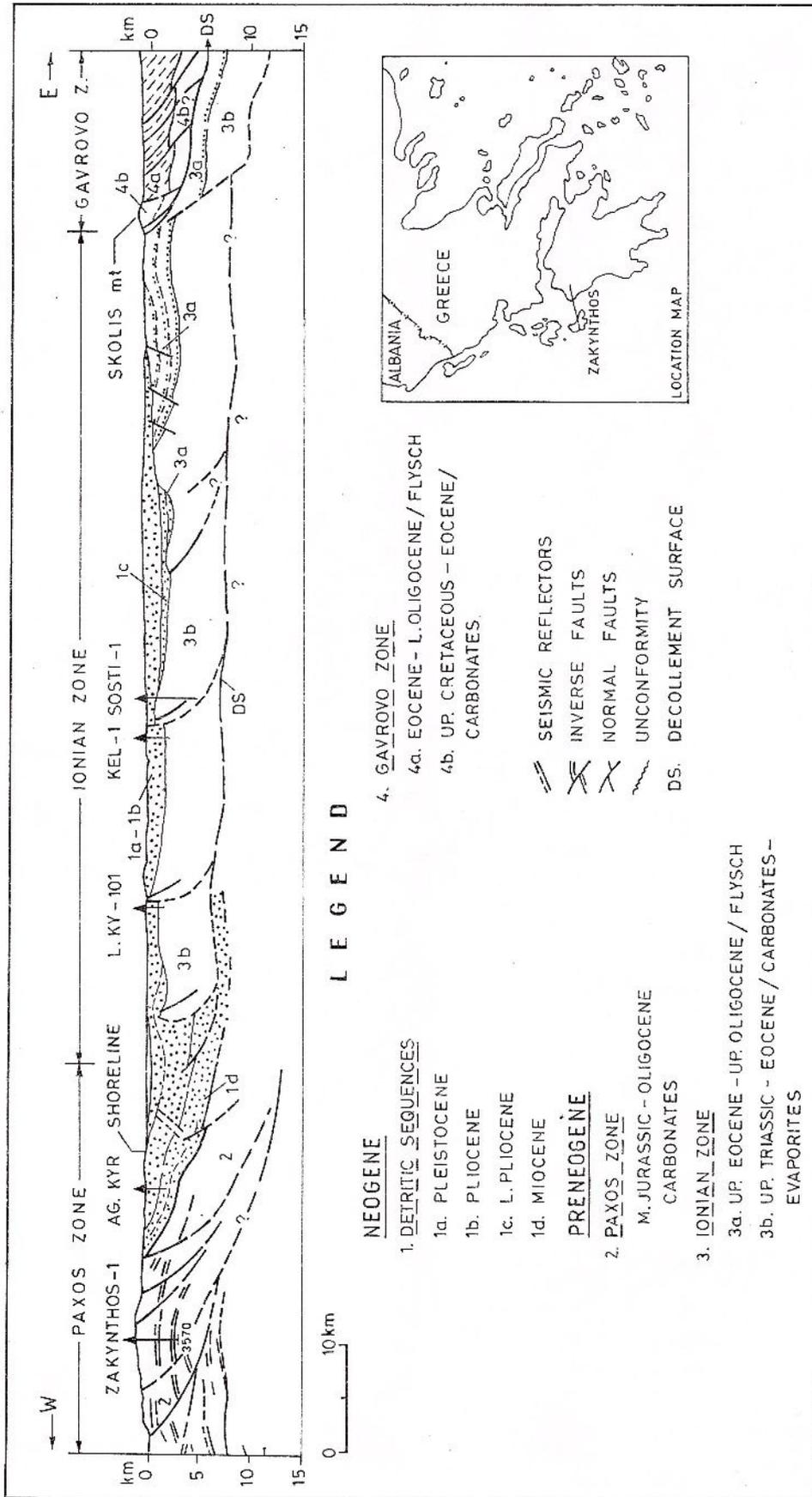


Fig. 7 - Geological section based on interpreted seismic profiles and well data

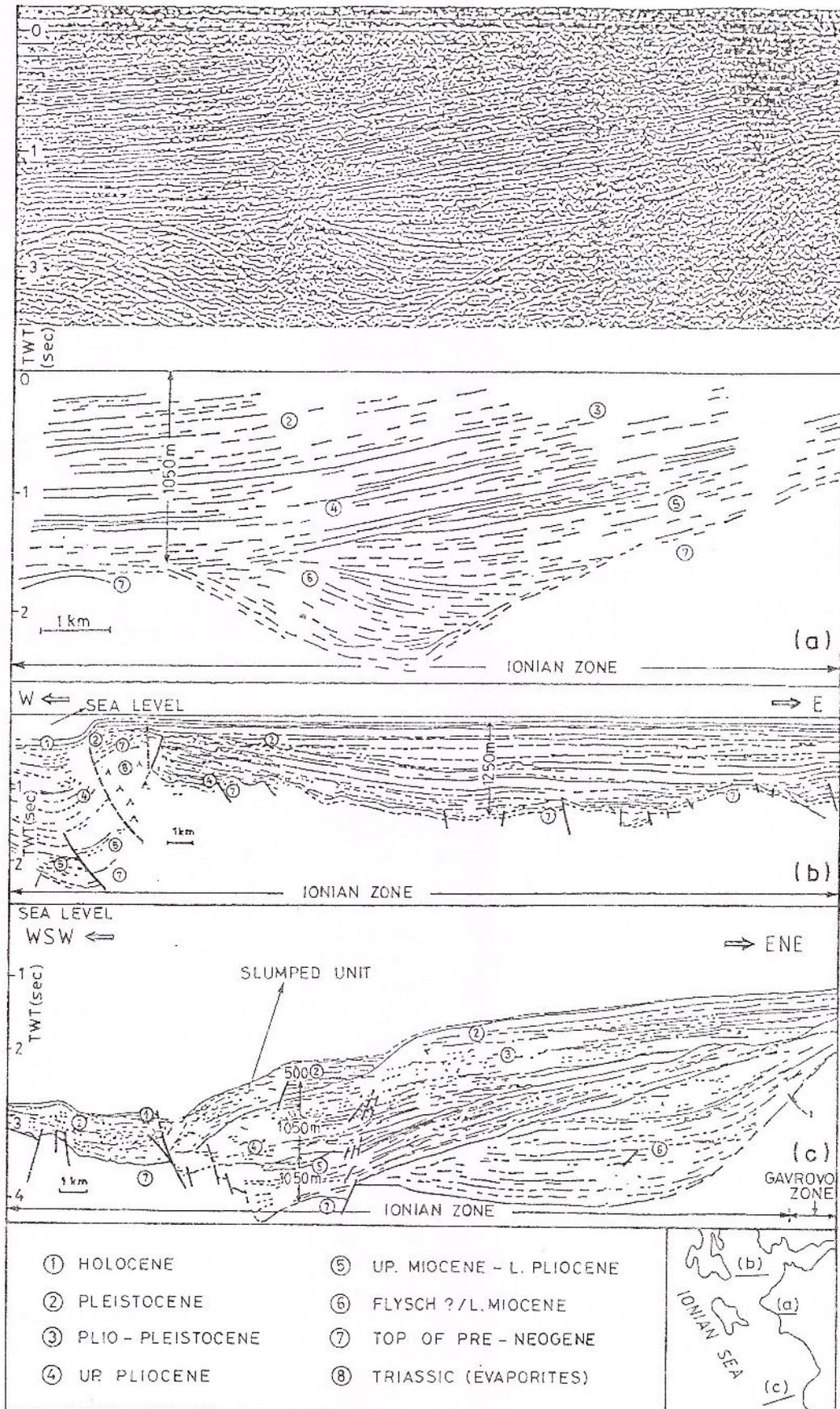


Fig. 8 - Line drawings based on unmigrated seismic profiles in Western Greece.

3. Structural evolution of Neogene basins

The structural evolution of the Neogene basins is well documented in the cross section of Figure 7. This section is based on seismic data and cuts through the external Hellenides, from Zakynthos island to Skolis mountain (NW Pelopónnesus). The development of the Neogene basins is strongly controlled by thrust sheet propagation to the west. In the eastern part of this section the imbricated Gavrovo zone overthrusts the flysch of the Ionian zone.

Flysch deposits in the front of the Gavrovo thrust corresponds to a foredeep sedimentation which was taking place in the Ionian zone from Upper Eocene to Upper Oligocene. The decollement surface is situated in the upper part of the underlying flysch of the Ionian zone. A new foredeep (in Zakynthos channel) was formed since Miocene with remarkable thickness of deposits. This event could be attributed to an important subsidence due to flexural depression of the eastern part of Preapulian zone in front of the newly formed Ionian thrust. At the same period, piggy back basins were formed behind the orogenic front as is the eastern basin of Sosti-1 well. In the Plio-Quaternary period the Preapulian zone (western part of this section) has been involved in the orogenic thrust belt, while further east, recent basins have been developed on the eroded Ionian zone.

Based on the above mentioned data, we could suggest that the character of the Neogene basins is mostly related to the thrust migration to the west. Therefore, more external areas are involved progressively in a typical geometry including foredeep basins in the front of the thrust belt and piggy back basins behind it.

The above-mentioned lithostratigraphic units as well as the main unconformities from the Neogene to Quaternary period are well recognized in many seismic profiles in Western Greece (Fig. 8).

The Upper Pliocene sediments are overlying earlier Neogene or Alpine sediments. This unconformity is well determined in many profiles. Furthermore, the unconformities of Lower Pliocene and Middle Miocene are observed in several profiles (Fig. 8). In profile (a), the uplift of the western part of the section due to thrust propagation to the west during the Miocene gave rise to back tilting of sediments and to depocentre migration to the east. In the Upper Pliocene an opposite depocentre migration started due to the general uplift of NW Pelopónnesus (eastern part of profile a, Fig. 8). Similar uplift movements have also been observed in SW Pelopónnesus and Parga area.

Offshore NW Pelopónnesus, the depth of the top of pre-Neogene formations varies greatly across the diapiric bodies. In front (west of the diapirs) we observe

a great thickness of Neogene sediments. Behind (east of the diapirs) the thickness of the Neogene sections is greatly reduced and the top of pre-Neogene formations is mapped in higher position. This is probably related to thrust sheet emplacement. Later, diapiric movements distorted the Plio-Quaternary horizons and local unconformities are observed (Fig. 8, profile b). In the offshore area east of Kefallinia and Zakynthos islands a depocentre migration to the east is observed. This migration is related to uplift movements in the Preapulian zone before the emplacement of the Ionian thrust sheet in early Pliocene. After this important tectonic event, the Preapulian zone was involved in the deformation.

In the last profile (c), Upper Miocene to Lower Pliocene sediments in the eastern part outcropping further to the east in the Kyparissia basin (Kamberis *et al.*, 1992) are overlying unconformably flysch and Lower Miocene deposits. This unconformity can be related to early Langhian compressional phase. Above this unconformity, tilted sediments related to uplift movements at the eastern part of this section, as well as margin propagation effects. At the western part, transverse strike slip faults seem to affect the sedimentary thickness.

4. Conclusions

The study of a combination of geological data and seismic information led us to distinguish three sedimentary sequences in Western Greece: (a) Lower Miocene (Burdigalian), (b) Middle Miocene to early Lower Pliocene and (c) Pliocene to Pleistocene.

These sequences are separated by regional unconformities related to major compressional events. Local unconformities have been recognised in the offshore and onshore area of Western Pelopónnesus, associated with diapiric movements of Triassic evaporites during the Pleistocene.

The development of the Neogene basins and their sedimentation are strongly controlled by the progressively thrust sheet propagation to the west. As a result of this orogenic front migration, external areas are gradually involved in a typical foreland basin evolution. This evolution is characterized by the development of an asymmetric basin (foredeep) in front of the thrust belt as well as piggy back basins on propagated thrust sheets. Finally the most recent basins are affected by back arc extension.

Seismic and well data confirm the instability of Preapulian zone during the Miocene, as also documented by other authors.

Backtilting and depocentre migration are observed in seismic profiles in Neogene sediments due to either the trust uplift of the Preapulian zone or the Upper



Pliocene uplift of more internal areas (NW Peloponnesus).

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NEOGENE/QUATERNARY MICROMAMMALS FROM MYGDONIA BASIN (MACEDONIA, GREECE)

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Abstract: The preliminary results from the study of the micromammalian fauna from the Neogene/Quaternary deposits of Mygdonia basin (Macedonia, Greece) are given. Several mammalian localities have been found and abundant micromammals have been unearthed and studied. The oldest fauna is that of Chrysavgi dated to late Aragonian (MN 7+8). The Gerakarou fauna is dated to the transition between Pliocene/Pleistocene and the Fauna of "Ravin of Vulgarakis" is dated to Biharian. The datation of several faunas allowed also the dating of the various formations of Mygdonia basin. Some palaeoecological remarks coming from the studied species are also given.

Introduction

Since the beginning of the 80's very few pieces of information were known about Neogene/Quaternary fossils from the Mygdonia basin (Psarianos, 1958; Dimopoulos, 1972; Sakellariou et al., 1979; Zamanis et al., 1980; Koufos, Melentis, 1983). At that time a team from the Laboratory of Geology and Palaeontology (University of Thessaloniki) led by G. Koufos began to investigate the area. During the last fifteen years a number of mammalian localities have been discovered and a great amount of fossils has been unearthed, while the stratigraphy of the Neogene/Quaternary deposits has been studied (Koufos et al., 1989, 1992, 1995). The macromammals have partially been studied (Koufos, 1986 a, b, 1992 a, b); the artiodactyls are on study (Kostopoulos, Koufos, 1994; Kostopoulos, 1996). The study of the micromammalian faunas is the subject of a doctoral thesis realized by K. Koliadimou under the scientific supervision of G. Koufos. Preliminary data about the micromammals were presented earlier (Koufos et al., 1995), while the results of their study will be presented in this article.

Stratigraphy and localities

Mygdonia basin (basin of the lakes Langada-Volvi) is situated north of Thessaloniki (Fig. 1). The Neo-

gene/Quaternary deposits of the Mygdonia basin have been divided in two groups, the Premygdonian Group and the Mygdonian Group. Premygdonian Group was deposited in a former and larger basin, called Premygdonian basin. Three formations have been distinguished in the Premygdonian Group (Psilovikos, 1977; Koufos et al., 1995):

Chrysavgi Fm is the oldest one and consists of alternating lenses and lens-shaped beds of grey-white, unconsolidated, coarse conglomerates, grey-white sands with silty and silty-clayey intercalations.

Gerakarou Fm consists of alternating lenses and lens-shaped beds of unconsolidated gravel, coarse sand and reddish-brown silt and clay.

Platanochori Fm is the uppermost formation and consists of typical fluvial-fluviolacustrine sediments (sands, sandstones, conglomerates, silty-sands and marly limestones).

Mygdonian Group was deposited in a new basin, the Mygdonian basin, formed at the end of the early Pleistocene after a tectonic event in the area. The deposits of the Mygdonian Group consist mainly of lacustrine, fluviolacustrine and limnodeltaic sediments.

Two localities with micromammals have been found in the Mygdonian basin.

a) "Chrysavgi" (CHR). It is situated in the northwestern part of the basin, in the older Chrysavgi Formation. The fossiliferous site is located about 2 km north of the village of Chrysavgi (Fig. 1).



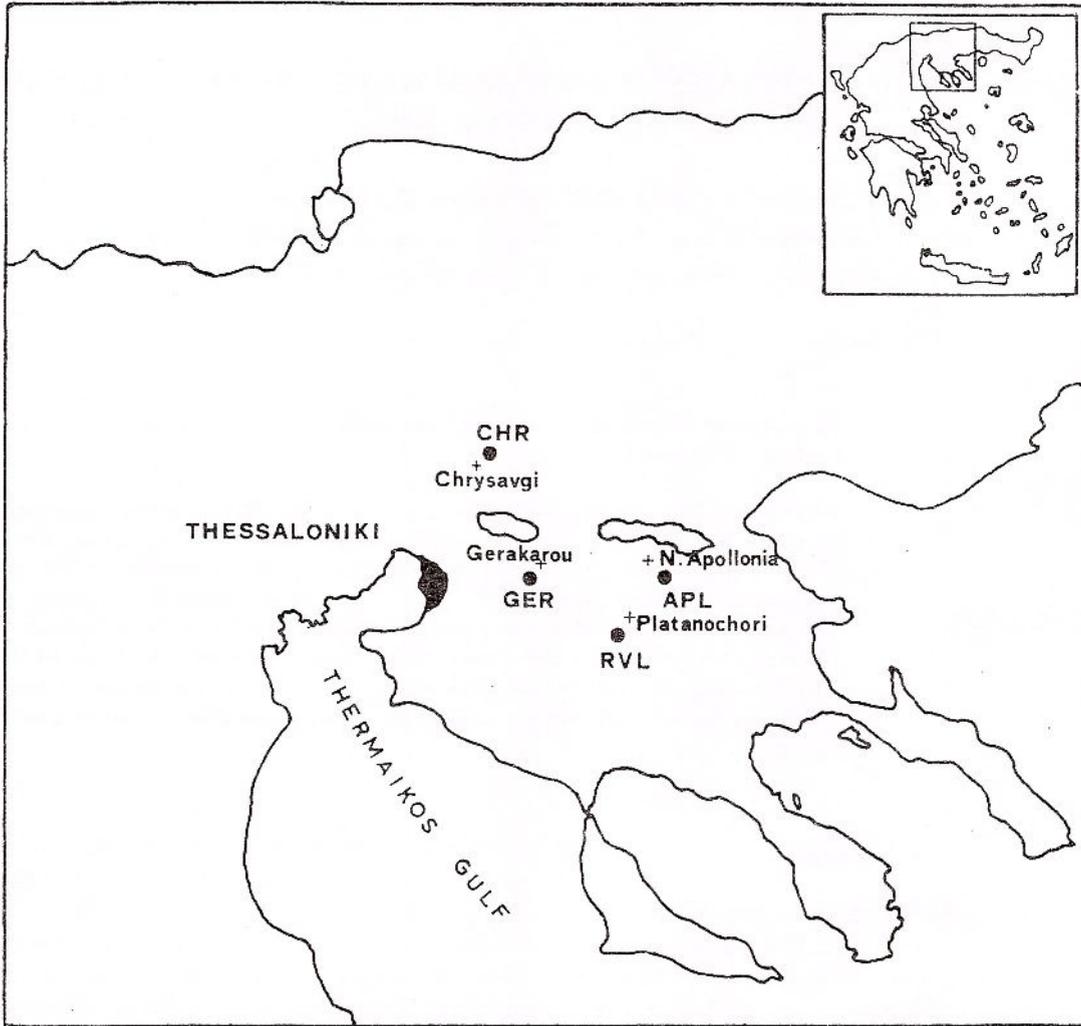


Fig. 1 - Sketch map indicating the micromammalian localities of Mygdonia basin. CHR=Chrysavgi-1; GER=Gerakarou-1; APL=Apollonia-1; RVL=Ravin of Voulgarakis.

b) "Ravin of Voulgarakis" (RVL). It is situated in the Marathousa basin, in the southern part of the area in the younger Platanochori Formation. The locality is about 2 km south of the village of Platanochori (Fig. 1).

Smaller micromammalian faunas have also been found in the following localities.

a) "Gerakarou 1" (GER). It is situated 1.5 km south of the village of Gerakarou and in the uppermost layers of the Gerakarou Formation (Fig. 1).

b) "Apollonia 1" (APL). It is situated about 1 km southeast of the village of Nea Apollonia, in the Platanochori Formation (Fig. 1).

LOCALITY "CHRYSAVGI-1" (CHR)

The micromammalian fauna from the locality CHR includes the species cf. *Schizogalerix* sp., *Desmanodon minor*, *Alloptox* sp., *Myomimus* sp., *Byzantinia*

bayraktepensis and *Megacricetodon minor*. Such faunas are well known from Turkey, while in Greece only one fauna is known from Plakia locality, Crete (De Bruijn, Meulenkamp, 1972). The fauna is poor and the determinations are quite doubtful for a comparison. Thus, the CHR fauna is compared with the nearest known ones from Turkey.

Byzantinia bayraktepensis is one of the most primitive species of the genus and it is known from the late Astaracian locality Bayraktepe-1 (MN 7+8) (Ünay, 1980). On the contrary, the genus *Megacricetodon* is absent in the Bayraktepe-1 fauna. Its last appearance in the Turkish faunas is referred from Kalamis locality where it was found together with a form intermediate between *Cricetodon* and *Byzantinia* (Ünay, De Bruijn, 1984). *Desmanodon minor*, in association with the last *Cricetodon*, is known from the locality Saricay (Engesser, 1980). In the younger

| M. Y. | SERIES | FAUNAL UNITS | MN ZONES | LOCALITIES | | GENERA | | | | | | |
|-------|----------------|--------------|----------|---|--------------------------------|---------------|------------|----------|----------|------------|----------------|-------------------------------|
| | | | | S.E. EUROPE | W. ASIA | | | | | | | |
| 9 | UPPER MIOCENE | VALLESIAN | 10 | Lefkon R. Zouste 1 Biodrak R. de Pluie Kastellios | Berislav Eldar | Schizogalerix | Desmanodon | Alloptox | Prolagus | Byzantinia | Myomimus | |
| 10 | | | | | Dendil Bayraktepe 2 | | | | | | | Karaözii |
| 11 | | | | Mahmutköy Comănești 2 | Sebastopol Varnista | | | | | | | Esmek Akcatöy Yassrorem |
| 12 | MIDDLE MIOCENE | ASTARACIAN | 8 + 7 | CHR Kalamis Pismankoy | Korethi Sofca Mikulov | Schizogalerix | Desmanodon | Alloptox | Prolagus | Byzantinia | Megacricetodon | |
| 13 | | | | | Yeni Eskihisar Bayraktepe 1 | | | | | | | |
| 14 | | | | | Saricay | | | | | | | |
| 15 | | | | | Bielometshevskaya Candir | | | | | | | Prebreza |
| 16 | ? | ? | ? | Pasalar | | | | | | | | |
| 16 | ORLEANIAN | | 5 | Chios | Ksylbulak | | | | | | | |

Fig. 2 - Biostratigraphic position and correlation of Chrysavgi-1 (CHR).

locality of Yeni Eskihisar *D. minor* disappears while the evolved *D. major* appears associated with *Byzantinia* (Engesser, 1980; Sen, Ünay, 1978). *Alloptox* is a common element in the Astaracian faunas of Asia Minor (Ünay, Sen, 1976; Sen, 1990). The coexistence of *Desmanodon minor*, *Alloptox* sp., *Byzantinia bayraktepensis*, and *Megacricetodon minor* indicates a late Astaracian age (MN 7+8) for the CHR fauna. The presence of *Byzantinia bayraktepensis* indicates for the CHR fauna an age younger than that of Saricay and Kalamis fauna. The CHR fauna is also older than the Yeni Eskihisar one because of the presence of *Desmanodon minor*. The genus *Prolagus* is represented in the CHR fauna by a new species;

however, the few available and fragmentary material cannot allow at the moment the certain definition of this species (Fig. 2).

The genera *Desmanodon* and *Byzantinia* predominate in the eastern Mediterranean region. Only small-sized forms of *Megacricetodon* species are known from this region. *Megacricetodon* disappeared much earlier in this area than in Western Europe (Ünay, De Bruijn, 1984). *Alloptox* is an Asiatic genus, till now unknown in the European continent. These elements indicate the strong relationships of the CHR fauna with the Asiatic ones.

Chrysavgi-1 is the second late Astaracian fauna referred from Greece. The fauna from the local-

| M. Y. | SERIES | FAUNAL UNITS | MN/MQ ZONES | RODENTS SUPERBIOZONES FEJFAR & HEINRICH 1990 | RODENTS ZONES FEJFAR & HEINRICH 1990 | RANGES OF GENUSES & SPECIES IN EUROPE | | | | | | |
|-------|-------------|--------------|-------------|--|--------------------------------------|---------------------------------------|---------|----------|---------|----------|----------|---------------|
| 1 | PLEISTOCENE | TORINGIAN | MQ 2 | Arvicola | 2 | A. terrestris | Hystrix | Apodemus | Mimomys | Borsodia | H. major | A. mystacinus |
| | | | | Microtus | 1 | A. cantiana | | | | | | |
| | BIHARIAN | MQ 1 | Microtus | 2 | M. savini | | | | | | | |
| | | | Mimomys | 1 | M. savini M. pusillus | | | | | | | |
| 2 | PLIOCENE | VILLANYIAN | MN 17 | Borsodia | 2 | Mimomys pliocaenicus | | | | | | |
| | | | | Villanyia | | | | | | | | |
| | | | | | | | | | | | | |
| 3 | | | MN 16 | | 1 | M. polonicus M. hajnackensis | | | | | | |

Fig. 3 - Biostratigraphic position of Gerakarou-1 (GER).

ity of Plakia (Pandanassa Fm, region of Rethymon, Crete) is also referred as a latest Astaracian one with forestrial characters (De Bruijn, Meulenkamp, 1972). The two faunas have not any common genus or species. This is probably due to the different palaeoenvironmental conditions. In addition, a strong European influence is proposed for the Plakia fauna (De Bruijn, Meulenkamp, 1972) while an Asiatic one for the CHR fauna is more probable.

LOCALITY "GERAKAROU-1" (GER)

The locality GER includes a rich macromammalian fauna (Koufos, Melentis, 1983; Koufos, 1986 a,b; Koufos, 1992 a, b; Kostopoulos, Koufos, 1994; Koufos et al., 1995). According to the macromammals,

the fauna has been dated to late Villafranchian; more precisely, the faunal assemblage of GER is situated between Senèze and Olivola Faunal Units (Kostopoulos, Koufos, 1994). The first micromammalian indication from GER was a *Mimomys* tooth (Zamanis et al., 1980). Some specimens of *Hystrix major* from the same locality were described by Koliadimou, Koufos (1991). The micromammalian fauna, unearched later from GER, includes the rodents *Apodemus* cf. *mystacinus* and cf. *Borsodia* sp. The genus *Mimomys* has a great biochronological range covering Pliocene and early Pleistocene. The last presence of *Hystrix* sp. in the Eastern Mediterranean is referred from the early Villanyian locality of Damatria (Rhodes, Greece) (Van der Meulen, Van

| M. Y. | SERIES | FAUNAL UNITS | FASES | RODENTS SUPERZONES FEJFAR & HEINRICH 1990 | RODENTS ZONES HEINRICH 1990 | LOCALITIES | "Arvicolid" species found in RVL FEJFAR & HORACEK 1986 FEJFAR & HEINRICH 1990 |
|-------|-------------|------------------------|---|--|--------------------------------|--|---|
| 1 | PLEISTOCENE | TORIN-GIAN BIHARIAN | Nagyhar- sanyhegy Bettia Mokra | Arvicola Microtus | Arvicola terrestris | FRANCE CENTRAL EUROPE FEJFAR & HORACEK 1983 HORACEK 1990 | Ranges of "Arvicolid" species found in RVL FEJFAR & HORACEK 1986 FEJFAR & HEINRICH 1990 |
| | | | | | A. cantiana | Mosbach Villany 8 Villany 6 West Runton Kozi Grzbiet Nagyharsanyhegy 4 Bourgade Vcelare 4D Monte Peglia 2 Vallerots Monte Peglia 1 Bettia 2 Nagyh. 2 Cava Sud Kamyk Villany 5 | |
| 2 | PLIOCENE | VILLANYIAN | MN 17 | Borsodia Villanyia | Mimomys plicatensis | Villany 3 Tegelen TC6 Kisláng | Microtus pitymyoides Lagurodon aranka Mimomys savini |

Fig. 4 - Biostratigraphic position and correlation of Ravin of Voulgarakis (RVL).



Kolfchoten, 1986). In spite of that *H. major* is known in Western Europe from the early Biharian localities of Ratonneau (Gervais, 1852) and Venta Micena (Agusti et al., 1986). *Apodemus mystacinus* appears in Greece in the fauna of the Uppermost Villanyian locality of Langada, Island of Kos (Van der Meulen, Van Kolfchoten, 1986). The general characters of the late Villanyian-early Biharian *Borsodia* from Tiligul and Kryzhanovka-4 localities (Tesakov, 1993) fit well with the morphology of *Borsodia* from GER but the scarcity of material cannot allow certain determination. Although the micromammalian fauna from GER is so poor, it agrees with the age given by the large mammals (Fig. 3).

LOCALITY "RAVIN OF VOULGARAKIS" (RVL)

The locality RVL is situated in the Platanochori Fm and yielded abundant material. The rich micromammalian fauna of RVL includes the following species: *Talpa* sp., *Crocidura kornfeldi*, *Sorex* cf. *minutus*, *Drepanosorex* cf. *praeearaneus*, *Asoriculus* cf. *castellarini*, *Beremendia fissidens*, Leporidae indet., *Spermophilus nogaici*, *Pliospalax* n. sp., *Glis* aff. *glis*, *Miomimus* sp., *Sicista subtilis*, *Apodemus mystacinus*, *Apodemus sylvaticus*, *Cricetinus* n. sp., *Miomimus savini*, *Microtus pitymyoides*, *Lagurodon arankae*, *Kislangia* sp.

The species *Miomimus savini* is a typical Biharian one. *Lagurodon arankae* is very common in the early Biharian faunas of Eastern Europe although it disappears much earlier in Central Europe. On the contrary, *Microtus pitymyoides* is known from few Biharian faunas of Western and Central Europe (Bourgade, France; Vcelare 4D, Slovakia). The morphological characters of *M. pitymyoides* from RVL fit very well with those from the last localities. The morphological and metrical processing of *M. pitymyoides* from RVL (according to Chaline, 1972 and Van der Meulen, 1973) proves also that it is slightly older than *M. pitymyoides* from Bourgade and Vcelare 4D. Using the same methods the RVL material is clearly older than the advanced forms of *M. gregaloides* and *M. arvalidens* from Villany 8,6 (Hungary). The coexistence of *M. savini*, *M. pitymyoides* and *L. arankae* indicates a late Biharian age for the RVL fauna (Fig. 4). The insectivores found in RVL, as well as the presence of *Spermophilus nogaici* agree with this age.

Some peculiar faunal elements have been recognized in the RVL material. A large-sized *Glis* similar to *Glis glis* is present and gives a younger idea for the age of the fauna; usually the small-sized *Glis sackdillingensis* is present in the early Biharian. The presence of *Sicista subtilis* in the RVL fauna indicates a steppic character and differentiates it from those of

Central and Western Europe, which are usually forestal. The genus *Pliospalax* is common in the Miocene and Pliocene faunas of Greece and Turkey (Sen, 1977; Ünay, 1978, 1981, 1990; De Bruijn, Van der Meulen, 1975; De Bruijn et al., 1970). Its presence, with a new species, in the Biharian fauna of RVL extends clearly its stratigraphic range. Finally the genus *Cricetinus* appears for the first time in the Biharian of Europe; the genus was known from the Pliocene of Europe (Csarnota 2, Beremend 15; Hungary) and from the Middle Pleistocene of China. The RVL *Cricetinus* is different from the known material of the genus and represents a new species.

The known Biharian localities from Greece are few. The locality of Kalymnos (Greece) is an exceptional one including species of "Eastern Mediterranean character with close relationships to Israel" (Kuss & Storch, 1978). The only common species with the RVL faunas is *A. podemus mystacinus* and the genus *Miomimus* (*M. roachi* in Kalymnos), and thus a correlation between both localities is not possible. *Miomimus savini* is known from the Early Biharian localities of Kaiafa and Zeli 2, 2A+B (Greece). *Lagurodon arankae* is also known from Kaiafa and Zeli 2. *Kislangia* cf. *rex* has been found in Megalopolis 2 (Greece) dated to Biharian (Van der Meulen, Van Kolfchoten, 1986). According to the arviculids the RVL fauna is considered younger than that of Tourkobounia-2 (Greece) with *Microtus nutiensis* and younger than that of Volos in which the species *Microtus arvalidens* appears. The small-sized *Glis sackdillingensis* is known from the older RVL locality of Tourkobounia-2 (Van der Meulen, Van Kolfchoten, 1986). Thus the presence of the *Glis* aff. *glis* in the RVL fauna seems to be the first appearance of the large forms of *Glis* in Biharian. The presence of *Jardanomys majori* in the southern Greek localities of Kalymnos, Kaiafa and Tourkobounia-2 suggests relationships with the Middle East ones. The absence of these species in the RVL fauna and the presence of *S. permophilus nogaici*, *Cricetinus*, as well as the abundant lagurids suggest close relationships with the Biharian faunas of the Eastern Europe.

A well-known Biharian fauna from the Balkan peninsula is that of Podumci-1, Croatia (Malez, Rabeder, 1984). The composition of Podumci-1 fauna is completely different from that of RVL. *Microtus thenii* from Podumci-1 belongs to a different evolution lineage from that of *Microtus pitymyoides* from RVL. The glirids are common in Podumci-1 and very rare in RVL, while *Pliomys*, found in Podumci-1 is absent in RVL. However, the presence of three different *Miomimus* in Podumci-1 (*M. cf. savini*, *M. malezi* and *M. pusillus*) is an indication that its fauna is older than the RVL one.



LOCALITY "APOLLONIA-1" (APL)

The locality APL is situated in Platanochori Fm and it is very rich in macromammals. According to the macromammalian fauna the locality APL is dated to latest Villafranchian (Koufos, 1992 a,b; Koufos et al., 1992; Kostopoulos, Koufos, 1994). The micromammalian fauna from APL consists of *Erinaceus europaeus*, Leporidae ind. and *Lagurodon arankae*. This poor fauna suggests a general Biharian age for the locality and completes the total faunal list.

Conclusions

The micromammalian faunas determined from the Mygdonia basin (Macedonia, Greece) allow the dating of the localities and deposits.

a) "Chrysavgi-1" (CHR) fauna is dated to late Astaracian (MN 7+8). The locality is supposed to be younger than those of Saricay and Kalamis and older than Yeni Eskihisar (Turkey).

b) "Gerakarou-1" (GER) fauna is dated to late Villanyian-early Biharian.

c) "Ravin of Voulgarakis" (RVL) fauna is dated to late Biharian. The locality is supposed to be slightly older than Vcelare 4D and Bourgade (Central Western Europe) and younger than Podumci-1 (Croatia). Comparatively to the Greek Biharian faunas Ravin of Voulgarakis is considered younger than Tourkobounia-2 and older than Volos.

d) "Apollonia-1" (APL) fauna is very poor and can only suggest a general Biharian age.

According to the above statements Chrysavgi Fm can be dated to Middle Miocene. Since now Chrysavgi Fm was considered as late Miocene (Psarianos, 1958; Dimopoulos, 1972). The new material indicates an older age for the formation and also suggests an early Miocene to early Middle Miocene age for the origin of the Premygdonian basin. The uppermost layers of the Gerakarou Fm are dated to late Pliocene-early Pleistocene; thus, a Pliocene age is possible for Gerakarou Fm. Finally, the Platanochori Fm is dated to the early Pleistocene.

The study of the micromammalian faunas from the Mygdonian basin gives some interesting palaeontological and palaeoenvironmental data. Chrysavgi-1 fauna includes the first appearance of the genus *Alloptox* in continental Europe, while a possible new *Prolagus* is present in this fauna. Moreover, the presence of the species *Desmanodon minor*, *Byzantina bayraklepis* and *Megacricetodon minor* is indicated for first time in the late Astaracian of Greece. The faunal assemblage of Chrysavgi (ochotonids, cricetids) indicates an open environment with humid and warm conditions (insectivores, *M. minor*). The humid and warm character is also confirmed by the presence of lignitic-xylitic remains.

The Ravin of Voulgarakis fauna gives some interesting biogeographic and stratigraphic data. The large-sized form of *Glis* (aff. *glis*) and *Sicista subtilis* are recorded for the first time in the Biharian of Europe. In the Biharian of Greece the species *Sorex* cf. *minulus*, *Asoriculus* cf. *castellarini* and *Spermophilus nogaici* appear for the first time. Moreover, in the RVL fauna the last appearance of the genus *Pliospalax* is recorded by a new species. Finally, the genus *Cricetinus* is found for the first time in the Pleistocene of Europe with a new species. According to the large mammals and the sedimentological data a dry and warm climate and an open environment is proposed for the latest Pliocene and the beginning of Early Pleistocene in the Mygdonia basin (Kostopoulos, Koufos, 1994). The presence of abundant lagurins in the RVL fauna suggests a colder climate for the early Pleistocene of the same area. Moreover, the majority of the insectivores indicates an increase of the humidity. The presence of sporadic water ponds in the area is indicated by the appearance of *Drepanosorex* and *Hippopotamus*, as well as by pharyngeal teeth of fishes, freshwater mollusks and sedimentological data. Finally, the abundant arvicolidids, the ground sciurels, the steppic zaptodids and the spalacids suggest an open environment during that time.

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A NEW CONSIDERATION FOR THE DIVISION OF VILLAFRANCHIAN

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Abstract: The term Villafranchian has been extensively used with various meanings, providing a great confusion. Recent studies on the Villafranchian faunas of Greece has provided new data for a better definition and division of Villafranchian. The comparison of the Greek faunas with those from Europe, as well as the data coming from bibliography allow some suggestions about the division of Villafranchian. Some new questions have also arisen and they are proposed for future discussion.

Introduction

Since 1865, when the Italian geologist L. Pareto introduces the term Villafranchian, several different opinions about the meaning and the chronological-stratigraphical limits of Villafranchian have been suggested. The completely different biochronological-biostratigraphical use of the term by geologists and/or paleontologists provided a great confusion. The necessity of a detailed chronology allowed many authors to propose several subdivisions of Villafranchian, intensifying the initial confusion.

The study of Villafranchian faunas from Macedonia (Greece) began in 1980 by G. K. (Koufos, Melentis, 1983; Koufos, 1986 a, 1986 b, 1987, 1992 a, 1992 b, 1993; Koliadimou, Koufos, 1991; Koufos, Kostopoulos, 1993; Koufos et al., 1989, 1991, 1992; Kostopoulos, Koufos, 1994). In the frame of this work he dedicated the study of the Villafranchian artiodactyls of Macedonia to D.K. for his thesis. Based on the new data from the artiodactyls and on the older ones from the Villafranchian faunas of Macedonia, as well as on the numerous recent articles about European faunas, a revision of Villafranchian is proposed. This revision is only a base for discussion in order to eliminate the problems, coming from the use of the term and to give as much as possible detailed chronology.

Historical aspects on Villafranchian divisions

In the geological time scale, Villafranchian occupies the time span between 3.3 and 0.8 M.y.; that means today, Middle-Late Pliocene and Early

Pleistocene (Bonnadonna, Alberdi, 1987; Azzaroli et al., 1988; Steininger et al., 1990; Masini, Torre, 1990). The fact that this time span includes the Plio-Pleistocene boundary (or Neogene/Quaternary boundary) explains the great interest of scientists about Villafranchian (e.g. Grichuk et al., 1965; Tobien, 1970; Delgado, 1979; Azzaroli, 1977, 1983; Agusti, 1986; Bonadonna, Alberdi, 1987). In the present article, Villafranchian is considered as a biochronological term, opinion which predominates among paleontologists, at least.

Since 1950, numerous studies on the Villafranchian mammalian faunal assemblages have allowed several authors to propose the subdivision of Villafranchian in smaller units (subunits). After Bout (1960) and Bourdier (1961), the most important effort was that of Heintz (1967, 1970). Studying the associations of cervids from France and Spain, the latter author proposed a subdivision of Villafranchian in three subunits: Early, Middle and Late. Early and Middle Villafranchian are considered equivalent to the biozones of Etouaires and St. Vallier respectively, while the Late Villafranchian is subdivided in two biozones: Seneze and Peyrolles. According to Heintz (1970), the lower limit of Villafranchian is situated between the associations of cervids from Perpignan-Montpellier and Etouaires. Heintz et al. (1974) generalized this scale, including the rest taxa of macro-mammals as well as rodents.

Azzaroli (1967), using the first idea of Heintz (1967) and concerning faunal elements from a larger part of Europe, proposes a similar subdivision of Villafranchian in Early, Middle and Late a and b. The "typical Villafranchian" of Villafrancha d'Asti is rep-



resented only by the Early Villafranchian of this scale (Azzaroli, 1967).

Although the subdivisions proposed by Heintz (1967, 1970) and Azzaroli (1967) seem to be similar, an attentive observation shows that there are some important differences. According to Heintz (1970), the faunal assemblage of St. Vallier belongs to Middle Villafranchian, while Azzaroli (1967) considers it as belonging to Early Villafranchian. The faunal assemblages of Seneze and Chillac, represent for Heintz (1967, 1970) the lower part of the Late Villafranchian, while Azzaroli (1967) put them in the upper part of the same subunit. The Olivola faunal assemblage is considered by Azzaroli (1967) as the typical one for the lower part of Late Villafranchian (before Seneze), opinion which was later revised by Azzaroli (1983, 1988).

The introduction of MN Zones by Mein (1976) eliminated apparently the above-mentioned problems. Mein (1976, 1990) divided Villafranchian (or "Villanyian" as it is referred by this author) in two parts: a lower one, corresponding to the biozone MN 16 (a and b), and an upper one, corresponding to the biozone MN 17. MN 16 is correlated with the Early Villafranchian of Heintz (1970) and Azzaroli (1967), while MN 17 includes the Middle and Late Villafranchian of the latter authors. According to Steininger et al. (1990) the limit between MN 16/17 is situated at the beginning of the Matuyama magnetic chron, while Azzaroli et al. (1988) and Torre et al. (1992) consider this date as the limit between MN 16 a and 16 b. The top of MN 17 zone is situated after the faunal assemblage of Tegelen (Mein, 1990; Steininger et al., 1990). Mein (1990) refers that, after the recognition of the Neogene/Quaternary limit at the top of Olduvai subchron (1.81 M.y.; Hilgen, 1991), some faunas, like Seneze, must be placed within Neogene. The unification of Middle and Late Villafranchian into MN 17 zone (Mein, 1990) helps an initial biochronological estimation of the faunas but turns aside a detailed chronology. Moreover, important faunas, like Olivola, against the current opinion, are situated out of Villafranchian and the term Villafranchian is referred as "Villanyian".

In the same period (1970-1990) the great development of micromammalian studies gave rise to a micromammalian biochronological scale. In the established micromammalian scale the time span covered by the zones MN 16 + MN 17 is named Villanyian. Villanyian underlies Biharian, which ends at about 0.5 M.y. In many cases the micromammalian Villanyian is considered homologous with Villafranchian, this providing a greater confusion (round table meeting, Madrid, 1976).

A new division of Late Pliocene-Early Pleistocene

interval (2.5-0.5 M.y.), without reference to the term Villafranchian, has been proposed by Agustí et al. (1987). Concerning micro and macromammals, four zones have been recognized:

- MN 17 zone, including the faunas of La Puebla, St Vallier and Roccaneyra (MN 17 a), as well as those of Seneze and Tegelen (MN 17 b).

- MmQ 1 zone, including the faunas of Olivola (MmQ 1 a) and Casa Frata (MmQ 1 b).

- MmQ 2 zone, including the faunas of V. Micena, Fuentesnuevas 2, Casablanca 1 & 4, Sainzailles.

- MmQ 3 zone, including the fauna of Mosbach; it is divided in MmQ 3 a and MmQ 3 b.

The scale of Agustí et al. (1987) is mainly based on the micromammalian evolutionary stages but also tries to include the macromammalian assemblages, as the first effort to unify or to correlate the two biochronological scales.

Guérin (1980, 1982, 1990) extends the MN Zones and proposes the MNQ Zonation. In the revised version of MNQ Zones (Guérin, 1990), Pliocene and Early Pleistocene are divided in 6 (? 7) different biozones (MNQ 14 - MNQ 19, ?MNQ 20), representing characteristic evolutionary stages of the mammalian associations. The use of Guérin's scale offers a more detailed and complete biochronology than others, for "Villafranchian" at least. However, the scale of Guérin (1990), as well as that of Agustí et al. (1987) ignore the term Villafranchian, which in any case is strongly connected with the mammalian chronology.

Another division of Villafranchian, based on the combination of Local Faunas/Faunal Units/Mammal Ages has been proposed by Azzaroli (1977), Azzaroli et al. (1988), Masini, Torre (1990), Torre et al. (1992). The most recent version of the Faunal Units scale (Torre et al., 1992) divides Villafranchian in three parts:

- Early, including the Faunal Units (F.U.) of Triversa and Montopoli

- Middle, including the F.U. of St. Vallier and Seneze and

- Late, including the F.U. of Olivola, Tasso and Farneta.

According to this division, the faunas of Olivola, Tasso and Farneta are situated in the "Late Villafranchian", against the above-mentioned opinions that this part of Villafranchian corresponds to the Faunal Unit of Seneze (e.g. Heintz, 1970). Moreover, the position of Olivola at the end of MN 17 is still being discussed; in a later article, Torre et al. (1992) put Olivola at the beginning of Early Pleistocene. The problem has been partly surpassed by the introduction of the term "Latest Villafranchian" for the Farneta Units (e.g. De Giuli et al., 1986).



The "Late-Latest Villafranchian" problem

If the lower limit of Villafranchian is (or was considered) more or less known and generally accepted, the upper one, as was already mentioned by Guérin, Faure (1982), is still being discussed. Concerning all the available data from the Greek and European macromammalian Villafranchian localities, we can observe that:

1. The transition from Middle to Late Villafranchian faunas is more or less progressive. The major part of genera and species is preserved, while some new arrivals and differentiations at subspecific level are not very significant.

2. On the contrary, a clear break of the faunal composition is observed between the so-called "Late" and "Latest" Villafranchian assemblages of macromammals. This break is expressed by the replacement of all the known genera of artiodactyls, while carnivores perissodactyls change at specific or subspecific level. The more significant changes from Seneze F.U. to Farneta F.U. (or between MNQ 18 and MNQ 20 according to Guérin's scale) are:

1. *Pliohyaena perrieri* is replaced by *Pliohyaena brevirostris*
2. *Nyctereutes megamastoides* and *Chasmaporthetes lunensis* disappear
3. *Canis arnensis* and *Panthera gombaszoegensis* appear
4. *Stephanorhinus etruscus etruscus* is replaced by *S. etruscus brachycephalus*
5. The association *Croizetoceras*-*Cervus* (= *Pseudodama*; Azzaroli, 1992)-*Eucladoceras* is broken
6. *Eucladoceras* is replaced by a primitive form of *Megaloceras* (= *Praemegaceros*)
7. The genera *Gazella*, *Gazellospira*, *Gallogoral* and *Megalovis* are replaced by new forms like *Sogerelia*, *Pontoceros*, *Capra* and *Praeovibos*
8. *Leptobos etruscus* is replaced by primitive bisons of the (sub)genus *Eobison*
9. *Sus strozzi* disappears while *Hippopotamus* appears.

In addition, from "Late" to "Latest" Villafranchian macromammalian assemblages, the number of cervids decreases, while bovids predominate.

According to Guérin, Faure (1982) the changes between MNQ 18 and MNQ 19 are not so significant as they are between MNQ 19 and MNQ 20. However, the F.U. of Olivola and Tasso, situated between the F.U. of Seneze and Farneta (or between the MNQ 18 and MNQ 20), probably represent transitional mammal assemblages from the "typical Villafranchian" faunas to the mid-Pleistocene ones (or Galerian according to Azzaroli et al., 1988; Sala et al., 1992). Contrary to the above mentioned opinion of Guérin, Faure (1992), there are some important differences

between the F.U. of Seneze and Olivola:

1. Most of the above-mentioned changes in the carnivores occur at that time.

2. The characteristic for southern Europe (from Spain to Greece), Middle-Late Villafranchian association of *Eucladoceras senezensis*-*Cervus philisi*-*Croizetoceras ramosus* disappears after the Seneze F.U.

3. The common *E. senezensis* in Seneze F.U. is replaced in the Olivola F.U. by *E. tetraceros* (France) or *E. dicranios*/*E. ctenoides* (Italy).

4. *Cervus philisi* is replaced by *Pseudodama nestii* (Italy) or *Cervus perolensis-ischnoceros* (France)

5. The genera *Croizetoceras*, *Gazellospira* and *Gazella* disappear. The latter genus is present in W Europe until the end of Middle Villafranchian, while in Greece, it still existed in Late Villafranchian.

The faunal changes initially observed at Olivola F.U. are strongly increased in Farneta F.U., where all the Villafranchian genera of artiodactyls have been already replaced. The fact that Mein (1990) considers Olivola out of MN 17 zone, while Guérin (1990) and Agusti et al. (1987) put it in a separate zone (MNQ 19 or MnQ 1, respectively), supports the above-mentioned observations.

The Greek localities of Gerakarou (GER) and Apollonia (APL). The study of the Macedonian Villafranchian fauna of "Gerakarou 1" (GER) (Kostopoulos, Koufos, 1994; Kostopoulos, Koufos, this volume) has provided some significant data. The association of artiodactyls and perissodactyls from GER is very similar to the West European Late Villafranchian faunas (especially to Seneze I). Nevertheless, the association of carnivores seems to be mixed: *Pliohyaena perrieri* coexists with *P. brevirostris*, while *Canis etruscus*, *Canis arnensis* and *Panthera gombaszoegensis* are present in the GER fauna. Concerning the general opinion of their Asiatic origin, an earliest arrival of some "Latest Villafranchian" forms of carnivores in Greece seems to be possible. The fauna of GER is situated at the end of Seneze F.U. (end of MNQ 18 according to Guérin, 1990) and it is considered as intermediate between Seneze and Olivola (Kostopoulos, Koufos, 1994; Kostopoulos, 1996).

The fauna of "Appollonia 1" (APL) has strong similarities with that of Venta Micena (Spain). The macromammalian assemblages from both localities are identical at generic level at least. This fact could be useful for a more extensive chronogeographical correlation. Both faunas represent the southernmost parts of N Mediterranean region and it seems evident that they cannot be far away from each other from the chronological point of view.



Table
Correlated subdivisions of European Villafranchian, based on
macromammalian faunas.
(*)= data coming from Greek localities

| Agosti et al. 1987 | Guarin 1990 | Mein 1990 | HERTZ 1970 | Cabi & Palompo 1985 r. press. | Torre et al. 1992 | Micro mam mals | KOSTOPOULOS & KOUFOS | | Greek Local Faunas | APPEARANCES | DISAPPEARANCES |
|--------------------------|----------------|--------------|---------------|-------------------------------------|----------------------|----------------------|----------------------|------------------------------|--------------------------|---|---|
| | | | | | | VILLAFRANCHIAN | | F. U. / Zones | | | |
| | | | | | | BIHARIAN | | FARNETA MNO,20 | APL, RVL | | |
| | | | | | | VILLANYIAN | | TASSO | | | |
| | | | | | | VILLAFRANCHIAN | | OLIVOLA | | | |
| | | | | | | VILLAFRANCHIAN | | GERAKAROU SENEZE MNT7b | 1/2 KRL LIB GEP, VSL | | |
| | | | | | | VILLAFRANCHIAN | | SI VALLIER MNT7a | DFN, VOL | | |
| | | | | | | VILLAFRANCHIAN | | MONTEPOLI MNT8 | | | |
| | | | | | | VILLAFRANCHIAN | | TRIVERSA MNT8a | | | |
| MNO2 | MNO20 | | | | | | | | | Serrapia, Capra, Pontoceros, Stephanochirus mercki, S. eltusius brachycephalus | |
| MNO1 | MNO19 | | | | | | | | | Epiborn, Praebornos, Megalobornos (Planingobornos) | Lepidobos Sis strozzi, Eucladoceros |
| MNT7b | MNO18 | | | | | | | | | Hypopitamus, C. (X) fasciatus | Procampoceros, Callagoral, Ch. lunatus, S. eltusius eltusius, Gazalopora torticornis |
| MNT7a | MNO17 | | | | | | | | | Eurodiceros terraceros/ daceros, Pseudodama nesii | Gazella bouvardi*, E. sinuatus*, C. ghisli-Crozeloceros, Pl. pernan*, Megalobos ?M. megamastoides |
| | MNO16 | | | | | | | | | Pl. brevicornis*, C. arvensis*, C. eltusius*, Pl. gombasztegenius* | G. borbanica, Lepidobos stenoelapton |
| | | | | | | | | | | Martencrotherium nivalium, Gazella bouvardi* | Procopreolus cusanius, Sus minor |
| | | | | | | | | | | ?Sis strozzi, Cervus* | |
| | | | | | | | | | | Lepobos, Equus, Gazellopsina, Stephanorhinus etruscus | |

Final remarks-suggestions

From the above-mentioned observations, it is clear that the use of the terms "Late" and "Latest" Villafranchian cannot express the important faunal changes, occurred between Seneze and Farneta F.U. Moreover, the use of the term "Villafranchian" cannot give back the homogeneity of Middle-Late "Villafranchian" faunas, as well as the transitional character of Olivola and Tasso F.U. Thus, a revision of the term "Villafranchian", as well as of its subdivisions, is necessary in order to have a more representative and reliable time scale of the Plio-Pleistocene period. It is also evident that the European mammalian paleontologists need a common time scale for the Plio-Pleistocene, where micro- and macromammalian stages, substages, biozones and Mammal Units shall be correlated and simplified. In the frame of this effort, some important questions must be discussed (•) while several suggestions could be proposed (◊). A large part of the selected data comes from unpublished articles, which are presented in several congresses, during 1994/1995.

• The term Villafranchian, as it is already mentioned by Azzaroli (1992), "has come to lose much of its intrinsic value". However, it could be preserved for historical reasons and for the sake of stability in nomenclature, even without real meaning (see also Azzaroli, 1992; p. 2). If this opinion becomes generally accepted, we must define or redefine the upper and lower boundaries of Villafranchian. Concerning the lower limit of Villafranchian, there are two different aspects in the recent literature:

- Steininger et al. (1990) suggest that the "Triversa faunas" of Italy are better correlated with the lower Matuyama magnetic chron, and Agusti et al. (1995) seem to agree with this idea.

- Torre et al. (1992), Azzaroli et al. (1988), Azzaroli (1992), Aguirre, Morales (1990), Palombo (1994), Opdyke et al. (1995), based on magnetostratigraphic, radiometric and oxygen isotope data, put the lower limit of Villafranchian between the Mammoth and Kaena reserved magnetic subchrons of Gauss, at about 3.3 M.y.

In the first case Villafranchian is correlated almost directly with Matuyama magnetic chron and covers the time span from 2.5-0.8 M.y. If, in addition, the Plio-Pleistocene boundary is transferred from 1.8 to 2.5 M.y., as proposed by several authors (see Suc, 1995; Pillans, 1995; Partridge, 1995; in abstracts INQUA, Berlin 1995), the "Villafranchian" will be correlated only with the lower Pleistocene.

The end of Matuyama magnetic chron (at about 0.8 Ma) is considered today as the upper limit of Villafranchian (e.g. Torre et al., 1992; Azzaroli et al., 1988; Azzaroli, 1992). Concerning magnetostrati-

graphic studies the micromammalian faunas of Orce-7 and Venta Micena are correlated with the Olduvai subchron (Agusti et al., 1995; Agusti, pers. comm.), while the macromammalian fauna of Venta Micena could be dated at 1.6 M.y. and cannot be younger (Agusti, pers. comm.). Consequently, that means that a similar age must be suggested for the homologous faunas of Apollonia (Greece)¹, Pirro Nord (Italy), which were considered until now as Latest Villafranchian and were dated indirectly at about 1.0-1.2 M.y. Nevertheless, the data of Agusti provide more problems which will be discussed below.

◊ If we decide to preserve the term Villafranchian, it is also better to keep its initial chronological boundaries from 3.3 to 0.8 M.y., according to the Italian school (Table). In this case the beginning of Villafranchian is correlated with the establishment of "Mediterranean double seasonality" (Suc et al., 1995). The end of Villafranchian can be correlated with the beginning of Brunhes magnetic chron and the elimination of thermophilous trees caused by the temperature decrease related to the beginning of the Pleistocene glacial cycles (Steininger et al., 1990; Suc et al., 1995; Shingleton et al., 1984).

Two additional important chronological points can be referred in Villafranchian:

- the 2.5 M.y., related to the end of Gauss magnetic chron, the earliest glacial-interglacial cycles in the northern hemisphere, the establishment of "Artemisia-steppe deciduous forest replacement" in the N Mediterranean region and the "Equus-Mammoth event" (Suc et al., 1995; Shingleton et al., 1984; Steininger et al., 1990; Turner, 1995; Azzaroli et al., 1988).

- the 1.8 M.y., related to the end of Olduvai subchron and the beginning of Pleistocene (Steininger, 1990; Hilgen, 1991; Zijderveld et al., 1991; Shingleton, 1995).

◊ The time span covered by MN 16 and MN 17 zones (3.3-1.8 M.y.) could be considered as a subunit of "Villafranchian" or as a distinct mammalian stage, if the term "Villafranchian" becomes contemptible. The lower limit of this period is correlated with that of "Villafranchian", while the upper one with the upper one with the end of Pliocene (or better with the top of Olduvai subchron, 1.8 M.y.). This idea exists already in the scale of Mein (1990), where the above-mentioned period is referred as "Villanyian", as well as in the proposal of Caloi, Palombo (1994, 1995), where it is problematically referred as "Villafranchian". There are two real possibilities. The

¹The content of the term Villanyian is not so clear, because the unit is based on the fissure filling faunas from Villany area, Hungary. Thus, the stage status of Villanyian cannot be assumed (Sen, pers. comm.).



period "MN 16 + MN 17" (or 3.3-1.8) could be referred as Villanyian¹ in respect of the micromammalian chronology or under a new name, which will avoid another confusion or disagreement (Table).

• The biozone MN 17 can be divided in two subzones MN 17a and MN 17b, in respect to the scale of Agusti et al., 1987. The period "MN 16 + MN 17" contains four Faunal Units:

Triversa F.U., corresponding to MN 16a or MNQ 16 lower part

Montopoli F.U., corresponding to MN 16b or MNQ 16 upper part

St.-Vallier F.U., corresponding to MN 17a or MNQ 17

Seneze F.U., corresponding to MN 17b or MNQ 18

Several faunal elements can be defined each of this subzones or F.U. (see Azzaroli, 1988; Torre et al., 1992; Guérin, 1990).

◊ The fauna of GER can be also correlated as well as with the end of MN 17b biozone or with the end of MNQ 18. The Gerakarou fauna possesses a particular transitional character which places it between Seneze and Olivola F.U. It could be also considered as a distinct Faunal Unit which marks the end of "MN 16 + MN 17" period (Table).

• The upper part of the referred "Villafranchian", or better the period 1.8-0.8 M.y. (top of the Olduvai subchron - beginning of the Brunhes chron) could be also considered as a subunit of "Villafranchian" or as a distinct mammal stage. Caloi, Palombo (1994, 1995) have proposed the name Protogalerian for this subunit, used in local level. This name is given to express the transitional character of the includes faunas. The Protogalerian of Caloi, Palombo (1995) includes the F.U. of Olivola, Tasso and Farneta as well as the fauna of Colle Curti considered as F.U. However, the last fauna is placed by the Florentine team and its co-operators, at the base of Galerian, while between Tasso and Farneta F.U. it is added the F.U. of Casa Frata (Kotsakis, pers. comm.). Caloi, Palombo (1995) divide Protogalerian in Early and Late, which can be correlated with the previous subdivisions as follows:

-Early Protogalerian - MNQ 19/Olivola-Tasso F.U./MmQ 1

-Late Protogalerian - MNQ 20/Casa Frata-Farneta F.U./MmQ 2

The Protogalerian of Caloi, Palombo (1995) also corresponds to the lower part of Biharian in micromammalian scale.

• Torre et al. (1993), using magnetostratigraphic data, place the fauna of Matassino (included in Olivola F.U.) above the top of Olduvai subchron or within the reverse episode above the Olduvai event (Zijderveld et al., 1991). The fauna of Casa Frata

is situated higher and possibly below the Jaramito subchron. On the other hand, Agusti (1995 and pers. comm.), correlate the micromammalian fauna Orce-7 with Olduvai subchron and suggests that the macromammalian fauna of Venta Micena cannot be younger than 1.6 M.y. Concerning all these data, it seems that there is a time interval of only 200.00 years between Matassino and Venta Micena. During this time interval a complete renewal of macromammalian fauna took place! If we accept both paleomagnetic calibration, the macromammalian fauna of Venta Micena must be situated almost at the same level with Casa Frata or Tasso and not with Farneta. Anyway, the data of Agusti allow to a dramatic compression of the time span referred to/or as Protogalerian and it is necessary to re-examine more critically all the regional magnetic calibration in relation with the macro- and micromammalian biozones.

• Concerning the problem of F.U. versus biozones there are no comments. Azzaroli (1992) notes that "a biozone is represented by a rock section fulfilling definite palaeontological conditions, while a mammal F.U. is represented by a collection of fossils from a known site" and from this point of view it is really easier to modify a sequence of mammalian faunas (F.U.) from a sequence of biozones. However, in the sequence of "Villafranchian" F.U. there is a particularity. The famous Olivola F.U. is present only in the Italian territory. Homologous macromammalian faunas are still unknown from Spain, France and Greece, where the faunal assemblages of Seneze F.U. are replaced more or less directly by those of Farneta F.U. If this absence is not accidental, then a different approach of the F.U. scale could be - or even must be - done. Thus, it is better to have a combination of both type scales for the moment, till we end in a general agreement.

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PALAEOECOLOGICAL REMARKS ON THE VILLAFRANCHIAN FAUNAS OF MACEDONIA, GREECE

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Key words: Plio-Pleistocene. Villafranchian. Europe. Macedonia. Greece. Palaeoecology.

Abstract: The Plio-Pleistocene faunas of Macedonia, Greece, are analysed for the definition of the palaeoenvironmental and palaeoecological conditions. The faunal composition and the faunal similarity of the various localities is computed and compared. Cenograms and multivariate analysis are also used. The different methods allow the definition of three more or less accessible palaeoenvironments. It is suggested: a) an open savannah-like environment with relatively arid sub-conditions for Middle Villafranchian; b) an open forest-savannah woodland environment with temperate conditions for Late Villafranchian; c) a savannah grassland, relatively humid environment with colder conditions for Latest Villafranchian.

Introduction

The study of Villafranchian in Macedonia was begun fifteen years ago by a team of the Laboratory of Geology and Paleontology, led by Prof. G.K. Several Villafranchian mammalian localities have been found and abundant material has been unearthed. The collected material includes more than 1,500 specimens, coming from seven different localities (Fig. 1). The carnivores and perissodactyls, as well as part of the artiodactyls and micromammals have already been studied by Sickenberg (1967, 1968), Koufos (1986 a, b; 1987, 1992 a, b; 1993 a, b), Koufos and Kostopoulos (1993; 1995 in press), Koliadimou, Koufos (1991), Koliadimou (1996). The artiodactyls from all these localities are being studied by D.K. for his thesis (Kostopoulos, Koufos, 1994; Kostopoulos, 1996). The referred localities belong to three chronological levels, corresponding to Middle, Late and Latest Villafranchian (Koufos et al., 1991, 1992; Kostopoulos, Koufos, 1994; Kostopoulos, 1996).

- Middle Villafranchian: "Dafnero-1" (DFN), "Volakas" (VOL)

- Late Villafranchian: "Gerakarou-1" (GER), "Krimi-1" (KRI), "Vasiloudi-1" (VSL)

- Latest Villafranchian: "Apollonia-1" (APL), "Ravin of Voulgarakis" (RVL)

Using all the available data from the Macedonian Villafranchian localities, the palaeoecological changes and relationships will be defined for that period. For

this purpose the faunal composition, faunal similarity indices and cenograms will be used. The comparison of the Macedonian faunas with the recent and European Villafranchian ones, by multivariate analysis, will also provide significant data for the palaeoenvironmental conditions.

Faunal lists

The revised and completed faunal lists of the Villafranchian localities of Macedonia are given in Plate. Some of the determinations come from unpublished data (Koufos, Kostopoulos, in press; Kostopoulos, 1966) and personal observations.

The Villafranchian localities of Macedonia

Two different methods are used for the palaeoecological study of the Greek Villafranchian mammalian assemblages: the faunal composition expressed by pie diagrams and the cenograms. The localities of DFN, GER and APL with the richest faunas, have been selected as characteristic of each chronological horizon (Middle, Late and Latest Villafranchian respectively).

A. Faunal composition. To compare faunal composition of the different Macedonian sites, it is necessary to compute the number of individuals in each fauna. The Minimum Number of Individuals (MNI) has been calculated from the dental remains (M_1) or from the postcranials (metapodials). Based on the frequencies of the different taxa in the faunas



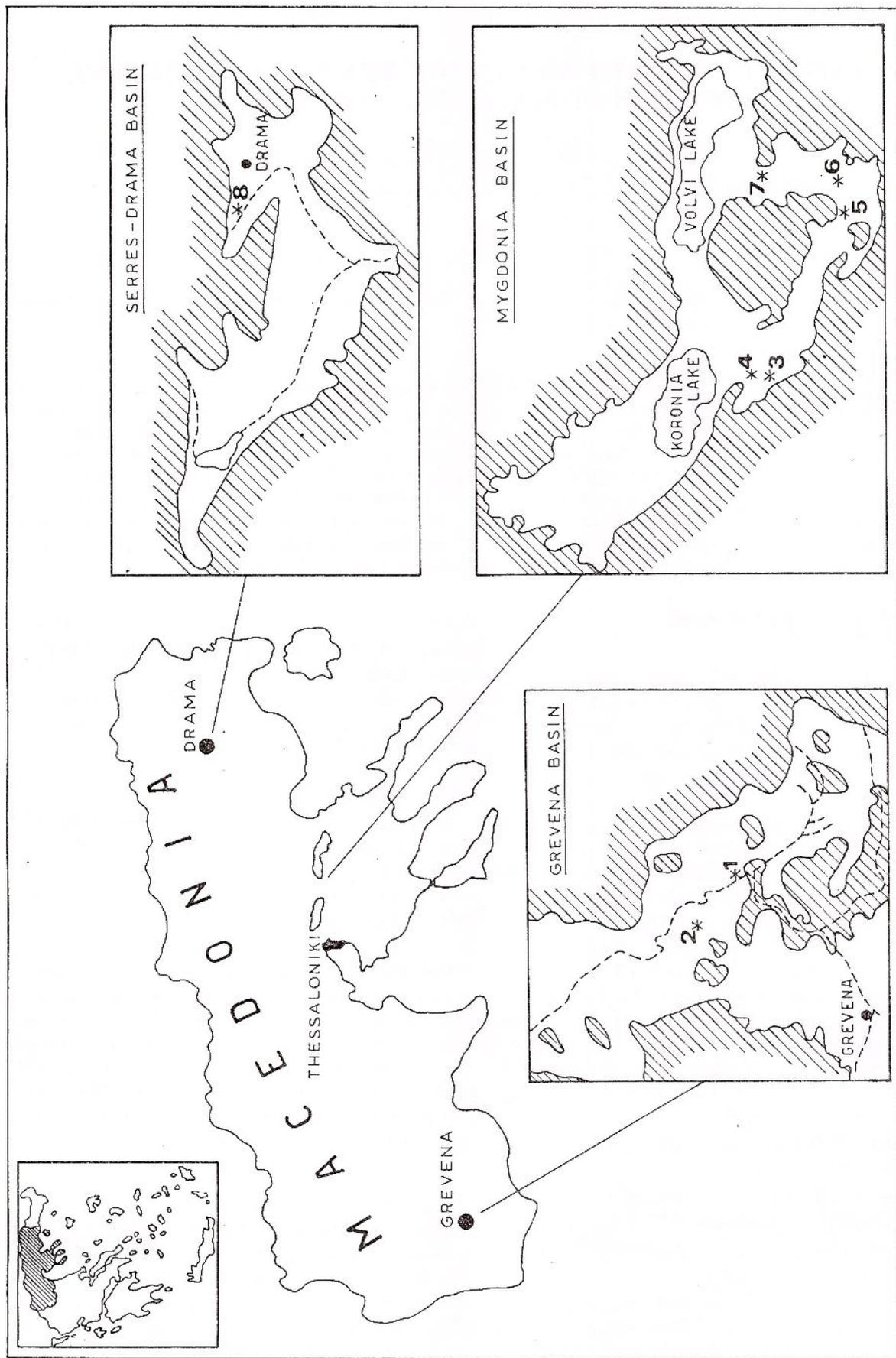


Fig. 1 - Sketch map of Macedonia (Greece) with the mammalian fossiliferous sites. 1, "Dafnero"; 2, "Libakos"; 3, "Gerakarou I"; 4, "Vassiloudji"; 5, "Ravin of Voulgarakis"; 6, "Krimni I"; 7, "Apollonia I"; 8, "Volakas".

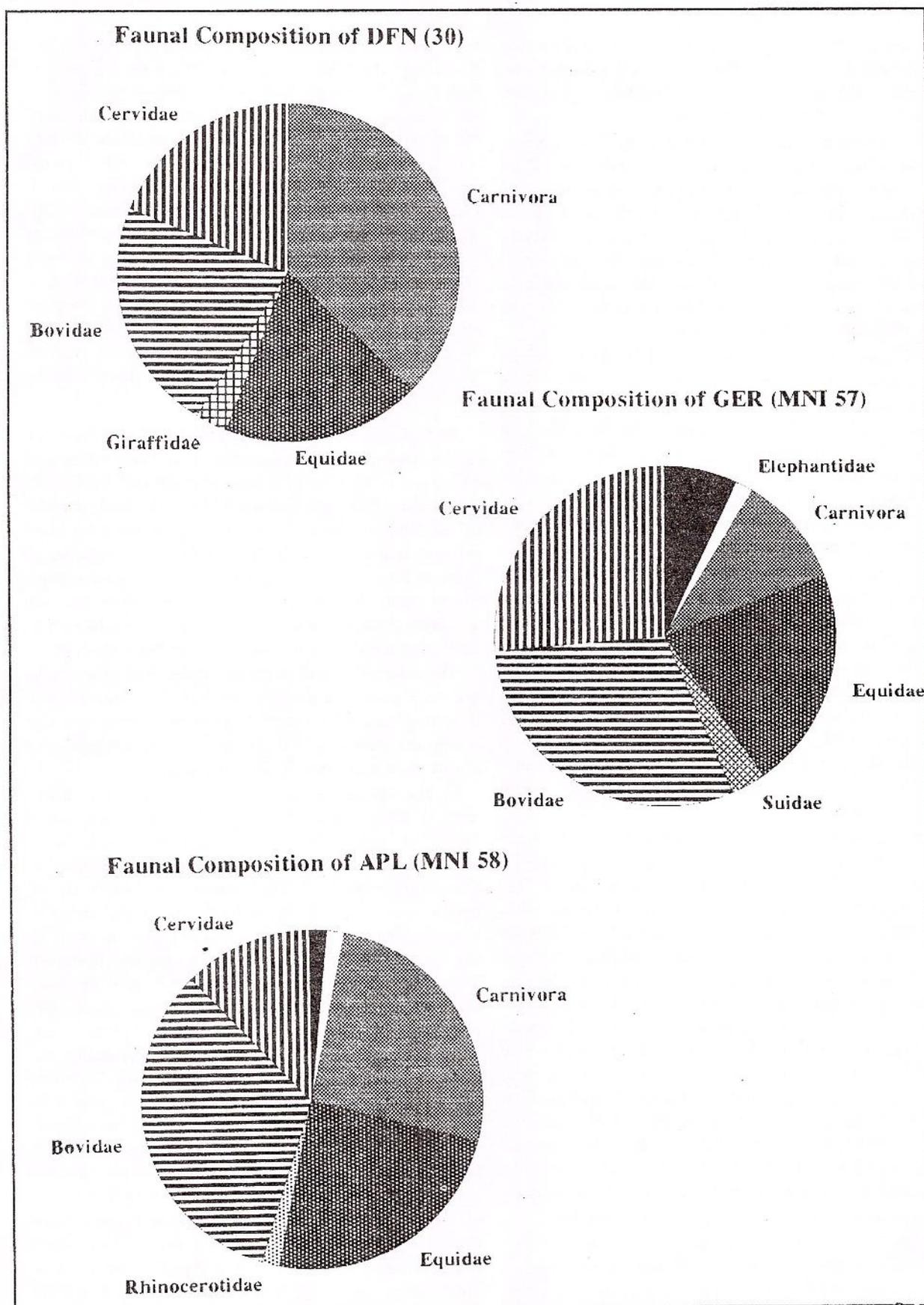


Fig. 2 - Faunal composition of the Greek Villafranchian assemblages DFN, GER, APL. Black area: micromammals.



the pie-diagrams of Figure 2 have been constructed. The selected taxa are: Micromammalia, Carnivora, Equidae, Rhinocerotidae, Elephantidae, Bovidae, Cervidae, Giraffidae and Suidae.

The faunal assemblage of DFN (Fig. 2 a) is characterized by a high percentage of carnivores (37.08 %), which, however, seems to depend on taphonomic conditions. The artiodactyls are also frequent (44.06 %), while bovids, cervids and equids are relatively balanced. The presence of giraffids, the large-sized equid (*E. stenomis* cf. *vireti*), and the slight predominance of bovids versus cervids, are good evidences for a mixed to open environment.

The faunal assemblage of GER (Fig. 2 b) is richer. The coarnivores are less frequent than in DFN while the frequency of equids is more or less similar in both faunas. The artiodactyls possess more than 50 % of the faunal composition. The cervids are more frequent in GER than in DFN, while suids replace giraffids and *E. stenomis mygdoniensis* (small-sized) replace *E. stenomis* cf. *vireti*. These features indicate mixed to forest environment.

Finally, the APL faunal assemblage (Fig. 2 c) is characterized by a smaller frequency of the cervids presence, than in the other two faunas. The cervids number of species is also strongly decreased; there is only one species in APL versus 3 in GER. The carnivores frequency is smaller than that of DFN, but the number of species is higher (5 species in DFN versus 8 in APL). The predominance of bovids versus cervids, the presence of rhinos and large-sized horses and the decreased number of cervids suggest a more open environment.

The environmental changes from Middle to Late Villafranchian seem to be not very important. The major part of genera or/and species are preserved. The frequencies of equids are similar between DFN and GER but the family is represented by different forms in each level. Thus, a large form (*Equus stenomis* cf. *vireti*) is present in DFN, while a smaller one (*Equus stenomis mygdoniensis*) in GER. The artiodactyls give more information. Bovids and cervids are relatively balanced, with a slight increase of cervids in Late Villafranchian. This fact, as well as the presence of giraffids in Middle Villafranchian (DFN, VOL), versus suids in Late Villafranchian (GER, VSL), are clear evidences that the environment is being more forestial in Late Villafranchian of Macedonia, while an open, savannah-like, environment is more possible for Middle Villafranchian.

From Late to Latest Villafranchian the ecological changes are stronger. All the known genera and species of Middle and Late Villafranchian artiodactyls from Macedonia are replaced by new forms (Kostopoulos, Koufos, 1994). The number of carni-

vores and bovids increases. Among the bovids, the most important form is a primitive bison. Comparatively to the recent faunas, the bison prefer open environments. In the Latest Villafranchian faunas of Macedonia, the cervids decrease remarkably, both in number of individuals and species. The only known cervid is a form of a primitive *Megaloceros*, which probably prefers open environments (Heintz, 1970). The equids are represented by a very large form of *Equus stenomis* and their frequency remains more or less constant. Rhinos are also present in the Latest Villafranchian fauna of APL, while in an isochronous site (RVL) from the same region (Mygdonia basin) *Hippopotamus* appears. All these elements suggest an open, grassland-like, environment, clearly different from that of Late Villafranchian.

B. Cenograms. The distributions of body weight in the mammalian communities have been estimated using the method of cenograms developed by Legendre (1986, 1989) and Cohen (1988). The body weight of the different species in the Greek faunas have been estimated using the standard values for each taxa, proposed by Legendre (1989). The obtained results are given in Figure 3. If sometimes there are not available measurements of M_1 in the Greek material, bibliographical data are also used (white symbols).

The selected Villafranchian localities of Macedonia are very poor in micromammals and in many cases it is impossible to know if there is a break on the cenogram (fauna of DFN) or to have a clear idea about the lower part of the cenogram.

In the cenogram of DFN (Fig. 3 a) the lower part is absent and it is impossible to have direct data but only relative in comparison with those of the other sites. The cenogram for GER gives more information. The absence of steps in its upper part (Fig. 3 b) and the smoothly inclined, almost parallel, upper and lower parts, as well as the break for the medium-sized species (between 5-10 or 100 gr-17 Kgr), indicate a mixed environment of the type "open forest-savannah woodland" (Legendre, 1986, 1989; Cohen, 1988; Perez, Soria, 1989/1990; Bonis et al., 1992). Comparing the cenogram for GER with those given by Legendre (1986, 1989) for the recent faunas, it fits well with the "sub-humid open forest-savannah woodland" faunas like Zinave (Africa). The cenogram for APL (Fig. 3 c) is slightly different. The steeply inclined upper part and the large break (larger than that of GER) in the area of the medium-sized species (500 gr-100 Kgr) are evidences for a more open environment of "savannah-grassland" type (Cohen, 1988; Legendre, 1986, 1989; Perez, Soria, 1989/1990; Bonis et al., 1992). Looking at the cenograms of Legendre (1986, 1989) for the recent faunas, the APL



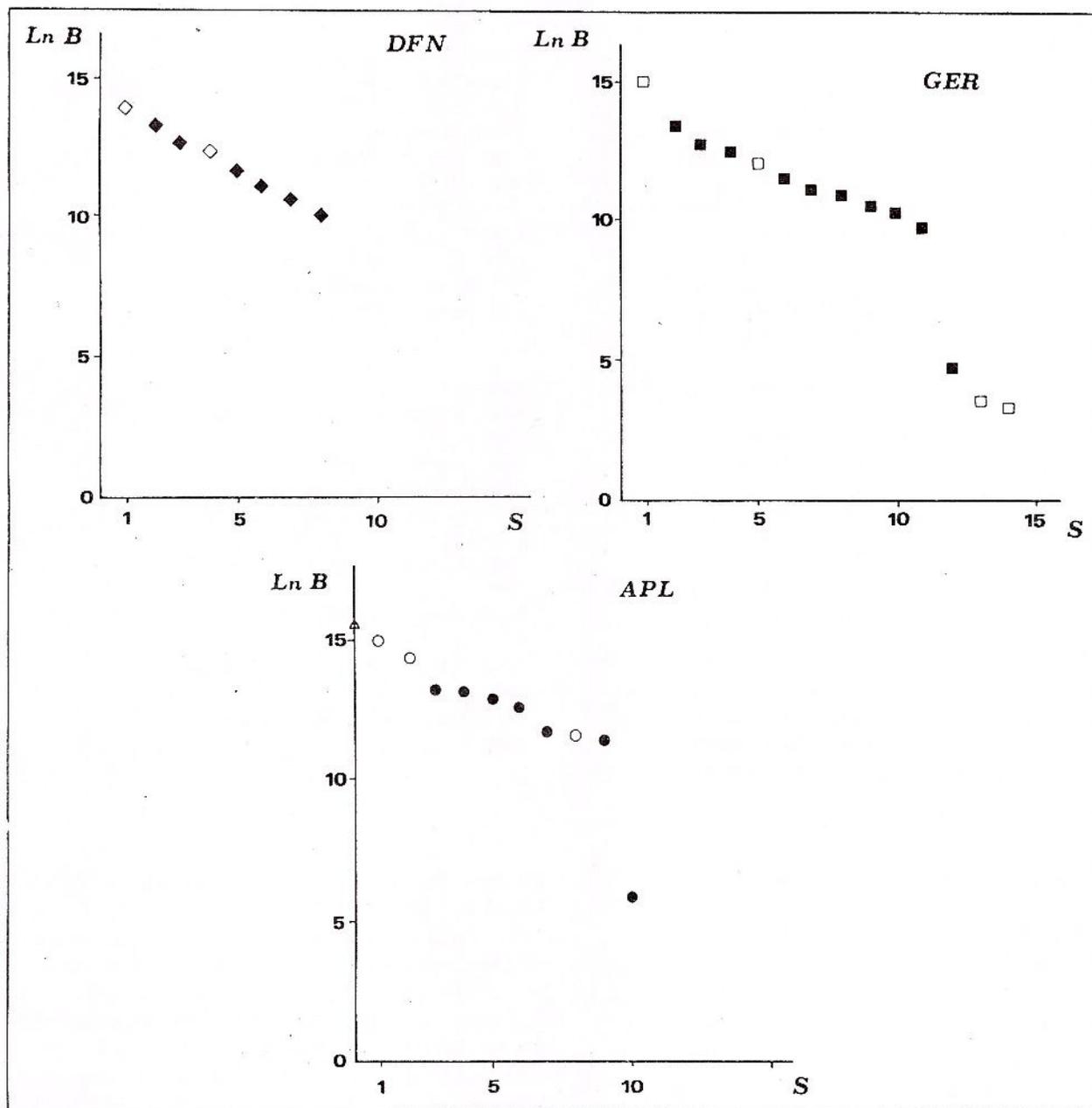


Fig. 3 - Cenograms of the Greek Villafranchian assemblages, DFN, GER, APL.

one fits well with those of "sub-arid/sub-humid savannahs" like Rwindi Rutschuru and Funtio Pecuario Masaguaral. Moreover, the APL cenogram is similar to that for the Greek Late Miocene locality of "Ravin de la Pluie", which represents a savannah (Bonis et al., 1992).

Comparison with the European and other localities

The methods of similarity indices and multivariate analysis have been applied in order to clarify the

possible palaeoecological relationships between the Greek and west-European Villafranchian faunas, as well as with faunas from Turkey, Asia Minor, Georgia and Siberia. The compared faunas are listed below.

A. Faunal Similarity. The similarity between two faunal assemblages can be expressed by several indices. The most common indices are those of Simpson (similarity index) and Pickford (index of distance), used by various authors. These indices indicate maximum similarity (100, Simpson's index) or maximum difference (100, Pickford's index).

Greece

VOL: Volakas;
 DFN: Dafnero-1;
 SES: Sesko, Central Greece,
 Symeonidis-1991/1992;
 GER: Gerakarou-1;
 LIB: Libakos, Western Macedonia,
 Steensma-1988;
 APL: Apollonia-1;
 APR: "Chronological fauna",
 including the elements of
 APL and RVL

France (Heitz et al., 1974)

VIA: Viallette
 ETO: Etouaires
 PAR: Pardines
 STV: St. Vallier
 CHI: Chillac
 COU: Coupet
 SEN: Seneze-1;
 BLG: Blassac la Girondie, Boueuf
 et al., 1992

Spain (Aguirre, Morales, 1990)

PUE: La Puebla de Valverde
 CAB: Casablanca-1
 VMI: Venta Micena
 CVI: Cueva Victoria

Italy (Azzaroli et al., 1986,

Masini et al., 1991)
 TRI: Triversa
 MON: Montopoli
 CSG: Costa st. Giacomo
 OLI: Olivola
 TAS: Il Tasso
 CFR: Casa Frata
 PND: Pirro Nord
 SLI: Slivia, Bon et al., 1992
 ISE: Isernia la Pineta

Others

GUL: Guliazi, Turkey, Sickenberg,
 Tobien, 1971
 SOL: Sogoturu, Turkey,
 Sickenberg, Tobien, 1971
 OUB: Oubeidiyeh, Israel
 Chernov, Guerin, 1985
 TER: Ternifne, Algeria, Geraads
 1981
 HAN: Hanjaka, Slovakia, Fejfar et
 al., 1990
 AKH: Akhalkalaki, Georgia, Vekua
 1986
 CHU: Chukochyan, Siberia, Sher,
 1986

$$\text{Simpson's index} = \frac{X \wedge Y}{X \vee Y} \times 100$$

$$\text{Pickford's index} = \frac{(X - X \wedge Y)(Y - X \wedge Y)}{XY} \times 100$$

where X = number of taxa in the first fauna, Y = number of taxa in the second fauna and X∧Y = number of common taxa.

Both indices are computed at specific level for 20 European Villafranchian faunal assemblages (Tab. 1, 2) and for 25 associations of artiodactyls (Tab. 3, 4).

The results, obtained for the complete faunal lists are not opposed to the results of artiodactyls associations (compare Tab. 1 to Tab. 3 and Tab. 2 to Tab. 4). On the contrary, the indices calculating for artiodactyls express more clearly the differences or similarities between the faunas. This is due to the fact that *Equus stenonis*, *Stephanorhinus etruscus*, *Archidiscodon meridionalis*, as well as numerous carnivores, present a high stability during Villafranchian. So the main differences between the faunas can be expressed by the changes in the associations of artiodactyls.

The VOL and DFN assemblages present the highest faunal similarity (Tabs. 1-4) with those of PAR, STV, CHI, COR, CAB and PUE, as well as with CSG (Simpson's index=50) from the Italian assemblages.

All these localities are dated to Middle or Middle-Late Villafranchian.

The GER assemblage is more similar to those from Middle-Late and Late Villafranchian of southern Europe, while it seems to be somewhat different from the Italian assemblages (Tabs. 1-4). The Pickford's index for the GER artiodactyls is <10 for COR, COU, CHI, and SEN. The latter three are dated to Late Villafranchian. The same index is >20 for all the Italian artiodactyl's associations, which are used and >15 for the Spanish ones of Early-Middle Villafranchian.

Concerning the total faunal lists (Tabs. 1, 2), the APL, CFR, PND, VMI and CVI assemblages of Latest Villafranchian present a relatively moderate similarity among them but higher comparatively to the other Villafranchian assemblages. Looking at the artiodactyls, this similarity seems to be stronger (Tabs. 3, 4).

A great similarity is also clear among the assemblages of Early Villafranchian. The similarity indices of VIA, ETO, TRI, MON and HUE assemblages are generally higher than 50 (Tab. 1). The distance among the artiodactyl associations of ETO, VIA, TRI, MON and HUE is also small (Pickford's index is generally lower than 50) (Tab. 4).



Table 1
Simpson's index of the total faunal assemblages from
4 Greek and 16 European Villafranchian localities,
calculated at specific level

| S-i | VOL | DFN | GER | APL | VIA | ETO | STV | COR | COU | SEN | MON | CSG | OLI | CFR | PND | HUE | PUE | CAB | VMI | CVI |
|-----|-----|------|------|------|------|------|------|-----|------|------|------|------|------|------|------|------|------|------|------|------|
| VOL | 100 | 61.5 | 37.5 | 25 | 13.3 | 31.3 | 62.5 | 80 | 58.3 | 62.5 | 16.6 | 37.5 | 31.3 | 12.5 | 12.5 | 37.5 | 62.5 | 42.8 | 18.7 | 12.5 |
| DFN | | 100 | 38.5 | 15.4 | 7.8 | 38.4 | 69.2 | 77 | 36.4 | 69.2 | 8.3 | 46.2 | 23 | 7.7 | 7.7 | 25 | 69.2 | 57 | 15.4 | 7.7 |
| GER | | | 100 | 29.4 | 13.3 | 11.1 | 33.3 | 70 | 58.3 | 44.4 | 16.6 | 37.5 | 28 | 12.5 | 12 | 50 | 33 | 57 | 18.7 | 22.2 |
| APL | | | | 100 | 6.7 | 17.6 | 17.6 | 10 | 8.3 | 17.6 | 0 | 12.5 | 17.6 | 25 | 35.3 | 0 | 29.4 | 14.3 | 56.3 | 41.2 |
| VIA | | | | | 100 | 86.7 | 33.3 | 20 | 25 | 33.3 | | | | | | | | | | |
| ETO | | | | | | 100 | 61.5 | 60 | 33.3 | 45.8 | 66.6 | 50 | 30 | 12.5 | 17.6 | 25 | 52.4 | 14.3 | 12.5 | 5 |
| STV | | | | | | | 100 | 100 | 75 | 70.8 | 75 | 81 | 45 | 31.3 | 23.5 | 50 | 90.5 | 57.1 | 19 | 25 |
| COR | | | | | | | | 100 | 50 | 80 | | | | | | | | | | |
| COU | | | | | | | | | 100 | 83.3 | | | | | | | | | | |
| SEN | | | | | | | | | | 100 | 41.6 | 68.8 | 50 | 25 | 23.5 | 37.5 | 81 | 57.1 | 25 | 20 |
| MON | | | | | | | | | | | 100 | 33.3 | 16.6 | 16.6 | 16.6 | 50 | 50 | 0 | 8.3 | 8.3 |
| CSG | | | | | | | | | | | | 100 | 43.7 | 19 | 12.5 | 50 | 68.8 | 28.6 | 18.8 | 18.8 |
| OLI | | | | | | | | | | | | | 100 | 56.3 | 35.3 | 0 | 35 | 14.3 | 31.3 | 30 |
| CFR | | | | | | | | | | | | | | 100 | 56.3 | 0 | 37.5 | 14.3 | 50 | 38 |
| PND | | | | | | | | | | | | | | | 100 | 0 | 29.4 | 0 | 56.3 | 45 |
| HUE | | | | | | | | | | | | | | | | 100 | 50 | 28.6 | 12.5 | 0 |
| PUE | | | | | | | | | | | | | | | | | 100 | 50 | 31 | 20 |
| CAB | | | | | | | | | | | | | | | | | | 100 | 43 | 14 |
| VMI | | | | | | | | | | | | | | | | | | | 100 | 56 |
| CVI | | | | | | | | | | | | | | | | | | | | 100 |

Table 2
 Pickford's index of the total faunal assemblages from
 4 Greek and 16 European Villafranchian localities,
 calculated at specific level

| P-i | VOL | DFN | GER | APL | VIA | ETO | STV | COR | COU | SEN | MON | CSG | OLI | CFR | PND | HUE | PUE | CAB | VMI | CVI |
|-----|-----|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|
| VOL | 0 | 19.2 | 41.6 | 57.3 | 75.8 | 56.5 | 23 | 10 | 23.4 | 22 | 73 | 39 | 51.5 | 76 | 77.2 | 51 | 19.6 | 46.4 | 66 | 79 |
| DFN | | 0 | 37.6 | 74.6 | 86 | 50.5 | 14.2 | 30.8 | 46.2 | 28.6 | 84.6 | 33.6 | 65.4 | 86.5 | 87 | 63.4 | 17.5 | 39.6 | 74 | 87.7 |
| GER | | | 0 | 51.9 | 77 | 82.5 | 51.2 | 18.3 | 25.4 | 37 | 74 | 41.6 | 54 | 78 | 78 | 39 | 47.6 | 33.3 | 68 | 62.2 |
| APL | | | | 0 | 88 | 73.5 | 73 | 85 | 86.3 | 72 | 100 | 77.2 | 70 | 57 | 42 | 100 | 54 | 81 | 20.5 | 38 |
| VIA | | | | | 0 | 7.1 | 66.6 | 69.3 | 60 | 53 | | | | | | | | | | |
| ETO | | | | | 0 | 0 | 16.5 | 31.4 | 67 | 33 | 24 | 36 | 55 | 81.3 | 73.5 | 69.6 | 29 | 82.6 | 81.3 | 91.6 |
| STV | | | | | | | 0 | 0 | 16.3 | 10 | 16.3 | 9.3 | 36 | 55.5 | 65 | 42.3 | 2.5 | 36.2 | 72 | 60.5 |
| COR | | | | | | | | 0 | 29.1 | 13.3 | | | | | | | | | | |
| COU | | | | | | | | | 0 | 10 | | | | | | | | | | |
| SEN | | | | | | | | | | 0 | 46 | 17 | 29 | 62.5 | 64 | 54.6 | 5.5 | 35.7 | 62.5 | 66.5 |
| MON | | | | | | | | | | | 0 | 50 | 75 | 73 | 73.5 | 33.3 | 35.7 | 100 | 86 | 87 |
| CSG | | | | | | | | | | | | 0 | 36.5 | 66 | 77.2 | 37.5 | 15 | 62.5 | 66 | 69 |
| OLI | | | | | | | | | | | | | 0 | 24 | 45.3 | 100 | 43.4 | 81.4 | 51.6 | 49 |
| CFR | | | | | | | | | | | | | | 0 | 20 | 100 | 44.6 | 80.3 | 39 | 44 |
| PND | | | | | | | | | | | | | | | 0 | 100 | 54 | 100 | 20 | 38 |
| HUE | | | | | | | | | | | | | | | | 0 | 40.4 | 53.6 | 82 | 100 |
| PUE | | | | | | | | | | | | | | | | | 0 | 40.4 | 52 | 65 |
| CAB | | | | | | | | | | | | | | | | | | 0 | 46 | 81 |
| VMI | | | | | | | | | | | | | | | | | | | 0 | 24 |
| CVI | | | | | | | | | | | | | | | | | | | | 0 |

Table 3
Simpson's index of the actiodactyls associations from
4 Greek and 21 European Villafranchian localities,
calculated at specific level

| S-i | VOL | DFN | GER | APL | VIA | ETO | PAR | STV | COR | COU | CHI | SEN | TRI | MON | CSG | OLI | TAS | CFR | PND | SLI | HUE | PUE | CAB | VMI | CVI | |
|-----|-----|-----|------|-----|-----|------|------|------|------|------|-----|------|-----|------|------|------|------|-----|-----|-----|-----|------|------|-----|-----|---|
| VOL | 100 | 86 | 62.5 | 0 | 20 | 12.5 | 83.5 | 71.5 | 83.5 | 66.5 | 100 | 66.5 | 25 | 33.3 | 50 | 33.3 | 0 | 25 | 0 | 0 | 75 | 83.3 | 100 | 0 | 0 | |
| DFN | | 100 | 71.5 | 0 | 0 | 0 | 50 | 57 | 50 | 80 | 80 | 57 | 25 | 16.5 | 57 | 16.6 | 0 | 0 | 0 | 0 | 50 | 66.5 | 100 | 0 | 0 | |
| GER | | | 100 | 0 | 20 | 12.5 | 66.5 | 57 | 83.5 | 100 | 80 | 75 | 25 | 33.3 | 50 | 33.3 | 66.5 | 0 | 0 | 0 | 75 | 66.5 | 100 | 0 | 0 | |
| APL | | | | 100 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 33.3 | 25 | 25 | 50 | 0 | 0 | 0 | 57 | 40 | |
| VIA | | | | | 100 | 100 | 20 | 20 | 20 | 20 | 20 | 40 | 25 | 40 | 20 | 0 | 0 | 0 | 0 | 0 | 0 | 25 | 20 | 0 | 0 | 0 |
| ETO | | | | | | 100 | 33.3 | 28.5 | 33.3 | 16.6 | 20 | 25 | 50 | 50 | 37.5 | 0 | 0 | 0 | 0 | 0 | 0 | 50 | 33.3 | 0 | 0 | 0 |
| PAR | | | | | | | 100 | 83.5 | 100 | 66.5 | 80 | 66.5 | 25 | 50 | 66.5 | 0 | 0 | 25 | 0 | 0 | 100 | 83.3 | 100 | 0 | 0 | |
| STV | | | | | | | | 100 | 83.5 | 66.5 | 100 | 71 | 50 | 16.5 | 71.5 | 16.6 | 0 | 0 | 0 | 0 | 100 | 100 | 100 | 0 | 0 | |
| COR | | | | | | | | | 100 | 83.5 | 80 | 83.5 | 25 | 50 | 66.5 | 16.6 | 33.3 | 0 | 0 | 0 | 100 | 83.3 | 100 | 0 | 0 | |
| COU | | | | | | | | | | 100 | 80 | 100 | 25 | 16.5 | 66.5 | 33.3 | 66.6 | 0 | 0 | 0 | 75 | 66.5 | 100 | 0 | 0 | |
| CHI | | | | | | | | | | | 100 | 100 | 25 | 40 | 80 | 20 | 0 | 0 | 0 | 0 | 75 | 100 | 100 | 0 | 0 | |
| SEN | | | | | | | | | | | | 100 | 25 | 33.3 | 62.5 | 66.5 | 66.5 | 0 | 0 | 0 | 25 | 83.3 | 100 | 0 | 0 | |
| TRI | | | | | | | | | | | | | 100 | 50 | 25 | 0 | 0 | 0 | 0 | 0 | 25 | 25 | 0 | 0 | 0 | |
| MON | | | | | | | | | | | | | | 100 | 66.5 | 0 | 0 | 0 | 0 | 0 | 75 | 50 | 0 | 0 | 0 | |
| CSG | | | | | | | | | | | | | | | 100 | 33.3 | 33.3 | 0 | 0 | 0 | 100 | 83.3 | 50 | 0 | 0 | |
| OLI | | | | | | | | | | | | | | | | 100 | 66.5 | 50 | 25 | 25 | 0 | 16.6 | 0 | 0 | 0 | |
| TAS | | | | | | | | | | | | | | | | | 100 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | |
| CFR | | | | | | | | | | | | | | | | | | 100 | 25 | 25 | 0 | 0 | 0 | 25 | 0 | |
| PND | | | | | | | | | | | | | | | | | | | 100 | 75 | 0 | 0 | 0 | 25 | 0 | |
| SLI | | | | | | | | | | | | | | | | | | | | 100 | 0 | 0 | 0 | 50 | 25 | |
| HUE | | | | | | | | | | | | | | | | | | | | | 100 | 100 | 50 | 0 | 25 | |
| PUE | | | | | | | | | | | | | | | | | | | | | | 100 | 100 | 0 | 0 | |
| CAB | | | | | | | | | | | | | | | | | | | | | | | 100 | 0 | 0 | |
| VMI | | | | | | | | | | | | | | | | | | | | | | | | 100 | 60 | |
| CVI | | | | | | | | | | | | | | | | | | | | | | | | | 100 | |

Table 4
 Pickford's index of the actiodactylis associations from
 4 Greek and 21 European Villafranchian localities,
 calculated at specific level

| P-i | VOL | DFN | GER | APL | VIA | ETO | PAR | STV | COR | COU | CHI | SEN | TRI | MON | CSG | OLI | TAS | CFR | PND | SLI | HUE | PUE | CAB | VMI | CVI |
|-----|-----|-----|------|-----|-----|------|------|------|------|------|-----|------|------|------|------|------|------|------|------|------|------|------|-----|------|-----|
| | 0 | 5 | 16.5 | 100 | 71 | 78 | 7.5 | 12.7 | 7.5 | 18.5 | 0 | 13.5 | 66.5 | 52 | 28 | 52 | 100 | 66.5 | 100 | 100 | 16.6 | 7.5 | 0 | 100 | 100 |
| | | 0 | 11 | 100 | 100 | 100 | 28.5 | 18.3 | 28.5 | 28.5 | 9 | 30 | 64.3 | 71.5 | 21.5 | 71.5 | 100 | 100 | 100 | 100 | 36 | 14.3 | 0 | 100 | 100 |
| | | | 0 | 100 | 70 | 76.5 | 16.5 | 21.5 | 6 | 0 | 10 | 10 | 65.5 | 50 | 25 | 50 | 25 | 100 | 100 | 100 | 15.6 | 16.5 | 0 | 100 | 100 |
| | | | | 0 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 57 | 64.3 | 64.3 | 36 | 100 | 100 | 100 | 6 | 43 |
| | | | | | 0 | 0 | 66.5 | 68.5 | 66.5 | 66.5 | 64 | 48 | 60 | 40 | 70 | 100 | 100 | 100 | 100 | 100 | 60 | 66.5 | 100 | 100 | 100 |
| | | | | | 0 | 0 | 50 | 53.5 | 50 | 73 | 70 | 60 | 37.5 | 31.3 | 39 | 100 | 100 | 100 | 100 | 100 | 37.5 | 50 | 100 | 100 | 100 |
| | | | | | | 0 | 0 | 5 | 0 | 11 | 9 | 20 | 62.5 | 25 | 16.5 | 100 | 100 | 62.5 | 100 | 100 | 0 | 3 | 0 | 100 | 100 |
| | | | | | | | | 0 | 5 | 14.3 | 0 | 14.3 | 36 | 14.3 | 10.7 | 71.5 | 100 | 100 | 100 | 100 | 0 | 0 | 0 | 100 | 100 |
| | | | | | | | | | 0 | 3 | 7 | 8 | 62.5 | 25 | 16.5 | 69.5 | 55.5 | 100 | 100 | 100 | 0 | 3 | 0 | 100 | 100 |
| | | | | | | | | | 0 | 7 | 0 | 0 | 62.5 | 69.5 | 16.5 | 44.5 | 22.2 | 100 | 100 | 100 | 12.5 | 11 | 0 | 100 | 100 |
| | | | | | | | | | 0 | 0 | 0 | 0 | 60 | 40 | 10 | 60.5 | 100 | 100 | 100 | 100 | 10 | 0 | 0 | 100 | 100 |
| | | | | | | | | | | | | 0 | 67.5 | 53.3 | 19 | 20 | 26.5 | 100 | 100 | 100 | 17.5 | 8.3 | 0 | 100 | 100 |
| | | | | | | | | | | | | 0 | 0 | 33.3 | 65.5 | 100 | 100 | 100 | 100 | 100 | 56 | 62.5 | 100 | 100 | 100 |
| | | | | | | | | | | | | 0 | 0 | 0 | 16.5 | 100 | 100 | 100 | 100 | 100 | 12.5 | 25 | 100 | 100 | 100 |
| | | | | | | | | | | | | | | 0 | 50 | 58.3 | 100 | 100 | 100 | 100 | 0 | 6.3 | 44 | 100 | 100 |
| | | | | | | | | | | | | | | | 0 | 0 | 22.2 | 33.3 | 62.5 | 62.5 | 100 | 69.5 | 100 | 100 | 100 |
| | | | | | | | | | | | | | | | | 0 | 0 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 |
| | | | | | | | | | | | | | | | | | | 0 | 56.3 | 56.3 | 100 | 100 | 100 | 64.3 | 100 |
| | | | | | | | | | | | | | | | | | | 0 | 0 | 6.3 | 100 | 100 | 100 | 64.3 | 60 |
| | | | | | | | | | | | | | | | | | | | 0 | 0 | 100 | 100 | 100 | 36 | 60 |
| | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 50 | 100 | 100 |
| | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 100 | 100 |
| | | | | | | | | | | | | | | | | | | | | | | | 0 | 100 | 100 |
| | | | | | | | | | | | | | | | | | | | | | | | 0 | 23 | 0 |

The relationships of OLI assemblage are more complicated. The Simpson's index between OLI and SEN is 66.5 (50 for artiodactyls) and between OLI and GER only 33.3 (28 for artiodactyls) (Tabs. 1, 3). However, the artiodactyls associations of GER and SEN, are more similar to those of Middle Villafranchian and differ from that of OLI and TAS (Tab. 2). The artiodactyl association of OLI presents a great distance (Pickford's index >40) from all the Early, Middle and Late Villafranchian associations of southern Europe, except that of SEN. Thus, it seems possible that the faunal changes characterizing the Latest Villafranchian have already been started from the OLI assemblage.

Finally, the study of the values (Tabs. 1-4) shows that the faunas of each chronological level, Early, Middle, Late and Latest Villafranchian, present a high similarity, which probably reflects a relative homogeneity of the environmental conditions.

B. Multivariate analysis. The programs ANCORR (ADDAD library) have been applied for multivariate analysis. The method has been used by Bonis et al. (1992 a, b, 1994) for the late Miocene Greek and Turkish mammalian faunas. 19 recent mammalian faunas from Africa are selected as basis for the comparison (for more information see Bonis et al., 1992 a):

TRA: Transvaal 10, Tropical forest
 KIW: Mont Kiwu, Mountain forest
 LMT: Lamto, Tropical forest and Savannah
 MAK: Makokou, Tropical forest
 LMB: La Maboke, Tropical forest
 TAI: Tai, Tropical forest
 TRN: Transvaal 7, Savannah woodland
 LOK: Lokori, Savannah woodland
 TRV: Transvaal 2, Mountain grassland
 KAF: Kafue, Flood plain grassland

Cervidae (v 8, 43 %), Primates (v 11, 27 %) and Equidae (v 3, 11 %) have the most important role in this division. The recent faunas, in the left part of the diagram, are separated by the second factor (23 %) in "forestrial" and "non forestrial". This is mainly due to the variables v 11 (Primates, 30 %), v 15 (Lagomorpha, 15.6 %) and v 7 (Tragulidae, 8 %). In the right part of the diagram, the Greek Villafranchian localities are also separated in two groups: the first one includes the faunas of DFN, SES, GER, VOL and LIB, while the second includes the fauna of APL, as well as the chronological assemblage APR. The family of Cervidae (v 8, 29 %) has the most important role in this division. Although a good parallelism on the second factor, between recent and fossil faunas is visible, an ecological identification seems to be uncertain. However, it is clear that the Greek localities of Middle and Late Villafranchian could correspond to a more mixed or forestrial environment than the Latest Villafranchian localities, which seems to be closer to open environments.

D. Multivariate study of recent and Villafranchian faunas (Fig. 5). The multivariate analysis of the recent and all the selected Villafranchian fauna gives more palaeoecological information. The first factor (41 %), with the main variables of Cervidae (v 8, 32 %) and Primates (v 11, 26 %), separates recent

RUV: Rukwa Valley, Flood plain grassland;
 TVL: Transvaal 5, Arid bushland
 SRG: Serengeti, Savannah woodland
 TSV: Transvaal 4, Savannah woodland
 TNL: Transvaal 9, Savannah woodland
 ZIN: Zinave, Savannah woodland
 GAB: Gabiro, Wooded grassland
 RDR: Rwindi-Rutschuru, Arid bushland
 ANB: Amboseli, Arid bushland

The mammals of both fossil and recent localities, are classified according to taxonomy. The 15 selected taxa represent the corresponding variables (v), used in the analysis: Proboscidea (v1); Rhinocerotidae (v 2); Equidae (v 3); Suidae (v 4); Hippopotamidae (v 5); Camelidae (v 6); Tragulidae (v 7); Cervidae (v 8); Bovidae (v 9); Giraffidae (v 10); Primates (v 11); Pholidota (v 12); Tapiridae (v 13); Carnivora (v14); Lagomorpha (v 15).

C. Multivariate study of recent and Villafranchian Greek faunas (Fig. 4). The elaboration of the data coming from the recent and the Greek Villafranchian faunas gave interesting information. The first factor (48 %) separates recent from Villafranchian faunas.

from fossil faunas. The second factor (17 %), with the main variables of Primates (v11, 34 %), Lagomorpha (v 15, 8.5 %) and Bovidae (v 9, 8.5 %), divides the recent faunas in "forestrial" and "non forestrial". In the right part of the diagram the fossil faunas have a more continuous distribution. However, most of the Early, Middle and Late Villafranchian localities are situated below or on the horizontal axis, while majority of the Latest Villafranchian localities are above this axis. The most important variables for this division are Tapiridae (v 13), Bovidae (v 9), Cervidae (v 8) and Lagomorpha (v 15). Thus, a more forestrial character could be supposed for the Early, Middle and Late Villafranchian faunas. The Latest Villafranchian faunas seem to correspond better to



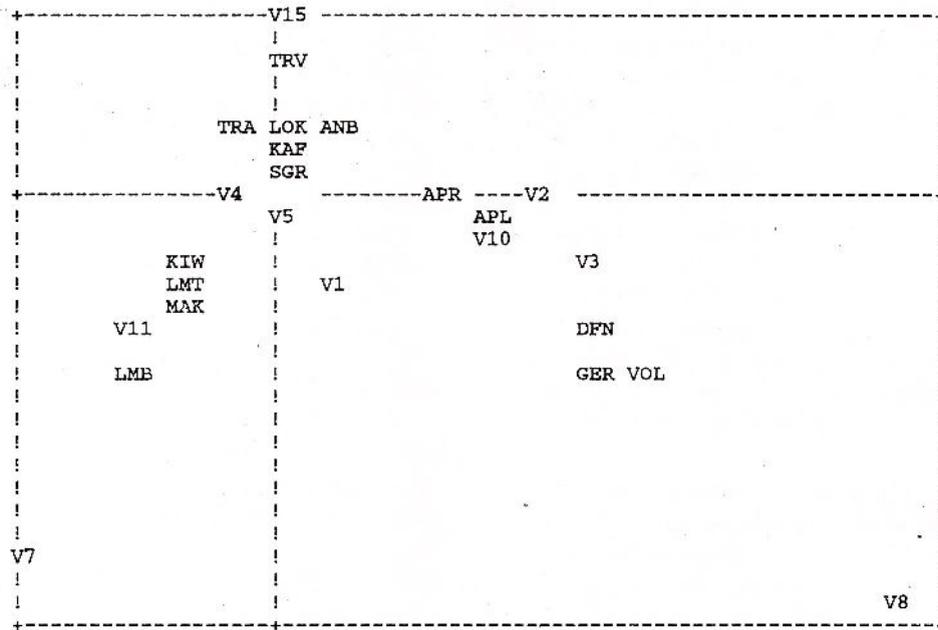


Fig. 4 - Multivariate analysis of recent and Greek Villafranchian localities [superposed points = TRN 9TRA), TVL (LOK), TSV (KAF), TNL (KAF), RDR (KAF), ZIN (SGR), GAB (SGR), V9 (SGR), V14 (SGR), TAI (MAK), LIB (DFN), V12 (V11)].

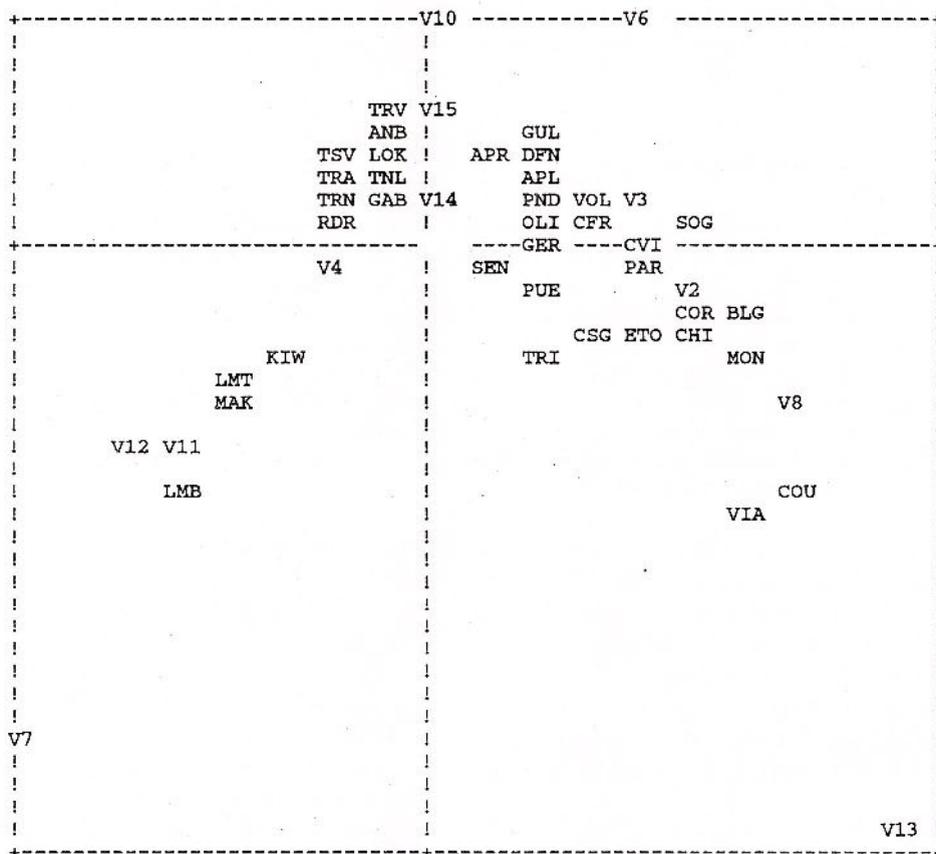


Fig. 5 - Multivariate analysis of recent and Villafranchian faunas [superposed points = TER (ANB), TVL (LOK), VMI (DFN), SLI (DFN), OUB (APR), KAF (TRA), SES (APL), AKH (APL), V5 (TNL), V9 (TNL), RUV (TRN), SGR (TRN), ZIN (TRN), LIB (VOL), STV (GER), ISE (CHI), V1 (ETO), TAI (LMT), HAN (VIA)].



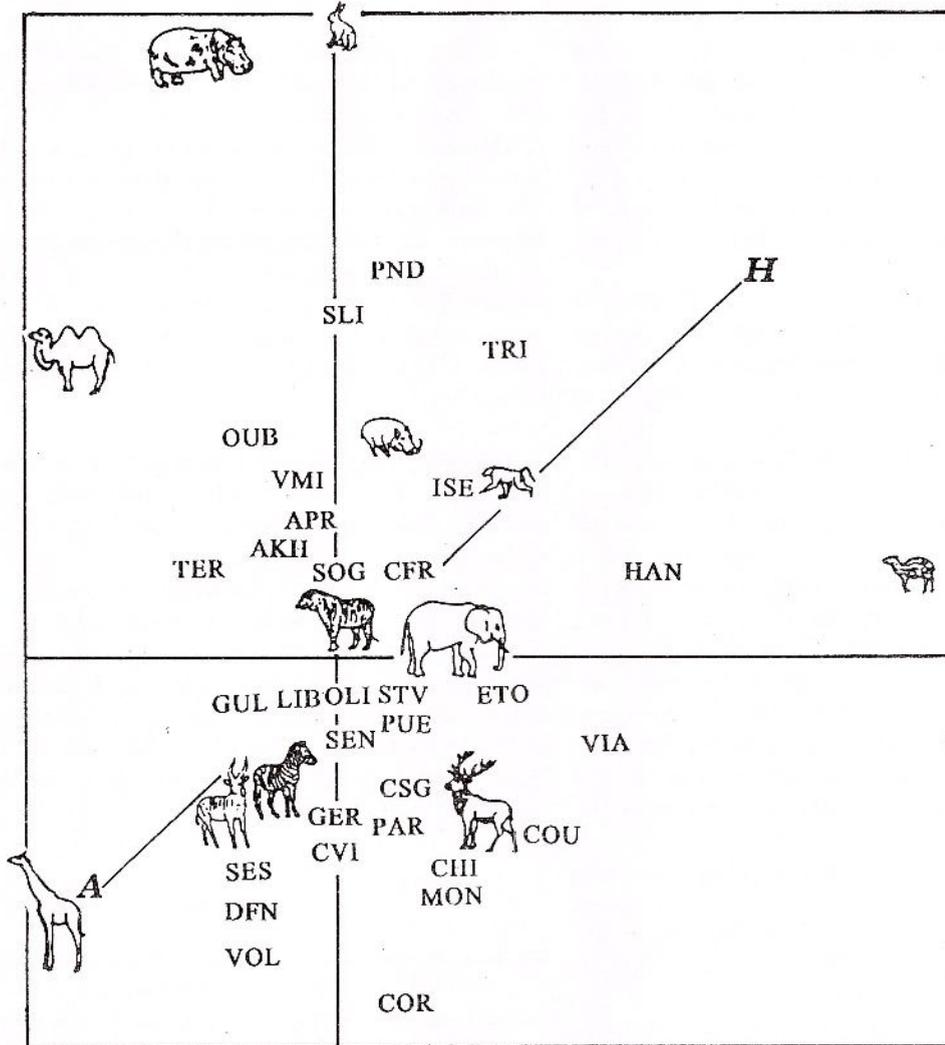


Fig. 6 - Multivariate analysis of the selected Villafranchian faunas.

an open environment. Concerning the age of the fossil faunas, a progressive reduction of the forestrial and humid character from Early to Latest Villafranchian faunas could be suggested. This observation fits also well with those given by other authors (e.g. Faure, Guerin, 1984; Bonifay, 1993; Palombo, 1994; Suc, 1995).

The faunas DFN, VOL, SES and GUL of Middle Villafranchian age are situated in the upper part of the diagram, as an exception of the previous observations. The variable of Giraffidae (v 10) is the main reason for this position. The hypothesis that all these faunas correspond to an open environment must be possible. The fact that the above-mentioned faunas are geographically situated at the eastern part of the Mediterranean region probably suggests a different zoogeographical province in eastern Mediterranean during Middle Villafranchian.

E. Multivariate study of Villafranchian faunas (Fig. 6). In this case of multivariate analysis, where the recent faunas have been excluded, a good separation on the second factor is observed. The faunas of Early, Middle and Late Villafranchian are situated below the horizontal axis, while the Latest Villafranchian ones, as well as HAN and TRI, are situated in the upper part of the diagram. The Middle Villafranchian faunas of DFN, VOL, SES and GUL are situated along a vertical line in the left side of the lower part of the diagram, separated from the contemporaneous ones of western Europe. The position of the variables in the diagram gives interesting information. The open environment taxa (Iagomorphs, hippopotamids, camelids, bovids and giraffids) are situated in the left part of the graph, while the closed environment taxa (suids, primates, tapirids and cervids) in the right one. This probably reflects an arrangement of the faunas, from those of open en-

vironment (left side) to those of closed environment (right side) (Fig. 6). Moreover, the taxa indicating arid conditions are concentrated in the lower-left corner of the graph, while the taxa indicating humid conditions, occupy the upper-right side. An arrangement of the faunas from those of arid to those of humid conditions along the axis A/H is also possible (Fig. 6).

If this hypothetical model is right, then the Earliest Villafranchian faunas of TRI and HAN correspond to a close (forest) but more humid environment than those of Early Villafranchian ETO, VIA, MON. In the Middle and Late Villafranchian faunas the forestal character decreases. The west-European Villafranchian faunas of this period, as well as GER, correspond to a mixed temperate environment. Azzaroli et al. (1988) and Turner (1995) agree with such a similar environment, while the pollen record suggests an "Artemisia steppe-deciduous forest replacement", which predominates in the N. Mediterranean region from 2.5-1.0 Ma (Suc et al., 1995). Moreover, according to this model, the Latest Villafranchian faunas correspond to an open and humid environment. The faunas of VMI (macromammalian), APR and AKH seem to have a less humid character than the isochronous Italian sites.

In the above-mentioned interpretations it is difficult to clarify if the referred "humidity" is correlated (or even confused) with coolness. Concerning data, coming from other authors (Shackleton et al., 1984; Turner, 1995; Suc et al., 1995), it seems that the referred decrease of humidity from Early to Middle Villafranchian (or from 2.5 Ma upwards) is correlated with the first major cool event of the northern hemisphere. In addition, the apparent increase of humidity in the Latest Villafranchian faunas is probably confused with the significant coolness, which took place at about 1.0 Ma (e.g. Suc et al., 1995). The presence of large-sized horses, *Praeovibos*, *Soergelia* and *Pontoceros* in the APL and VMI faunas, indicates open habitats with cold climates (about horses see also Alberti et al., 1995).

Conclusions

All the above-mentioned data from different palaeoecological methods are in good agreement. According to these data, the palaeoecological study of the Villafranchian faunas of Macedonia suggests:

- an open, savannah-like, environment with relatively sub-arid conditions, in Middle Villafranchian (DFN, VOL faunas),

- an open forest-savannah woodland environment with temperate conditions, in Late Villafranchian (GER, KRI, VSL faunas) and

- a savannah grassland, relatively humid environment with colder conditions, in Latest Villafranchian (APL, RVL faunas).

Moreover, a reduction of the forestal and humid character is observed from the Earliest to the Late Villafranchian faunas of southern Europe. Nevertheless, humidity and/or coolness increase in Latest Villafranchian. In addition, the climatic conditions of Middle Villafranchian seem to be different between western and eastern Mediterranean, while in Late and Latest Villafranchian there is a relative climatic homogeneity.

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Plate

Revised and completed faunal lists of the Villafranchian mammalian localities from Macedonia (Greece). The determination of the carnivores from VOL is according to Sickenberg, 1968.



| | DFN | VOL | GER | VSL | KRI | APL | RVL |
|--|-----|-----|-----|-----|-----|-----|-----|
| Elephantidae indet. | | | + | | | + | |
| <i>Nyctereutes megamastoides</i> | + | + | | | | | |
| <i>Vulpes alopecoides</i> | + | + | | | | + | |
| <i>Vulpes praecorsac</i> | | + | | | | | |
| <i>Alopex praeglacialis</i> | | + | | | | | |
| <i>Canis etruscus</i> | | | + | | | | |
| <i>Canis arnensis</i> | | | + | | | | + |
| <i>Canis apolloniensis</i> n.sp. | | | | | | + | |
| <i>Canis (Xenocyon)</i> sp. | | | | | | + | |
| <i>Ursus etruscus</i> | + | | | + | | + | |
| <i>Bosdagius felinus</i> | | + | | | | | |
| <i>Meles dimitrius</i> | | | + | | | + | |
| <i>Baranogale</i> cf. <i>helbingi</i> | + | | | | | | |
| <i>Putorius</i> sp. | | | | | | | + |
| <i>Pliohyaena perrieri</i> | | | + | | | | |
| <i>Pliohyaena brevirostris</i> | | | + | | | + | |
| <i>Chasmaporthetes lunensis</i> | + | | | | | | |
| <i>Panthera gombaszoegensis</i> | | | + | | | | |
| <i>Megantereon megantereon</i> | | + | | | | | |
| <i>Megantereon cultridens</i> | | | | | | + | |
| <i>Lynx issiodorensis</i> | | ? | | | | + | |
| <i>Equus stenonis</i> | | + | | + | | + | + |
| <i>Equus stenonis mygdoniensis</i> | | | + | | cf. | | |
| <i>Equus stenonis</i> cf. <i>vireti</i> | + | | | | | | |
| <i>Stephanorhinus etruscus</i> | | | | | + | | |
| Rhinocerotidae indet. | | + | | | | + | |
| <i>Sus strozzii</i> | | | + | + | | | |
| <i>Hippopotamus amphibius antiquus</i> | | | | | | | + |
| <i>Mitilanotherium martinii</i> | + | + | | | | | |
| <i>Eucladoceros senezensis</i> | + | + | | | + | | |
| <i>Eucladoceros senezensis</i> cf. <i>senezensis</i> | | | + | | | | |
| " <i>Cervus</i> " <i>philisi</i> cf. <i>valliensis</i> | + | | | | | | |
| " <i>Cervus</i> " <i>philisi</i> | | + | | | | | |
| " <i>Cervus</i> " sp. | | | + | | | | |
| <i>Croizetoceros ramosus gerakarensis</i> n.ssp. | | | + | | | | |
| <i>Croizetoceros ramosus</i> | | + | | | | | |
| <i>Megaloceros</i> sp. | | | | | | + | |
| <i>Leptobos etruscus</i> | | | | + | | | |
| <i>Leptobos</i> cf. <i>etruscus</i> | | | + | | | | |
| <i>Leptobos</i> | | cf. | | | | | |
| " <i>Leptobos</i> " sp. | | | | | + | | |
| <i>Bison (Eobison)</i> n.sp. | | | | | | + | |
| Bovinae indet. | | | + | | | | |
| <i>Gazellospira torticornis</i> | + | cf. | | | cf. | | |
| <i>Gazellospira</i> sp. | | | ? | | | | |
| <i>Gallogoral meneghinii sickenbergii</i> n.ssp. | | + | | | | | |
| <i>Gallogoral</i> cf. <i>meneghinii</i> | + | | | | | | |
| <i>Gazella bouvraini</i> n.sp. | + | sp. | + | | | | |
| <i>Gazella</i> cf. <i>bouvraini</i> | | | | + | | | |
| " <i>Prostrepsiceros</i> " <i>koufosi</i> n. sp. | | | + | | | | |
| <i>Procampoceras</i> sp. | | ? | | ? | | | |
| <i>Pontoceros ambiguus mediterraneus</i> n.ssp. | | | | | | + | |
| <i>Soergelia brigittae</i> n.sp. | | | | | | + | |
| <i>Praeovibos</i> sp. | | | | | | + | |
| Caprinae indet. | | | | | | + | |
| Antilopinae indet. | + | | | | | | |

TIMING OF MAJOR EVENTS IN THE NEOGENE GEOLOGIC EVOLUTION OF THE BETIC CORDILLERA (SOUTH SPAIN)

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Key words: Neogene. Alboran Sea. Betic Cordillera. Sedimentary record. Geodynamic evolution. Southern Spain.

Abstract: According to the main sedimentary discontinuities in the Neogene sedimentary record of the Betic Cordillera, it is possible to consider seven time spans during the Neogene (Oligocene-Early Aquitanian, Late Aquitanian-Early Burdigalian, Late Burdigalian-Langhian, Serravallian, Tortonian, Messinian and the Pliocene-Quaternary) in which different tectono-sedimentary events occurred. Each tectono-sedimentary event is characterized by the installation of a different palaeogeography and by the destruction of sedimentary basins and the formation of others.

1. Introduction

The Betic Cordillera (Fig. 1) is generally divided in two main domains: the Internal and the External Zones. The Internal Zones are formed, from bottom to top, of three complexes tectonically superposed: the Nevado-Filabride, the Alpujarride and the Malaguide. These complexes are composed mainly of Palaeozoic and Triassic rocks, and the two lower ones are affected by Alpine metamorphism. The External Zones, which formed the southern and southeastern margin of the Iberian Plate, are made up of Secondary and Tertiary sediments and are divided in Subbetic and Prebetic. The Campo de Gibraltar Complex is part of the former Magrebian-Masyllian trough originally extending from Calabria to the Rif. From the Early Miocene, the collision of the Internal Zones with the southern palaeomargin of Iberia caused its rotation, disarticulation and expelling towards the foreland (especially the Subbetic); coetaneously some Neogene basins were formed (most were destroyed later). Two of these remained during the Neogene, although strongly transformed: the Guadalquivir Basin, in a northwestern position and with a foreland basin character, and the Alboran Basin, south of the Betic chain, which can be considered to be the back-arc basin.

2. Major events

The main discontinuities recognized in the Betic Neogene sedimentary record divide it into seven time

spans. Each discontinuity corresponds to tectonic deformation (Fig. 2).

2.1. Oligocene to Early Aquitanian

The structuring of the Betic Internal Zones, which are estimated to have occupied (during the Late Mesozoic and Early-middle Tertiary) an original position some 400 or 500 km to their present position (Sanz de Galdeano, 1990; Sanz de Galdeano et al., 1990), was completed during this period. Towards the end of this time (Early Aquitanian) no sediments occur in the Internal Zones, with the exception of the Malaguide Complex, that received detrital sediments.

In the External Zone, the Prebetic received detrital sedimentation, with marine and continental intercalations, ending towards the Late Oligocene-Aquitanian as marine deposits (Jerez, 1979). In the Subbetic, which shows clear signs of tectonic instability, such as some olistostromal deposits in the Southern Subbetic trough (Comas, 1978), the sedimentary record is composed mainly of sandstones, limestones and marls.

The Alboran Basin had probably not yet begun to open, although Mulder, Parry (1977) propose this time for the beginning of its opening. The Internal Zones of the Betic Cordillera formed part of the Eoalpine chain (belonging to Europe or at least situated in an intermediate position between Europe and Africa), which ran down through the south of Sardinia and the west of Sicily. On the Malaguide Complex, several basins with detrital sediments started to



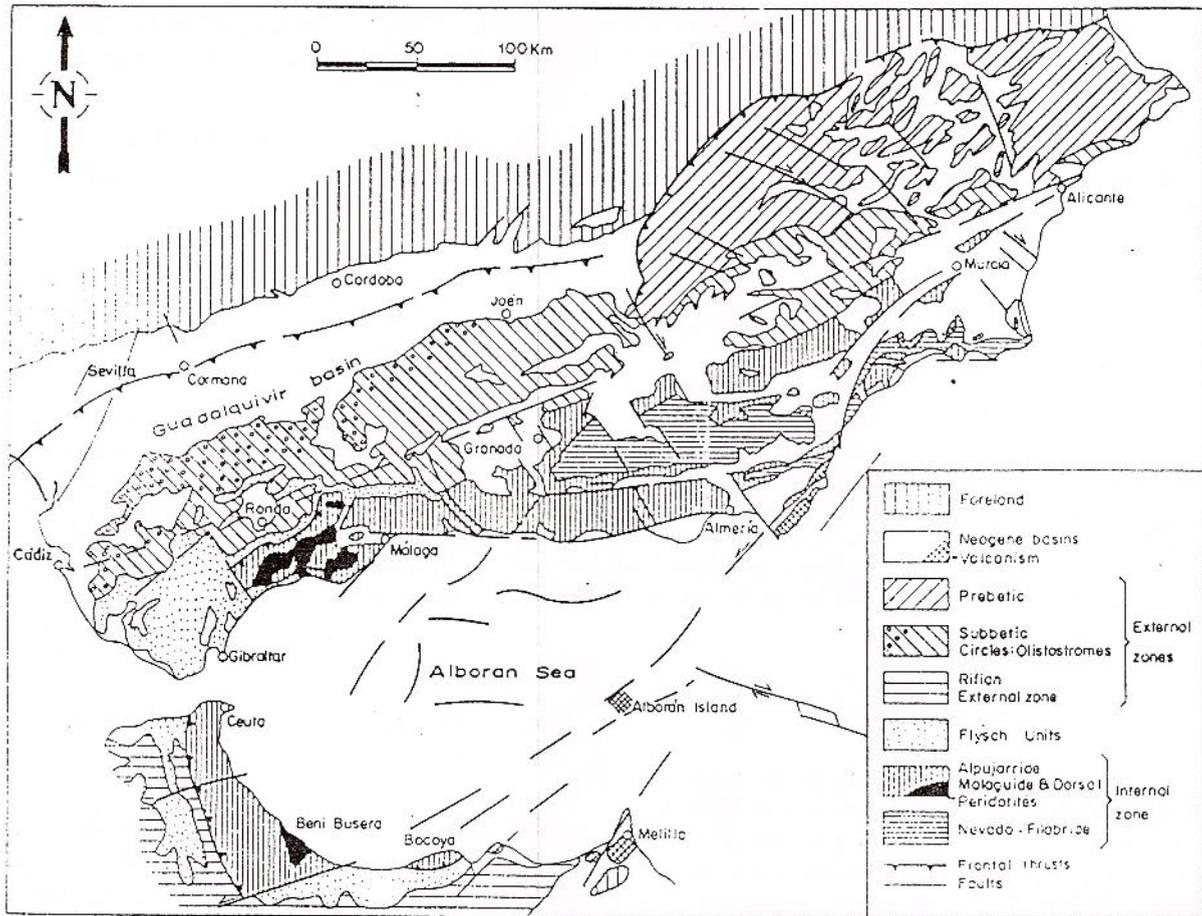


Fig. 1 - Geological sketch of the Betic Cordillera.

form. These are represented today, for example, in the Pantano de Andrade (Bourgeois, 1978; Gonzáles Donoso et al., 1983) and Alosaina (Bourgeois et al., 1972) or Ciudad Granada formations (Mac Gillavry et al., 1963; Jutson, 1980). These formations are grouped in the Ciudad Granada Group by Martín-Algarra (1987) and are partially contemporaneous with the Numidian Flysch deposits that accumulated in the Magrebian-Masyllian trough.

This period coincides with the first detected cooling of the Nevado-Filabride complex (which does not outcrop at this time) according to De Jong (1991).

2.2. Late Aquitanian to Early Burdigalian

In the Internal Zone and along its contact with the External Zone several unconformable sedimentary formations (the Alamo Fm., the Espejos Fm., the Murchas Fm., the Alamillos Fm., the Viñuela Fm., the Almogia, Casabermeja and Las Millanas Fm., grouped in the Viñuela Group by Martín-Algarra,

1987), developed on the Malaguide and Alpujarride basement.

During the Early Burdigalian, or towards the end of it, the suture between the Internal and External zones was formed. Thus, the Internal Zones, moving westwards from their original position, locally overthrust the southern edge of the Subbetic in a dextral transpressional contact (westward displacement of the Internal Zone, Andrieux et al., 1971; Tapponier, 1977; Sanz de Galdeano, 1983; Leblanc, Olivier, 1984; Olivier, 1984; Maldonado, 1985; Hermes, 1985). For this reason, some of the previously cited contemporaneous formations were affected by these horizontal movements (Mac Gillavry et al., 1963; Bourgeois, 1978; Sanz de Galdeano, 1983; Martín-Algarra, 1987; García Dueñas et al., 1988; Galindo Zaldívar et al., 1989; Jabaloy et al., 1992).

This event strongly affected the Subbetic Zone, to the extent that the units of its eastern edge are almost completely absent, its central part have under-

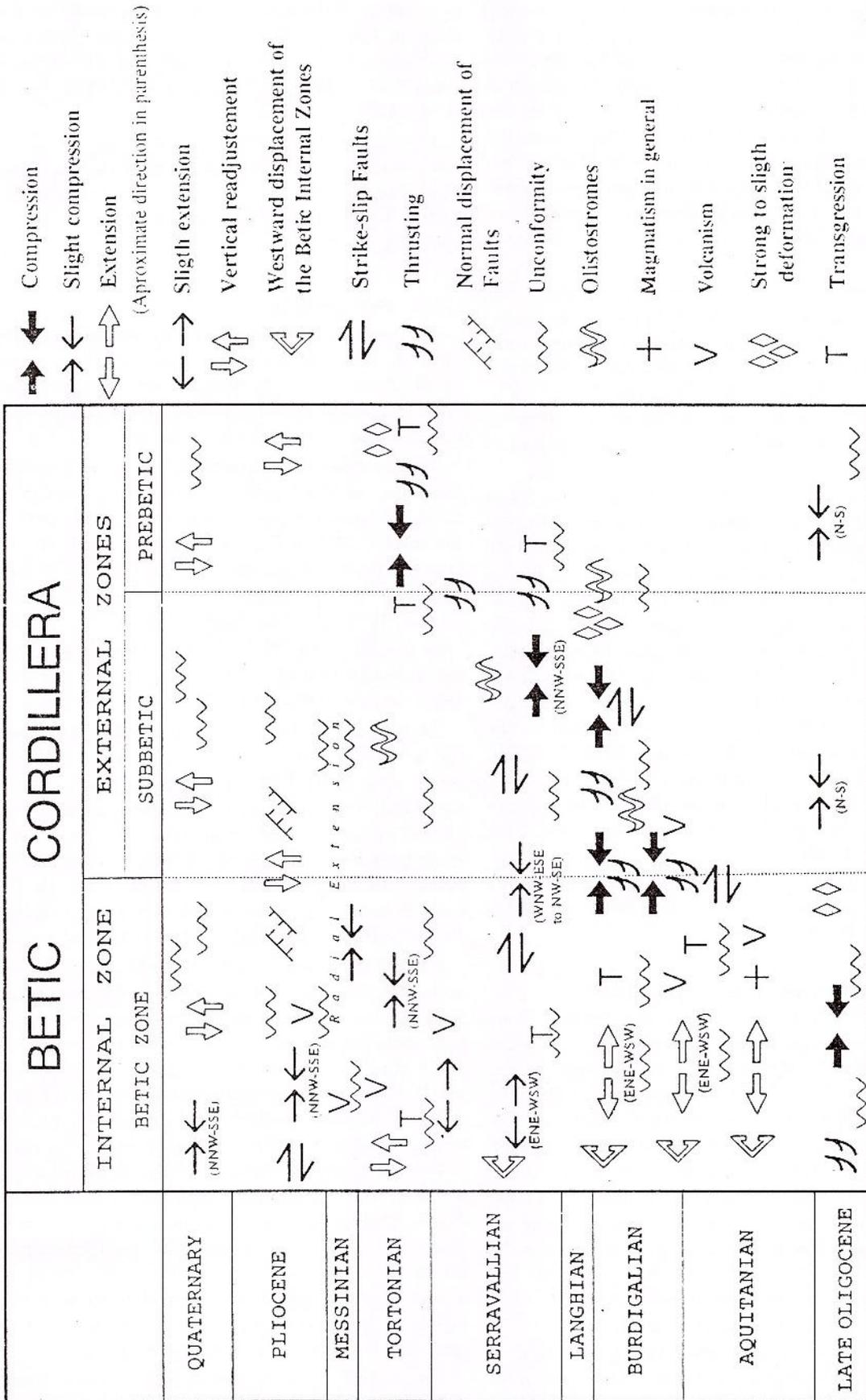


Fig. 2 - Correlation sketch of the main geological events occurred in the Betic Cordillera.

gone considerable northwestward displacements and especially the units of its northwestern front form the olistostromal masses deposited in the foredeep (the North Betic Strait formed at this time). Transcurrent movement also developed in a band of the External Zones and, according to De Smet (1984), produced a positive mega-flower structure in the Subbetic Zone. This N70E band of transcurrence has been named the Cadiz-Alicante dextral strike-slip fault zone (Sanz de Galdeano, 1983), moving actively to the Lower Langhian.

In the Prebetic a transgression took place and the North Betic Strait (or Guadalquivir proto-basin) was rapidly filled by the olistostromal complex, composed mainly of Triassic rocks; also diatomitic marls (Albarizas-type facies) were deposited.

Further to the southwest, in the western front of the chain, tectonic flakes developed in a westward direction.

Extensive movements affecting the Internal Zones (García Dueñas *et al.*, 1988, 1992; Galindo Zaldívar *et al.*, 1989; Galindo Zaldívar, 1983), from this time till the beginning of the Late Miocene, are responsible for the refting of the Alboran Basin as a western prolongation of the opening of the Algero-Provençal basin (Mac Gillavry *et al.*, 1963; Bourgois, 1978; Martín-Algarra, 1987; García Dueñas *et al.*, 1988; Galindo Zaldívar *et al.*, 1989; Sanz de Galdeano, 1990; Jabaloy *et al.*, 1992). This extension is linked to the magmatic processes that Torres Roldán *et al.* (1986) dated as 22-23 m.a. Immediately later appeared cinerites in the Viñuela Group and also in contemporaneous sediments located in the Subbetic Zone (Soria *et al.*, 1992).

These periods coincide with the second cooling peaks of the micas in the Nevado-Filabride complex as shown by De Jong (1991).

2.3. Late Burdigalian to Langhian

During the Late Burdigalian, in the External Zone began a transgression mainly recorded by the deposition of white marly facies (Albarizas-type facies, Jerez, 1979). These facies became part of the olistostromal formations in the North Betic Strait, because the Subbetic structuring had not yet concluded (Hoedemaeker, 1973), even thrusting on the Prebetic.

To the west, the Flysch units continued moving westwards, although some of these units, organized as olistostromal masses, subsequently rode over the Internal Zone (Bourgois, 1978; Rodríguez-Fernández, 1982; Olivier, 1984) and even over the Alboran Basin (Jurado, Comas, 1992; Comas *et al.*, 1992).

The tectonic processes during this period lead to the clear distinction of a curvature which prefigured the Gibraltar Arc. In fact, the Penibetic (which may

be considered the internal part of the Subbetic Zone) units in the area of the Gibraltar Arc show major clockwise rotations (between 30° and 80° during the Early-Middle Miocene) Osete *et al.* (1988); Villalain *et al.* (1995).

The end of this period seems to coincide with the last recorded cooling peak of the micas in the Nevado-Filabride complex. It also marks the beginning of a brittle deformation of this complex (De Jong, 1991; Monié *et al.*, 1991).

2.4. Serravallian

This event was characterized by a significant transgression, which caused the development of a bioclastic platform and marly basinal deposits upon almost all of the External Zone and important clastic wedges in some places of the Internal Zones.

The processes of thrusting (towards the NNW) continued in the Subbetic Zone. The North Betic Strait continued trapping large amounts of olistostromal masses. Their front migrated towards the foreland, with development of some piggyback basins (Roldán García, Rodríguez-Fernández, 1991; Roldán García, 1995) over the Subbetic Zone.

In the SE of the Betics a considerable amount of volcanic material (Bordet, 1985; Hernandez *et al.*, 1987; Serrano, 1990 and 1992) began to outflow.

At the end of the Serravallian, the thrusting of the Subbetic over the Prebetic ended. In the Internal Zones the E-W faults, particularly those in the Alpujarra Corridor, became active, with right-lateral movements (Sanz de Galdeano *et al.*, 1985; Rodríguez-Fernández *et al.*, 1990). Coetaneously, a substantial sea-level fall is recorded in all the Betic Cordillera (Rodríguez-Fernández, Sanz de Galdeano, 1990; Rodríguez-Fernández, Martín-Penela, 1993).

During the Early and Middle Miocene the mechanism causing the displacement of the Betic-Rifian Internal Zone is linked with the formation process of new oceanic crust in the western Mediterranean, especially in the Algero-Provençal Basin, accompanied by the thinning of the continental crust in the Betic-Alboran domain (Docherty, Banda, 1995). This process of extension produced, among other effects, the westward expulsion of the Betic-Rifian Internal Zone, which in turn deformed the Betic and Rifian External Zones. During this westward movement, the Internal Zone also underwent pronounced internal displacements among and even within its complexes, in such a way that the upper ones moved westwards in relation to the lower ones. These displacements are generally considered of extensive character.

The end of the westward displacements of the Internal Zone dramatically changed the geodynamic sit-



uation from the Late Miocene.

2.5. Tortonian

A major palaeogeographic change occurred at the beginning of the Tortonian, appearing several intermountain basins on the Internal Zone, such as the Almansora (Guerra Merchán, 1992), Tabernas (Pascual Molina et al., 1991), Sorbas and Vera (Ott d'Estevou, 1980; Montenat, Ott d'Estevou, 1990), and in the contact area between the Internal and External Zones, such as the Granada (Rodríguez-Fernández, 1982), Guadix (Rodríguez-Fernández et al., 1992) and Lorca basins (Montenat et al., 1987; Sanz de Galdeano, Vera, 1992). These basins were flooded by the Tortonian Sea, the best documented transgression in the Neogene record of Betic Cordillera. At this time appears the first detritus inherited from the Nevado-Filabride complex, evidence of the beginning of its outcropping (Volk, 1966; Rodríguez-Fernández, Sanz de Galdeano, 1992).

A not very important Intratoronian stage of deformations is recorded in several sectors of the Betics (Montenat, 1973; Jerez, 1973; Estévez et al., 1982; Rodríguez-Fernández, 1982), for which the direction of compression was approximately NNW-SSE.

The opening of the Alboran Basin ended (Jurado, Comas, 1992; Comas et al., 1992) and the Betic Cordillera began to form notable reliefs. The NE-SW Palomares-Carboneras set of faults, which traversed the Alboran Basin (Larouzière et al., 1988) and continued roughly into the Nekor fault system in Morocco, were considerably active and the volcanism along this fault zone contributed to the formation of the Alboran Ridge. This volcanism is also clearly visible in the SE of the Betics.

During the Late Tortonian the Gibraltar Arc finally attained its present structure. At the same time, the final structuring of the Prebetic took place with the formation of folds and tectonic slices, moving northwards and northwestwards. The Prebetic emerged and closed the North Betic Strait, transformed into the Guadalquivir Basin, which still progressed towards the foreland and where the olistostromal complex continued its advance (Roldán García, 1995).

Towards the end of the Tortonian a clear N-S to NNW-SSE compressive stage is detected (Weijermars, 1987; Cloething et al., 1992; Rodríguez-Fernández, Martín-Penela, 1993), as well as an E-W extension, contributing to the uplift of some of the most prominent sierras of the Internal Zone (Sierra Nevada, Filabres, etc.). During this time the eustatic fall, combined with a progressive uplift of the Betic and Rif Cordilleras, produced the isolation of the Mediterranean Sea. So, the communication between

the Atlantic and the Mediterranean through the Betic Cordillera was reduced to a few narrow straits, as in the Málaga Basin (López-Garrido, Sanz de Galdeano, 1991, and in press).

2.6. Messinian

The complete isolation of the Mediterranean provoked its desiccation. Then a bull's-eye type facies distribution developed in the Messinian peripheral basins of the Alboran Sea. In the Guadalquivir Basin the olistostromal complex ended its advance (Roldán García, 1995).

The displacement of the Palomares and Carboneras NE-SW sinistral strike-slip faults was tectonically important (Bousquet, Philip, 1976; Coppier et al., 1989), related with a N-S to NNW-SSE compression and an approximately perpendicular extension. In this area and in other northeastern areas, especially in the sector of Cartagena and Mazarrón (Murcia province), the volcanic outflow continued (Larouzière et al., 1988).

Many of the intermountain basins formed at the beginning of the Tortonian became continental (in several, this occurred by the end of the Tortonian, Rodríguez-Fernández et al., 1984). This process was facilitated by a radial extension affecting especially the central part of the Betic Cordillera, combined with a strong uplift (Sanz de Galdeano, López-Garrido, 1991).

2.7. Pliocene and Quaternary

A notable eustatic rise, and probably the movement of E-W faults, resulted in the invasion of water from the Atlantic through the present Strait of Gibraltar, causing widespread flooding of the Mediterranean.

Thus, the Pliocene began in the Betic Cordillera with a quick and sudden marine inundation of the peripheral basins of the Alboran Sea (Groupe de Recherche Néotectonique, 1977; Sanz de Galdeano, López-Garrido, 1991), while the continental basins of the interior developed lacustrine areas changing to fluvial systems which drained the surrounding areas either longitudinally or transversally, depending on the morphology.

This period, especially the Late Pliocene, is characterized by a highly active tectogenesis in the Betic basins, with the reactivation of old fault systems, which in many cases now moved as normal faults. The radial extension (coexisting somewhat with a N-S to NNW-SSE compression) continues, producing important vertical displacements mostly of the ancient fault sets, with subsidence processes in the basins and strong uplifting in the surrounding reliefs (Rodríguez-Fernández, Martín-Penela, 1993).



In the Late Pliocene a regression took place and the physiography of the Betic Cordillera was by now very similar to that of today.

3. Conclusions

The Neogene stratigraphic record of the Betic Cordillera can be divided in seven time intervals, separated by major discontinuities, with palaeogeographical changes reflecting the destruction of sedimentary basins and the formation of others.

In the first, which occurred during the Oligocene-Early Aquitanian, the structuring of the Betic Internal Zones ended, at the same time unconformable Neogene sediments appeared in the Malaguide Complex. Initial extensive movements led to the opening of the Algero-Provençal Basin. Coetaneously, during the Early Miocene the Numidian Flysch sediments coming from the African continent were deposited in the Masyllian Trough, which continued through the Magred to the southern Apennines.

During the second and third periods (Late Aquitanian-Langhian) the Internal Zones were affected by an important extension process, and the formation of the Alboran Basin began as a western prolongation of the Algerine Basin. In this same process, the westward displacement of the Internal Zones gave rise to the dextral transgressive contact with the External Zones. As a consequence, the Subbetic Zone underwent severe deformation, resulting in the first olistostromal masses accumulating in the new North Betic Strait. The Gibraltar Arc began to be formed.

An important transgression during the Langhian and Serravallian was marked by the predominance of marly pelitic deposits. The Betic Internal Zones continued the extension with crustal thinning and the Alboran area suffered a pronounced subsidence. In the Betic External Zones the same features persisted as in the previous interval, and piggy-back basins related to movements of notable olistostromal masses appeared in its northern front.

During the Tortonian in the Betic Internal Zones and in the Alboran Sea the ENE-WSW extension ended, and numerous intermountain basins appeared. A minor intra-Tortonian stage of NNW-SSE compressions occurred in the Betics.

Towards the end of the Late Tortonian, the Betics were affected by an approximately NNW-SSE compression, linked to the Africa-Iberia collision. Prominent reliefs with antiformal shape developed, and the Gibraltar Arc acquired its present structure. From this time a process of radial extension began to affect the Betic Cordillera, bringing about great vertical displacements that would continue more or less

without interruption till the Pleistocene. This situation, and especially the important eustatic fall, resulted in the isolation of the Mediterranean Sea.

During the Messinian, in the Betic Cordillera the most notable tectonic feature was the displacement of the Palomares and Carbonates sinistral strike-slip faults, in line with a N-S to NNW-SSE compression.

Finally, during the Early Pliocene, a eustatic rise, and possibly the movement of faults, lead to the invasion of water from the Atlantic, through the present Gibraltar Arc, causing a widespread transgression in the Mediterranean area.

The most compelling point of overall interest is that all these tectono-sedimentary episodes are directly related to the phenomena of extension which occurred in the western Mediterranean and especially with the N-S compression during the collision of the European and African plates.

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MEDITERRANEAN CALCAREOUS NANNOPLANKTON IN THE DACIC BASIN

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Key words: Calcareous nannoplankton. Sarmatian. Meotian. Pontian. Dacian. Romanian. Dacic Basin. Pannonian Basin. Euxinic Basin.

Abstract: The presence of the calcareous nannoplankton at different stratigraphic levels in the Middle Miocene-Pliocene deposits of the Dacic Basin proves a series of connections, as a rule of short standing, between the Dacic and Mediterranean basins. The nannofossils, that arrived concomitantly with the marine water levels in a medium with higher or lower salinity, either died or tried to adapt themselves to the new conditions through sizes or morphostructure changes.

On the Romanian territory, the Sarmatian-Romanian deposits of the Dacic Basin cover the following structural units: the Scythian and Moesian Platforms, the southern part of the Carpathian Fore-deep and, partly, the Subcarpathian Nappe (Fig. 1).

If in the Volhynian-Early Bessarabian time span this basin constituted a connection zone between Central and Oriental Paratethys, beginning with the Late Bessarabian it becomes a constitutive part of the Oriental Paratethys.

The paleogeographical connections of the Dacic Basin with the neighbouring basins (Pannonian and Euxinic Basins) as well as with the Mediterranean are well recorded in the nannoplankton assemblages.

Volhynian Early Bessarabian

In this time-interval, the Dacic Basin deposits contain two nannoplankton assemblage types, each of them with specific features.

The first type, that characterizes the Central Paratethys (Baldi-Beke, 1960, 1964; Lehotayova, 1974; Stradner, Fuchs, 1979; Müller, 1974; Nagymarosy, Müller, 1988; Mărunțeanu, 1992; Mărunțeanu, Papaianopol, 1994; etc.), contains abundant *Braarudosphaera bigelowii*, *Calcidiscus leptoporus*, *Calcidiscus macintyreii* and *Reticulofenestra pseudoumbilicus*. Generally, the nannofossils are smaller in size in comparison with the holotype-species, and the *Calcidiscus* species show numerous morphotypes with various structures in the central areas. The assemblages also contain some endemic species, such as *Reticulofenestra gelida*, *Rhabdosphaera pannonica* and *Rhabdolithus poculi*.

In the Dacic Basin, this type was identified in the Mohrensternia beds (Fig. 2, I); the clay levels underlying the Ervilia and Inaequicostata beds (Fig. 2, VIII); the Plicatiforma beds (Fig. 2, VII, VIII); the basis of the Late Volhynian successions of the Valea Morilor Formation (Fig. 2, II); the Criptomactra pesansensis beds, situated over the Mactra and Inaequicostata beds (Fig. 2, VIII); the middle part of the Valea Morilor Formation (Fig. 2, II); the uppermost part of the Valea Morilor Formation (Fig. 2, II) and of the Dara Member (Fig. 2, VII).

The second type of the nannoplankton assemblages, very well developed in the Mediterranean area (Theodoridis, 1984; Müller, 1978) in the same time-interval, was identified at different stratigraphic levels of the Volhynian-Early Bessarabian deposits from the Dacic Basin. The nannofossils, with normal sizes, are rich in species but poor in specimens and belong to the *Discoaster kugleri* (NN7), *Catinaster coalitus* (NN8) and *Discoaster hamatus* (NN9) Zones, of the Standard Nannoplankton Zonation (Martini, 1971).

In the Dacic Basin, the assemblages with *Discoaster kugleri* (NN 7 Zone) were recognized only in the Abra and Mohrensternia (Fig. 2, VIII) or Inaequicostata and Ervilia (Fig. 2, VIII) beds. Nannoplankton with *Catinaster coalitus* and *Catinaster calyculus* (NN 8 Zone) exists in: Plicatiforma beds (Fig. 2, III); the lowermost and the middle part of the Dara Member (Fig. 2, VII); the Mactra and Obsoletiforma beds (Fig. 2, IV); the Mactra and Paphia beds (Fig. 2, V); the basis of the Sușița Formation (Fig. 2, X). The nannofossil assemblages, with *Discoaster hamatus*, *Discoaster calcaris* and *Discoaster challengerii* (NN 9 Zone) were identified



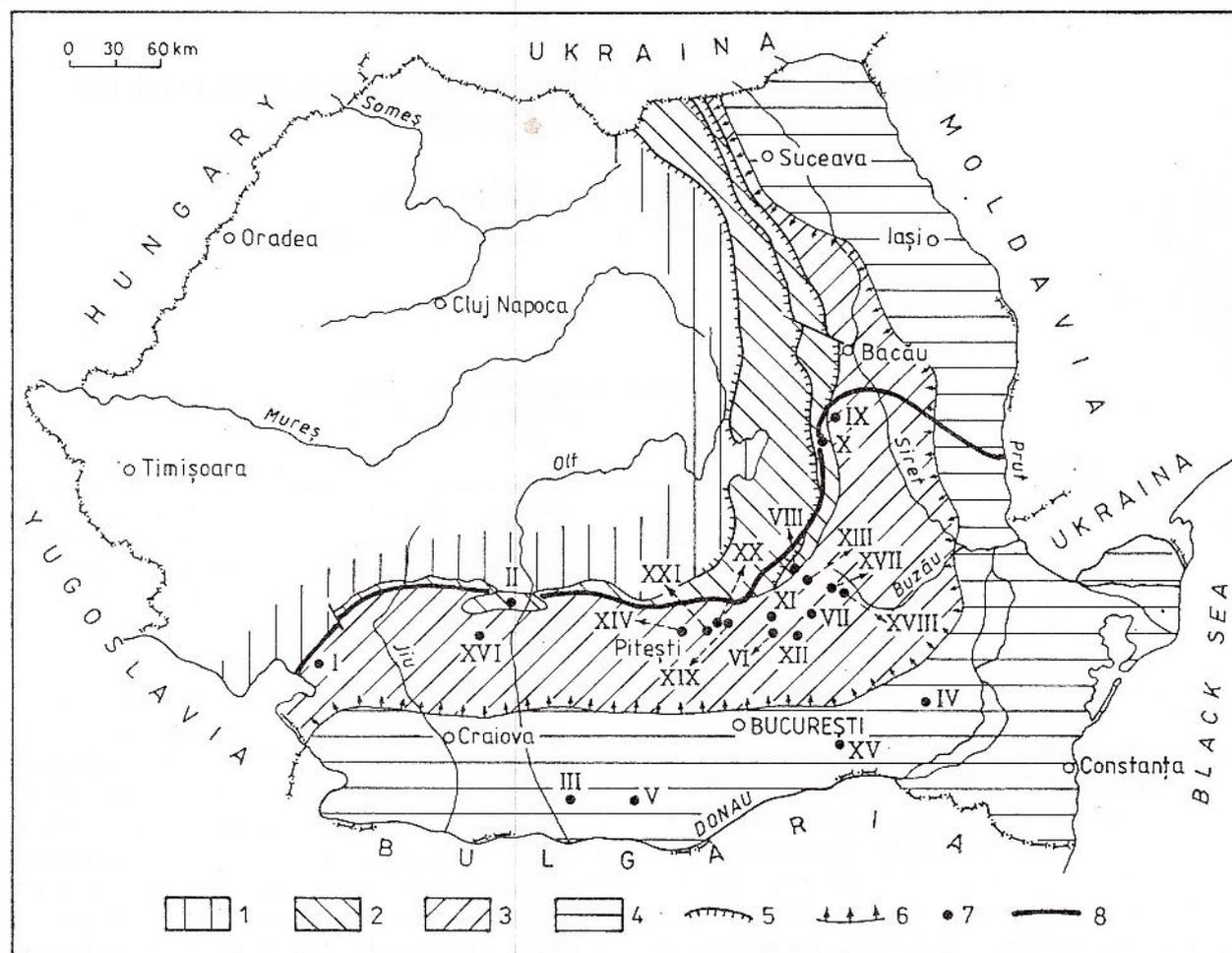


Fig. 1 - Location of the geological sections in the Dacic Basin. I, Morilor Valley; II, Căzănești; III, Stoicănești; IV, Sudiți; V, Roșiori; VI, Călugăreni; VII, Dara; VIII, Bozioru; IX, Trotuș Valley; X, Haloșu Mare Valley; XI, Tătaru; XII, Tohani; XIII, Pânătau; XIV, Doicești; XV, Argova Valley; XVI, Cucești; XVII, Berca; XVIII, Berca; XIX, Cosmina; XX, Cosmina; XXI, Edera.

1, dacides; 2, moldavides; 3, foredeep; 4, platforms; 5, overthrust nappes; 6, flexure; 7, location of geological sections; 8, faults

in: the upper part of the Dara Member (Fig. 2, VII) and of the Valea Morilor Formation (Fig. 2, II), as well as in the Mactra and Inaequicostata beds (Fig. 2, VIII).

The alternation of these two nannoplankton assemblage types in the Volhynian-Early Bessarabian deposits from the Dacic Basin proves its temporary connections with the Pannonian and Mediterranean Basins.

2. Late Bessarabian-Meotian

The Late Bessarabian, Chersonian and Early Meotian deposits of the Dacic Basin contain rare nannoplankton assemblages, developed discontinuously in time and space and dominated by the *Thoracosphaera* and *Scyphosphaera* species. The variegated beds (Fig. 2, VI a), Răpile Formation

(Fig. 2, IX), *Congeria* and *Teodoxus* beds of the Pârăul Sărat Member (Fig. 3, VI b), *Congeria* and *Helix* beds of the Valea lui Cernat Member (Fig. 3, XII) contain calcareous nannoplankton with *Scyphosphaera amphora*, *S. proceră*, *S. conica*, *S. campanula*, *Thoracosphaera deflandrei*, *T. fossata*, *T. saxes* and rare occurrences of *Calcidiscus leptoporus*, *Coccolithus pelagicus* and *Reticulofenestra pseudoumbilicus*.

A rigorous calcareous nannoplankton zonation after these assemblages with mostly facies indicator species is impossible to achieve. We suppose that the nanofossils belong to the *Discoaster calcaris* (NN 10) Zone, because *Scyphosphaera campanula* seems to occur in the NN 10 Zone.

On the basis of this assemblage, it is very difficult to say whether the Dacic Basin was or was not in connection with the Mediterranean. However, the

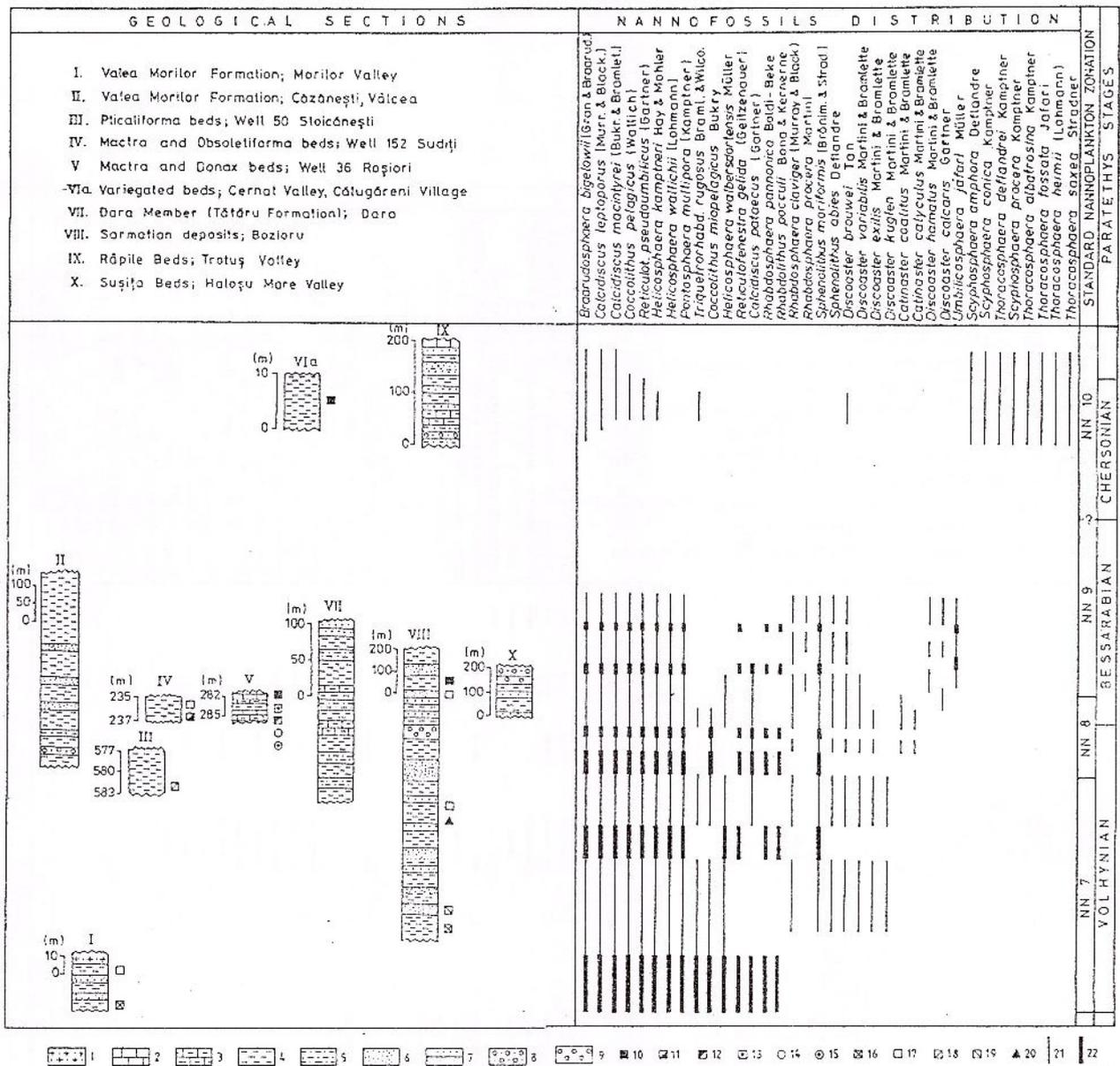


Fig. 2 - Distribution of the calcareous nannofossils in the Vollynian, Bessarabian and Chersonian deposits. 1, tuffites; 2, limestones; 3, marls; 4, clays; 5, silty clays; 6, sands; 7, sandstones; 8, pebbles; 9, conglomerates; 10, *Mactra*; 11, *Obsoleliforma*; 12, *Donax*; 13, *Paphia*; 14, *Calliostoma*; 15, *Duplicata*; 16, *Mohresternia*; 17, *Inaequicostata*; 18, *Plicatiforma*; 19, *Abra*; 20, *Ervilia*; 21, Mediterranean influences; 22, Pannonian influences.

endemic Pannonian nannoplankton with *Noelaerhabdus* and *Bekeithella* of the Central Paratethys (Jerkovic, 1970; Bona, Gal, 1987; etc.) shows a total interruption of the connections between the Dacic and Pannonian Basins in the Late Bessarabian, Chersonian and Early Meotian time.

In the Mytilaster, *Ervilia* and *Abra* levels of the Late Oltenian (Early Meotian) *Dosinia* beds (Fig. 3, XI a, VI c, VI d), the nannofossils content with *Discoaster quinqueramus*, *Amaurolithus primus* and *A. delicatus* characterizes the lower part of the *Discoaster quinqueramus* (NN 11) Zone.

Calcareous nannoplankton that belongs to the NN

11 Zone was also discovered in the Moldavian (Late Meotian) deposits, e.g.: *Psilunio* beds (Fig. 3, XIII - were *Discoaster calcaris* appears as well); Slatina Formation, intercepted from the 50 Stoicânești well, between 613-618 m (Fig. 3, III b - where the assemblage also contains *Discoaster berggrenii* and *Discoaster loeblichii*); Dreissenomia beds (Fig. 3, VI e - where the occurrences of the *Amaurolithus bizzarus*, *A. amplificus* and *Anguloithina arca* were also identified).

The presence of the Tethysian nannoflora only in a few stratigraphic levels of the Late Oltenian-Moldavian deposits proves the episodic and short

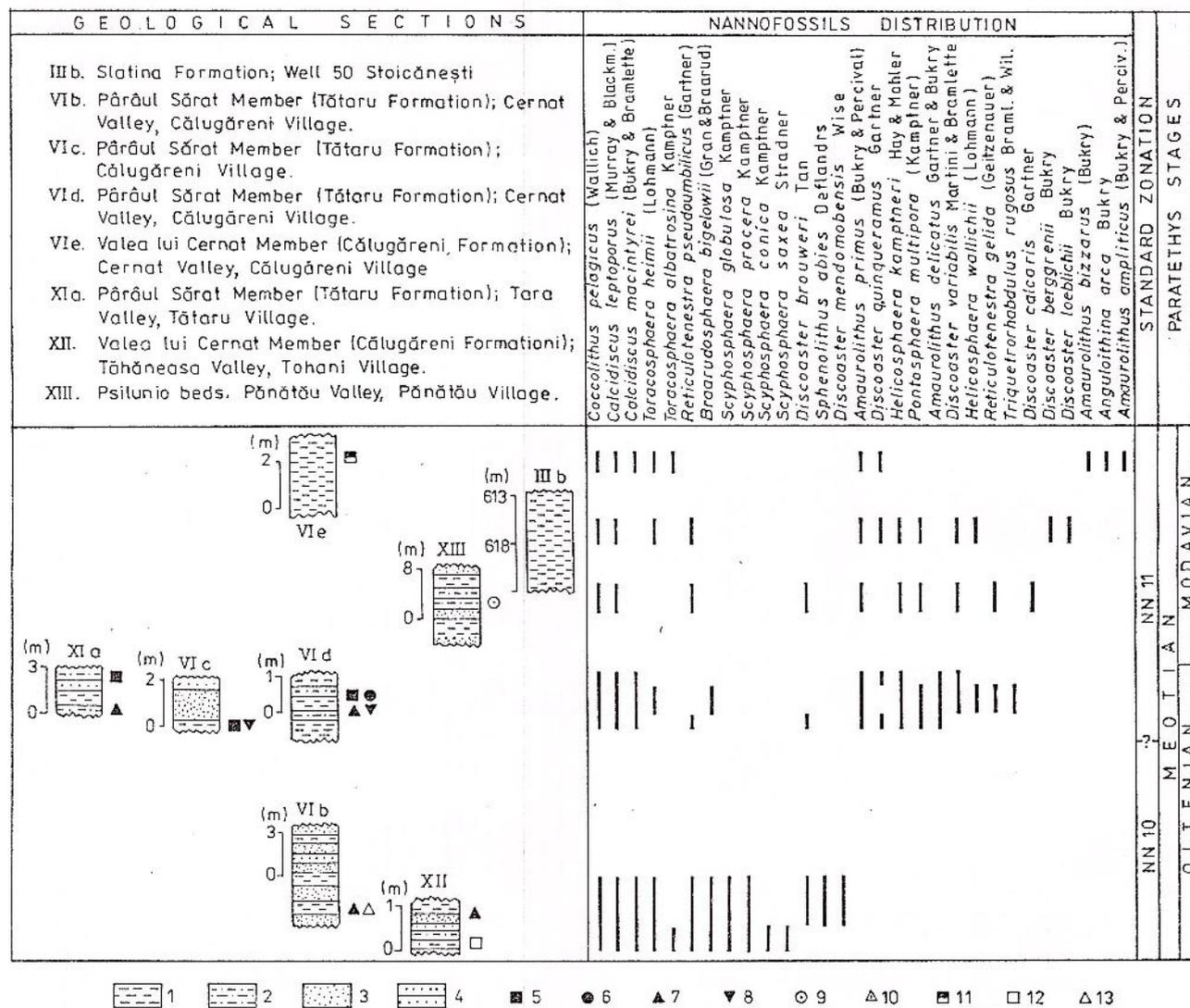


Fig. 3 - Stratigraphical distribution of the calcareous nannofossils in the Meotian deposits. 1, clays; 2, silty clays; 3, sands; 4, sandstones; 5, *Dosinia*; 6, *Abra*; 7, *Ervilia*; 8, *Mytilaster*; 9, *Psilunio*; 10, *Congerina*; 11, *Dreissenomya*; 12, *Helic*; 13, *Tcodorus*.

connections between the Dacic and Mediterranean Basins, probably through the Egean Sea and the Romanian shelf of the Black Sea.

3. Pontian

Starting with the Pontian to the end of the Neogene, the calcareous nannoplankton assemblage was mentioned only in the Dacic and the Aralo-Ponto-

Caspian Basins. Thus, in the Pontian formations of the Dacic Basin, the nannofossils belonging also to the NN 11 Zone were discovered in: the Early Odessian Paradacna beds (Fig. 4, XI b); the Early Portafferian Caladacna and Paradacna beds (Fig. 4, VI f); the Early Bosphorian Chartoconcha and Paradacna beds (Fig. 4, VI g); the Late Bosphorian Chartoconcha and Lunadacna (Fig. 4, XIV) or Lunadacna and Caladacna (Fig. 4, XV) beds. In the



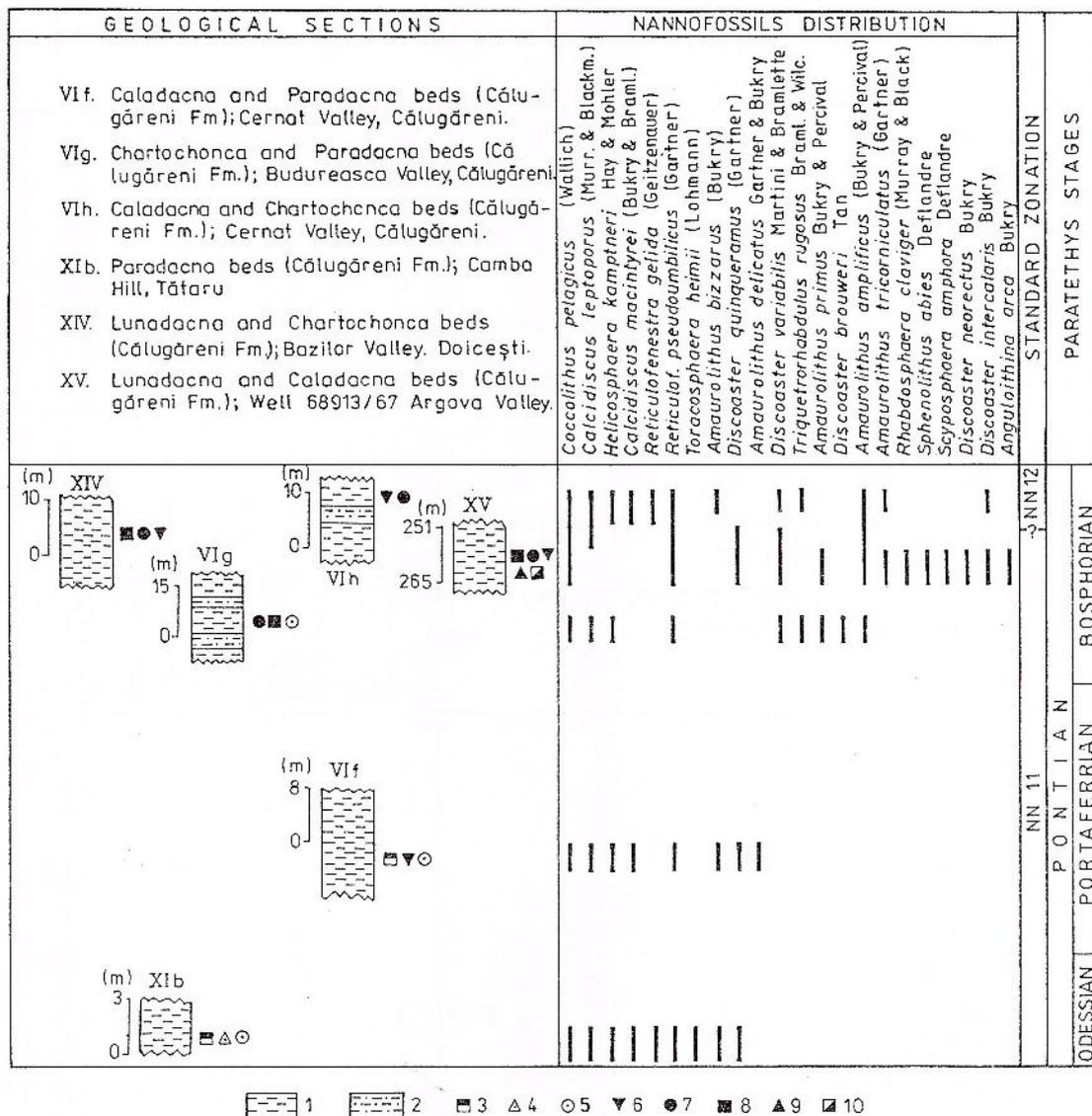


Fig. 4 - Stratigraphical distribution of the calcareous nannofossils in the Pontian deposits. 1, clays; 2, silty clays; 3, Paradacna; 4, Congeria; 5, Valenciennius; 6, Caladacna; 7, Chartoconca; 8, Lunadacna; 9, Dreissena; 10, Pontalmira.

Caladacna and Chartoconca beds (Fig. 4, VI h), Late Bosphorian in age, the nannoplankton assemblage, that contains *Amaurolithus tricorniculatus* but less *Discoaster quinqueramus*, may belong to the *Amaurolithus tricorniculatus* (NN 12) Zone.

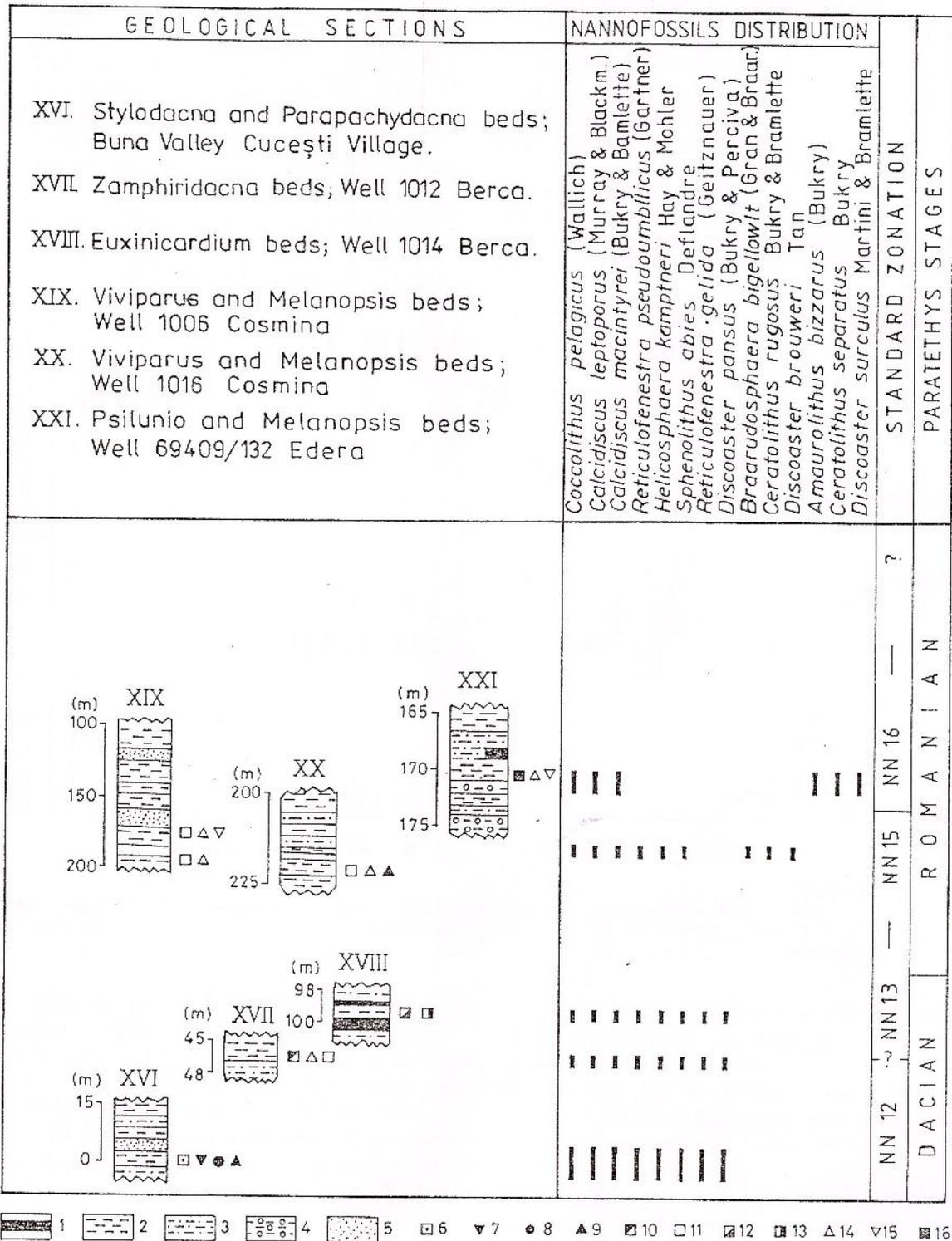
Lately several researches have erroneously considered that the first occurrence of the *Amaurolithus tricorniculatus* corresponds to the boundary between NN 11 and NN 12 Zones. Our studies demonstrated that in the Late Pontian (Lunadacna and Caladacna beds), *Amaurolithus tricorniculatus* occurs in the same assemblage with *Discoaster quinqueramus*. We concluded that the first occurrence of *Amaurolithus tricorniculatus* can be recorded in the uppermost

part of the NN 11 Zone. The same assemblage was mentioned, approximately at the same stratigraphic levels, in the Pontian deposits of the Ponto-Caspian Basin (Semenenko, Liulieva, 1978).

We may conclude that in Pontian time, the short connections between Tethys and Oriental Paratethys existed at least in the basal Odessian, Early Portafferrian, Early and Late Bosphorian.

4. Dacian

The Dacian deposits of the Dacic Basin are very poor in the calcareous nannoplankton; it has been identified only in a few stratigraphic levels in: the



1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16

Fig. 5 - Stratigraphical distribution of the calcareous nannofossils in the Dacian and Romanian deposits. 1, coal; 2, clays; 3, silty clays; 4, clay with concretions; 5, sands; 6, Stylodacna; 7, Prosodacna; 8, Pseudocatillus; 9, Dreissena; 10, Zamphiridacna; 11, Viviparus; 12, Eurinocardium; 13, Pseudocatillus; 14, Melanopsis; 15, Teodorus; 16, Psilunio.

Stylodacna and Parapachydacna beds (Fig. 5, XVI); the Zamphiridacna beds (Fig. 5, XVII); the Euxinocardium beds (Fig. 5, XVIII). The assemblage with *Coccolithus pelagicus*, *Calcidiscus leptoporus*, *C. macintyreii*, *Reticulofenestra pseudoumbilicus*, *Helicosphaera kamptneri* and very rare *Discoaster pansus* can indicate the Amaurolithus tricorniculatus (NN 12) Zone, as suggested by the first occurrences of *Discoaster pansus*.

The boundary between NN 12-NN 13 Zones is situated probably in the uppermost part of the Dacian stage, because it was identified in the Late Kimmerian deposits of the Black Sea shelf (Mărunțeanu, unpublished data).

According to the specific features of the Dacian nannoplankton assemblages, we may consider that, in the Dacian time, the Dacic Basin was not connected with the Mediterranean. However, the intermittent water changes between the rest of the Oriental Paratethys and Tethys existed in this time interval because the Kimmerian deposits (Semenenko, Liulieva, 1978) contain occasionally Mediterranean assemblages.

5. Romanian

As a rule, the fresh water deposits cannot contain marine calcareous nannoplankton. However, the Middle Romanian deposits from the Carpathian Fore-deep contain the nanofossils belonging to the *Reticulofenestra pseudoumbilicus* (NN 15) and *Discoaster surculus* (NN 16) Zones. Thus, in the Viviparus and Melanopsis beds (Fig. 5, XIX, XX), intercepted from 1006 Cosmina (between 150-200 m) and 1018 Cosmina (between 200-225 m) wells, an assemblage with *Reticulofenestra pseudoumbilicus* and *Ceratolithus rugosus*, belonging to the NN 15 Zone, was discovered. Above these deposits, the Psilunio and Melanopsis beds (Fig. 5, XIX, XXI), intercepted from 69409/132 Edera (between 170-175 m) and 1006 Cosmina (between 127-147 m) wells contain a nanofossils assemblage with *Ceratolithus sepparatus* and *Discoaster quinqueramus*, belonging to the NN 16 Zone.

The presence of the marine nannoplankton in the Romanian fresh water deposits can be explained only by a short connection between the Dacic Basin and the Tethys in this time-interval.

The discontinuous distribution in time and space as well as the specific features of the calcareous nanofossils assemblages of the Sarmatian-Romanian deposits in the Dacic Basin prove the intermittent connections between this basin and the Mediterranean. Many species, arrived through the short Tethysian ingressions into brackish-fresh waters of the Dacic

Basin, died immediately, marking exactly the moment of the connections. A lot of nanofossils, as *Coccolithus pelagicus*, *Reticulofenestra pseudoumbilicus*, *Helicosphaera kamptneri* and *Calcidiscus leptoporus*, resisting to salinity exchanges, survived a longer time, through their adaptation by the new aquatic biota conditions. Only the presence of these nanofossils in sediments cannot indicate the direct connections between their depositional areas and the world ocean.

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BIOSTRATIGRAPHIE ET INTERPRETATION PALÉOGÉOGRAPHIQUE DE LA SUCCESSION OLIGOCÈNE MOYEN-MIOCÈNE MOYEN, AU NORD DU BASSIN MÉSOHELLÉNIQUE (DÉPRESSION DE KORÇA, ALBANIE DU SE), BASSÉES SUR LES FORAMINIFÈRES ET NANNOFOSSILES CALCAIRES

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Key words: Biostratigraphy. Paleogeography. Foraminifera (planktonic, small and large benthic). Calcareous nanofossils. Mesohellenic Basin. Korça Depression. SE Albania.

Abstract: *Biostratigraphy and Paleogeographic Interpretation of the Mesohellenic Basin (Korça Depression, SE Albania) on the Basis of Foraminifera and Calcareous Nanofossils.* This study is concerned with the Oligo-Miocene mollasic succession of Korça Depression (SE Albania). Biostratigraphically, these deposits are divided into several zones based on planktonic, larger foraminifera and calcareous nanofossils. The base of the series, transgressively overlying the ophiolites and Cretaceous-Eocene limestones, belongs to Middle Oligocene (respectively *Globorotalia opima* zone and *Sphenolithus distentus* - NP 24 zone of planktonic foraminifera and calcareous nanofossils). At the top of the succession, Langhian in age, the *Globigerinoides bisphaericus* zone and *Sphenolithus heteromorphus* - NN 5 zone have been recognised. Some transitional forms between *Globigerinoides bisphaericus* and *Pracorbulina*, similar to *P. sicana* are to be noted. Four distinguished assemblages of larger foraminifera are characteristic of: the Middle Oligocene, Late Oligocene, Aquitanian and lower part of Burdigalian. The acquired data of smaller benthic, larger and planktonic foraminifera as well as calcareous nanofossils allowed us to give the paleogeographical and paleoecological interpretations.

I. Introduction

Les dépôts du bassin Mésohellénique présents au Sud-Est d'Albanie, connus sous le nom de la Dépression de Korça, ont fait l'objet de diverses études géologiques. Ces travaux ont été ponctuels, traitant en particulier les niveaux charbonneux ou des horizons tel que le Calcaire Corralien (Petro & Dodona, 1976). La biostratigraphie était basée surtout sur la macrofaune.

Les campagnes menées entre les années 1990-1992 par nous et nos collègues de l'Institut Géologique du Fer, dans la Dépression de Korça avaient pour but essentiel d'établir une biostratigraphie la plus fine possible et de donner les conditions paléogéographique et les paléomilieus.

Nous avons effectué un échantillonnage serré dans les sédiments favorables pour l'étude des foraminifères (grands et petits benthiques et planctoniques), nanofossiles calcaires et palynologie. Tous les représentants de la macrofaune (coraux,

échinides et mollusques) ont été recoltés par le spécialiste en question. Les résultats obtenus pour les foraminifères et nanofossiles calcaires sont présentés ici, alors que la palynologie et la macrofaune aient leur propre objet.

II. Cadre géologique

La Dépression de Korça représente l'extension vers le Nord du bassin Mésohelléniques au territoire albanais (Fig. 1). C'est une arrière-fosse orientée NNO-SSE respectant l'alignement général des Hellénides-Albanides. A l'Est, elle est limitée par la zone de Korabi (=zone Pélagonienne) et, à l'Ouest, par les zones de Mirdita (=zone Subpélagonienne) et Krasta-Cukali (=zone de Pinde).

La succession oligo-miocène est transgressive sur le socle composé de roches ultrabasiqes, calcaires du Crétacé-Eocène et le flysch de Krasta-Cukali. Dans cette série, Kumati et al. (1995) ont distingué un nombre de formations (Fig. 7) dont la valeur est locale.



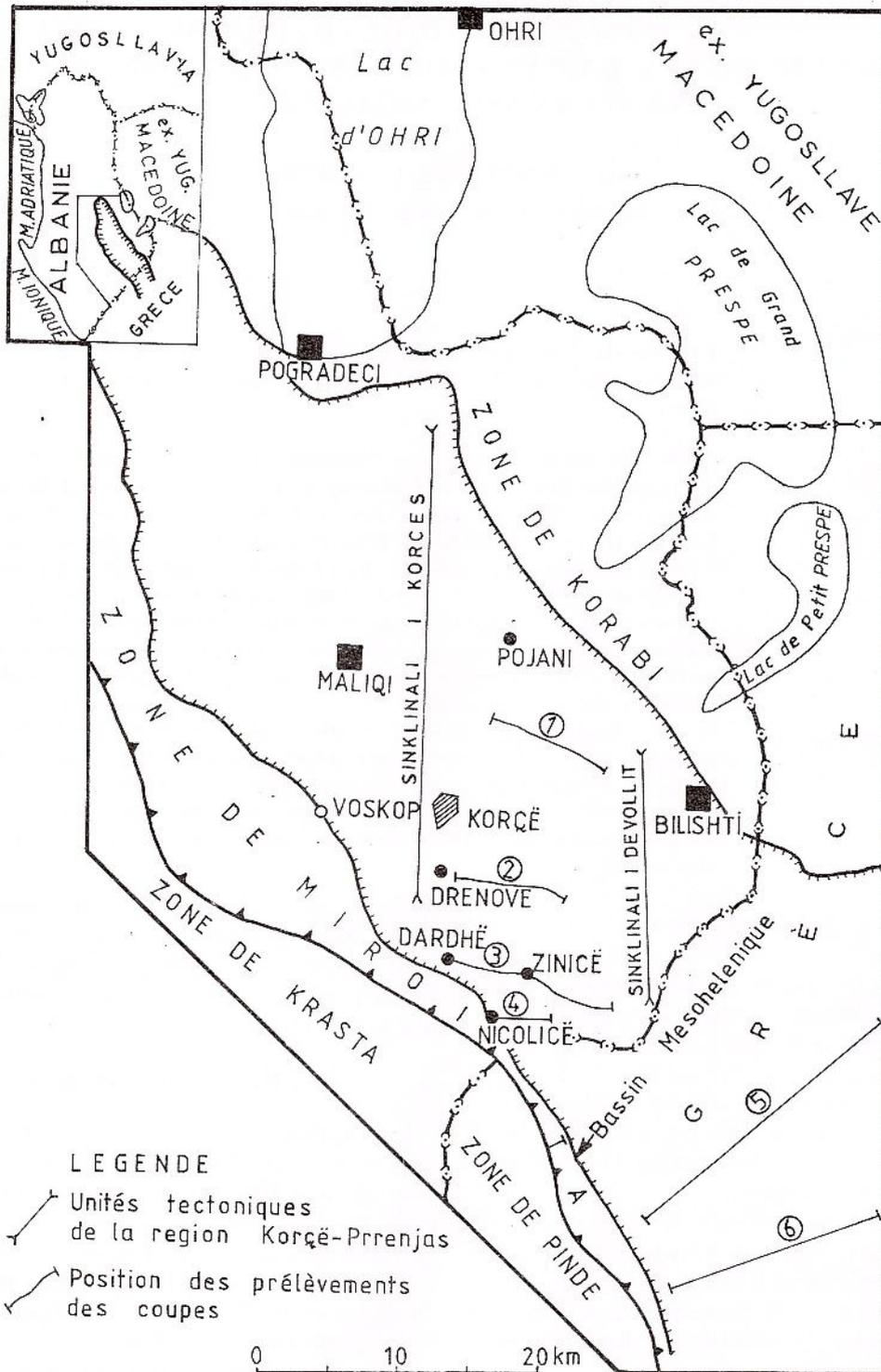


Fig. 1 - Situation géo-géographique de la dépression de Korça et des coupes étudiées. Dans la région de Moravë-Devoll: 1, coupe de Plasë; 2, coupe de Drenovë; 3, coupe de Dardhë-Sinicë; 4, coupe de Nicolicë. Dans le territoire grec (Soliman et Zygoyanis, 1979): 5, coupe d'Ondria; 6, coupe d'Eptahorion.

| ZONE IONIENNE | | | | DÉPRESSION DE KORÇA | | | | | | | | |
|------------------------------------|--|---|---|--|---|--|---|--|---|--|---|------------------------------|
| NANNO- PLANCTON | PALYNO- LOGIE | FORAMINIFÈRES | | AGE | FORAMINIFÈRES | | PALYNOLOGIE | NANNO- PLANCTON | | | | |
| | | BENTOS (Grands) | PLANCTON | | PLANCTON | BENTOS (Grands) | | | | | | |
| <i>Sph. hetero.</i> | Kompleksi ⑧ | | <i>Pracorbulina</i> | LAN- GHIAN | | | <i>Hystrichosphaeraceae</i> + <i>Chytroeisphaerida</i> + <i>Tasmanaceae</i> | <i>Sph. hetero.</i> | | | | |
| <i>Helicosphaera ampliapertura</i> | | | <i>Globigerinoides bisphaericus</i> | | <i>Globigerinoides bisphaericus</i> | | | | <i>Helicosphaera ampliapertura</i> | | | |
| <i>Helicosphaera scissura</i> | | Kompleksi ⑦ | | | <i>Globorotalia scitula</i> + <i>Globorotalia acrostoma</i> | <i>Globorotalia scitula</i> + <i>Globorotalia acrostoma</i> | | | <i>Gonyaulacysta</i> + <i>Pxydiella</i> <i>Dapsidinium</i> + <i>Chiropteridium</i> | <i>Helicosphaera scissura</i> | | |
| <i>Helicosphaera carteri</i> | | | | | <i>Miogyssina globulina</i> + <i>Mioplepidocyclina burdigalensis</i> | <i>Globigerinoides trilobus</i> + <i>Catapsydrax gr. dissimilis</i> | | <i>Globigerinoides trilobus</i> + <i>Catapsydrax gr. dissimilis</i> | | | <i>Hystrichosphaeraceae</i> + <i>Tasmanaceae</i> | <i>Helicosphaera carteri</i> |
| | | Kompl. ⑥ | | | | | | | | | | |
| | | Kompleksi ⑤ | <i>Miogyssins tani</i> | | <i>Globoquadrina dehiscens</i> | <i>Globoquadrina dehiscens</i> | | <i>Miogyssina tani</i> | <i>Miogyssina gunteri</i> | <i>Deflandrea aranitensis</i> | | |
| | | | <i>Miogyssins gunteri</i> | | | | | | | <i>Hystrichosphaeridiaceae</i> | | |
| <i>Sphenolithus delphix</i> | | Kompleksi ④ | <i>Lepidocyclina s.l.</i> + <i>Miogyssinoides s.l.</i> | | <i>Globorotalia kugleri</i> | <i>Globorotalia kugleri</i> | | <i>Globorotalia kugleri</i> | <i>Miogyssinoides complanata</i> + <i>Lepidocyclina (N) morgani</i> | <i>Deflandrea phosphoritica</i> + <i>Deflandrea albania</i> + <i>Hystrichosphaeridiaceae</i> | <i>Sphenolithus ciproensis</i> | |
| <i>Sphenolithus ciproensis</i> | <i>Globigerina Ciproensis Ciproensis</i> | | | <i>Globigerina Ciproensis Ciproensis</i> | <i>Globigerina Ciproensis Ciproensis</i> | | | | | | | |
| <i>Sphenolithus distentus</i> | Kompleksi ③ | <i>Numulite s.l.</i> + <i>Lepidocyclina s.l.</i> | <i>Globorotalia opima opima</i> | OLIGOCÈNE MOYÉN Pg3 | <i>Globorotalia opima opima</i> + <i>Globig. ciper. angulifurcalis</i> | <i>Nummulite s.l.</i> + <i>Lepidocyclina s.l.</i> | <i>Deflandrea oligocaenicus</i> + <i>Hystrichosphaeridiaceae</i> | <i>Sphenolithus distentus</i> | | | | |
| <i>Sphenolithus praedistentus</i> | | | <i>Globigerina ciproensis angulifurcalis</i> | | <i>Globigerina ciproensis angulifurcalis</i> | | | | | | | |

Fig. 2 - Schéma de corrélation des zones de foraminifères planctoniques, grands foraminifères, nannoplancton et palynologie entre la zone ionienne et la dépression de Korça.

Les dépôts de l'Oligocène moyen ont un caractère grossier, étant formés de conglomérats de base, grès, calcaires biodétritiques et peu de silts. Leur origine peu profonde (delta-plateau continental) est aussi prouvée par la présence de la macrofaune et la microfauve néritique.

L'Oligocène supérieur suit normalement par des dépôts de turbidites où prédominent les grès massifs et les conglomérats.

Successivement, la même sédimentation se poursuit à l'Aquitaniien et une partie du Burdigalien. Il faut noter ici la présence des calcaires biodétritiques riches en grands foraminifères déplacés.

Le reste du Burdigalien et le Langhien sont marqués par des marnes bleutées riches en plancton.

Les dépôts lacustres plio-pleistocènes sont transgressifs et ne contiennent que de rares ostracodes et charophytes.

III. Biostratigraphie

III. 1. Les foraminifères planctoniques et petits benthiques

La biozonation adoptée pour les foraminifères planctoniques est celle définie dans la zone ionienne par Kici et al. (1972) et Dalipi et al. (1975).

Dans la coupe de Plasë et d'autres points de la Dépression de Korça, la partie inférieure de la succession a livré de rares individus de *Globorotalia opima opima* et *Globigerina ciperoensis angulisuturalis*. Ces espèces permettent de dater l'Oligocène moyen (zone à *Globorotalia opima opima*) indiquant ainsi l'âge de transgression. L'association des foraminifères planctoniques est très pauvre par suite du milieu littoral.

Plus haut, dans les coupes de Plasë et Nikolicë (Fig. 3) est identifiée la zone à *Globigerina ciperoensis ciperoensis*, riche en foraminifères planctoniques et petits benthiques. L'association de planctoniques est composée de: *Globigerina ciperoensis angulisuturalis*, *G. ciperoensis ciperoensis*, *G. ciperoensis angustiumbilocata*, *G. tripartita*, *G. trilocularis*, *G. venezuelana*, *Globorotalia opima nana*, *Globorotaloides suteri*, *Catapsydrax* gr. *dissimilis*, *C. unicus* etc. Les petits benthiques sont représentés par *Rhabdamina subcylindrica*, *R. robusta*, *Dorothia brevis*, *Cyclamina latidorsata*, *Spiroplectamina carinata*, *Robulus rotulatus*, *Almaena escornebovensis*, *A. hieroglyphica*, *Anomalina acuta*, *A. pseudogrosserugosa*, *Bulimina ovata*, *Uvigerina mexicana*, *Planulina renzi*, *Cassidulina subglobosa*.

Cet assemblage est similaire à celui de la zone à *Globigerina ciperoensis ciperoensis* dans la zone ionienne, mais ici les foraminifères benthiques sont plus nombreux et plus développés. La disparition de *Globigerina ciperoensis angulisuturalis*, dans les deux

bassins, a lieu avant l'apparition de *Globorotalia kugleri*.

La zone à *Globorotalia kugleri* est reconnue dans la coupe de Nikolicë. Les dimensions de l'espèce *Globorotalia kugleri* sont plus réduites que dans la zone ionienne. La présence de *G. kugleri* et l'absence de *G. ciperoensis angulisuturalis* sont presque les seules différences de la zone à *Globigerina ciperoensis ciperoensis*.

Il faut noter qu'autour de l'Oligocène supérieur-Miocène inférieur existe un intervalle épais de grès et conglomérats qui rend difficile la mise en évidence de la limite Oligocène-Miocène.

Quelques rares échantillons (coupe de Nikolicë et prélèvements ponctuels) ont fourni une association riche en foraminifères planctoniques, bien préservée, datant la zone à *Globoquadrina dehiscens* de l'Aquitaniien. Cette association est comparable à celle de la zone ionienne où l'apparition de *Globoquadrina dehiscens* a été employée pour tracer la limite Oligocène-Miocène.

La zone à *Catapsydrax* gr. *dissimilis*-*Globigerinoides trilobus* a été identifiée dans la coupe de Sinicë (éch. 13-23) (Fig. 3) datant la partie inférieure du Burdigalien. L'apparition de *Globigerinoides trilobus* a servi à définir la base de la zone et, en même temps, la limite Aquitaniien-Burdigalien. L'assemblage de foraminifères est riche et bien développé. Les planctoniques sont représentés par *Globigerina* gr. *ciperoensis*, *G. venezuelana*, *Catapsydrax* gr. *dissimilis*, *Globigerinoides trilobus*, *G. altiapertura*, *Globorotalia* aff. *mayeri*, *G. opima nana* etc. et les petits benthiques par *Rhabdamina robusta*, *Spiroplectamina carinata*, *Robulus rotulatus*, *Cibicides pseudoungerianus*, *Anomalina grosserugosa* etc.

Plus haut, dans la coupe de Sinicë (éch. 24-53) est reconnue la zone à *Globorotalia scitula*-*Globorotalia acrostoma* du Burdigalien. L'association des foraminifères planctoniques est très riche, composée de *G. acrostoma*, *G. opima nana*, *Globigerinoides trilobus*, *G. quadrilobatus*, *G. sacculifer*, *Globoquadrina dehiscens*, *Globigerina* gr. *ciperoensis*, *G. venezuelana*, *G. trilocularis* etc. Il est à noter qu'à la base de cette zone disparaissent *Catapsydrax* gr. *dissimilis*. Parmi les petits foraminifères benthiques on cite *Cyclamina latidorsata*, *Valvulina pennatula*, *Robulus rotulatus*, *Pleurostomalla alternans*.

En dessus, dans la coupe de Sinicë, suit la zone à *Globigerinoides bisphaericus* (éch. 54-68), qui représente la dernière zone du Burdigalien. L'association des foraminifères planctoniques diffère très peu de celle de la zone précédente, toujours très riche et très bien conservée.

Il faut noter que dans la partie sommitale de la



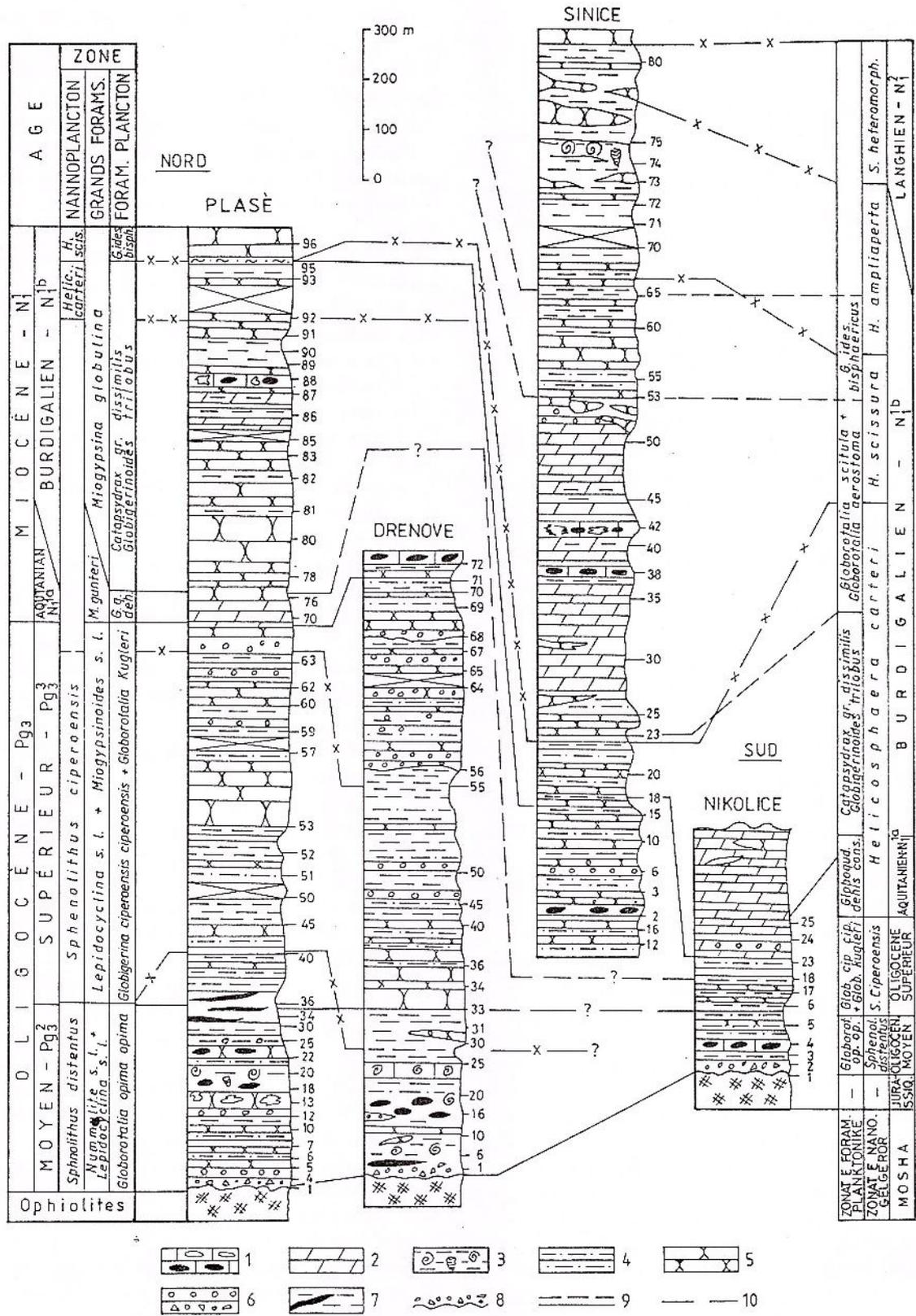


Fig. 3 - Corrélation biostratigraphique des coupes étudiées.
 1, calcaires à grands foraminifères, *Lithothamnium*; 2, marnes; 3, argiles à macrofaune; 4, silt; 5, grès; 6, conglomérat; 7, lignite; 8, transgression; 9, ligne de corrélation des zones des foraminifères planctoniques; 10, ligne de corrélation des zones du nannoplankton.

coupe de Sinicë (éch. 75-81) existent des formes transitionnelles de *Globigerinoides bisphaericus* à g. *Præorbulina*, très liées à *P. sicana*.

III. 2. Les grands foraminifères

Dans les dépôts de l'Oligocène moyen, à divers endroits (Lavdar, Plasë, Drenovë, Bellovodë, Floq, Kamenicë, Nikolicë), on a trouvé de nombreuses *Nummulites fichteli* associées à *Lepidocyclina* (*Nephrolepidina*) *praemarginata* et *L. (Eulepidina) elephantina*.

À l'Oligocène supérieur (coupe de Plasë, Drenovë, Nikolicë) les grands foraminifères sont représentés par *Nummulites bouillei*, *Lepidocyclina (N.) morgani*, *L. (Eulepidina) dilatata*, *Operculina complanata*, *O. gr. alpina*, *O. ammonoidea*, *Austotrillina striata*, *Archais hensoni*, Soritidae, Victoriellidae etc.

Dans la partie supérieure de l'Aquitaniien (Bitinckë, Nikolicë) se trouve un assemblage de *Mioegypsinia gunteri*, *M. gunteri-tani*, *M. tani*, *M. tani int. tani-globulina* et *M. globulina*.

Les calcaires biodétritiques de la partie inférieure du Burdigalien renferment *Mioegypsinia globulina*, *M. intermedia*, *Amphistegina haueriana*, *Pararotalia lithothamnica* etc.

Les grands foraminifères de l'Oligocène moyen sont *in situ*, alors que ceux de l'Oligocène supérieur et Miocène inférieur soient transportés des parties périphériques peu profondes.

III.3. Les nannofossiles calcaires

Les associations des nannofossiles calcaires observées dans le bassin Mésohellénique sont très similaires à celles de la zone ionienne. La richesse varie en fonction du milieu et de la nature du sédiment.

Les biozonations de Martini (1971), Bukry (1975) et Okada & Bukry (1980) sont bien applicables, exception faite pour le Miocène inférieur où l'absence de *Discoaster druggii* et *Triquetrorhabdulus carinatus* rend difficile la distinction entre les zones NN 1, NN 2 et NN 3. L'extrême rareté de ces espèces tropicales a été observée aussi dans la zone ionienne (Vathi, 1985, 1987, 1993). Cela nous a obligé d'utiliser d'autres marqueurs pour subdiviser cet intervalle (Vathi, 1993). Ainsi, ont été introduites la zone à *Helicosphaera carteri*, renfermant l'intervalle entre les apparitions de *H. carteri* et *H. ampliaperta*, et la zone à *Helicosphaera scissura*, l'intervalle entre les apparitions de *Helicosphaera ampliaperta* et *Sphenolithus heteromorphus*, ont été introduites. La zone à *H. carteri* est partiellement corrélée aux zones NN 1-NN 2 de Martini (1971), ou aux sous-zones à *E. deflandrei*-*E. druggii* de Theodoridis (1984). La zone à *H. scissura* comprend l'intervalle NN 2 (à part) - NN 3 ou les sous-zones à *H. vedderi*, *T. martinii*, *S. belemnoides* de Theodoridis (1984).

Les données de nannofossiles calcaires montrent que la base de cette succession appartient à l'Oligocène moyen (zone NP 24) et le sommet au Langhien (zone NN 5).

Zone à *Sphenolithus distentus* (NP 24)

Peu d'échantillons des formations grossières de la partie inférieure des coupes de Plasë et Drenovë (formations de Mborja, Drenova, Drenica, Argiles à Chama et Grès de Plasë) (Figs. 4, 5, 7) ont livré des associations pauvres jusqu'à assez riches (Argiles à Chama). La coupe de Drenova s'est avérée relativement plus riche que celle de Plasë. Les nannofossiles calcaires sont en général de petite taille, ce qui s'explique par les conditions littorales, peu propices à leur mode de vie.

La nannoflore de cette zone est composée de *Cyclicargolithus floridanus*, *C. abisectus*, *Coccolithus pelagicus*, *Sphenolithus moriformis*, *S. predistentus*, *S. distentus*, *S. ciperoensis*, *Zygrhabdulus bijugatus*, *Dictyococcites bisectus*, *D. heslandii*, *Braarudosphaera bigelowii*, *Micrantolithus* sp., etc.

On y retrouve des espèces remaniées du Crétacé-Eocène telles que *Watznaueria biporta*, *W. barnesae*, *Arhangelskiella cymbiformis*, *Eiffelithus turrisseiffeli*, *E. eximius*, *Cretarhabdus crenulatus*, *Micula staurophora*, *Ellipsolithus macellus*, *Cruciplacolithus tenuis*, *Pontosphaera latoculata*, *Lanternithus minutus*, etc.

Zone à *Sphenolithus ciperoensis* (NP 25)

La Formation de Bozdovec s'est révélée plus favorable à l'étude des nannofossiles calcaires. L'espèce marqueur de la zone NP 25, *Sphenolithus ciperoensis*, bien qu'en peu d'individus, est présente dans tous les échantillons. Les formations "Guri i Cjapit", "Merget e Dardhës" et Morava sont pauvres en nannofossiles calcaires, mais toutefois l'appartenance à la zone NP 25 est assurée.

L'association, bien préservée, est formée de *Cyclicargolithus floridanus*, *C. abisectus*, *Dictyococcites bisectus*, *D. heslandii*, *Sphenolithus moriformis*, *S. ciperoensis*, *S. conicus*, *Helicosphaera recta*, *H. perchnielseniae*, *H. euphatis*, *Ericsonia fenestrata*, *Zygrhabdulus bijugatus*, *Coccolithus pelagicus*, etc. Le remaniement du Crétacé-Eocène est présent presque par les mêmes espèces que dans la zone NP 24.

La formation conglomérato-gréseuse de Bradavicë et la partie inférieure de la formation marneuse de Sinicë, dans la coupe de Plasë, ne contiennent pas de nannofossiles calcaires, alors que dans la coupe de Sinicë on ait trouvé de rares individus de *Cyclicargolithus floridanus*, *Coccolithus pelagicus*, *Dictyococcites heslandii*, *Sphenolithus moriformis*, *Helicosphaera euphratis*. Ces espèces ont une large répartition stratigraphique, ne permettant pas de préciser l'âge.



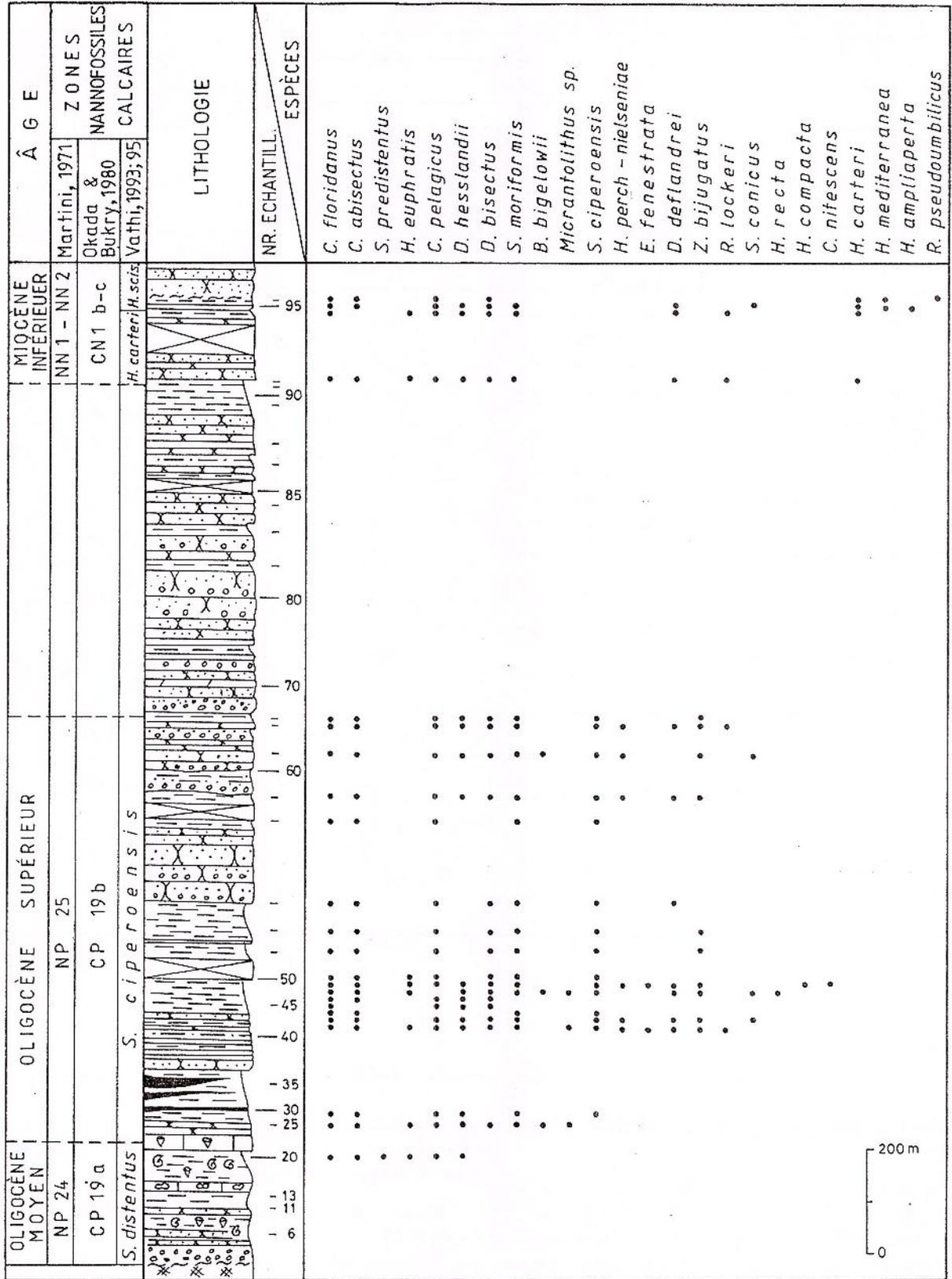


Fig. 4 - Répartition stratigraphique des espèces de nannofossiles calcaires dans la coupe de Plasă.

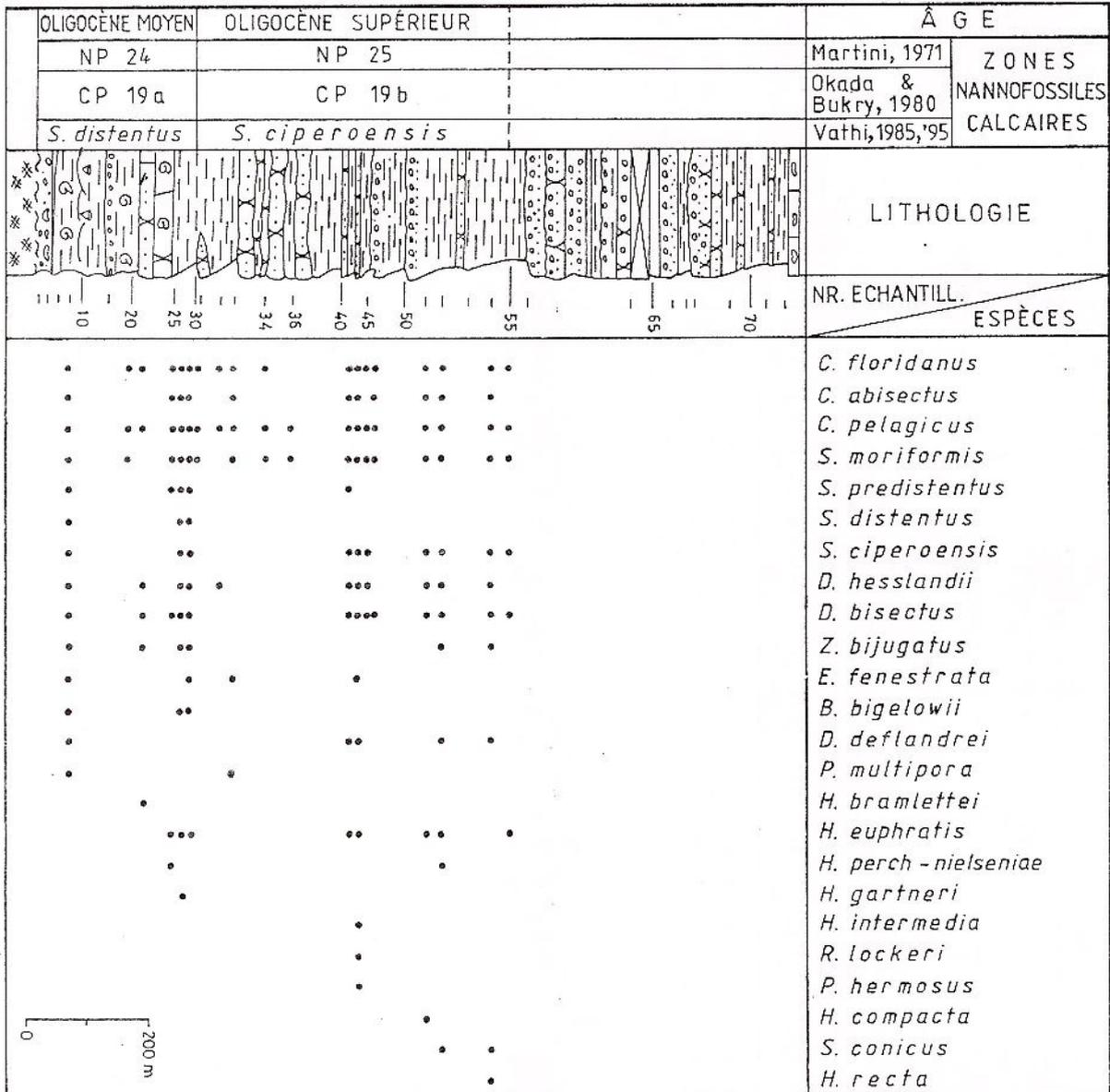


Fig. 5 - Répartition stratigraphique des espèces de nannofossiles calcaires dans la coupe de Drenovë.

Zone à *Helicosphaera carteri* (NN 1-NN 2 à part)

Cette zone a été reconnue dans peu d'échantillons dans les coupes de Plasë (éch. 92-94) (Fig. 4) et Sinicë (éch. 19-22) (Fig. 6). L'association de nannofossiles calcaires est assez riche, bien conservée, composée de *Cyclicargolithus floridanus*, *C. abisectus*, *Coccolithus pelagicus*, *Dictyococcites hesslandii*, *Helicosphaera carteri*, *H. mediterranea*, *H. euphratis*, *Reticulofenestra lockeri*, *Pontosphaera*

multipora, *Sphenolithus moriformis*, *Discoaster deflandrei*, etc. Dans la coupe de Sinicë se trouvent aussi *Sphenolithus delphis* et *S. belemnus*.

Zone à *Helicosphaera scissura* (NN 2 à part - NN 3)

Elle a été identifiée dans la partie terminale de la coupe de Plasë et dans l'intervalle d'échantillons 24-71 de la coupe de Sinicë. L'association est riche et assez bien préservée. On y trouve *Helicosphaera ampli-aperta*, *H. scissura*, *H. mediterranea*, *H. carteri*, *Coc-*

colithus pelagicus, *Cyclicargolithus floridanus*, *C. abisectus*, *Pontosphaera multipora*, *Sphenolithus moriformis*, *S. belemninos*, *Discoaster deflandrei*, *Dictyococcites hesslandii*, etc.

Zone à *Helicosphaera ampliapertura* (NN 4)

Cette zone a été reconnue seulement dans la coupe de Sinicë (éch. 71-77). L'association, riche et bien conservée, est composée de *S. moriformis*, *Helicosphaera ampliapertura*, *H. scissura*, *H. euphratis*, *H. intermedia*, *H. carteri*, *Geminilithella rotula*, *Reticulofenestra pseudoumbilicus*, *Coccolithus pelagicus*, *Cyclicargolithus floridanus*, *C. abisectus*, *Braarudosphaera bigelowii*, *Discoaster deflandrei*. Le remaniement est très faible, représenté par *Watznaeria barnesae*, *Cribrosphaerella ehrenbergi* du Crétacé et *Ericsonia formosa*, *E. fenestrata*, *Dictyococcites bisectus*, *Reticulofenestra umbilicus*, *Zygrhablithus bijugatus* de l'Eocène-Oligocène.

Zone à *Sphenolithus heteromorphus* (NN 5)

C'est la dernière zone identifiée dans la Dépression de Korça et dans tout le bassin Mésohellénique. Au Nord de la Grèce, la dernière zone reconnue par Muller (in Formili & Iokim, 1990) a été la zone NN 4.

L'existence de la zone NN 5 au sommet de la coupe de Sinicë montre bien la persistance du bassin jusqu'au Miocène moyen (Langhien).

L'association est riche et bien conservée. Elle est formée de *Coccolithus pelagicus*, *Helicosphaera carteri*, *Cyclicargolithus floridanus*, *C. abisectus*, *Dictyococcites hesslandii*, *Geminilithella rotula*, *Micrantolithus* sp., *Braarudosphaera bigelowii*, *Pontosphaera multipora*, *Sphenolithus heteromorphus*, *Calcidiscus leptoporus*, *C. macintyreii*, *Lithostromation perdurum*, *Discoaster exilis*, etc. Il est à noter l'abondance des genres *Braarudosphaera* et *Micrantolithus* dans les échantillons 80 et 81. En général, il est admis que les associations avec *Braarudosphaera*, *Micrantolithus* et *Pemma* caractérisent les milieux littoraux (Fuchs et Strander, 1977).

IV. Paléogéographie

Les données acquises par les observations en terrain et l'étude des foraminifères et nannofossiles calcaires nous permettent d'envisager le panorama général de l'évolution du bassin.

N'insistons pas sur le mécanisme de mise en place du bassin Mésohellénique, il est prouvé que les premiers dépôts appartiennent à l'Eocène moyen. D'une épaisseur réduite, ils sont transgressifs sur différentes zones tectoniques (flysch de Krasta-Cukali, calcaires

ou ophiolithes de Mirdita) représentés par des calcaires bréchiques, grès marneux et conglomérats.

À l'Eocène supérieur, la région émerge et la mer se retire vers le Sud-Est.

La transgression de l'Oligocène moyen s'est étendue sur un territoire plus vaste, par suite des déplacements vers l'ouest des zones de Krasta-Cukali et Mirdita. La mer était peu profonde, exprimée par le caractère grossier des sédiments du littoral et l'absence des foraminifères et nannofossiles calcaires dans les niveaux plus argileux. Le milieu néritique persiste pendant la sédimentation de la formation de Mborja (où localement existent d'eaux saumâtres et la formation des niveaux de lignite), et du Calcaire Corallien (coraux et bryozoa en position biologique et une grande richesse en grands foraminifères).

Au fur et à mesure que la bathymétrie s'accroît, on voit l'arrivée de rares foraminifères planctoniques et nannofossiles calcaires. À l'Oligocène supérieur les conditions du littoral externe-bathyal supérieur s'expriment par l'influence des turbidites et l'augmentation du plancton. Ensuite, l'apport intensif des sédiments assez grossiers au talus a donné naissance aux Grès de Plasë. Les conditions n'étaient pas favorables au développement des microorganismes. La macrofaune trouvée est déplacée des parties côtières. L'activité des courants s'accroît autour de la limite Oligocène-Miocène en formant les conglomérats de "Guri i Cjapit".

Les mêmes conditions du bassin se maintiennent de l'Aquitainien jusqu'à la partie inférieure du Burdigalien. L'existence des niveaux à calcaires biodétritiques, le mélange des associations profondes avec de grands foraminifères et la macrofaune prouvent la nature turbiditique des dépôts.

À partir de la zone à *Globorotalia scitula*-*G. acrotoma* du Burdigalien, le bassin devient plus profond. La sédimentation marneuse, les associations des nannofossiles calcaires et foraminifères très riches sont des aspects en commun avec le bassin ionien. Le haut niveau de la mer a été observé aussi dans divers points de la Méditerranée, exprimé par la transgression qui, d'après Bizon & Muller (1977), s'est produite dans les zones NN 3 - NN4. Cette transgression a été accompagnée par une augmentation de la température, ce qui a mené à une homogénéisation des associations de la nannoflore et microfaune de l'Atlantique, la Méditerranée et l'Indo-Pacifique (Bizon & Muller, 1977).

L'abondance des genres *Braarudosphaera* et *Micrantolithus* dans les derniers échantillons de la coupe de Sinicë est un indice du littoral, qui annonce ainsi le comblement du bassin au Langhien. Les grès sommitaux sont azoïques.



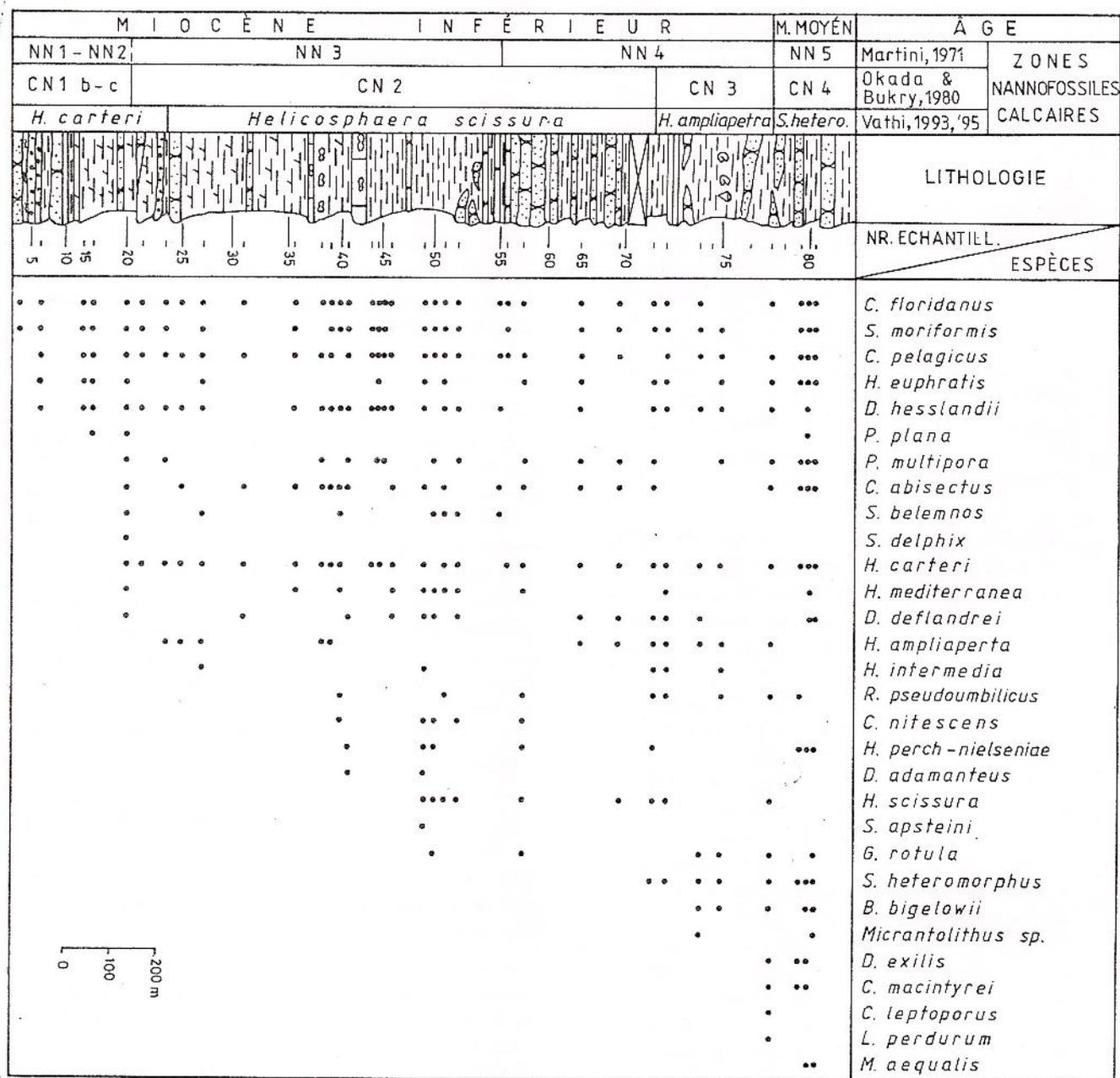


Fig. 6 - Répartition stratigraphique des nannofossiles calcaires dans la coupe de Sinicë.

La transgression du Plio-Pleistocène, venant du SSE, a donné naissance à un bassin peu profond aux eaux saumâtres. Les dépôts sont gréseux, avec des formations de lignite dans certains endroits. Dans telles conditions n'ont pas pu survivre que de rares ostracodes et des charophytes.

Vue d'ensemble sur les études en Grèce

Les résultats obtenus en Albanie et en Grèce donnent la possibilité de présenter un panorama général

du bassin Mésohellénique.

Résultat de l'orogénèse alpine, le bassin Mésohellénique s'individualise à l'Eocène moyen. Les dépôts marneux ou calcaires reposent sur les ophiolithes ou les calcaires de la zone de Mirdita (zone Subpélagonienne), ou les calcaires de la zone de Korabi (zone Pélagonienne).

À l'Eocène supérieur la région émerge et la mer se retire, vers le Sud-Est.

De nouveau, la mer s'installe à l'Oligocène moyen.

| ÂGE | | FORMATIONS DE BASSIN MESOHELLÉNIQUE (GRECE) | BIOHORIZONS FORAMINIFÈRES PLANCTONIQUES | FORMATIONS DEPRESSION DE KORÇA | ZONES DE FORAMINIFÈRES PLANCTONIQUES (Depression de Korça) |
|---------------------------------------|--------------------------------------|---|---|--------------------------------|--|
| OLIGOCÈNE MOYEN - Pg ₃ | MIIOCÈNE SUPÉRIEUR - Pg ₃ | | | | |
| OLIGOCÈNE MOYEN - Pg ₃ | Talioris / Baspaj | Eptahori | ↑ <i>Globigerina ciproensis angulisuturalis</i> | Mborja | <i>Globigerina ciproensis angulisuturalis</i> |
| | | | | | |
| OLIGOCÈNE SUPÉRIEUR - Pg ₃ | Voion / Petkos | Kridari | ↓ <i>Globorotalia opima opima</i> | Dardha | <i>Globorotalia opima opima</i> |
| | | | | | |
| MIIOCÈNE MOYEN LANGHIEN | Burdigaliens | Tsofilion | ↑ <i>Globorotalia kugleri</i> | Bozdovec | <i>Globigerina ciproensis ciproensis</i> |
| | | | | | |
| MIIOCÈNE MOYEN LANGHIEN | Ondria | S i n i c a | ↑ <i>Globorotalia kugleri</i> | Dardha | <i>Globigerina ciproensis ciproensis</i> |
| | | | | | |
| MIIOCÈNE MOYEN LANGHIEN | Burdigaliens | Tsofilion | ↑ <i>Globoquadrina dehiscens</i> | Brada v i c a | <i>Globoquadrina dehiscens</i> |
| | | | | | |
| MIIOCÈNE MOYEN LANGHIEN | Ondria | S i n i c a | ↑ <i>Globorotalia kugleri</i> | S i n i c a | <i>Globorotalia scifula - Globorotalia acrostoma</i> |
| | | | | | |
| MIIOCÈNE MOYEN LANGHIEN | Ondria | S i n i c a | ↑ <i>Globigerinoides trilobus</i> | S i n i c a | <i>Cafapsydrax gr. dissimilis</i> |
| | | | | | |

Fig. 7 - Corrélation entre les formations décrites dans les parties grecque et albanaise du bassin mésohellénique.



Les dépôts sont transgressifs sur les marno-calcaires de la phase précédente, les ophiolithes et les calcaires du Crétacé de Krasta-Cukali. Le bassin persiste jusqu'au Miocène moyen, exprimé par une succession mollassique.

A l'Oligocène moyen le bassin était plus profond dans le territoire grec où les sédiments sont plus pélitiques et rythmiques, alors que dans la région de Korça les sédiments soient plus grossiers. Ainsi la formation d'Eptahorion (Grèce du Nord) est caractérisée par des dépôts argileux, riches en foraminifères planctoniques, contrairement aux formations analogues d'Albanie (Mborja, Drenova, Sinica) où les dépôts sont plus grossiers, pauvres en foraminifères planctoniques et nannofossiles calcaires, englobant de grands foraminifères indicateurs d'eaux peu profondes.

A partir de l'Oligocène supérieur, une sédimentation turbiditique a lieu dans les coupes grecs et aussi dans celles albanaises. La profondeur s'accroît et atteint le maximum au Burdigalien supérieur-Langhien, avec une sédimentation marneuse et une grande richesse en organismes planctoniques.

La mer se retire de la région de Korça au début du Miocène moyen vers le Sud-Sud-Est. Cependant, dans les coupes grecques apparaissent les formations gréseuses d'Orlia et Ondria.

La transgression du Plio-Pleistocène a formé de petits bassins plutôt lacustres.

V. Conclusions

Avec la transgression de l'Oligocène moyen, le bassin Mésohellénique a vécu la deuxième phase, la plus importante de son histoire. Dans la région de Korça le bassin a persisté jusqu'au Langhien, formant ainsi une série mollassique successive de grande épaisseur. Les conglomérats de base sont transgressifs sur les ophiolithes et les marno-calcaires du Crétacé-Eocène.

La répartition des organismes dans la série est contrôlée par les faciès. Ainsi, le milieu néritique de l'Oligocène moyen et peu propice aux organismes planctoniques, alors qu'il soit favorable aux grands foraminifères. Le nombre des foraminifères planctoniques et nannofossiles calcaires augmente dans les turbidites de l'Oligocène supérieur-Miocène inférieur et ceux-ci deviennent abondants dans les marnes du Burdigalien supérieur-Langhien.

Les données offertes par les foraminifères planctoniques nous ont permis d'inclure la série dans l'intervalle des zones à *Globorotalia opima* à la base, et *Globigerinoides bisphaericus* au sommet. Des

formes transitoires entre *G. bisphaericus* et *G. Praeorbulina*, similaires à *P. sicana*, sont présentes dans les dépôts sommitaux.

Les nannofossiles calcaires prouvent que la base de la succession appartient à la zone à *Sphenolithus distentus* - NP 24 et le sommet à la zone à *S. heteromorphus* - NN 5, mettant ainsi en évidence le Langhien.

L'étude des grands foraminifères a identifié quatre assemblages distincts caractérisant l'Oligocène moyen, l'Oligocène supérieur, l'Aquitainien et la partie inférieure du Burdigalien.

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THE LARGER FORAMINIFERA OF MIDDLE OLIGOCENE – LOWER MIOCENE OF IONIAN ZONE AND KORCA DEPRESSION (SOUTH ALBANIA)

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Key words: Larger foraminifera. Biostratigraphy. Paleogeography. Middle Oligocene-Lower Miocene. Ionian zone. Korca depression (South Albania).

Abstract: In the Middle Oligocene and Lower Miocene of South Albania (Ionian zone and Korca depression) larger foraminifera are found studied for biostratigraphic and paleogeographic purposes. In both studied zones the Oligocene-Miocene boundary was determined by the larger foraminifera. In these zones deposits are represented by the successive terrigenous ones. In the first part of the upper short information was given on the geographic-geologic position and the facies of the regions. It must be stressed that larger foraminifera are studied on thin sections, which are prepared for the detritic limestones. These foraminifera are also studied by the materials which are dug up from clays and slumps. The study shows the spread of the larger foraminifera in South Albania and the paleogeographic data which testify for the correlation of both basins with the Mediterranean ones during the Middle Oligocene-Lower Miocene. In the second part of the paper information was given on genera and species of larger foraminifera which are found in South Albania. In both regions are found the same species and genera but the difference between them is in the manner of their place. In the Ionian zone species and genera are reworked inside the flysch-flyschoidal deposits of the Middle Oligocene. In the lateral parts (Tirane, Valesh, Tervoll) the foraminifera are found *in-situ*. In the Korca depression they are found in deposits of Middle Oligocene, the same as to lateral parts. In the deposits of Middle Oligocene-Lower Miocene, the foraminifera are reworked inside slumps which have derived from the lateral parts of the basin. The phylogenetic development and the manner of placing of larger foraminifera inside the rock reflect exactly the sedimentary-tectonic evolution of both basins. These data give the information for parts having the same evolution and the changes between them. The third part of this study presents the comparison of data which are gathered during the study of the larger foraminifera in South Albania with data known from literature for this organism group to Mediterranean regions (Italy, Greece). The paper was accompanied by plates of larger foraminifera where their study was performed on thin sections and in forms which are dug up from clays and slumps of both regions. This work is being accompanied by reflective literature and figures showing position of larger foraminifera on studied superficial geological sections and correlation of sections on the basis of the planctonic faunistic zones with the zones based on the larger foraminifera in South Albania.

Introduction

Larger foraminifera are used to decipher the stratigraphy of Middle Oligocene-Lower Miocene deposits. These foraminifera are encountered in the Ionian zone and Korca depression (South Albania). Based on the successive terrigenous deposits, it has been possible to distinguish some biostratigraphic events and consequently to determine Miocene-Oli-

gocene boundary and these between the stages in both represented zones.

In the Ionian zone many superficial sections are carried out on Lower Miocene-Middle Oligocene deposits studied by planktonic foraminifera. It is concluded that these deposits belong to upper-middle bathyal, while large foraminifera encountered in these deposits are of reworked nature. In lateral parts of Ionian zone larger foraminifera are encoun-



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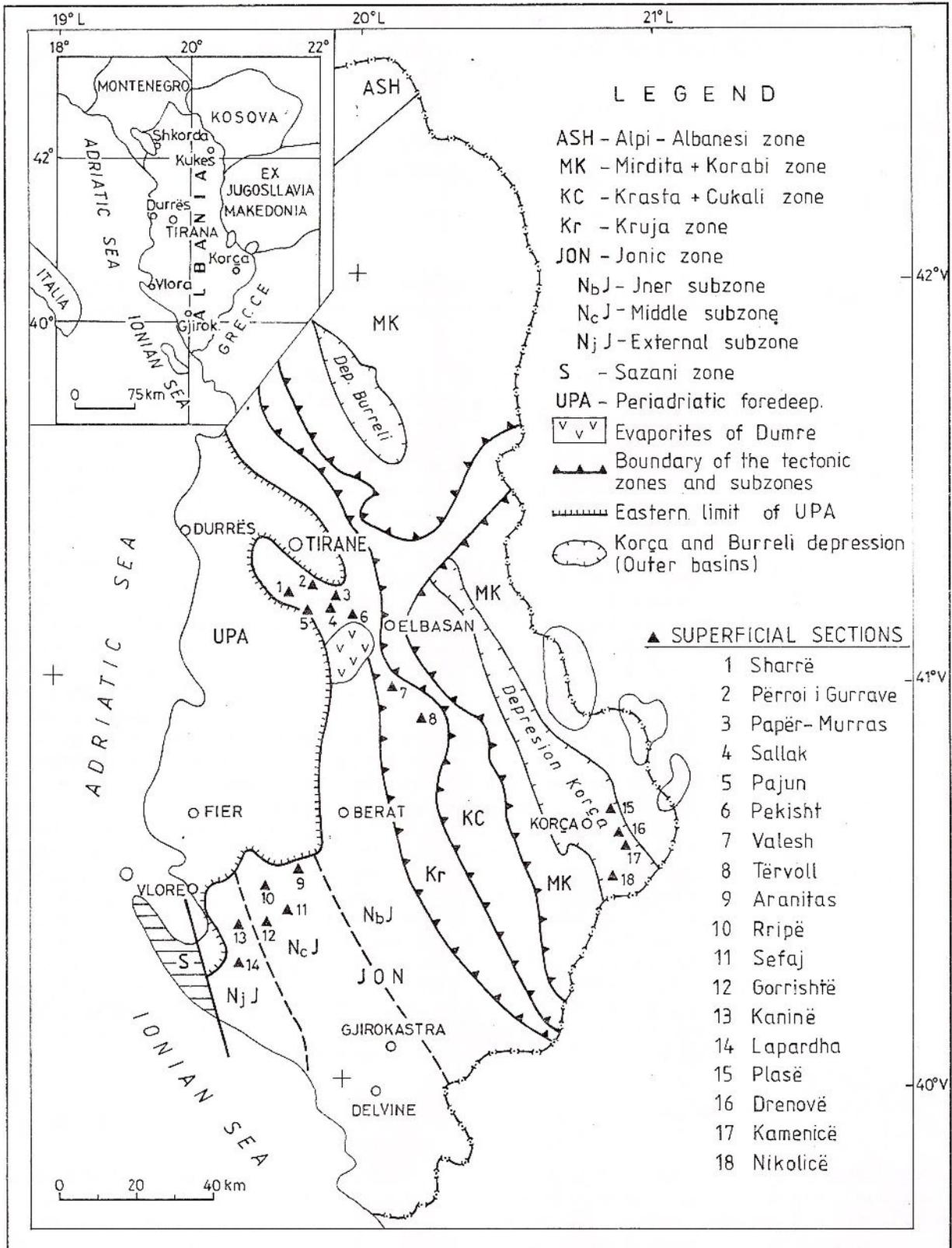


Fig. 1 - Principal tectonic elements and geographic position of the sections studied in South Albania.

tered in shallow deposits ranging from Middle-Upper Oligocene (Valesh) to Lower Miocene (Tervoll).

In the Korca depression, the larger foraminifera of Middle Oligocene are encountered in littoral deposits (Plase, Drenove) and they are of autochthonous nature. Planktonic foraminifera can be absent or appear rarely in these deposits. The larger foraminifera can be distinguished in flyschoidal and detrital-limestone deposits of Oligocene-Lower Miocene age. Development of larger foraminifera and manner of their pacing inside rock reflect exactly the evolution of sedimentation and tectonic factors of both basins. They reflect also resemblances and changes that exist between them.

I. Geographic-geologic position of Ionian zone and Korca depression on occasion of Albanides

A) Ionian zone

In the South Albania paleogeographic zones are separated in section Albanide-Dinaride (Fig. 1) Ionian tectonic zone is bordered with Kruja zone eastwards and relations between them are of tectonic and overthrust nature. Ionian zone is bordered with Sazani zone and periadriatic foredeep to the west. Based on the structural-geological data the Ionian zone is divided into three units: inner subzone, central subzone, external subzone.

The deposits of this zone are represented by the evaporitic ones and (limestones, flysch-flyschoidal formation, molasse) of Cenozoic-Mesozoic age encountered on the surface in the vast part of Ionian zone the cover evaporites are encountered on the surface (Dunre) and near tectonic faults in south west of Albania. In the low part (Triassic to Middle Jurassic) sealed deposits are represented by neritic limestones while up to Upper Eocene by pelagic ones. The quick change of sedimentation (from carbonatic to terrigenous in Eocene-Oligocene boundary) is a very visible geological fact in Ionian zone. At the bottom of Oligocene a marly horizon was deposited and upon it, the flysch deposits of Lower Oligocene intercalated with horizons of detrital limestones. In continuity the deposits of Mid-Upper Oligocene are represented by the thin bedded flysch are intercalated with detritic limestone beds up to the bottom of Upper Oligocene and slumps with reworked large foraminifera which are of different genera and species (Ylli, 1980; Sadushi, 1982; Myftari, 1981, 1989, 1990). From top of the Upper Oligocene up to low part of Burdigalian the deposits are represented by thick sandy facies which is placed successively on underlying deposits. In this manner the Oligocene-Miocene boundary presents a successive lithological

interval and it is possible to remark some changes on the composition of planktonic foraminifera and the large ones accured inside these deposits (Myftari, Kurteshi, 1995) (Figs. 2, 3). Burdigalian-Aquitian boundary is also represented by successive deposits while microfauna and larger foraminifera, other than the Oligocene ones, are enriched with new genera and species. Planktonic foraminifera (Figs. 2, 3) testify that deposits of low Oligocene up to low part of Burdigalian bathyal. At the begining of middle and upper part of Burdigalian (Globorotalia scitula-G. acroctoma zone) massive marls are found which are deposited transgressively on underlying deposits at the eastern part. The deposits of Langhian up to Pliocene ones are placed successively on underlying ones to the tectonic Sazani zone having considerable thickness.

B) Korca depression

It represents an extended depression parallel with Dinarido-Albanide-Helenide direction in the south-east of Albania forming the northern part of the Mesohellenic basin (northern Greece).

Deposits of Middle Oligocene of littoral nature are deposited transgressively on ultrabasic rocks of Jurassic age and Cretaceous-Eocene limestones and flysch of Krasta. There is a thick horizon (10-30 m thickness) of detritic limestone enriched with different neritic fauna. Deposits of Upper Oligocene are sedimented successively on the Middle Oligocene. Sedimentation environment has been a shallow sea. Gradually sediments towards up are represented by turbidite, thick bedded sandy flysch and gravels. The boundary of Upper Oligocene-Aquitian age is successively but it is difficult to determine it because the fossils are absent and related with depositional sequence. In section of Drenove and Nikolice deposits of Aquitian and Burdigalian age are enriched with planktonic foraminifera and represented by detritic limestone beds. These beds contain larger foraminifera reworked from shallow lateral parts of the basin. Langhian deposits are not verified by foraminifera while deposits of Serravallian up to Messinian are absent. Fluvial and lacustrine deposits of Pliocene-Pleistocene age (Devoll, Korce) are deposited transgressively on underlying deposits. They have thickness up to 1000 m. The geologic data testify for an overthrown basin, where the filling of basin occurred at the same time with basement shifting towards the west (Pigy-back type basin). This shifting has started after Lower Oligocene (in Korca depression the deposits of Lower Oligocene are missing). Succession of ages from north and north-east towards south-east testify for gradual regression of sea from northern territories (Albania)



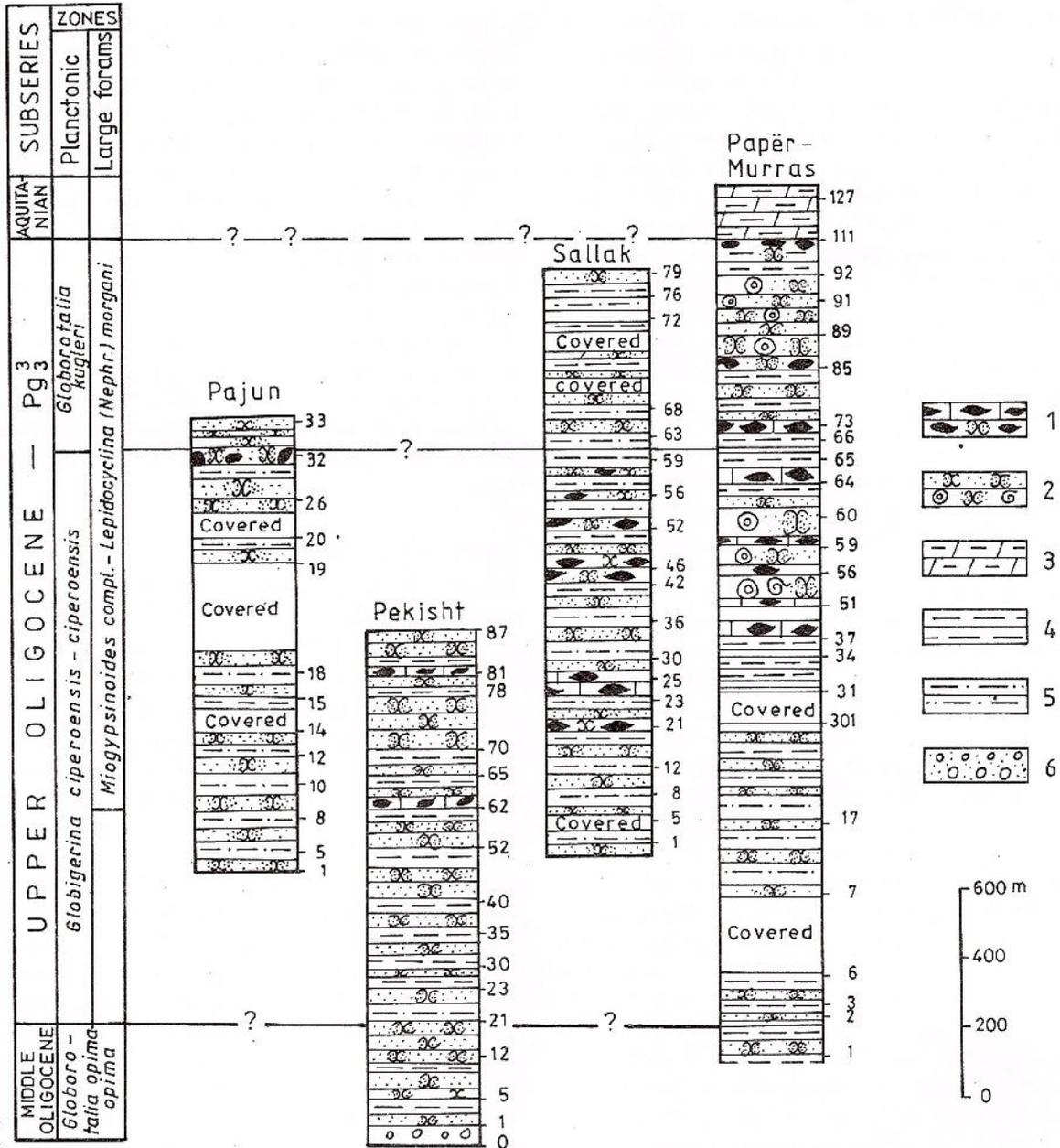


Fig. 2 - Chronostratigraphic correlation between sections with Upper Oligocene-Aquitanian deposits in the north-eastern part of the Ionian Zone based on planktonic and larger foraminifera. 1, detritic limestones with larger foraminifers; 2, sandstones; 3, marls; 4, clays; 5, slumps; 6, conglomerates.

towards south-east (Mesohellenic basin, northern Greece) (Kumati et al., 1995).

II. Genera and species of larger foraminifera observed in South Albania

In both studied units larger foraminifera are observed as in Oligocene flysch and Lower Miocene flyschoids (Ylli, 1980; Sadushi et al., 1982). Reworking of Oligocene larger foraminifera into central and external Ionian subzones is visible because the bio-

detritic material is mainly represented by foraminifera of Eocene age accompanied by the planktonic foraminifera of shallow zones and large foraminifera living in this zones and are not reworked. In the lateral parts of the Ionian zone (Valesh, Tervoll) and in Middle Oligocene deposits of the Korca depression the foraminifera are not reworked (Kumati, Vathi, 1995). In both cases this group of foraminifera has a great importance for deciphering of stratigraphy and for correlation of regions which are at great distance in Albania (Figs.

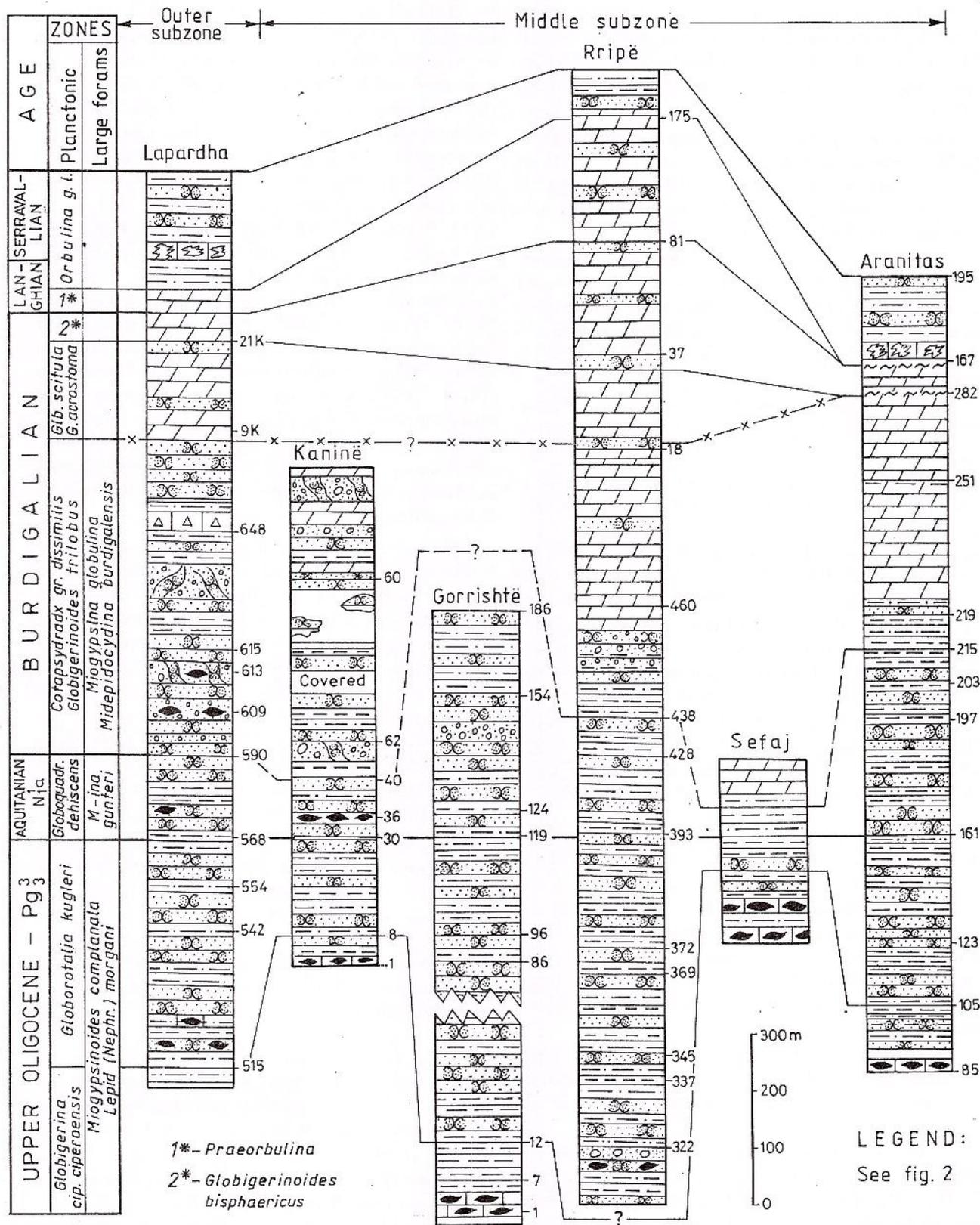


Fig. 3 - Chronostratigraphic correlation between sections with Upper Oligocene-Lower Miocene deposits western and south-western part of the Ionian Zone based on planktonic and larger foraminifera.

2, 3) between them and they also serve to correlate the Albanian and Mediterranean regions (AGIP Mineraria, 1988; Matteuci, Schiavinotto, 1977; Schiavinotto, 1979; Drooger, Loagland, 1986; Mulder, 1975).

There is a difficulty because larger foraminifera in pelagic deposits are not encountered in all sections but they are present at separated horizons (Figs. 2, 3) far from each other according to turbidic portions derived from shallow zones into basin.

1. *Genus Nummulites*. In deposits of Lower-Middle Oligocene in Ionian zone are encountered *Nummulites* represented by Eocene and Oligocene species. They are accompanied by *Nummulites fichteli* (reticulate) (Pl. IV, Fig. 2) (Ylli, 1980; Myftari, 1981).

Presence of this species in slumps and detritic limestones testify the Oligocene age of flysch in the Ionian zone.

In Korca depression the presence of *N. fichteli* is ancient (Cizancourt, 1931; Petro, Dodona, 1976). Cizancourt has observed *N. intermedius* near village Pulaha at white spring (Bellovode). These are accompanied by *Lepidocyclina* which determine Latorfian-Rupelian age. We have observed this species in many sections of Ionian zone as in Valesh, Plasa section, Drenove, Floq etc. It can be dug out from the Middle Oligocene deposits. We have observed *N. boulei* and *N. vascus* (Plase, Drenove). In all cases the species of this genus are found accompanied by specimens of genus *Lepidocyclina* which testifies for Middle-Upper Oligocene age (Mulder, 1975; AGIP Mineraria, 1988; Drooger, Laagland, 1986).

2. *Genus Lepidocyclina*. This genus was encountered in Middle Oligocene (Pl. I, II, IV; Pl. V, Figs. 1, 2, 4). The more primitive species is *L. (Nephrolepidina) praemarginata* (Pl. I, Figs. 1-6; Pl. II, Figs. 4, 7) which is encountered in the lateral parts of Ionian zone (Valesh) and in reworked complexes of younger ages. It is testified by the biometric studies carried out on these specimens from the Ionian zone (Myftari, 1981).

This species was found in the Korca depression and it can be dug out of rocks and is accompanied mainly by specimens of species *N. fichteli* (Kumati et al., 1995).

In the Ionian zone (encountered since Middle Oligocene) are found forms A (Pl. I, Figs. 1-6) and B (Pl. II, Figs. 4, 7) of this species in layers that are accompanied by detritic limestones and in detritic limestones up to the bottom of Upper Oligocene.

In the above mentioned zone, in the Upper Oligocene species *L. (Nephrolepidina) morgani* (Pl. I, Figs. 10, 11, 12; Pl. II, Figs. 5-7, 11) (based on

De Mulder's classification) (Mulder, 1975; Drooger, Laagland, 1986) was found. This species ranges up to Lower Miocene and was widely found in Ionian zone and Korca depression. At the low part of the Burdigalian was found species *L. (Nephrolepidina) tournoueri* (Pl. I, Figs. 13, 14) indicating disappearance of genus *Lepidocyclina* in these intervals of South Albania. The subgenus *Eulepidina* was widely found in Ionian zone and Korca depression (Pl. I, Figs. 7, 15-18; Pl. II, Figs. 1-3, 8-10, 12; Pl. IV, Figs. 1, 3; Pl. V, Fig. 2) (Myftari, 1990). In the first zone are encountered forms A and B of species *L. (Eulepidina) dilatata* with small dimensions (Pl. I, Figs. 15-18; Pl. IV, Fig. 3). Grade of embrace of protoconch by deuterocoenon varies from visible tagentiality to forms with spherical protoconch in central position (Pl. I). In Tirana region and Korca depression are found specimens of *L. (Eulepidina) elephantina* (Pl. II, Figs. 1-3, 8-10, 12; Pl. IV, Fig. 1) having dimensions up to 8 cm diameter in biological position in Middle Oligocene deposits and in the low part of Upper Oligocene (Pl. IV, Fig. 1).

Difference of specimens of genus *Lepidocyclina (Eulepidina)*, according to size of test, is made because of their carriage by turbiditic currents from eastern parts of Ionian zone towards western parts of it. These specimens are extremely numerous in Korca depression. They are represented by *L. (Eulepidina) elephantina* accompanied by other fossils in biologic position (Corals, Gastropods, Bivalvia) (Pl. I, Fig. 7; Pl. IV, Fig. 3). The same appearance was observed at coral limestone which is encountered in deposits of Middle Oligocene in Korca depression (Perto, Dodona, 1976; Kumati et al., 1995).

It must be added that specimens of subgenus *Nephrolepidina* and *Eulepidina* are found together in Ionian zone and Korca depression (Pl. V, Figs. 1, 2). They range since Middle Oligocene. When they disappeared, it was observed that subgenus *Eulepidina* is not found in Burdigalian deposits but it ranges up to bottom of Aquitanian and then it dies out (Myftari, 1989, 1990).

3. *Genus Miogypsina s.l.* (Pl. III, Figs. 1-19; Pl. V, Figs. 3, 4). More primitive specimens are unispiral species of which are found in Ionian zone (Pl. III, Figs. 1-3). It was encountered after disappearance of reticulate *Nummulites* indicating Middle and Upper Oligocene boundary together with species *L. (N.) morgani*. The specimens of this species have a vast spread in south-west of Ionian zone (Myftari, 1981) and in north-north-east (Tirana region) (Myftari, 1981, 1990), and in Korca depression (Kumati et al., 1995). In the surficial sections of both territories the most elementary specimens are presented



by long whorls (up to 24 chambers). Those specimens decreased gradually until they form a phylogenetic line, completed from primitive, *Miogyssina* to the most evolved *Miogyssina*. At (Pl. III, Figs 1-19) one can observe the gradual reduction of elementary whorl from specie *M. complanata*-*M. formosensis* (Pl. III, Figs. 4, 5), *M. bantamensis* (Pl. III, Fig. 6; Pl. V, Fig. 3) through intermediate species and other species of subgenus *Miogyssina*.

Aquitanian-Upper Oligocene boundary was determined by the appearance of subgenus *Miogyssina s.s.* or more exactly by species *Miogyssina gunteri* (Pl. III, Fig. 7). During the Aquitanian, these species have been accompanied by specimens of species *Miogyssinoides bantamensis* in Ionian zone (sections Lapardha, Kanine) and larger foraminifera. At the top of the Aquitanian species *M. (Miogyssina) exemplum intercentrale gunteri tani* is found.

M. (Miogyssina) tani (Pl. III, Figs. 8-13, 15), *M. (M.) ex.int. tani-globulina*, *M. (M.) globulina* (Pl. III, Figs. 14, 16, 17), *M. (M.) ex. int. globulina-intermedia* (Pl. III, Fig. 19) which is the last individual of the evolution the line of *Miogyssina s.s.* in Ionian zone and Korca depression in south of Albania. Species *M. (Miolepidocyclina) burdigalensis* was encountered in Burdigalian and disappeared at the low part of it. This species characterizes the top of *Miogyssina s.s.* in Ionian zone and after this level it migrated to regions of northern Africa (Adams, 1981; Dooger, Laaland, 1986; Mulder, 1975).

4. Genus *Austrotrillina* (Pl. V, Fig. 2). At the levels of appearance of *Miogyssinoides complanata* in north and north-east of Ionian zone (Tirana region and Korca depression) specimens of *Austrotrillina striata* (Myftari, 1981, 1989) are found. This genus ranges in Upper Oligocene and is not found in Lower Miocene (Aquitanian).

5. Genus *Spiroclypeus* (Pl. III, Figs. 26-29). Specimens of this genus are found of reworked nature together with complexes of Upper Eocene if possible species *S. granulatus*. In Upper Oligocene species *S. blanckenhorni* and *S. tidoenganensis* are found (Globig. cip. ciperoensis zone according to the planktonic foraminifera). These species are not found in the Upper Oligocene.

6. Genus *Heterostegina* - specimens of genus *Spiroclypeus* are found together with individuals of genus *Heterostegina*. In Lower Miocene species *Heterostegina depressa* (Pl. III, Fig. 31) is found.

7. Genus *Pararotalia* (Pl. III, Figs. 22-24; Pl. V, Fig. 6). A diversity of species of this genus, as *P. viennoti* (Pl. III, Figs. 22-24), *P. lithothamnica*, *P.*

verriculata, are found from Middle Oligocene up to Lower Miocene without particular stratigraphic importance.

8. Genus *Operculina* (Pl. III, Fig. 30; Pl. V, Fig. 5). This genus is very frequent in deposits of Upper Oligocene-Lower Miocene of Korca depression and rarely in Ionian zone. *O. gr. alpina*, *O. complanata* are found but without any stratigraphic importance they have a little value to the paleogeographic data.

9. Genus *Amphistegina* (Pl. III, Figs. 20, 21; Pl. V, Fig. 4). This genus ranges in deposits of Middle and Upper Oligocene and Lower Miocene together with species *A. hauteriana*. It also range up to younger ages with specie *A. lessoni* (Pl. III, Figs. 20, 21) without any stratigraphic value.

10. Genus *Sphaeroigypsinina* (Pl. III, Fig. 25) specimens of *S. globulina* are found in complex of large foraminifera of Upper Oligocene-Lower Miocene in Ionian zone without stratigraphic value.

11. *Peneroplidae-Soritidae* specimens of *Peneroplis ormaricus* in micro and macro foraminiferic specimens are found in deposits of Upper Oligocene of Korca depression. It is not found in the younger deposits.

12. Other genera together with complex of large foraminifera in Ionian zone are found other smaller benthonic foraminifera as *Victoriella aquitanica* (Pl. IV, Fig. 3), *Carpenteria*, *Borelis melo* and *Planolinderina escornebovensis*. Except the last species the others are of no stratigraphic value.

III. Comparisons with other regions of Mediterranean

All genera and species of larger foraminifera of Middle Oligocene up to Lower Miocene (South Albania) are found in central Mediterranean and mainly in adjacent places with Albania (north Italia and north Greece) (Matteucci, Schiavinotto, 1977; Schiavinotto, 1979; Drooger, Laagland, 1986; Mudler, 1976). They testify for the same ages and their vast spread in this bioprovince. Correlations which are realised on the basis of planktonic foraminiferic zones haven't any visible difference, except unprecision that results from the different volume of stages and local planktonic faunistic zones. These data testify that the studied region is a part of history of development, in general occasion of Mediterranean during Oligocene-Lower Miocene age.

Facts which are mentioned above testify this conclusion. Species of larger foraminifera of South Albania are the same as other places of Mediterranean



(AGIP Mineraria, 1988; Adams, 1981; Drooger, Laagland, 1986; Mulder, 1975). In Albania (Ionian zone) these foraminifera are found in flyschoidal deposits of deep and open sea of reworked nature. This is the only change between Mediterranean places and Ionian zone. These foraminifera are derived from lateral shallow parts of the same basin having had conditions for foraminifera living (Myftari, 1981; Sadushi, 1982; Myftari et Kurteshi, 1994). The eastern part of Ionian zone is overlapped by overthrust of Kruja Shallow zone, consequently a part of deposits containing specimens of large foraminifera (autochthonous nature) are overlapped by Kruja tectonic zone. Some fragments are not overlapped by this zone and are present (Valesh) (Myftari, 1981).

In Korca depression are found deposits of complexes of larger foraminifera in their biological position of Middle Oligocene and are accompanied by fauna of shallow waters. Consequently, Korca depression change from the history of paleogeographic development against Ionian zone. It serves as a connective bridge between the western regions, characterized by deep and open deposits and the south-east parts represented by shallow water ones (Kumati et al., 1995). One can say that genera and species of large foraminifera of Upper Oligocene-Lower Miocene age (in Albania) are the same with the central Apennines (Italia) and Greece ones (AGIP Mineraria, 1988; Matteucci, Schiavinotto, 1979; Mulder, 1975). They testify for the same paleogeographic conditions of these regions on the basis of the presence of the same groups of phylogenetic line of genus *Miogypsina* and *Lepidocyclina*. The same phenomenon was observed for specimens of planktonic foraminifera and smaller benthonic ones which are the same. In general, the specimens of large foraminifera of South Albania have more resemblance with large foraminifera ones of central Mediterranean. The same stratigraphic levels of their appearance and disappearance in South Albania and other places are the best evidence to do this comparison and to come to this conclusion.

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Plate I

- Figs. 1 ab, 2 ab, 3 ab, 4, 5, 6** - *Lepidocyclina (Nephrolepidina) praemarginata* (DOUVILLE). Forma A. Figs. 1, 2, 3, 4, 6, equatorial view, X 15; Figs. 1 b, 2 b, 3 b, details of the nukleokonk of the specimen figures 1 a, 2 a, 3 a; Fig. 5, axial view. Age: Middle Oligocene Sharre (Tirane) section.
- Fig. 7** - Clays with *Ulepidina* sepp. in biological position. Natural size. Age: Upper Oligocene. Valesh (Elbasan) section.
- Figs. 8, 9** - *Lepidocyclina (Nephrolepidina)* sp. X 20. External view. Age: Upper Oligocene. Aranitas (Fier) section.
- Figs. 10, 11, 12** - *Lepidocyclina (Nephrolepidina) morgani* LEMOINE and DOUVILLE R. Forma A, X 20. Figs. 10, 11, equatorial view; Fig. 12, axial view. Age: Upper Oligocene. Gorisht (Vlore) section.
- Figs. 13, 14** - *Lepidocyclina (Nephrolepidina) tournoueri* LEMOINE and DOUVILLE R. Forma A, X 20. Fig. 13, equatorial view; Fig. 14, axial view. Age: Early Miocene (Early Burdigalian). Velcan (Fier) section.
- Figs. 15, 16, 17, 18** - *Lepidocyclina (Eulepidina) dilatata* MICHELOTTI. Forma A, X 15. Fig. 7, clays with *Eulepidina*, X 1; Figs. 15, 17, 18, equatorial view; Fig. 18, *Eulepidina* with protoconch and deuteroconch in concentric position; Fig. 16, axial view (part). Age: Early Miocene (Aquitanian). Aranites (Fier) section.



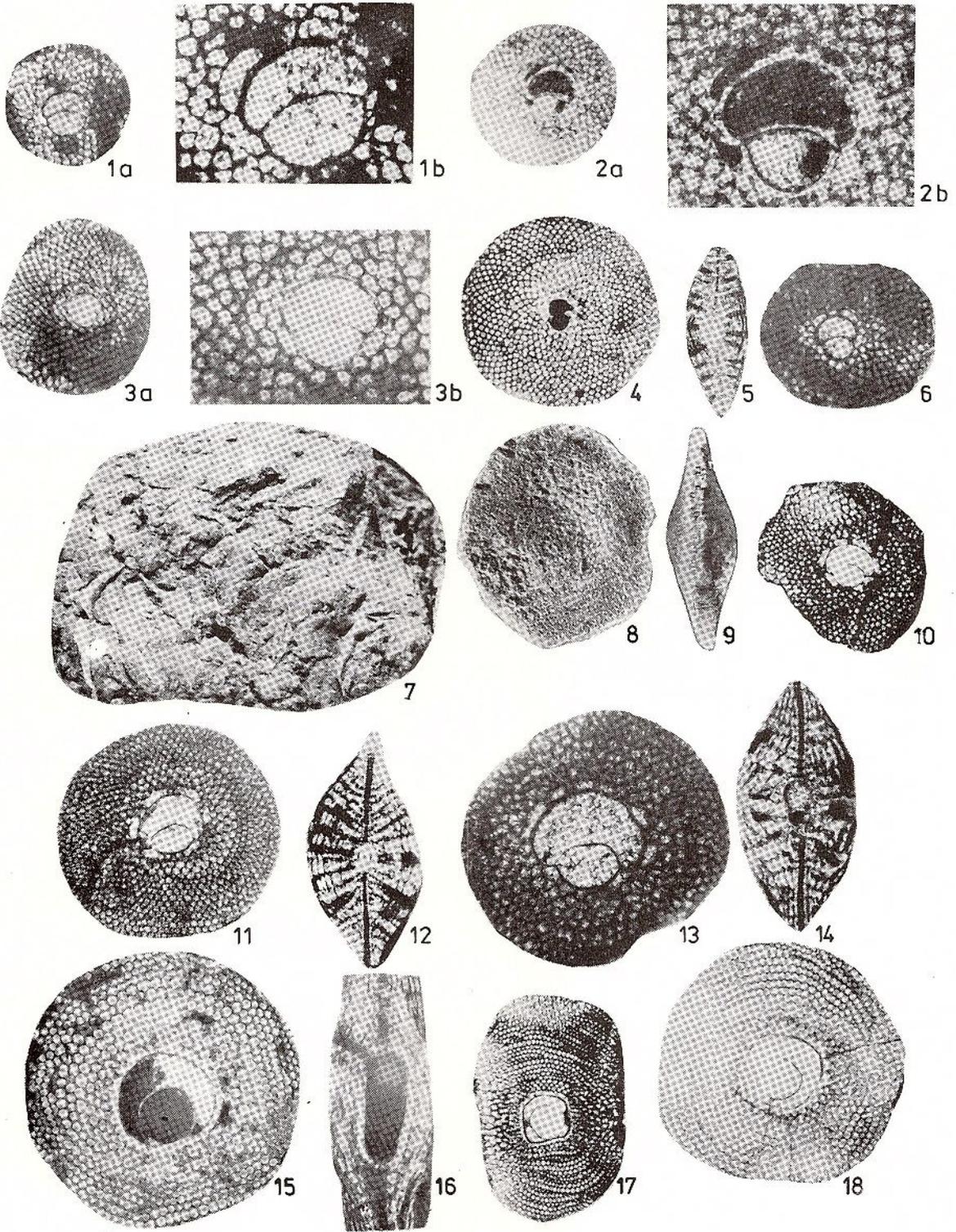
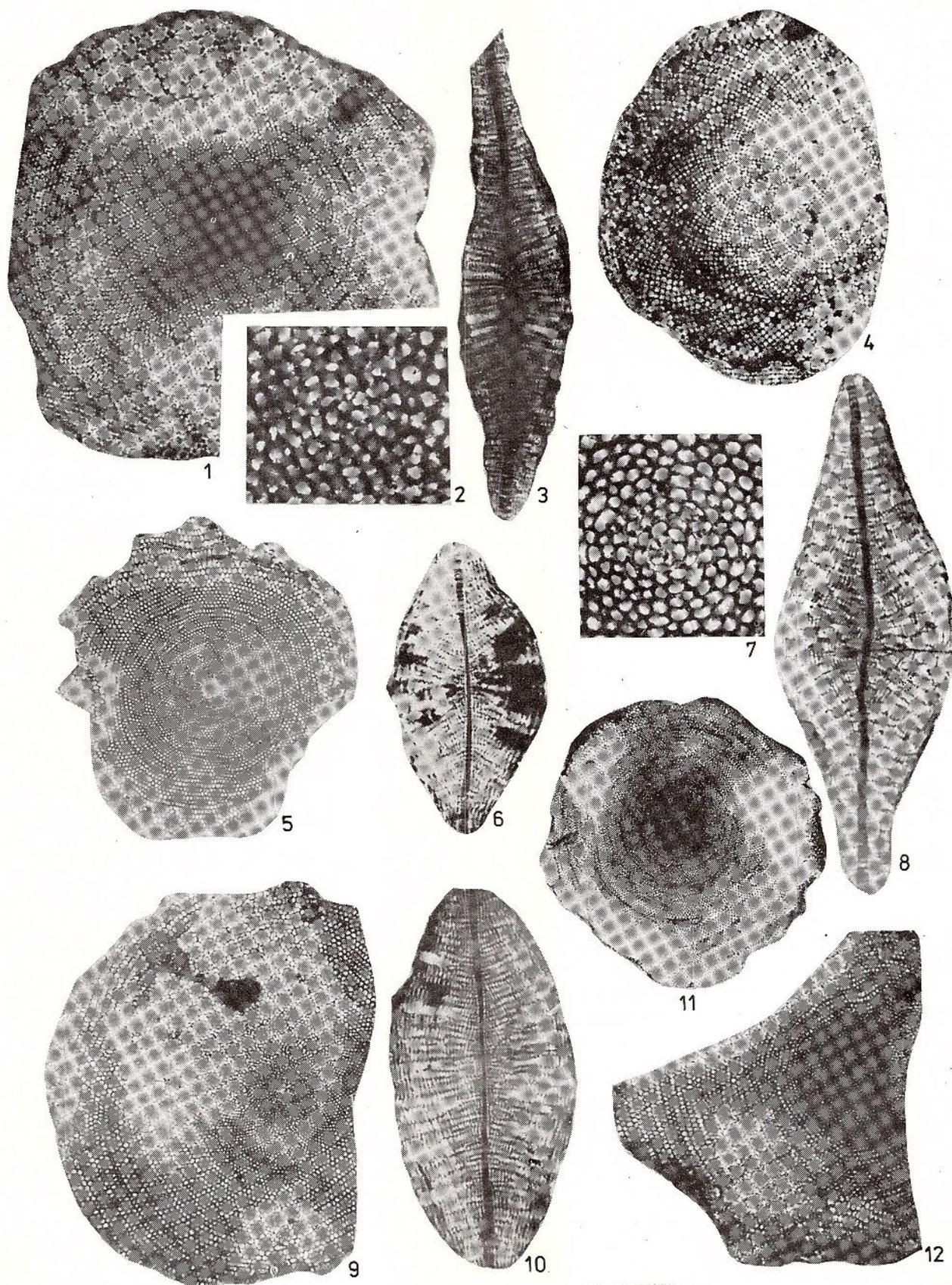


Plate II

- Figs. 1, 2, 3, 8, 9, 10, 12** - *Lepidocyclina (Eulepidina) elephantina* LEMOINE and DOUVILLE. Forma B, X 15. Age: Upper Oligocene. Valesh (Elbasan) section. Internal view forma B; Figs. 1, 9, 12, equatorial view; Figs. 3, 8, 10, axial view; Fig. 2, details of the nukleokonik specimen Figure 1, X 100.
- Figs. 5, 6, 7, 11** - *Lepidocyclina (Nephrolepidina) morgani* LEMOINE and DOUVILLE X 15. Age: Upper Oligocene. Plase (Korce) section. Internal view: Forma B. Figs. 5, 11, equatorial view; Fig. 6, axial view; Fig. 7, details of the nukleokonik specimen Figure 4, X 100.
- Figs. 4, 7** - *Lepidocyclina (Nephrolepidina) praemarginata* DOUVILLE. X 15. Age: Middle Oligocene. Kushtaj (Vlore) section. Internal view. Forma B, Fig. 4, equatorial view. Fig. 7, details of the nukleokonk specimen Figure 4.





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Plate III

- Figs. 1, 2, 3** – *Miogypsina (Miogypsinoides) complanata* SCHLUMBERGER, X 15. Fig. 1, external view; Fig. 2, equatorial view; Fig. 3, axial view. Age: Upper Oligocene. Aranites (Fier) section.
- Figs. 4, 5** – *Miogypsina (Miogypsinoides) formosensis* YABE and HANZAWA, X 15. Equatorial view. Age: Upper Oligocene. Aranites (Fier) section.
- Fig. 6** – *Miogypsina (Miogypsinoides) bantamensis* (TAN SIN HOK), X 15. Equatorial view. Age: Upper Oligocene. Paper Murras (Elbasan) section.
- Fig. 7** – *Miogypsina (Miogypsina) gunteri*, X 15. Equatorial view. Age: Early Miocene (Aquitanian). Kanine (Vlore) section.
- Figs. 8, 9, 10, 11, 12, 13, 15** – *Miogypsina (Miogypsina) tani* DROOGER, X 15. Figs. 8, 9, 10, 11, 13, equatorial view; Fig. 12, axial view. Age: Early Miocene (Early Burdigalian). Lapardha (Vlore) section.
- Figs. 14, 16, 17** – *Miogypsina (Miogypsina) globulina* MICHELOTTI, X 15. Fig. 14, axial view; Figs. 16, 17, equatorial view. Age: Early Miocene (Early Burdigalian). Lapardha (Vlore) section.
- Fig. 18** – *Miogypsina (Miogypsina) globulina ex int. intermedia*, X 15. Equatorial view. Age: Early Miocene (Early Burdigalian). Lapardha (Vlore) section.
- Fig. 19** – *Miogypsina (Miolepidocyclina) burdigalensis* (GUNBEL), X 15. Equatorial view. Age: Early Miocene (Early Burdigalian). Lapardha (Vlore) section.
- Figs. 20, 21** – *Amphistegina lessoni* D'ORB., X 15. Fig. 20, equatorial view; Fig. 21, axial view. Age: Early Miocene-Middle Miocene. Rripe (Fier) section.
- Figs. 22, 23, 24** – *Pararotalia viennoti* (GRIG), X 15. Fig. 23, external view; Fig. 22, equatorial view; Fig. 24, axial view. Age: Oligocene-Early Miocene. Sallak (Tirane) section.
- Fig. 25** – *Sphaerogypsina globula* (REUSS), X 15. Equatorial view. Age: Oligocene-Early Miocene. Rajun (Elbasan) section.
- Fig. 26** – *Spiroclypeus blanckenhorni* VAN DER VLERK, X 15. Equatorial view. Age: Upper Oligocene. Aranites (Fier) section.
- Figs. 27, 28, 29** – *Spiroclypeus tidoenganensis* VAN DER VLERK, X 15. Fig. 27, equatorial view; Figs. 28, 29, axial view. Age: Upper Oligocene. Aranites (Fier) section.
- Fig. 30** – *Operculina complanata* DE FRANCE, X 15. Equatorial view. Age: Oligocene-Middle Miocene. Sallak (Tirane) section.
- Fig. 31** – *Heterostegina depressa*, X 15. Equatorial view. Age: Early Miocene-Middle Miocene. Lapardha (Vlore) section.



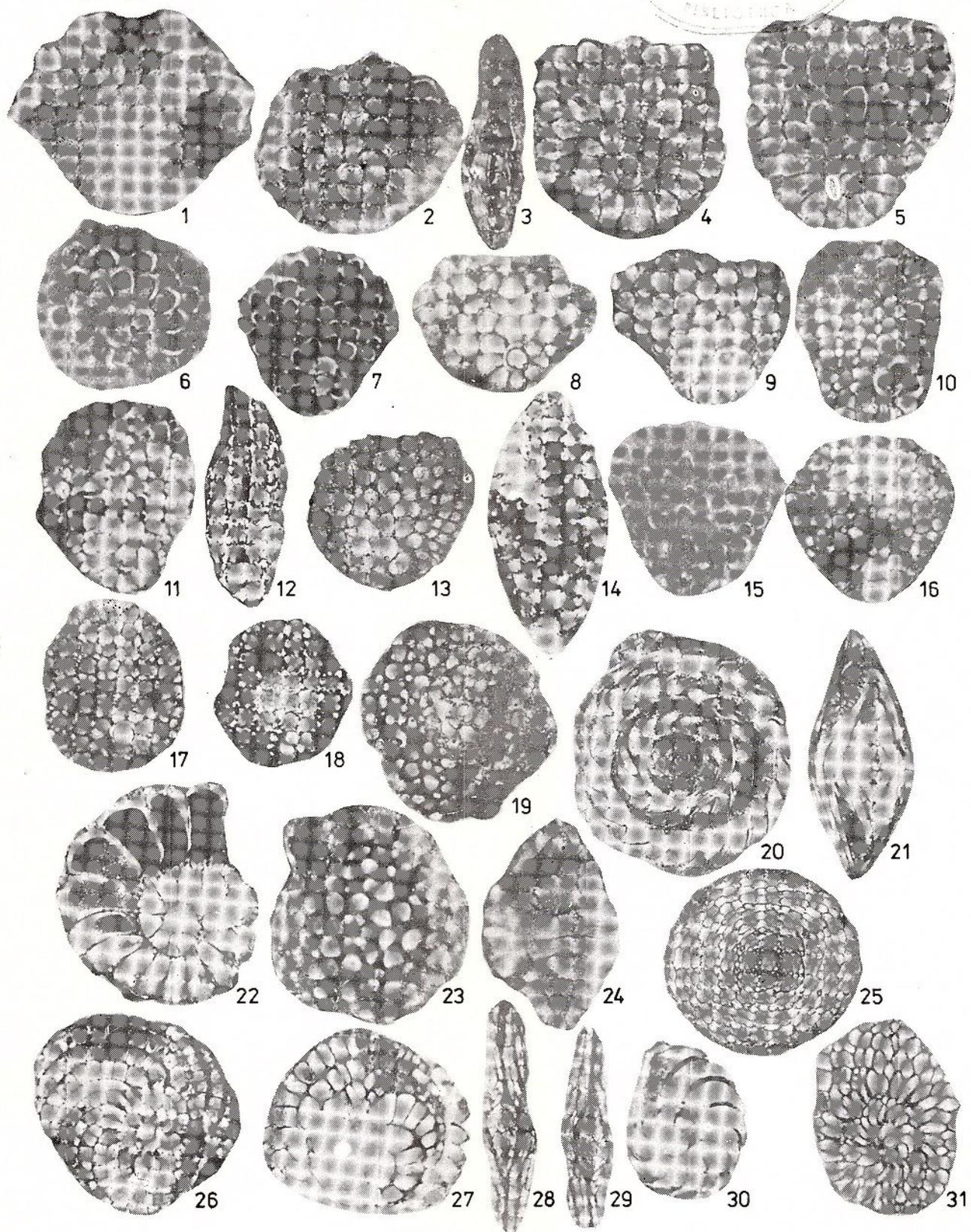
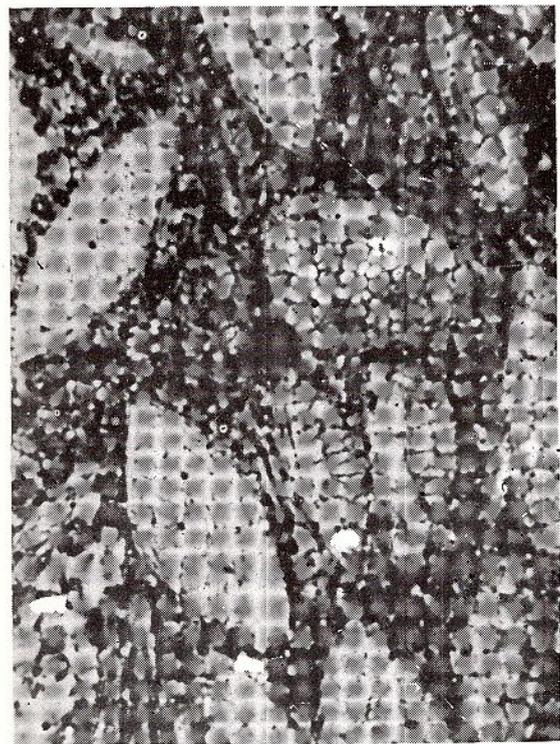
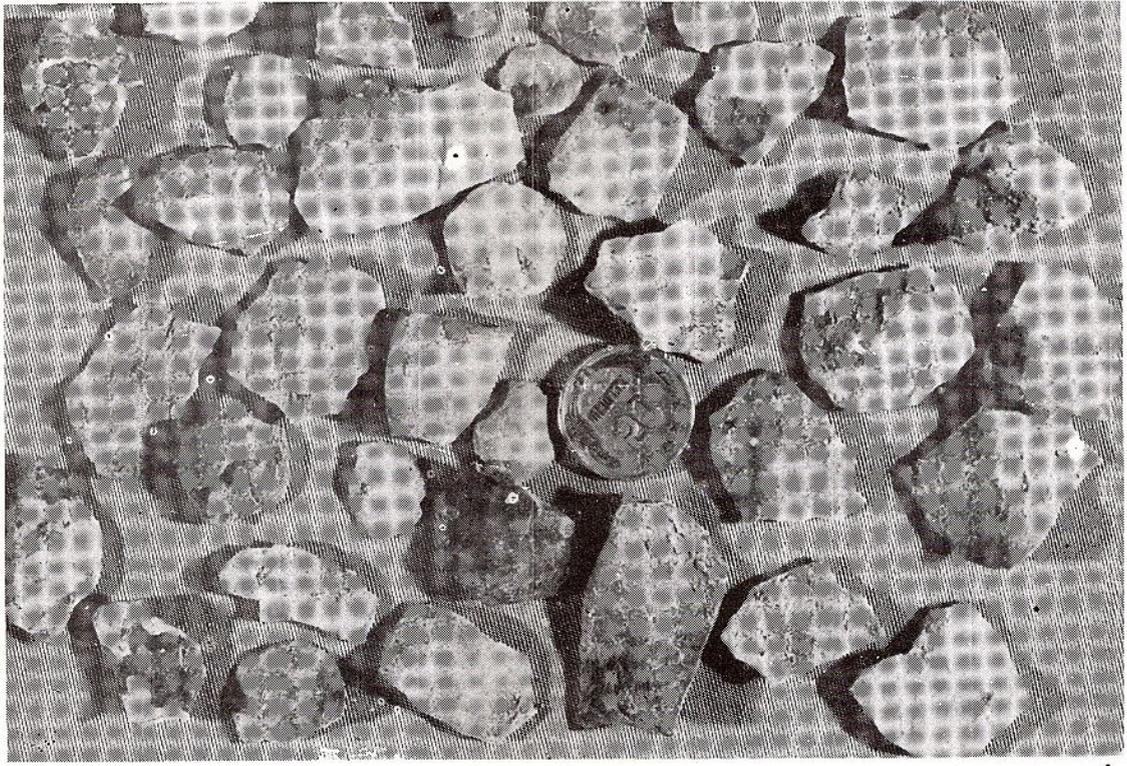


Plate IV

- Fig. 1** - *Lepidocyclina (Eulepidina) elephantina* LEMOINE and DOUVILLE, X 0,8. Forma B. Age: Middle Oligocene. Plase (Korce) section.
- Fig. 2** - Microfacies-sandstone with *Nummulites fichteli*, X 15, transversal and oblique view. Age: Middle Oligocene. Valesh (Elbasan) section.
- Fig. 3** - Microfacies-sandstone with larger foraminifera, X 15, *Lepidocyclina (Eulepidina) dilatata* (Forma A+B) and *Victoriellinae*, bentonic foraminifera. Age: Upper Oligocene. Perroi i Gurrave (Tirane) section.





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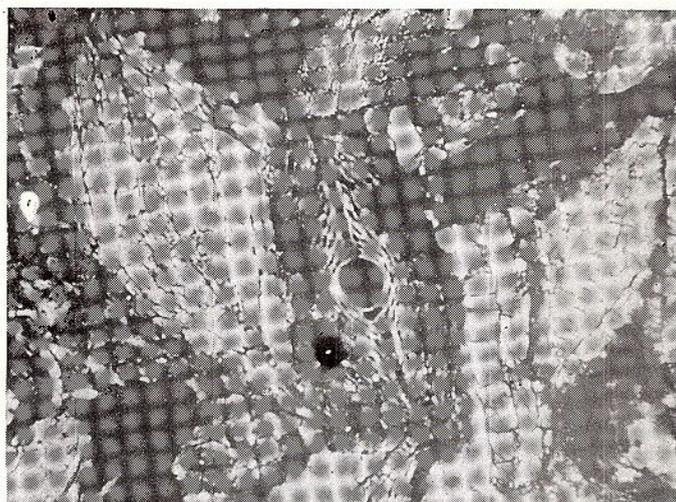


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Plate V

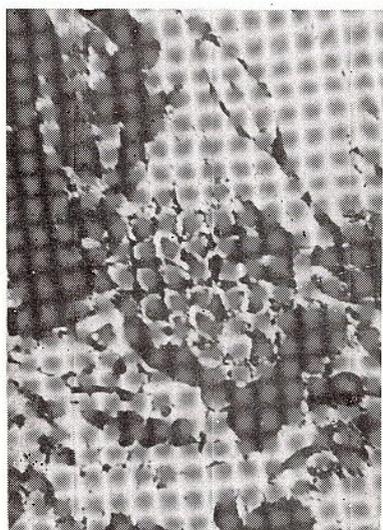
- Fig. 1** – Microfacies detritic limestones with *Lepidocyclina* (*Nephrolepidina*) spp., X 15. Age: Upper Oligocene. Paper-Murras (Elbasan) section.
- Fig. 2** – Microfacies detritic limestones with *Lepidocyclina* (*Eulepidina*) *dilatata*, *L.* (*Necrolepidina*) spp. and *Operculina*, *Victoriellinae*, bentic foraminifera, Bryozoa, etc. X 25. Age: Upper Oligocene. Sallak (Tirane) section.
- Fig. 3** – Microfacies with *Myogypsoides* (*Miogypsina*) *bantamensis* (equatorial view), *Operculina* sp., *Pararotalia* sp., X 25. Age: Upper Oligocene.
- Fig. 4** – Microfacies with *Miogypsinoidea* sp. (axial view), *Amphistegina* sp., *Operculina* sp., *Operculina* sp., *Lepidocyclina* (*Nephrolepidina*) sp., X 25. Age: Upper Oligocene. Sallak (Tirane) section.
- Fig. 5** – Microfacies with *Operculina* spp., bentic foraminifera, etc., X 25. Age: Upper Oligocene. Plase (Korce) section.
- Fig. 6** – Microfacies with *Pararotalia* spp., X 25. Age: Upper Oligocene. Kushtaj (Vlore) section.



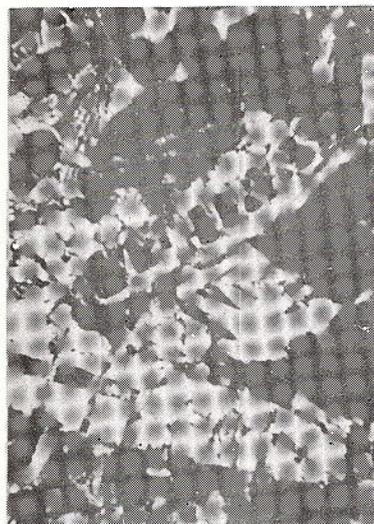
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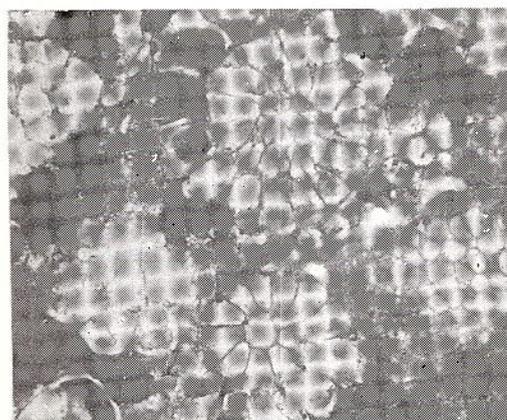
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DID THE MEDITERRANEAN DRY UP? - NEW INFORMATION FROM SEDIMENTATION PATTERNS AND PLANKTONIC FORAMINIFERA ASSOCIATIONS IN THE MESSINIAN OF SICILY

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Key words: Messinian. Pre-evaporitic deposits. Sicily. *Globorotalia margaritae*. *Globorotalia puncticulata*. Tripoli Formation.

Abstract: Recent work within the Sicilian Caltanissetta basin has established the existence of a complex ladder of thrust-top perched basins associated with thin-skin tectonics (Butler et al., 1995). Correlation between perched basins is difficult, however, first cycle evaporites were in place before the Palaeomediterranean basin-wide desiccation event. Post-desiccation onlap (culminating in the Pliocene, Trubi Formation maximum flooding surface) brought with it the diagnostic *Globorotalia margaritae* planktonic foraminifera community which traditionally has been used to zone the Pliocene of the Mediterranean. Our present work, however, indicates that considerable caution should be exercised in using these faunas as the southernmost Miocene perched basins in Sicily already contained the *Globorotalia margaritae* community early in the Messinian (i.e. before the first evaporites) In fact, *Globorotalia margaritae* and *Globorotalia puncticulata* occur in pre-evaporitic strata near Riesi town and further west at the Gallo d'Oro (Racalmuto) river section. The two species are semi-continuously present throughout the Messinian of the Riesi section. These new recordings cast uncertainty on the magnitude of the "Salinity Crisis" drawdown event, and on the validity of employing *Globorotalia margaritae*, *G. puncticulata* and *Sphaeroidinellopsis seminulina* as Pliocene biostratigraphic markers.

Introduction

The Late Miocene evaporite-related deposits of the island of Sicily are, in northern areas, developed above a predominantly siliciclastic Terravecchia Formation (Schmidt di Friedberg, 1962) built out into the northern margins of the Caltanissetta basin (Fig. 1). Further south they develop above deeper water, fully marine Tortonian marls, both around the western flanks of the Hyblean Mountains in SE Sicily (Rigo, Barbieri, 1959; Di Grande, Romeo, 1975), and further west around Agrigento. As regional tectonism increased towards the close of Tortonian time partial restriction of the areas south of the northern Sicilian fold belts occurred and led

ultimately to the development of evaporitic conditions throughout central Sicily.

It is generally considered in the literature that the diatomitic Tripoli Formation developed earliest. It frequently contains foraminifera and radiolarians in addition to diatoms, and fossil fish are also extremely common (see Decima, Wezel, 1971; Broquet et al., 1984; Grasso et al., 1990). This formation is closely succeeded by micritic to dolomitic carbonates of the Calcare di Base Formation which attain the thickest development in the north but generally have a ubiquitous distribution within the Caltanissetta basin (Decima et al., 1988; Pedley, Grasso, 1993). Frequently the formation is autobrecciated due to dissolution of intercalated evaporite beds and subse-



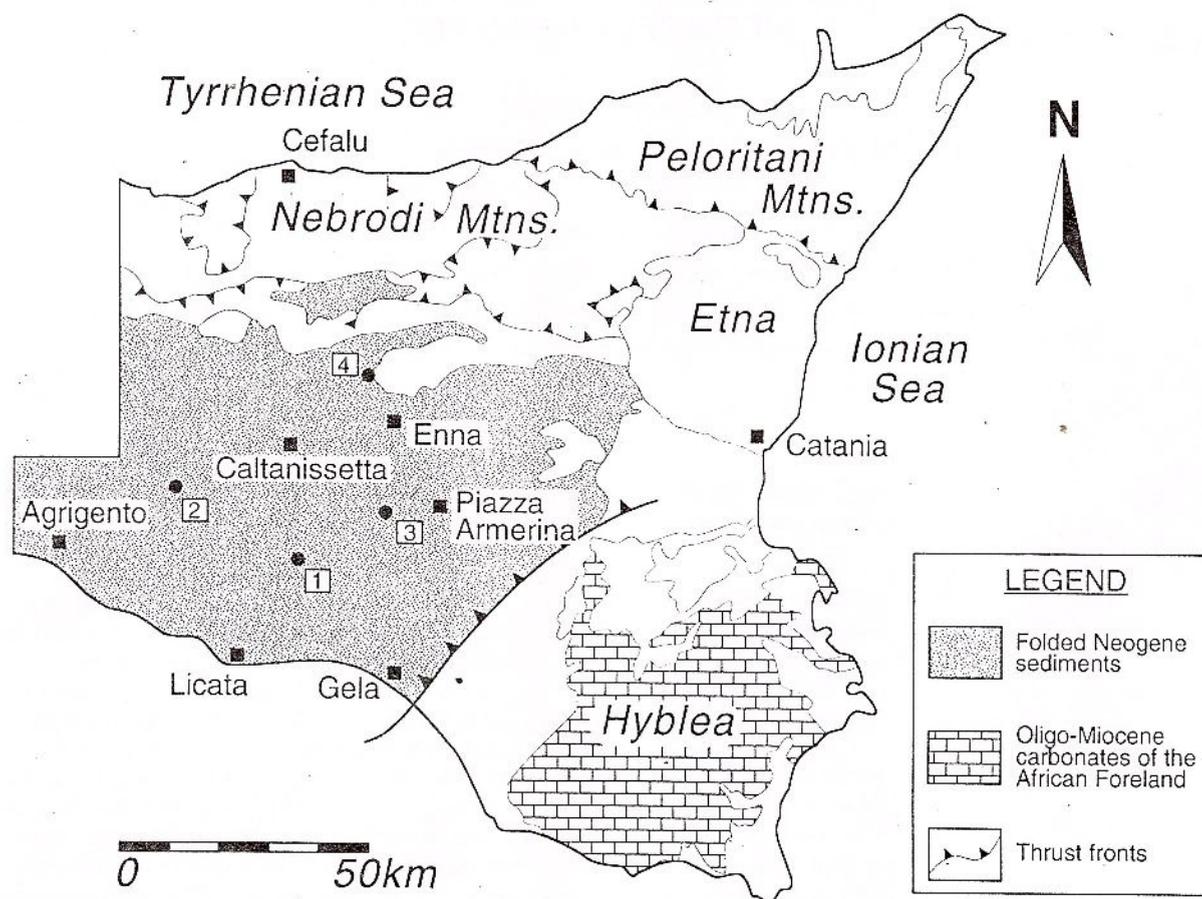


Fig. 1 - Geological sketch map to show the regional distribution of geological units surrounding the "Caltanissetta basin" and field sections figured in this work. 1, Riesi; 2, Racalmuto; 3, M. Torre; 4, Contrada Gaspa.

quent collapse of the part to fully lithified carbonates. Finally, these carbonates pass transitionally upwards into a dominantly evaporite complex of gypsum, halite and locally to iodine and potassium-rich salts, all intercalated with thin marl and clay bands (Ogniben, 1957; Decima, 1982). Decima and Wezel (1971) consider that these evaporites are subdivisible into a lower series (first cycle) mainly of *in situ* deposits, and an upper series (second cycle) dominated by evaporite resediments and terrestrial derived sediments. Collectively they are held to indicate widespread Late Miocene Mediterranean basin-wide desiccation associated with the "Messinian salinity crisis" (Cita, 1982).

Caltanissetta basin

Several hitherto unstudied field sections within the Tripoli Formation, Calcare di Base and associated

evaporites of central-south Sicily have been lithologically logged and important sections are indicated in Figure 1. All collected material has been sampled with extreme care at 1 m to 2 m intervals in order to avoid contamination from fallen debris within joints, neptunian dykes, and the effects of bioturbation. The field sections are located within syntectonic basins developed on the top of south verging thrust-sheets developed during the Afro-European collision event. These basins vary from tight "en echelon" complexes in north central Sicily to larger open basins in the external foreland region to the south. They are best considered as a diachronous association of perched basins (Butler et al., 1995). This view is compatible with earlier work which suggests that the Tripoli Formation is diachronous across the region (Bellanca et al., 1986; Richter-Bernburg, 1973).

Initial studies (Grasso et al., 1990; Pedley, Grasso,

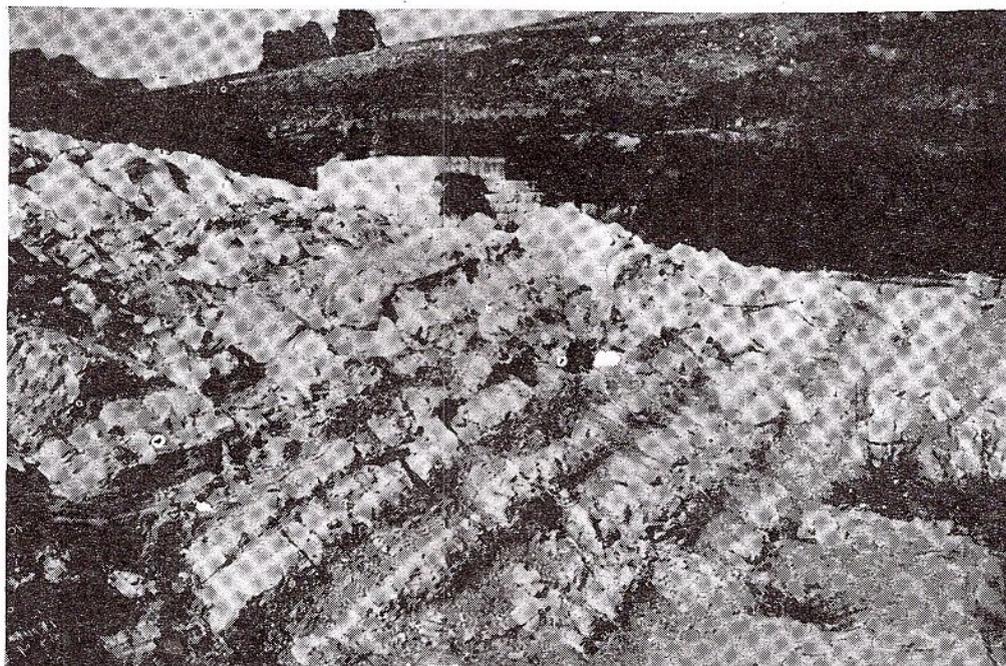


Fig. 2 - Field photograph of the Riesi section (Serra Pirciata), showing typical diatomite laminates low in the section (right) and more massive Calcare di Base lime mudstones towards the top (left hand side).

1993) focused predominantly on the north and central outcrops (e.g. Contrada Gaspa, see Fig. 1, site 4). Typically these contain relatively low diversity foraminiferal associations, dominated by near surface living planktonics associated with a limited range of benthic foraminifera (oligotypic assemblage of auct.). The planktonic foraminifera taxa are commonly dominated by *Neogloboquadrina multiloba*, *Turborotalia quinqueloba*, *Globigerinita glutinata* and *Globigerinita wuella*. Benthic foraminifera, although generally infrequent, are dominated by *Bulimina aculeata* and *Bulimina echinata*. These studies recognised cyclicity, well marked by lithological variability and faunal succession, within the foraminiferal population. This cyclicity was considered to have resulted from repeated local basin isolation and marine flooding events. These events appear to have been controlled by a combination of Palaeomediterranean eustasy and regional tectonism.

The present study considers the southern part of the "Caltanissetta basin" between the towns of Agrigento and Piazza Amerina (Fig. 1). Two recently studied localities, Serra Pirciata and Racalmuto, contrast with typical faunal associations from central and northern areas in that they contain many marl and diatomite-rich beds dominated by high diversity planktonic and benthonic foraminiferal associations, including some with mesopelagic planktonic foraminifera. This is all the more surprising on ac-

count of the association of the fossiliferous strata with interbedded evaporites. The Serra Pirciata section near Riesi town (see Fig. 2) well illustrates the interrelationships between microfaunas and pre-evaporitic sediments within this southern part of the so-called Caltanissetta Basin.

Riesi (Serra Pirciata) Section

This field section lies on the eastern limb of a tight isoclinal fold (Trabia-Tallarita syncline) crossed by the Riesi to Sommatino road at about 5.5 km to the west of Riesi town. The sub-vertical strata lie at the bottom of the Imera river gorge. Figure 3 illustrates the lithological log of this section. Detailed structural field mapping carried out around the sample site by Drs H. Lichorish and R. W. Butler (pers. comm.) conclusively demonstrates that the sample section is uncomplicated by thrusting, slides or other phenomena which might lead to indiscriminate mixing of faunas during Pliocene, or indeed, during recent times.

In many ways the sequence is typical of the Sicilian Messinian as it commences above dark grey Messinian Terravecchia Formation clays containing a normal fully marine microfauna. The succeeding Tripoli Formation is dominated by diatomaceous laminates with grey clay alternations and contains scattered fragmentary and entire teleost fish remains. At the top of the formation several beds show internal contorted laminae and a thin autobreccia is present.

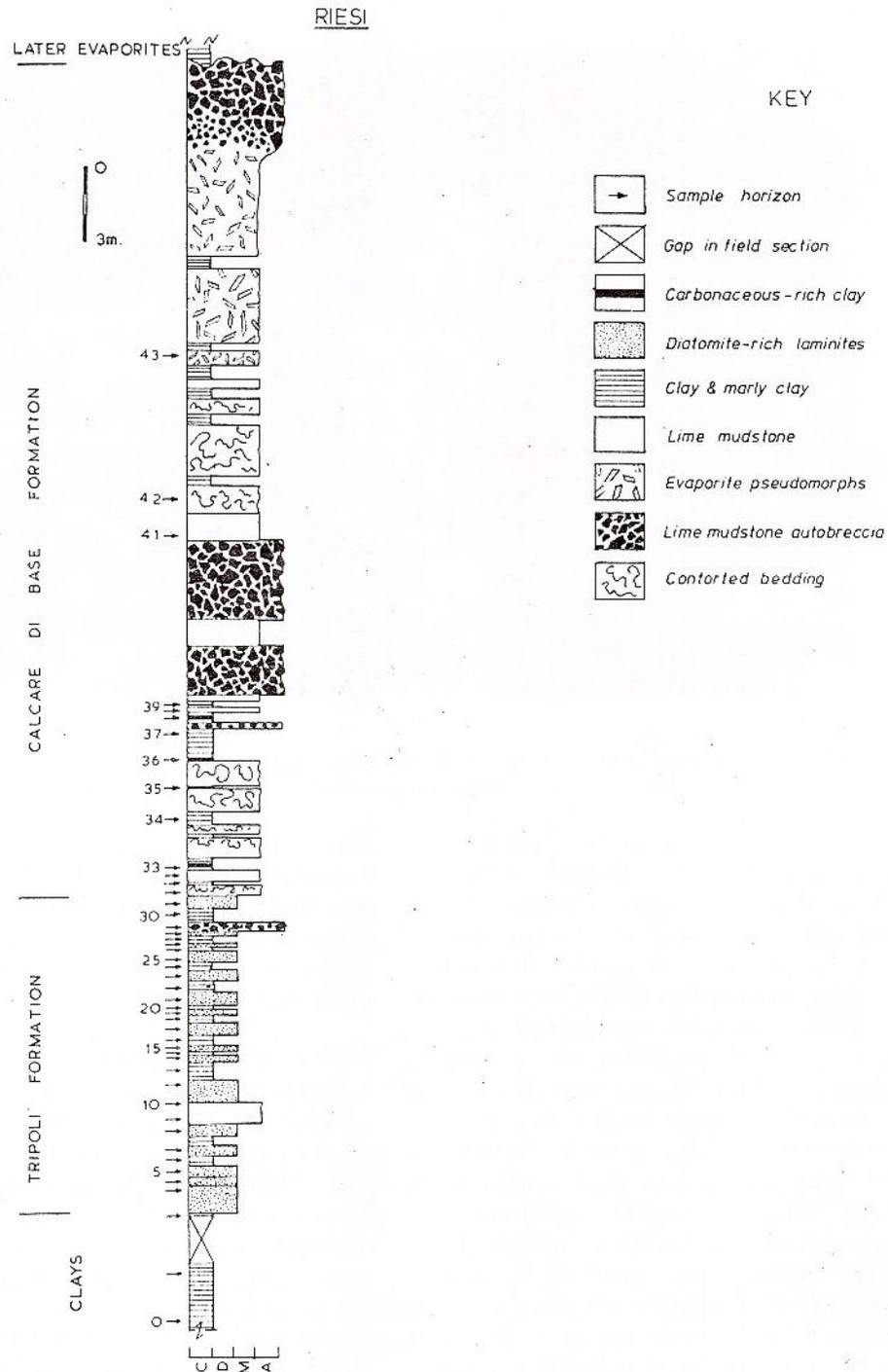


Fig. 3 - Lithological log of the Riesi section at Serra Pirciata. The numbers down the left-hand side of the column refer to sampling locations indicated in Figures 4 and 5.

These latter beds mark the transition up into the overlying Calcare di Base Formation.

The Calcare di Base typically is an evaporitic limestone formation composed predominantly of autobrecciated lime mudstones, together with lime mudstones showing soft sediment deformation fabrics associated with evaporite dissolution. No evaporites

crop out in the section, however, their former presence is confirmed by gypsum pseudomorphs present in the upper beds. The few thin grey clays intercalated with the carbonates may in part represent insoluble residues remaining after local evaporite dissolution. The sedimentology is considered in detail in Pedley and Grasso (1993).

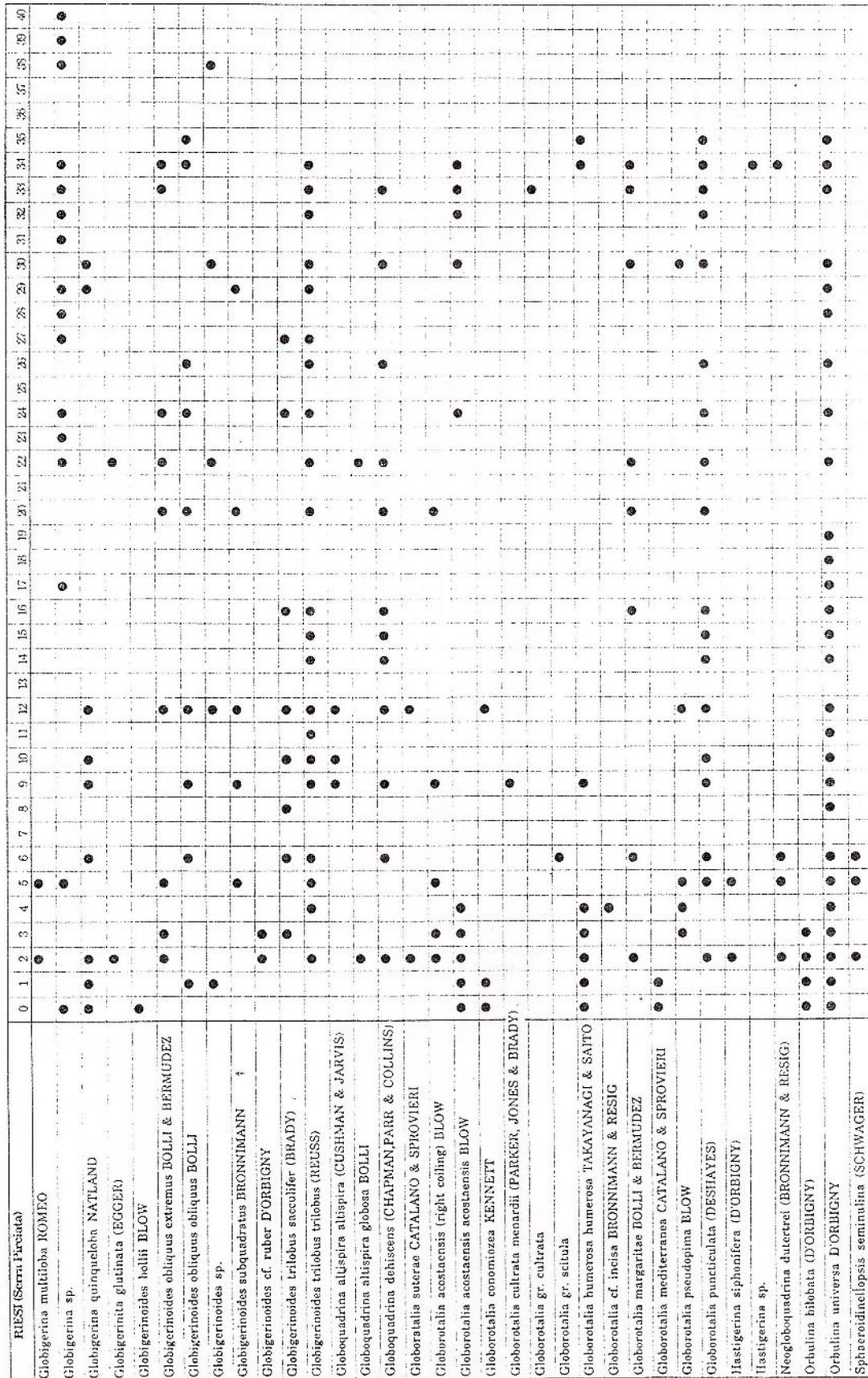


Fig. 4 - Planktonic faunal distribution ranges of the Riesi (Serra Pirciata) section. N.B. *Globorotalia margaritae*, *G. puncticulata* and *Sphaeroidinellopsis semimulina* occur throughout the Tripoli Formation and in the pre-diatomaceous clays at the base.

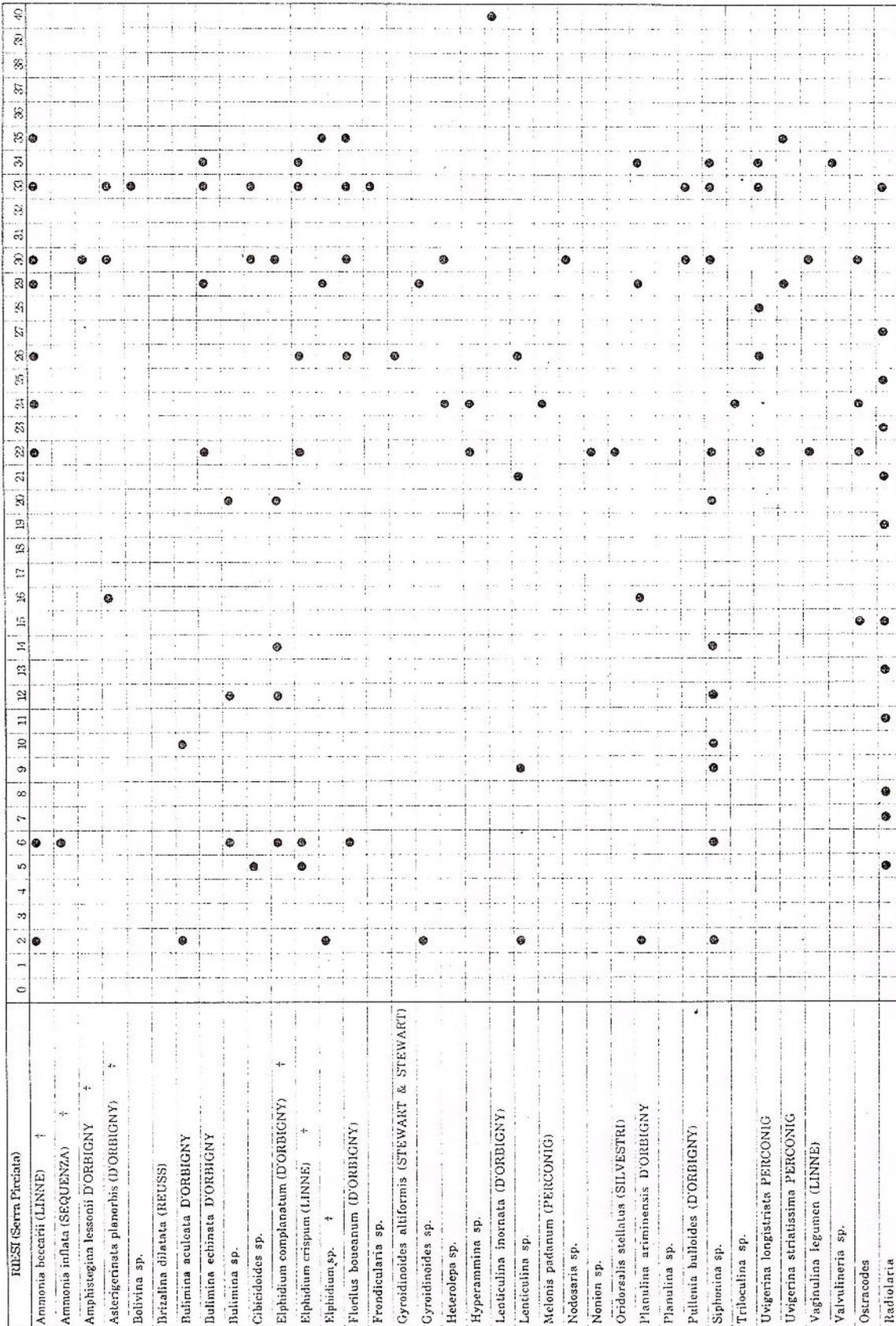


Fig. 5 - Benthonic foraminifera range chart for the Piesi (Serra Pirciata) section. Note the diversity of species compared with published lists from other Messinian sections in the caltanissetta region. N.B. species with a cross against them are reworked.

Microfauna

Microfaunal range charts of the Serra Pirciata section are provided in Figures 4 and 5. Forty-one detailed micropaleontological analyses of diatomaceous laminites, grey clays and lime mudstones from this section were carried out. The samples showed the occurrence of a rich and diverse planktonic and benthonic foraminifera, diatom, radiolarian, ostracode and fish assemblage. Foraminifera were well preserved with only a few reworked benthonics being present. The lowest part of the section (samples 0 and 1 in Fig. 4) is characterised by a typical Messinian association of *Globorotalia conomiozea*, *G. mediterranea*, *G. acostaensis*, *G. humerosa*, *Globigerina quinqueloba*, *Orbulina* sp. and *Globigerinoides* spp.

We emphasise the presence of planktonic and benthic communities with high variability and in particular the planktonic elements are represented not only by forms typical of surface waters (e.g. *Orbulina universa*) but by deeper water, mesopelagic, foraminifera such as *Globorotalia margaritae*, *G. puncticulata* and *Sphaeroidinellopsis seminulina*. *Globigerinoides* spp., *Globorotalia acostaensis acostaensis* (sinistral and dextrally coiled), *G. humerosa*, *G. incisa*, *Globiquadrina altispira* are also present. The benthics are also characterised by high diversity and abundance and by relatively few reworked individuals.

Age of the Tripoli Formation deposits at Serra Pirciata

Previous foraminiferal studies (see discussion in Pedley and Grasso, 1993) confirm that the Tripoli Formation of Sicily commenced within the *Globorotalia mediterranea* Subzone of the Messinian *Globorotalia conomiozea* Zone (Colalongo et al., 1979; Iaccarino, 1985) and continued to accumulate for much of the succeeding Late Messinian *Globorotalia multiloba* Subzone.

The Riesi (Serra Pirciata) section, however, differs significantly from the other described accounts of comparable lithologies in measured sections within the Caltanissetta region (e.g. Contrada Gaspa, Grasso et al., 1990; Falconara, D'Onofrio et al., 1975), it is richer in absolute numbers of individuals both of planktonic and benthic foraminifera. Secondly, samples are much more diverse in terms of actual species. For example, the Contrada Gaspa section (Grasso et al., 1990) contains less than ten planktonic species and nine benthics at its most diverse whereas the Riesi section contains up to nineteen planktonic species and up to thirteen benthic species (see Figs. 4 and 5 in Pedley and Grasso, 1993).

In particular, the remarkable occurrence of three

species of foraminifera (*Globorotalia margaritae*, *G. puncticulata* and *Sphaeroidinellopsis seminulina*) at (Serra Pirciata) Riesi, and at Racalmuto (see later) requires detailed consideration. Typically these are part of a deeper water planktonic community which, it is generally believed, does not appear until the commencement of deep water chalk deposition (Trubi Formation) in Pliocene times. Nevertheless, on the basis of the typical Late Miocene planktonic foraminifera present in association with the problem species we have no hesitation in concluding that the Tripoli Formation of Serra Pirciata is of Messinian age.

Previous records in Sicily

Globorotalia margaritae has been recorded previously in the Upper Miocene both of the Atlantic and Palaeomediterranean (Adams, 1976; Montenant et al., 1976; Benson and Rakic el Bied, 1991; Sierro et al., 1993), and there is growing evidence to demonstrate that its first appearance datum (FAD) in the Atlantic is diachronous. The Palaeomediterranean occurrences, recorded from Sicily by previous workers, appear to be from the highest beds within the Messinian. Of special consideration is the occurrence of both *Globorotalia margaritae* and *G. puncticulata* in the Arenazzolo (topmost Messinian) of Capo Rossello and Eraclea Minoa in south Sicily (Brotsma, 1978). Colalongo (1968) also recorded the occurrence of *Globorotalia margaritae*, *G. puncticulata* and *Sphaeroidinellopsis seminulina* in the uppermost Messinian of the Capodarso-Pasquasia section in central Sicily "within clays above the Second Cycle evaporites".

Our new recordings strengthen these earlier Sicilian records and push back the first Sicilian appearances of these three planktonic species within the palaeomediterranean to a pre-salinity crisis situation.

This recording of the earliest occurrence of *Sphaeroidinellopsis seminulina* in the Late Miocene of southern Sicily (early in the Messinian at Riesi) is also interesting as the *Sphaeroidinellopsis*. Acme Zone (Cita, 1973, 1975) is considered by many to be Zanclean (Early Pliocene), see Iaccarino (1985). However, the species has previously also been reported from the Upper Miocene in both Atlantic (Salvatorini and Cita, 1979; Weaver and Raymo, 1989) and the Mediterranean and therefore appears, as with *Globorotalia margaritae*, that the FAD of *Sphaeroidinellopsis seminulina* is also likely to be diachronous.

The principal bone of contention with previous workers is the occurrence of *Globorotalia puncticulata* within the Late Miocene, which can be seen clearly in Figure 4 to occur abundantly through-



out the entire Tripoli Formation of the Riesi section. This is particularly surprising as evolutionary trends of *Globorotalia puncticulata* have been demonstrated in both Atlantic and Pacific oceans (Kennett, 1973; Berggren, 1977; Wei and Kennett, 1988). Moreover, the first appearance of *Globorotalia puncticulata* in the Atlantic and Mediterranean is dated not before Early Pliocene (4.15 to 4.55 my), according to Weaver and Raymo (1983), Rio et al. (1984).

Confirmation of the new Sicilian occurrences comes from an additional field section along the Gallo d'Oro river, some 3 km north of the town of Racalmuto, and in the adjacent Racalmuto mine. Both these locations lie over 30 km west of Riesi. The river section has yielded abundant unworked individuals of all three species commencing low in the Tripoli. In addition, the problem species also occur in association with other planktonic foraminifera within later Messinian clays sandwiched between evaporites towards the top of the section. A fully marine planktonic foraminiferal assemblage with *Sphaeroidinellopsis* has also been retrieved from evaporite associated clays lying beneath 430 m of massive evaporites within the Racalmuto mine.

The problem is that these occurrences of fully marine and deeper water planktonic foraminifera occur in Late Messinian strata. Clearly they predate the Pliocene transgression, and indeed indicate that all three problem species were present within the Mediterranean region prior to the "Salinity Crisis". Although both *Sphaeroidinellopsis* and *Globorotalia margaritae* may with difficulty be accommodated into existing Messinian distribution models of foraminiferal occurrence given that the perched basins which contain them were very late in development, current distribution models cannot cope with the appearance of *Globorotalia puncticulata* before the onset of the Pliocene.

Iaccarino (1985) noted that the first appearance of *Globorotalia puncticulata* within the Mediterranean Basin could reflect a later migration into warmer waters after its initial evolutionary appearance in the cooler waters of the nearby Atlantic Ocean, thus suggesting a diachronous FAD for this species. We find great difficulty, however, in accepting the presence of this foraminifer as unequivocal evidence in favour of a basal Pliocene age as defined in the biostratigraphic scheme of Cita (1975 b).

Significance of a *Globorotalia-Sphaeroidinellopsis* community in classic Messinian lithologies

The strata containing these faunas are well below the first appearance of evaporites, in sediments containing a diverse range of other marine organisms.

Furthermore, the associations continue throughout the later evaporites in the Racalmuto sections.

The basin ladder model of Butler et al. (1995), adopted for our palaeoenvironmental interpretations, provides a predictive frame for better understanding our foraminifera distributions. It forecasts that the youngest, and least deformed basins will be furthest from the northern Sicilian fold-belts (i.e. will only be found in the southern part of the Caltanissetta basin), and will lie adjacent to the remaining fully marine Palaeomediterranean Sea.

Brief marine incursion into these southern semi-enclosed basins will provide a window on faunal developments within the Palaeomediterranean generally. From this it becomes clear that *Globorotalia margaritae* and *G. puncticulata*, in association with a rich associated planktonic microfauna, appear repeatedly within the Serra Pirciata basin throughout the Messinian episode. To do so they must already have been established within the Palaeomediterranean during the initial offlap stage of the Early Messinian (i.e. early in the first cycle).

From our data we suggest that both *Globorotalia margaritae* and *Sphaeroidinellopsis seminulina* migrated into the Palaeomediterranean during the earlier Messinian (Late Miocene) shortly after evolving in the open ocean (cf. Zachariasse and Spaak, 1983). They may have entered the Palaeomediterranean as a result of climatic cooling and concomitant expansion of the cool eastern Atlantic boundary current which permitted southward migration of the species. Once present within the Palaeomediterranean basin these faunas either remained resident, or were regularly reintroduced from the world oceans, throughout the Sicilian evaporite episodes.

A somewhat greater mystery is apparently attached to *Globorotalia puncticulata* with a first appearance datum low in the Tripoli Formation at Riesi; yet, according to Weaver and Raymond (1983), and Rio et al. (1984), the species does not appear before Early Pliocene times in the Atlantic. Further work, especially magnetostratigraphic studies, are needed within the Mediterranean to resolve this problem in order to fix precise age constraints for the Messinian-Pliocene boundary.

Conclusions

This study confirms observations by earlier workers that *Globorotalia margaritae* and *Sphaeroidinellopsis seminulina* are present generally within the classic Messinian successions of southern Sicily. The present work, however, demonstrates clearly that they occur at some southern Sicilian sites in pre-evaporitic diatomites and in clays sandwiched between evaporitic



limestones.

The presence, at multiple levels, of these and other planktonic and benthic species in the Messinian of southern Sicily demonstrates conclusively that a body of normal marine water existed within the confines of the Mediterranean throughout the Neogene evaporative event (Salinity Crisis). This conclusion strengthens current hypothesis that most of the Sicilian Messinian evaporitic strata were deposited in coastal embayments adjacent to a large relict, and possibly normal marine Palaeomediterranean water body.

Of paramount importance to this hypothesis, however, is the discovery of abundant *Globorotalia puncticulata* within these Messinian sediments. Elsewhere in the world the first appearance datum for *Globorotalia puncticulata* consistently post-dates the Upper Miocene. There is now an urgent need to re-examine and seek out other sections which might also show these same associations, and to modify the concept of total desiccation of the Palaeomediterranean during the latest Messinian.

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RELATIONSHIPS AND CORRELATION OF THE PLIOCENE FAUNAS OF THE DACIC BASIN, ROMANIA

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Key words: Pliocene. Mammalian associations. Faunal events. Biostratigraphy. Magnetostratigraphy. Dacic Basin (Paratethys). Romania.

Abstract: Pliocene mammalian faunas of the Dacic Basin span a time interval which ranges from below the Cochiti event of the Gilbert paleomagnetic epoch, at the base (Berești, southern Moldova, Early Pliocene, Late Dacian), to the Olduvai event of the Matuyama epoch at the top (Tetoiu-Valea Grăuceanului, Olteț valley, Oltenia, Late Pliocene, Late Romanian). Several faunal events mark the first (first appearance datum FAD) or last (last appearance datum LAD) occurrence of a characteristic taxon generally within a continuous stratified sequence of deposits of known magnetic polarity. The Early Pliocene is characterized by the presence of *Mimomys moldavicus* at Mălusești (southern Moldova) and *Mimomys rhabonensis* at Lupoaia and Drănic - 0 (Oltenia, Cochiti event). The Middle Pliocene is characterized by the successive occurrence in Oltenia of *Stephanorhinus elatus* (Covrigi, Early Gauss), *S. etruscus* (Mătășari-Brădețu, late Early Gauss, event consecutive to a climatic depression) and "*Mammuthus*" *rumanus* (Tuluțești, southern Moldova, Cernătești, Oltenia, Middle Gauss subchron). The Cernătești fauna and its equivalent of small mammals at Podari are indicative of a warm-temperate climate. Occurrence of *Mimomys stehlini/minor* (Milcovu din Vale, Oltenia, end of the Late Gauss). During the second part of the Late Pliocene (late Early Matuyama, Oltenia) occur dry steppe elements such as *Borsodia* (Slatina - 2) and *Alactaga* (Cherlești-Moșteni); extinction of *Dolomys* lineage and appearance of *Mimomys* ex gr. *plioaenicus*. The Late Pliocene macromammals include at Valea Grăuceanului girafid (*Mitilanotherium*), pangolin (*Manis*), primate (*Paradolichopithecus*) indicating warm-temperate climatic conditions. A few "pebble tools", found at Dealul Mijlociu (Tetoiu area), seem to be linked to a human activity at about 1.7-1.8 Ma.

1. Introduction

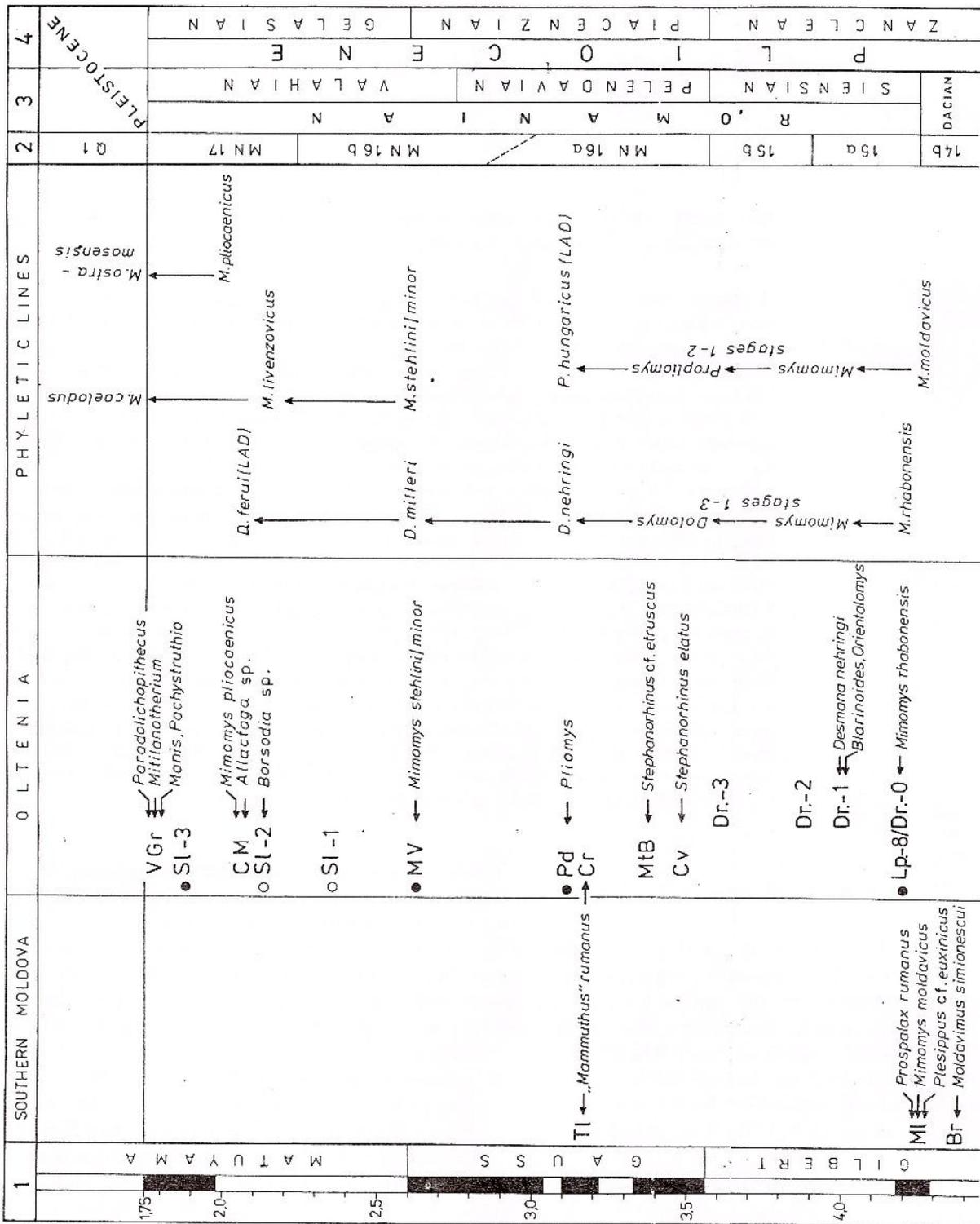
Paleontological investigations carried out since 1960 on the Pliocene fluvio-lacustrine deposits of the Dacic Basin supplied a rich and various material of macro- and micromammals indicating a series of biological events which appear to be of stratigraphic importance. In the two last decades the knowledge of small mammal associations has been considerably increased (Feru et al., 1978, 1979; Rădulescu & Samson, 1983, 1987; Rădulescu et al., 1993, 1995).

An overview of the Pliocene mammals of the Dacic Basin showed that the paleontological record is distinguished by certain discontinuities which can sometimes be interpreted in terms of abrupt changes due, on the one hand, to extinctions (last appearance datum LAD) and to the appearance of new types of animals (first appearance datum FAD) on the other.

In some sequences of deposits, where the paleontological record comes from numerous successive fossiliferous levels, some micromammalian lineages, especially of rodents, are characterized by a gradually progressing evolution. In contradistinction, several abrupt changes of the faunas can be explained by migrations and first appearance (FAD) of new invading types of mammalian species.

We adopted a three-fold subdivision of the Pliocene according to the chronostratigraphy of the Mediterranean area (Tethys). In this framework, the Early Pliocene corresponds to the Zanclean and most part of the Gilbert paleomagnetic epoch; the Middle Pliocene covers the Piacenzian and the Gauss paleomagnetic epoch, whereas the Late Pliocene is equivalent to the Gelasian and to the first part of Matuyama paleomagnetic epoch consisting of the Early Matuyama subchron and Olduvai event above which the Pliocene/Pleistocene boundary is deline-





Selected fossiliferous sites of Pliocene age in the Dacic Basin, Romania, based on specific events of Mammalian evolution. 1, magnetic polarity time scale; 2, Mediterranean Neogene mammal zones; 3, Central Paratethys stages; 4, Mediterranean stages. Br=Berești; CM=Cernătești; Cv=Covrigi; Dr=Drânc; Lp=Lupoaia; MI=Mălășeni; MtB=Mătășari-Brădețu; MV=Mîlcovu din Vale; Pd=Podari; SI=Slătina; TI=Tulucești; VGr=Valea Grănceanului.

ated. We indicated in the Table the equivalence between the Tethyan chronostratigraphic subdivisions and the biostratigraphic scheme of the Dacic Basin (Paratethys) established on the basis of molluscan faunas (Andreescu, 1982; Andreescu et al., 1981; Andreescu et al., 1986).

A correlation of the mammalian faunas with the biozonation of the Mediterranean Neogene (MN zones) (Mein, 1990) was also undertaken. Calibration of the mammalian assemblages is based on paleomagnetic determinations, correlation with molluscan faunas and evolutionary stages of mammals themselves. The faunal events discussed in this paper apply only to the Dacic Basin. It is worth mentioning, however, that similar faunal events have also been registered in other stratigraphic sequences in various regions of Eurasia.

This report is concerned with the Pliocene faunas of southern Moldova (particularly with the classical fossil sites of Berești, Mălușteni and Tulucești, located in the northeastern portion of the Dacic Basin and known from the beginning of this century) (Athanasiu, 1915; Simionescu, 1922, 1930, 1932) and Oltenia (fossil sites located in the valley of the Jiu and Olt Rivers and their tributaries in the western portion of the Dacic Basin) (Andreescu et al., 1981; Feru et al., 1983; Rădulescu et al., 1993, 1995).

Fossiliferous sequences of deposits examined here span a time interval of about 2.5 million years (Ma), from more than 4 Ma for the Berești fauna (assigned to the Late Dacian, equivalent to the late part of the MN - 14 zone) to about 1.8 Ma (age of the association from Valea Grăunceanului in the Olteț valley, Oltenia, belonging to the Late Romanian, equivalent to the late MN - 17 zone).

It is worth remembering that several faunal events correlate relatively well with periods of cooling and/or dryness having, very probably, a global significance. Processes of in-place evolution of morphogenetic lineages, especially among rodents, seem to be a response to various environmental constraints.

2. Moldova

2. 1. Early Pliocene

2.1.1. MN 14 zone. The Berești fauna, assigned to the Late Dacian, includes great amounts of leporids (*Trischizolagus dumitrescuae* RĂDULESCU & SAMSON) and ochotonids (*Ochotona ursui* SIMIONESCU, *Pliolagomys* sp.) corresponding to the so-called "lagomorphic complex" of eastern and southeastern Europe (Schevčenko, 1965). The absence of arvicolids is to be noted, the faunal assemblage from Berești spanning a time interval preceding the im-

migration of the first *Mimomys* forms (*M. moldavicus* type) and postdating the arrival of *Prosomys* (= *Promimomys*) *insuliferus* (KOWALSKI) in south-eastern Europe and Aegeo-Anatolian areas. Taking into account the evolutionary stage of most species and the absence of archaic *Mimomys* forms, we assigned the Berești fauna to the end of MN-14 (Rădulescu & Samson, 1989, 1995). It should be noted that the Berești rodent fauna includes, among other species, representatives of genera *Pliospalax* [*P. macoveii* (SIMIONESCU)] and *Microspalax* (*M. odessanus* TOPACHEVSKY). A characteristic species of the Berești assemblage appears to be a small-sized primitive cricetid species *Moldavimus simionescui* (SCHAUB) showing close relationship to the vast *Cricetulus* group (Samson & Rădulescu, 1973).

Trischizolagus represents an important component of the Berești fauna. The representatives of the genus were undoubtedly of southern (Mediterranean and Asiatic) origin and they were spread from Spain in the west to Afghanistan in the east. The occurrence of *Trischizolagus* at Pul-e-Charki (Afghanistan) is, for the time being, controversial, although the material from this locality, interpreted as belonging to the genus *Serengetilagus* (Sen & Erbajeva, 1995) reveals many affinities with primitive forms of *Trischizolagus*. Besides the Mediterranean area, *Trischizolagus* is known not only from the Dacic Basin, but also from eastern adjacent areas (the Republic of Moldova, southern Ukraine) (Averianov, 1995). To the north and west *Trischizolagus* is replaced by *Hypolagus* (Samson & Rădulescu, 1973) which was also spread in eastern Europe and Asia.

Among the macromammals, a big antelope, documented by a fragment of a lower molar, was recently identified as "*Parabos*" cf. *athanasiui* (Simionescu, 1922). This species was not mentioned in the preceding faunal list of Berești.

The Berești faunal assemblage shows great affinities with the "lagomorphic complex" of the Republic of Moldova and southwestern Ukraine on the one hand and with the fauna of the Aegean region on the other, particularly with the mammalian assemblage at Megalo Emvolon (Karaburun) (Macedonia, Greece) which contains *Trischizolagus dumitrescuae*, *Microspalax odessanus*, *Dolichopithecus rusciniensis* DEPÉRET and *Parabos macedoniae* ARAMBOURG & PIVETEAU (a big antelope very similar to "*P.*" *athanasiui* from southern Moldova).

On the basis of mammals (association of the genus *Parabos* with *Sus minor* DEPÉRET), Megalo Emvolon is generally situated within MN-15 zone, although according to pollen data it belongs to MN-14 zone (Benda & Meulenkamp, 1990). Despite of some non-negligible similarities, the Berești mammalian com-



plex could antedate the Megalo Emvolon fauna or it may correspond to the earliest (MEV) of the three different fossiliferous levels which have been recognized at Megalo Emvolon (Koufos *et al.*, 1991).

As a whole, the Berești fauna indicates that the climate was predominantly of continental type, in contrast to the more humid conditions prevailing in the Early Ruscinian (beginning of MN-14).

2.1.2. MN-15 a subzone. A renewal in the fauna is marked at Mălușteni by the presence of particular species of rodents such as *Mimomys moldavicus* KORMOS and *Prospalax rumanus* SIMIONESCU, which are here considered as northern immigrants. The arrival of *Mimomys* (FAD), undoubtedly from Asia was, very probably, coincident with a climatic depression (a cooling at global scale associated with more humid conditions than those at Berești). In this connexion it should be remembered that the limited number of arvicolids in the Mediterranean Ruscinian (MN-15) seems to have a climatic cause. The climate was too warm for voles (van der Meulen & Kolfshoten, 1986), the family Arvicolidae thriving best in the temperate zone. The penetration of *Prospalax rumanus* (FAD), known mainly from Middle European zones, to southeastern Europe may also indicate a climatic depression.

M. moldavicus from Mălușteni (Rădulescu & Samson, 1989) seems to have its counterpart in the vole material (mentioned as *Promimomys moldavicus*) at Grebeniki-2, a locality on the right bank of the Kuchurgan valley (Odessa region, Ukraine) (Vangengeim *et al.*, 1995). Grebeniki-2 is placed at the end of MN-14 (Pevzner *et al.*, 1996). In our opinion, the first appearance of primitive forms of *Mimomys* may be used as a marker for the beginning of MN-15.

Mălușteni is also the type-locality for the big antelope "*Parabos*" *athanasiui* represented by a damaged skull and various skeletal remains belonging to one individual (Simionescu, 1922, 1930). This antelope derives, very probably, from an Anatolo-Aegean stock which is poorly known. A close related species is represented at Megalo Emvolon by ? *Parabos macedoniae*. According to Gromolard (1981) the generic status of the species *macedoniae* is questionable.

Simionescu (1930) described and figured (l.c. Fig. 42) from Mălușteni a juvenile maxillary belonging to a large-sized horse of stenoid type. This material was attributed to *E. cf. robustus*. Samson (1976) assigned the same specimen to *Plesippus (Allohippus) cf. euzinicus* SAMSON. Macarovici (1972) considered that the Mălușteni horse was not contemporaneous with the bulk of the fauna of this locality. He suggested that the horse material could come from a later horizon, the so-called "Bălăbănești gravels". According to Ghenea (1968) the "Bălăbănești gravels" are ab-

sent from the Mălușteni area. Taking into account the aspect of the horse maxillary which does not differ from the rest of the fossils from Mălușteni, we are inclined to maintain a species of monodactyle horse in the association of this locality. Tentatively, *Plesippus cf. euzinicus* might belong to the first migratory wave of monodactyle horses coming from North America. This event could be dated to about 4 Ma.

2. 2. Middle Pliocene

2.2.1. MN-16a subzone. The "*Mammuthus*" *rumanus* (Ștefănescu, 1924) FAD marks the presence at Tulucești of a very primitive elephant in association with *Mammul borsoni* (HAYS), *Anancus arvernensis* (CROIZET & JOBERT), *Paracamelus* and *Cervus cf. perrieri* CROIZET & JOBERT (Athanasiu, 1915; Ștefănescu, 1924; Ghenea & Rădulescu, 1964; Samson, 1976). In addition, *Plesippus (Allohippus) euzinicus* occurs at Oasele, a neighbouring locality containing the same fossiliferous level (Samson, 1976).

The type-specimen of "*M.*" *rumanus*, a fragment of a last lower molar, is characterized by primitive diagnostic features: enamel thickness up to 5 mm, very low number of plates (5 plates + 1 talon for a length of the crown of 160 mm), maximum width 85 mm (Athanasiu, 1915, Table 17, Fig. 4; Ștefănescu, 1924, Plate).

In the Pliocene stratigraphic context of the Dacic Basin, the arrival of "*M.*" *rumanus* from Africa may be situated within the middle Gauss normal subchron.

3. Oltenia

3.1. Early Pliocene

3.1.1. MN-15a subzone. In the northwestern portion of Dacic Basin (Jiu-Motru interfluvium), *Mimomys rhabonensis* RĂDULESCU, SAMSON & ȘTIUCĂ FAD marks the presence of a large-sized primitive *Mimomys* at Lupoia Coal Quarry (fossiliferous level at the base of coal layer VIII) (Rădulescu *et al.*, 1989).

Paleomagnetic determinations showed that Lupoia/VIII can be situated within the Cochiti normal event and coal layers IX to XI belong to the Late Gilbert subchron. The Gilbert/Gauss boundary was placed above coal layer XI, whereas coal layers XI to XIV were assigned to the Early Gauss subchron (Andreescu *et al.*, 1986).

One of the most complete and interesting sequence of deposits containing fossiliferous levels rich in micromammalian remains is known at Drănic in the Jiu valley (Rădulescu *et al.*, 1993; 1995). The fossiliferous levels can be grouped into four main divisions from Drănic-0 to Drănic-3.



The lowest level with micromammals, called Drănic-0, yielded an association similar to that at Lupoia/VIII, both sites containing *Mimomys rhabonensis* in the same evolutionary stage (Rădulescu et al., 1995). The origin of *M. rhabonensis* is still unknown, but this species seems to represent an eastern (Asiatic ?) immigrant confined to eastern and southeastern Europe (Dacic Basin, the Republic of Moldova, southwestern Ukraine). At Drănic-0 is also present a representative of the *Mimomys moldavicus* group in association with *Apodemus dominans* and six insectivores among which *Allosorex stenodus* FEJFAR and *Blarinoides mariae* SULIMSKI are to be noted. *Allosorex stenodus* appears to be a species with Central European affinities (the type-locality for the species is Ivanovce in Slovakia (Fejfar, 1966)). *Allosorex* is recorded at Drănic-0 for the first time in Romania. We mentioned this particular taxon taking into account its extremely rareness.

Blarinoides mariae FAD marks the lowest stratigraphic occurrence of this species in the Dacic Basin. *Blarinoides* is also present at Drănic-3 and Podari where its last appearance was registered.

At Drănic-1 *M. rhabonensis* is also associated to *M. moldavicus*. It is interesting to remember that the former species is characterized by the presence of dolomyian morphotypes, whereas the latter displays propliomyian features in the dental structure (Rădulescu & Samson, 1996; Rădulescu et al., 1995). Some other elements such as *Desmanella* sp. among insectivores and a small castorid among rodents are also characteristic of this stratigraphic level. *Desmanella* belongs, very probably, to an Aegean-Anatolian stock which spread northward reaching the latitude of the Dacic Basin.

The faunal assemblage from Drănic-1 is constituted of three distinct fossiliferous levels. The lower level reveals many similarities to Drănic-0 micromammals. The middle level is distinguished by the arrival of some new taxa such as *Miocchinus* sp. (Insectivora, Erinaceidae) and *Orientalomys* sp. (Rodentia, Muridae). The former taxon is practically a relict form showing affinities with the Late Miocene fauna of Anatolia; the latter one, possessing primitive features, is also indicative of Anatolian relationships. The highest level yielded an impoverished fauna in which *Desmana nehringi* KORMOS was identified.

Desmana nehringi FAD marks the earliest occurrence of this taxon in the Drănic section. The species is considered a northern guest, its arrival indicating, very probably, a climatic depression. At Drănic-1, *D. nehringi* replaces *D. amutriensis* RĂDULESCU, SAMSON & ȘTIUCĂ known in Oltenia from Drănic-1 (lower level), Drănic-0 and Lupoia/VIII.

3.1.2. MN-15b subzone. The Drănic-2 fauna is dis-

tinguished by the first appearance of *Galemys sulimskii* RUMKE and *Deinsdorfia* cf. *kordosi* REUMER. As a whole, this faunal assemblage includes also *Desmana nehringi* and an unidentified species of the genus *Ruemkelia*. The large-sized voles of *Mimomys rhabonensis* type show a progressive accentuation of the *Dolomys* patterns associated with an increase in hypsodonty. The medium-sized mimomyian specimens, allocated to *Mimomys moldavicus*, display frequently a propliomyian aspect.

3.2. Middle Pliocene

3.2.1. MN-16a subzone. At the Gilbert/Gauss boundary, the micromammals of the Dacic Basin, as indicated by the Drănic-3 faunal assemblage, are characterized predominantly by an endemic evolution, the most part of the larger voles of the *Mimomys/Dolomys* lineage having already acquired a dolomyian morphology (especially in the structure of the last upper molar).

The macromammalian community during the Early Gauss subchron is distinguished by the appearance of new invading types of animals. The Covrigi faunal unit, situated within this time interval, includes a new type of rhinoceros *Stephanorhinus elatus* (CROIZET & JOBERT) (= *Brandtorhinus jeanvireti* GUERIN), representing, very probably, an Asiatic immigrant. In the faunal succession of the western area of the Dacic Basin, *S. elatus* makes its first appearance (FAD) above coal layer XII in the local stratigraphy (Feru et al., 1965, 1983).

A remarkable further event consists in the first occurrence (FAD) of a smaller rhinoceros, attributable to *Stephanorhinus etruscus* (FALCONER) group, in the sandy deposits above coal layer XIII at Mătăsari-Brădețu (Jiu-Motru interfluvium) (Feru et al., 1965, 1983; Rădulescu & Samson, 1995). On the basis of palaeomagnetic calibration, the *S. etruscus* FAD can be placed within the Early Gauss normal subchron, in a later phase, postdating the arrival of *S. elatus*. The origin of *S. etruscus* is still obscure and its first appearance in Europe is still controversial. If faunal evolution in Italy is taken into account, *S. etruscus* would have occurred earlier in southeastern Europe, because paleomagnetic data suggest an age of about 3.1 Ma for the Mătăsari-Brădețu fossiliferous deposits. An earlier occurrence of *S. etruscus*, which confirms our evidence, is also indicated in Spain at Las Higuieruelas, Alcolea de Calatrava (Ciudad Real), where magneto-stratigraphic studies indicated a reversed magnetic polarity for the fossiliferous section corresponding indistinguishably to Mammoth or Kaena events (Mazo, 1996).

3.2.2. MN-16a/b subzone. A further step in the



constitution of a new mammalian community is indicated by the occurrence of a very primitive elephant (assigned provisionally to the genus *Mammuthus*) "*M.*" *rumanus* (see above) at Cernătești in western Oltenia (Samson & Rădulescu, 1973; Schoverth *et al.*, 1963). In our collection, "*M.*" *rumanus* is represented by a last upper molar (M3/)293 mm in length, constituted of eight plates and two talons. The laminar frequency (3.07) and the index of hypsodonty (1.18) are very low. The enamel is very thick averaging 4.25 mm.

In Italy, the Montopoli faunal unit, calibrated with the transition from the Gauss to the Matuyama paleomagnetic epoch, yielded a relatively primitive elephant *M. gromovi* (ALEXEEVA & GARUTT) (Azaroli, 1977). In the Dacic Basin, "*M.*" *rumanus*, which is more primitive in comparison with the Montopoli elephant, can be placed, on paleomagnetic evidence, within the Middle Gauss subchron.

According to Garutt (1992), "*M.*" *rumanus* is a descendant of *M. subplanifrons* OSBORN from South and East Africa. At the same time, "*M.*" *rumanus* displays similarities to "*M.*" *africanus* (ARAMBOURG) from North and East Africa. "*M.*" *rumanus* appears to belong to the first wave of elephants immigrated from the Africa to southeastern Europe via the Anatolian and Aegean regions.

At Cernătești, the mammalian assemblage including "*M.*" *rumanus* consists of *M. borsoni*, *A. arvernensis*, *S. elatus*, *S. etruscus*, *Metacervocerus pardnensis* (CROIZET & JOBERT), Cervid indet. (small-sized) in association with a rich Middle Romanian molluscan fauna.

A broad equivalent biostratigraphic level containing micromammalian remains is known from Podari (Jiu valley). The rich and various small mammal association contains nine insectivores, nine rodents and one ochotonid lagomorph. Representatives of the genus *Mimomys* are completely absent. As a whole the Podari fauna includes many taxa from earlier faunas.

The *Pliomys* (represented by a small-sized species) FAD marks the earliest occurrence of this genus in the Dacic Basin. At the present stage of knowledge, the origin of this small-sized *Pliomys* (s.s.) remains still unknown.

Further it must be remembered that at the biostratigraphic level typified by Podari fauna, the evolutionary processes leading from *Mimomys rhabonensis* to *Dolomys nehringi* on the one hand and from *M. moldavicus* to *Propliomys hungaricus* (KORMOS) on the other are already accomplished. Besides *Pliomys* sp., which is an immigrant form, the arviculids are represented only by *D. nehringi* and *P. hungaricus*, two well - defined morphological entities which are

the result of an endemic evolution (Rădulescu & Samson, 1996).

Paleomagnetic investigations suggest that Cernătești and Podari should be placed within the Middle Gauss subchron. Judging from the evolutionary stage of the micromammals, Podari can be ascribed to a time interval later than Csarnota-2 (Hungary).

The unusual rich small mammal assemblage at Podari, indicating a substantial increase in biodiversity, appears to correspond to a shift to warmer climatic condition (a peak of warmth ?) around 3 Ma.

3.2.3. MN-16a subzone. The late Middle Pliocene, corresponding to the end of the Late Gauss subchron, is characterized by the presence at Milcovu din Vale (Olt valley) of a mimomyian species belonging to *Mimomys stehlini* KORMOS/minor FEJFAR group (Rădulescu & Samson, 1983; Rădulescu *et al.*, 1995) in association with *Dolomys milleri* (NEHRING) and *Trogotherium minus* NEWTON. An abundant molluscan fauna indicating the beginning of the Wallichian was also collected (Andreescu *et al.*, 1981). Paleomagnetic data showed that Milcovu din Vale section can be situated within the Late Gauss normal subchron.

The small mammal association at Slatina-1, relatively rich in number of individuals, displays only a limited number of taxa with *Dolomys milleri* as dominant species. This impoverished fauna might correlate with a climatic depression (Praetiglian stage). It is worth mentioning that the associated molluscan fauna reveals boreal affinities (Andreescu *et al.*, 1981). Paleomagnetic studies indicated that Slatina-1 belongs to the Early Matuyama subchron.

3.2.4. MN-17 zone. During the MN-17 the climate registered several oscillations of shorter duration. More humid at the beginning of this time interval, the climate becomes drier, as shown by the presence in the Dacic Basin of steppe markers (*Borsodia*, *Allactaga*).

The *Borsodia* sp. FAD marks the lowest stratigraphic occurrence of the steppe lemming lineage in the Slatina section. The Slatina-2 faunal associations include, besides *Borsodia*, three insectivores (*Talpa fossilis* PETENYI, *Desmana nehringi*, *Beremendia fissidens* KORMOS) and three rodent species (*Apodemus* sp., *Dolomys ferui* RĂDULESCU & SAMSON, *Mimomys* cf. *Livenzovicus* ALEZANDROVA). Macromammals are represented by elephant *Mammuthus gromovi* and cervid *Eucladoceras* sp.

Slatina-2 fossil locality is reversely magnetized belonging to the end of the Early Matuyama subchron. Paleomagnetic determinations detected a normal polarity zone that overlies Slatina-2 fossiliferous level. This normal zone was correlated as Olduvai event (Andreescu *et al.*, 1981).

The *Allactaga* FAD is based on the occurrence of a new type of rodent, unknown in earlier faunal communities, in the small mammal association from Cherleşti-Moşteni in the Slatina area. The *Allactaga* FAD marks the earliest occurrence of an arid steppe/desert inhabitant in the western portion of the Dacic Basin. A large-sized vole, identified as *Mimomys* ex gr. *pliocaenicus* (MAJOR), also appears at this biostratigraphic level.

It is interesting to note that *Allactaga* sp. at Cherleşti-Moşteni possesses a dental pattern showing a relationship to *Allactaga euphratica* THOMAS from the eastern Mediterranean area. On the basis of dental morphology, *Allactaga* sp. is considered as a southern element belonging to a new migratory wave of Anatolo-Aegean origin.

Dolomys ferui LAD marks the highest stratigraphic occurrence of a representative of *Dolomys* lineage in the Slatina area. At this level, a new invading vole species (*M.* ex gr. *pliocaenicus*) tends to replace the last *Dolomys*.

Although very poor, the Slatina-3 (placed within the Olduvai event) (Andreescu et al., 1981) contains a medium-sized castorid *Trogontherium dacicum* RĂDULESCU, which is also present in the rich mammalian association of the Valea Grăunceanului at Tetoiu (= Bugiuleşti). This latter site is located in the valley of the Olteţ River, a tributary of the Olt River (Rădulescu & Samson, 1990).

Preceded by the fossiliferous level with macromammals from "La Pietriş", the Valea Grăunceanului fauna includes again southern immigrants such as an extinct girafid *Mitilanotherium inerspectatum* SAMSON & RĂDULESCU, 1966 [syn. *Macedonitherium martinii* SICKENBERG, 1967 described from Volakas (type-locality), Macedonia, Greece] known otherwise from the Aegean (northern Greece) and Anatolian regions, pangolian *Manis* cf. *hungarica* KORMOS [present at Villany-3 (type-locality) in southern Hungary] (Kormos, 1934) and large-sized ostrich *Pachystruthio* sp., the two latter forms being, very probably, of African origin.

It is also interesting to note the presence of an extinct larger, more terrestrial macaque lineage (*Paradolichopithecus geticus* NECRASOV, SAMSON & RĂDULESCU). The same lineage is known from Seneze (type-locality for *P. arvernensis* DEPERET), France (Delson, 1974).

The mammals at Valea Grăunceanului indicate a warm temperate climate that favoured some species to extend their range northwards into the Dacic Basin.

An important biostratigraphic event in the Tetoiu area is indicated at 1.7-1.8 Ma by the possible appearance of a ?*Homo erectus* lineage documented through

the discovery at Dealul Mijlociu of a few flint pebbles which seem to be linked to human activity (Rădulescu & Samson, 1990, 1991; Samson & Rădulescu, 1963). Tentatively, these tools of "oldowan aspect" might be indicative of the presence of the same human type known from Dmanisi, East Georgia, Caucasus, on the basis of a mandible belonging to the species *Homo erectus*. According to Gabunia and Vekua (1995), the human fossil from Dmanisi could be a descendant of *Homo habilis* from Olduvai (Africa), a relationship that would imply a migration from Africa to Eurasia.

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SEDIMENTATION BREAKS IN UPPER MIOCENE IN SOME BOREHOLES OF THE KOLUBARA BASIN, NORTHWESTERN SERBIA

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Abstract: Discontinuities in the sedimentation cycle were indicated by detailed biostratigraphical analysis of fossil foraminifers and ostracodes in Upper Miocene formation of the Kolubara Neogene Basin. Particularly notable were the breaks which affected the nature of and the fossil associations at the transitions between Sarmatian and Pannonian, and Pannonian and Pontian. Besides biostratigraphical evidence, neotectonic and other geological data were used.

The studied boreholes are located east of the Kolubara River (Fig. 1) in two coal fields: Vcr-1/77, Veliki Crljeni; T-145/81 and J-612/81, Turija. Both fields are in the eastern part of the Kolubara Neogene Basin, one of several marginal basins to the large Pannonian aquatorium, 50 km south of Belgrade.

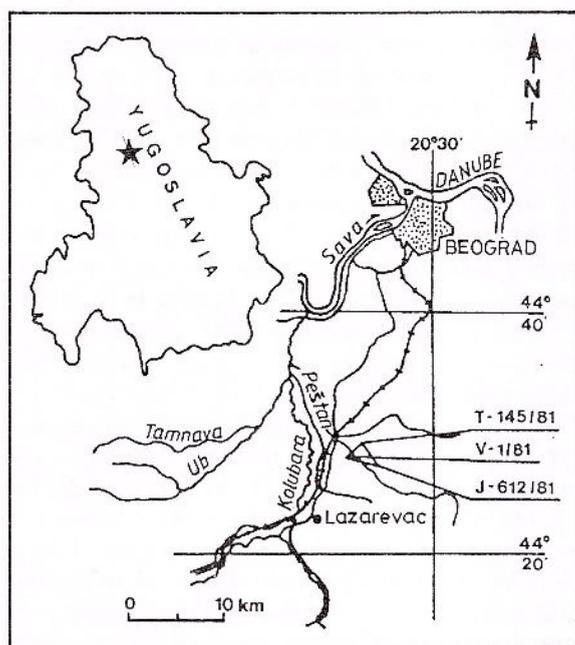


Fig. 1 - Physical setting of Kolubara Basin and borehole locations.

Investigation history

Sarmatian sediments in the Kolubara Basin lie over Cretaceous or Upper Badenian deposits, like else-

where on the margin of the Pannonian Basin (Stevanović, 1977; Petrović et al., 1987). Lithologically, Sarmatian is mainly represented by marly or sandy clays, and rarer by carbonate sandstones, sandy or organogenic limestones. These sediments, dated on the basis of the contained molluscan and foraminiferal fauna, belong to the Lower or Middle Sarmatian to Volhynian or Bessarabian respectively (Stevanović, 1977; Spajić et al., 1886; Petrović et al., 1987; Petrović & Mitrović, 1990).

The Sarmatian/Pannonian boundary has often been studied. A large discontinuity between the two stages was described by Jekelius (1943). On the basis of malacofaunal data from a section in the Sava bank, Stevanović (1985) infers that Pannonian levels are transgressive over Sarmatian deposits. Krstić (1973) confirms the hiatus in the town of Belgrade based on ostracod microfauna. Similar indications are given by Rundić (1989, 1995) who also mentions a minor break within the Pannonian.

The Carpatho-Balkanides bent in the early Pannonian, which resulted in the differentiation of Pannonian and Dacian Basins. The sedimentation regime west of the Carpathians was caspibrackish, quite similar to the lake sedimentation. This tectonic activity is associated with the Attican orogeny in the Volhynian-Pannonian period (Sikošek & Krstić, 1991). On the other side of the Carpathians, time equivalents of the Pannonian stage are developed: Bessarabian, Chersonian and Meotian. The mentioned tectonic event caused the Bessarabian transgression, but it had not a significant effect on the Sarmatian sequence in the Pannonian Basin, because the movements were probably synchronous with the basin division on either side of the mountain arc.



Several regressive phases during the Pannonian have been recognized in Belgrade area: the earlier in Lower Pannonian deposits, and the later in the early Upper Pannonian (Pannonian E in Papp, 1985) where the ostracod *Hemicytheria croatica* Zone is transgressive over the older deposits. This was noted in Belgrade area and in Croatia (Sokač, 1972). Similarly, a minor transgression was recognized at the beginning of late Upper Pannonian, when the last but one zone of ostracod development was deposited in the caspi-brackish Pannonian (a total of eight biozones by Krstić, 1985). This part of the Pannonian deposits is transgressive in Belgrade environs, commonly over Mesozoic or Sarmatian rocks (Krstić, 1973; Sikošek & Krstić, 1991).

The Pannonian/Pontian boundary is marked by Rhodanian orogeny which resulted in reestablishment of the communication between Pannonian and Dacian basins. It should be mentioned, however, that Upper Pontian deposits are transgressive and Lower Pontian ones are generally lacking in the southern, coastal parts of the Pannonian basin. Upper Pontian is transgressive in Dacian (Sikošek & Krstić, 1991).

Upper Miocene sediments as corresponding to the Lower and Middle Sarmatian, Pannonian and Pontian (Fig. 2). Although shortly spaced one from another, the boreholes indicated certain biostratigraphical dissimilarities.

Sarmatian

On the basis of foraminiferal microfauna from the examined boreholes, standard biostratigraphical zones were identified, known in the central Paratethys (Petrovic, 1971; Papp et al., 1974; Spajic et al., 1986). Sediments of the Porosonion granosum Zone were identified in all three boreholes (at 298.60 m in Vcr-1/77; 129.90 m in J-612/81; and at 182.20 m in T-145/81), which is the closing zone of the Lower Sarmatian or Volhynian substage represented by sands, carbonate sandstones and organogenic limestones. The Porosonion subgranosum Zone, earlier identified (Spajic et al., 1986; Petrovic et al., 1987; Petrovic & Mitrovic, 1990) in T-145/81, suggests by its biostratigraphical features the Middle Sarmatian, or the Lower Bessarabian.

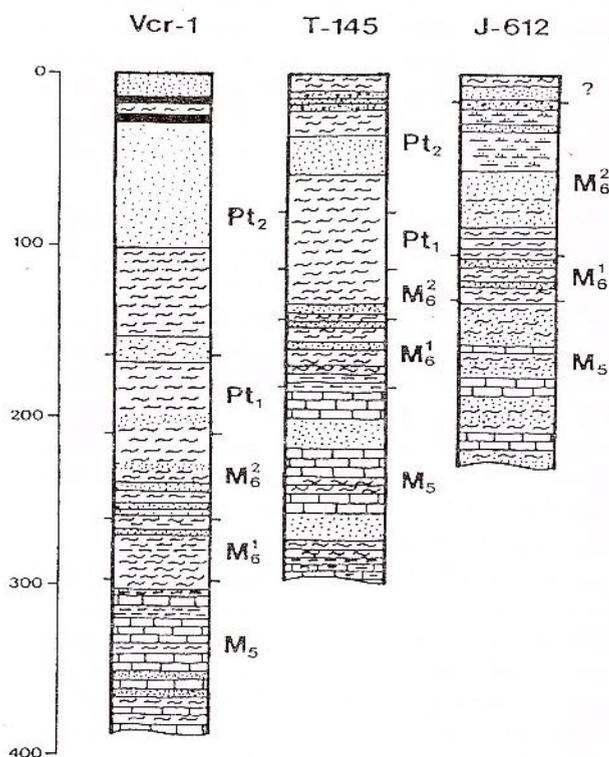


Fig. 2 - Comparative lithostratigraphical columns of the Upper Miocene in examined boreholes.

Biostratigraphical overview

Fossil foraminifers and ostracodes from holes Vcr-1/77, J-612/81 and T-145/81 were used to identify

Pannonian

Pannonian sediments were found in each of the examined boreholes (Fig. 2), generally represented by a succession of silty and clay sands, silstones and quartz sands. On the basis of the ostracod biozones, the formations were dated as Lower and Upper Pannonian. Applying the Pannonian division into eight biozones (Krstić, 1985), three zones of the Lower Pannonian were identified in the three boreholes: *Hemicytheria hungarica*, *H. tenuistriata*, or its equivalents, and *Propontoniella candeo*. The basal Lower Pannonian *Hemicytheria lorentheyi* Zone is lacking in each hole. The lithological boundary is quite distinct, as the carbonate component, dominant in Sarmatian formations, is gradually missing.

Consequently, the mentioned minor sedimentation break between Sarmatian and Lower Pannonian was found at the following depths: 294 m (Vcr-1/77), 176 m (T-145/81) and 125.70 m (J-612/81). The set of Lower Pannonian deposits varies in thickness (about 50 m in Vcr-1/77 and T-145/81 and down to 25-30 m in J-612/81). Zonal species *Hemicytheria tenuistriata* was not detected in T-145/81, but interval 155-145.20 m is its time equivalent.

Upper Pannonian deposits are transgressive and begin with the ostracod *Hemicytheria croatica* Zone. This age was not ascertained in J-612/81, as all samples more shallow than 104.10 m were negative in ostracod content and could only be taken for its equivalent. The basal Upper Pannonian *Amplocypris abscissa* Zone was not established. Besides

the mentioned Zone in holes Vcr-1/77 and T-145/81, a younger ostracod zone, *Serbiella sagittosa*, was identified (119.10-112.30 m in T-145/81 and 226.30-210.20 m in Vcr-1/77). Although Pontian deposits in the two holes are apparently natural and conformable over Pannonian sediments, the latest ostracod zone, *Typhlocyprilla lineocypriformis*, characteristic of the uppermost Pannonian, is everywhere lacking. The thickness of thus reduced Upper Pannonian is about 30 m in T-145/81 and about 40 m in Vcr-1/77.

Pontian

Pontian sediments are quite uniform, dominantly composed of a succession of gray sands, clays sands and siltstones, as found between 210.20-123.10 m in Vcr-1/77 wherefrom most of samples were extracted. The boundary could have been placed on the ostracod association more upward, but the purely Pontian molluscs (*Congeria zagrabiensis* Brusina) were decisive in determining the range between the given depth. The Pannonian/Pontian transition seems gradual and continuous, but the lack of the latest Pannonian biozone indicates a minor break in the late Pannonian. The presence of a mixed Pannonian/Pontian ostracod association was noted at the depth of 165 m, where typical Upper Pontian forms increasingly occurred and thus marked the boundary between Lower and Upper Pontian.

Pontian formations in T-145/81 were deposited in the depth interval 112.20-78.10 m. Like in Vcr-1/77, first were encountered transitional forms which suggested old Pontian levels, whereas Upper Pontian was proved from 81.40 m upward.

Final considerations

The Kolubara Basin acquired its general outline in the Middle Miocene (Fig. 3), and the lithospheric extension in the region gradually dwindled and ceased in the Upper Miocene. Since late Sarmatian to the present time, the subsidence has been gradual as a result of lithospheric cooling and terminal equilibration (Andjelkovic, 1987, 1989; Djokovic et al., 1988; Marovic et al., 1995, in press; Royden 1988; Royden & Baldi, 1988). This is supported by some geophysical observations on the eastern margin of the Kolubara Basin of its reverse character (Mladenovic & Kemenci, 1991). Associated with this is the difference in the depth of some stratigraphic units in boreholes (e.g. the difference in depth of Sarmatian and Pannonian sediments in J-612/81 and Vcr-1/77 is about 170 m) which is an evidence of blocks and shears along faults.

Note that all holes were drilled north of a conspicuous fault zone ("Medosevac dislocation", Maksi-

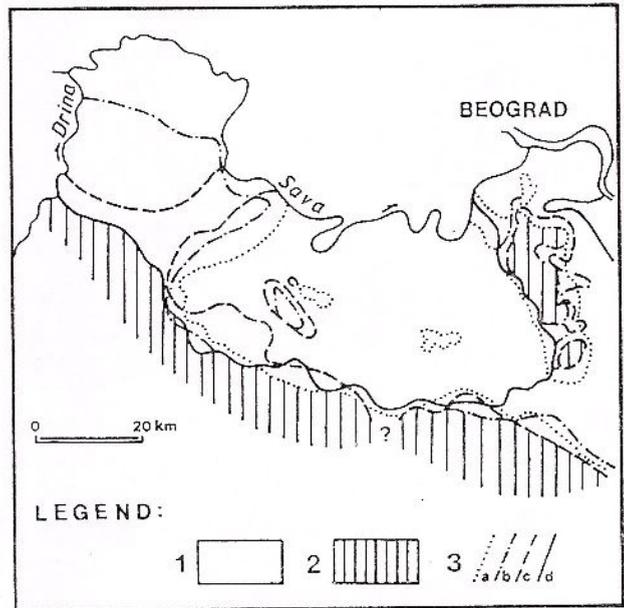


Fig. 3 - Location of Pannonian aquatorium coastlines in the Kolubara Basin: 1, Pannonian aquatorium; 2, land; 3, location of coastlines: a, Badenian; b, Sarmatian; c, Pannonian; d, Pontian. (Data and maps from Andjelkovic et al., 1991).

movic, 1993) which is extending E-W (Fig. 4). Transgressive over the paleorelief, Neogene development is complete in the area, whereas south of the dislocation, Upper Pontian coaliferous sediments directly overlie the Paleozoic complex. It seems indicative that the base relief in the south was penetrated by drilling at a depth of only 350 m, while the deepest hole in the north ended at 600 m in Sarmatian deposits and the paleorelief could be expected only below 800 m. Hence, the investigated area is a depositional environment of a continuous sedimentation from the earliest Miocene. This environment includes elements which are the evidence of a dynamic area. In respect of the short distance between the holes and of the proximity of a large positive structure (Volujak horst) in the southwest, the depth relationship of some stratigraphic units is a consequence of differential faulting and shears which indicate block features in the region. Slow downthrow prevailed over compressive movements and uplifts. Nevertheless, detailed biostratigraphic investigations revealed the compressions, which were significant at Sarmatian/Pannonian and Pannonian/Pontian boundaries and weaker during each of these divisions. Widely missing are basal parts of Lower Pannonian, i.e. the oldest ostracod biozones, and Upper Pannonian is generally incomplete. Moreover, late Pontian levels are mainly present,

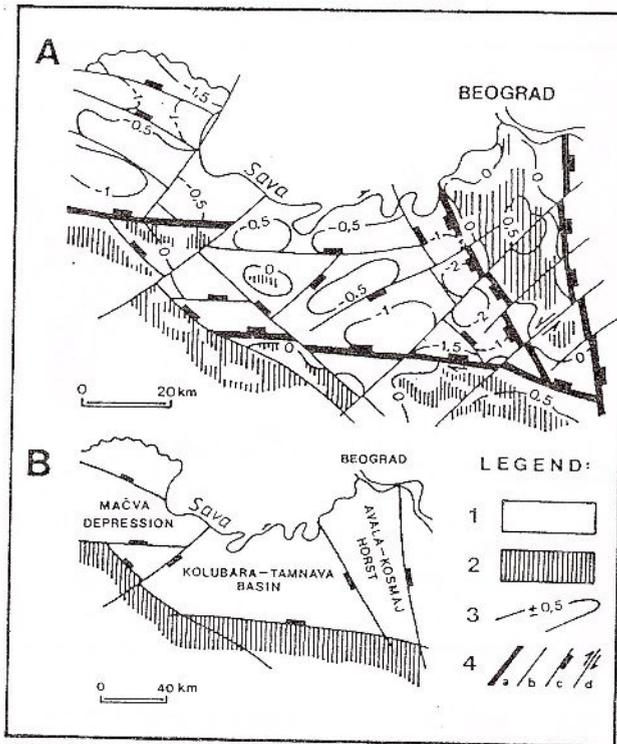


Fig. 4 - Neotectonic schematic map of the Kolubara Basin (A) and large neotectonic units of the Pannonian Basin south-eastern margin (B). 1, Neogene; 2, pre-Neogene; 3, total amplitude of vertical neotectonic movements (in km); 4, neotectonic faults: a, main marginal subsidence and trough; b, bounding minor neotectonic units; c, relatively downthrown block; d, transcurrent fault (from Marovic et al., 1995, in press).

whilst older ones are generally lacking.

It follows from the above stated that besides the dominant trend of continuous subsidence in this part of the Kolubara basin there were several strong compressive stages which affected the coastline causing occasional regression of waters from some blocks. This resulted in longer or shorter breaks in sedimentation of Upper Miocene formations and consequent "biostratigraphic gaps" as one of "witnesses" of the events.

Acknowledgments

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PARATETHYAN MOLLUSC FAUNAS FROM THE NEOGENE OF MACEDONIA AND THRACE, NORTHERN GREECE

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Abstract: Some Paratethyan mollusc faunas from the Neogene of northern Greece (Macedonia and Thrace) are studied. The faunas come from three different basins: a. Axios - Thermaikos basin (10 sites), b. Serres - Strymon - Strymonikos gulf basin (6 new sites), and c. Xanthi - Komotini basin (3 new sites). All faunas consist of a monotonous mollusc association including: Limnocypridae, Dreissenidae, *Parvivenerus* sp., *Maetra* sp., *Theodoxus* sp., *Melanopsis* sp., *Hydrobia* sp. The faunal stability, the lithological similarities and the stratigraphic position of the fossiliferous sites indicate a Paratethyan transgression in Northern Greece during Uppermost Miocene.

Introduction

Neogene of Northern Greece is characterised by the presence of two different types of mollusc faunas:

- A Paratethyan (Euxinic), characterised by low salinity molluscs faunas, and
- A Tethyan (Mediterranean), with mollusc faunas indicating normal salinity.

As a result, Northern Greece is considered as a transitional zone between Paratethyan and Tethyan faunal provinces (Steffens et al., 1979; Kojumdjieva, 1987).

A few years ago the presence of Paratethyan mollusc faunas, in the wider area of northern Greece, was known only by a few isolated fossiliferous sites. The "classical" sites of Alatini and Trilophos in Axios-Thermaikos basin keep the interest of several palaeontologists (Gillet, 1937; Gillet & Faugeres, 1970; Gillet & Geissert, 1971; Stevanovic, 1961, 1963, 1964; Papp, 1979; Steffens et al., 1979).

In the Neogene deposits of Serres-Strymon basin, Paratethyan faunas are briefly referred, by Gramann & Kockel (1969) as well as by Beratis & Kojumdjieva (1988).

Concerning the Paratethyan faunas there are no other references in northern Greece. Several new fossiliferous sites have been found in the Chalkidiki peninsula and new material has been unearthed and studied (Syrides, 1990). After that our efforts to locate new sites with Paratethyan faunas have been

continued eastwards in the Serres-Strymon-Strymonikos gulf basin and Xanthi-Komotini basin. During the last five years several new sites with Paratethyan molluscs have been found and abundant fossils have been collected. Some indications of Paratethyan faunas appear also in Ferres, east to Alexandroupolis (Syrides, pers. observ.), but research is still in progress.

In this article the preliminary data about the presence of the Paratethyan faunas and their significance to the Neogene stratigraphy of N Greece is given.

AXIOS - THERMAIKOS BASIN

Several fossiliferous sites with Paratethyan mollusc faunas have been located along the west part of the Chalkidiki peninsula, from the city of Thessaloniki up to the Kassandra peninsula. Besides the "classical" sites of Trilophos and Alatini several new fossiliferous localities named "Trilophos" (TRF), "Koumi Vrysi" (KVR), "Kardia-1" (KRD-1), "Kardia-2" (KRD-2), "Sholari" (SHL), "Lakkoma" (LKM), "Nea Gonia-1" (NGN-1), "Krini" (KRN), "Fokca" (FOK) and "Sani" (SANI) were found (Fig. 1 a), (Syrides, 1990). The faunal list of each locality is given below; data are coming from Syrides (1990) and new personal observations.

Locality "Trilophos" (TRF):
Theodoxus micans (GAUDRY, FISCHER)
Valvata sp.
Hydrobia sp.



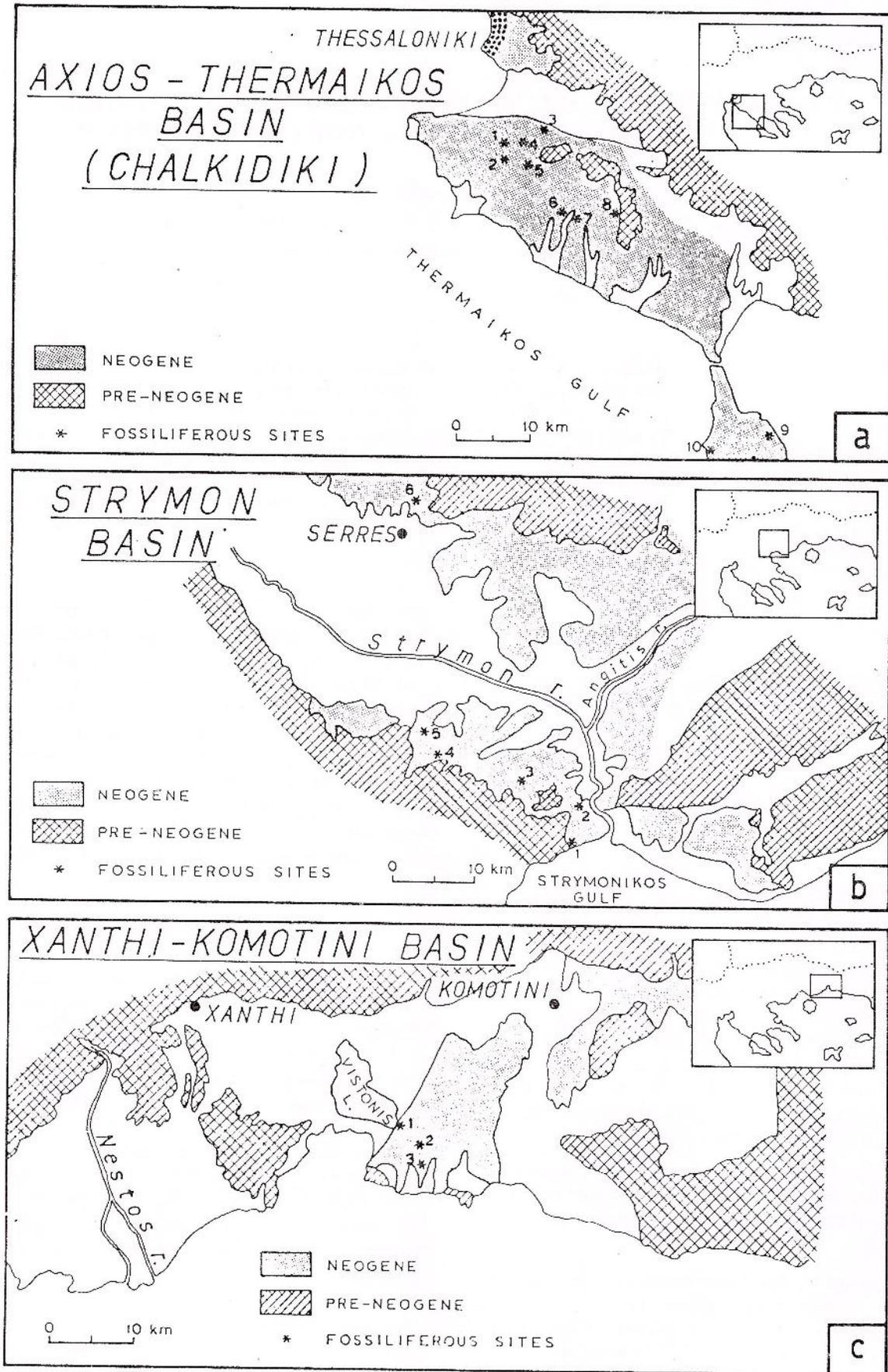


Fig. 1 - Fossiliferous sites with Paratethyan mollusc faunas in the Neogene of Macedonia and Thrace, N Greece.

164.114



Pyrgula boettgeri FUCHS
Pyrgula sp.
Clathrocaspia sp.
Micromelania sp.
Melanopsis bonelli MANZONI
Melanopsis sp.
Pseudoprosodacna littoralis littoralis (EICH.)
Pseudoprosodacna sp.
Limnocardium sp.
Congeria/Dreissena sp.
Mactra (Spisula) sp.
Parvivenus widhalmi (SINZOV)
Abra sp.
Unio sp.

Locality "Koumi Vrysi" (KVR):

Theodoxus sp.
Valvata obtusaeformis LOER.
Hydrobia sp.
Pyrgula boettgeri BRUS.
Melanopsis bonelli MANZ.
Melanopsis eulimopsis BRUSINA
Planorbis sp.
Pseudoprosodacna littoralis BARB.
Limnocardium sp.
Congeria/Dreissena sp.
Parvivenus widhalmi (SINZOV)

Locality "Kardia-1" (KRD-1):

Theodoxus sp.
Hydrobia sp.
Melanopsis bonelli MANZ.
Melanopsis sp.
Pseudoprosodacna sp.
Pseudocatillus sp.
Limnocardium sp.
Congeria/Dreissena sp.
Spisula (Pseudoxyperas) faugeresi (GILLET, GEIS-
 SERT)

Locality "Kardia-2" (KRD-2):

Theodoxus micans (GAUDRY, FISCHER)
Theodoxus sp.
Melanopsis sp.
Limnocardium sp.
Congeria/Dreissena sp.
Abra sp.

Locality "Sholari" (SHL):

Melanopsis sp.
Planorbis sp.
Pseudoprosodacna littoralis (EICH.)
Pseudoprosodacna sp.
Pseudocatillus sp.
Limnocardium sp.

Congeria/Dreissena sp.

Locality "Lakkoma" (LKM):

Melanopsis bonelli MANZ.
Limnocardium sp.
Congeria/Dreissena sp.

Locality "Nea Gonias-1" (NGN-1):

Melanopsis bonelli MANZ.
Pseudocatillus sp.
Limnocardium sp.;
Congeria/Dreissena sp.
Parvivenus widhalmi (SINZOV).

Locality "Krini" (KRN):

Theodoxus sp.
Melanopsis (Lyrcaea) bonelli MANZONI
Melanopsis sp.
Radix sp.
Pseudoprosodacna littoralis cf. *plicatolittoralis*
 SINZOV
Pseudoprosodacna sp.
Pseudocatillus pseudocatillus (BARBOT)
Pseudocatillus sp.
Pontalmyra sp.
Limnocardium sp.
Congeria/Dreissena sp.
Mactra (Spisula) sp.
Parvivenus cf. *widhalmi* (SINZOV)

Locality "Fokea" (FOK):

Theodoxus sp.
Melanopsis bonelli MANZ.
Melanopsis sp.
Pseudoprosodacna littoralis (EICHW.)
Pseudocatillus pseudocatillus (BARBOT)
Pontalmyra sp.
Limnocardium sp.
Congeria/Dreissena sp.
Mactra (Spisula) sp.
Parvivenus widhalmi (SINZOV)
Abra sp.

Locality "Sani" (SANI):

Theodoxus micans (GAUDRY, FISCHER)
Rissoa sp.
Cerithium sp.
Bittium sp.
Nassa sp.
Hydrobia sp.
Clathrocaspia sp.
Ostrea sp.
Limnardiidae ind.
Congeria/Dreissena sp.
Loripes lacteus (LIN.)
Venus sp.
Parvivenus widhalmi (SINZOV)



Corbula sp.

Abra sp.

Unio sp.

Paratethyan faunas also exist in the west margins of the Axios-Thermaikos basin (in the wider area of Katerini) but the research is still in progress, and there are no clear data and determinations until now.

SERRES-STRYMON BASIN

The references about the presence of Paratethyan mollusc faunas in Serres-Strymon basin are few. From the brackish "Choumnikon beds" Gramann & Kockel (1969) refer the presence of:

Congeria cf. *subcarinata*

Dreissenidae

Paradacna cf. *abichi*

Limnardiidae

A Pontian faunule is also known from a small outcrop near the bridge of the Angitis river (Beratis & Kojumdjieva, 1988) including the following species:

Paradacna abichi abichiformis

Congeria (Andrusoviconcha) navicula

Abra tellinoides

Six new fossiliferous sites, named "Akti Kerdylion-1", "Kato Kerdylion-1", "Aidonochori-1", "Skini-1", "Houmniko-1" and "Perdikari-1" (Fig. 1 b), were found during the last five years. Since the material is on study, preliminary faunal lists are given below.

Locality "Akti Kerdylion-1"

It is situated on the northern coast of the Strymonikos gulf (Fig. 1 b*¹), near the settlement of Akti Kerdylion. The outcrop is along a road cut, very close to the pre-Neogene basement. The fossiliferous beds have a thickness of more than 2 m, and consist of grey-white to yellowish coarse sandstones. The fossils are badly preserved and appear only as casts and moulds. The fauna includes:

Limnardiidae ind.; *Maetra (Spisula)* sp.; *Parvivenus* sp.

Locality "Kato Kerdylio-1"

It is situated near the ruins of the village of Kato Kerdylion (Fig. 1 b*²) on the top of a small hill. The fossiliferous beds consist of grey-white to yellowish sandstones with a thickness of 5-8m. The fossils are numerous and appear as casts and moulds. The first determined fauna consists of:

Theodoxus sp.

Melanopsis sp.

Pseudoprosodacna sp.

Pseudocatillus sp.

Limnardiidae ind.

Congeria/Dreissena sp.

Maetra (Spisula) sp.

Parvivenus sp.

Locality "Aidonochori-1"

It is situated about 1.5 km south of the village of Aidonochori on the sides of the ravine "Mylon Rema" (Fig. 1 b*²). Along a road cut beds of grey-yellowish fossiliferous fine sandstones are exposed. The preservation of the fossils is poor (casts and moulds), and the following genera have been determined:

Pseudoprosodacna sp.

Limnardiidae ind.;

Dreissena/Congeria sp.

Locality "Skini-1"

It is situated on the hill "Skini" (307 m a.s.l.) about 3 km SW to the village of Sitochori (Fig. 1 b*⁴). The hill "Skini" is a cuesta consisting of marly-sandy limestone beds inclined about 5-10° north-northwestwards. The limestones are about 3-5 m thick and overlie a series of continental red-beds (Fig. 2 b) which cover the pre-Neogene basement. The fauna includes only casts and moulds of:

Melanopsis sp.

Pseudoprosodacna sp.

Pseudocatillus sp.

Limnardiidae ind.

Congeria/Dreissena sp.

Parvivenus sp.

Locality "Houmniko-1"

It is situated 1 km west of the village of Houmniko, on the top of a hilly terrain (Fig. 1 b*⁵). The fossils have been found in a few meters outcrop of grey-green clayey sands alternating with grey-dark grey sandstones. The preservation of the fossils is very poor (casts and moulds). The fauna includes:

Theodoxus sp.

Hydrobia sp.

"*Helix*" sp. 1 (large size)

"*Helix*" sp. 2 (small size)

Pseudocatillus sp.

Limnardiidae ind.

Congeria/Dreissena sp.

Maetra (Spisula) sp.

Parvivenus sp.

Locality "Perdikari-1"

It is situated in the area of the old coal mines of Perdikari, close to the city of Serres (Fig. 1 b*⁶). In a small outcrop across a road, alternated fossiliferous beds of grey-white sands and grey-green silty-clayey sands are exposed. The fossils are quite well preserved but they are very fragile. The first determined fauna includes:

Theodoxus sp.



Hydrobia sp.
Melanopsis cf. *bonelli*
Melanopsis sp.
 "Helix" sp. 1 (large size)
Pontalmyra sp.
 Limnocypridae ind.
Congeria/Dreissena sp.
Mactra (Spisula) sp.
Parvivenus cf. *widhalmi*
Abra sp.

XANTHI-KOMOTINI BASIN

The presence of Paratethyan mollusc faunas has been referred for the first time to the Xanthi-Komotini basin. During the last two years our investigations in the basin gave three new fossiliferous sites, named "Glykoneri-1", "Mesi-1" and "Glyfada-1", with mollusc fauna (Fig. 1 c) (Syrides et al., in prep.). The sites are situated in the area of the villages Glykoneri, Mesi and Glyfada, respectively. The wider area reveals a very smooth low hilly terrain with mean altitude less than 30-40 m a.s.l.; as a result, the absence of outcrops and escarpments is considerable.

Locality "Glykoneri-1"

It is situated 1 km SW of the village of Glykoneri on the coast of the lake Vistonis (Fig. 1 c¹). The outcrop consists of the following beds from top to bottom (Fig. 2 c):

- 1-3 m of yellowish to very pale brown fossiliferous marly limestone.

- 1-2 m of red beds, consisting of coarse sand and pebbles intermixed with brown-red brown silt and clay. The contact between limestone and red beds is irregular, probably an old erosional surface.

The fossils are numerous but unfortunately they are badly preserved (casts and moulds). The first determination of the collected material yields the following fauna:

Theodoxus sp.
Melanopsis cf. *bonelli*
Melanopsis sp.
 "Helix" sp. 1 (large size)
 "Helix" sp. 2 (small size)
Pseudoprosodacna littoralis (EICHW.)
Pseudocatillus pseudocatillus (BARBOT)
 Limnocypridae ind.
Congeria/Dreissena
Mactra (Spisula) sp.
Parvivenus sp.

Locality "Mesi-1"

It is situated about 1 km N of the village of Mesi

(Fig. 1 c²). The fossils have been found on the top of a hill in a very restricted occurrence of a yellow sandy limestone (0.5 m thick). Only a few specimens (moulds and casts) have been collected. The first determined fauna includes:

Theodoxus sp.
Melanopsis sp.
Pseudoprosodacna sp.
 Limnocypridae ind.
Congeria/Dreissena sp.

Locality "Glyfada-1"

It is situated about 1 km S of the village of Glyfada (Fig. 1 c³). On the top of a hill the cultivation exposes blocks of pale yellow to grey-yellow fossiliferous sandy limestone. The collected material from these blocks consists of badly preserved casts and moulds. The first determination of the fauna yields the following genera:

Melanopsis sp.
Pseudoprosodacna sp.
 Limnocypridae ind.
Congeria/Dreissena sp.
Mactra (Spisula) sp.

Discussion on the fauna

In the studied Paratethyan faunas of N Greece the specific determination is not always possible because of the bad preservation of the fossils. In most cases they are represented by casts and moulds. However, the fauna yields several data for a first correlation between the localities, as well as between the basins.

It is remarkable that although the localities are geographically distributed in three different basins, they contain common genera (Table). The synthesis of the fauna includes pure Paratethyan (Euxinic) elements (*Pseudoprosodacna*, *Pseudocatillus*, Limnocypridae ind., *Dreissena/Congeria*), as well as genera of Mediterranean (Tethyan) origin [*Abra*, *Parvivenus widhalmi*, *Mactra (spisula)*] that were immigrated and adopted into the Euxinic basin (according to Stevanovic, 1963, 1964, 1966, this immigration happened during the Upper Miocene). The fauna also contains limnic elements such as *Theodoxus*, *Valvata*, *Hydrobia*, *Melanopsis*, *Planorbis*. Land snails ("Helix") also occur in some localities. This mollusc association is monotonous, but remains more or less constant in all localities of the three basins.

In the studied faunas mollusc genera indicating normal salinity (marine origin) are unknown until now, except in the locality of SANI (Axios-Thermaikos basin). The fauna of SANI contains Dresseiids, Limnocyprids as well as several genera indicating normal salinity (*Rissoa* sp., *Cerithium* sp.,



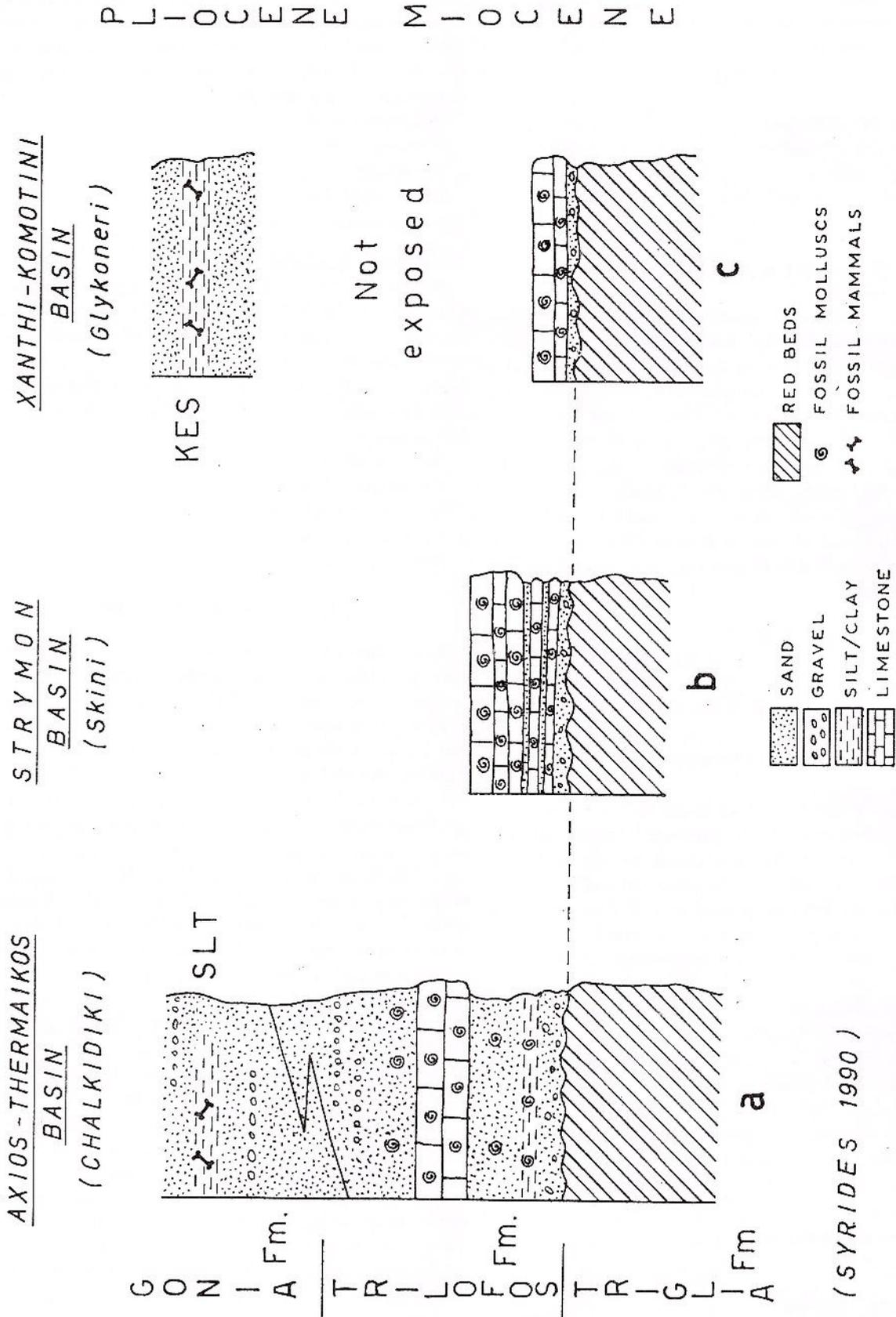


Fig. 2 - Lithostratigraphic correlation between the three studied basins of Macedonia and Thrace, N Greece.

Bittium sp., *Nassa* sp., *Ostrea* sp., *Loripes lacteus*, *Venus* sp., *Corbula* sp.). Although these genera indicate a marine origin they include euryhaline representatives too. Nevertheless the fauna from SANI is an exception and it is still under study. Papp (1979, p. 235) refers also to Trilophos some marine representatives (*Rissoa* sp., *Bittium greacense*, *Cerithium* sp., *Dosinia meotica*, *Donax* sp., *Sphenia* etc.) but further investigation is needed for this.

The faunas of Chalkidiki (Axios-Thermaikos basin) have been dated to Upper Miocene (Pontian) (Syrides, 1990). According to Papp (1979) the faunas of Trilophos (Chalkidiki area) and Trakhones (near Athens) are of Lower Pontian age. They contain a typical mollusc association consisting of *Pseudocatillus pseudocatillus*, *Pseudoprosodacna littoralis*, *Parvivenus widhalmi*, *Abra tellinoides* and *Spisula (Pseudoxyperas)*. The same mollusc association has also been found in the faunas of Strymon and Xanthi-Komotini basins (Table). A Pontian age can be also suggested for these localities. The palaeoecological characters of the fauna indicate an aquatic environment of low salinity with fresh water (*Theodoxus*, *Valvata*, *Hydrobia*, *Melanopsis*, *Planorbis*), and locally terrestrial (*Helix*) influence.

Stratigraphic data

The stratigraphy of the Neogene sediments of Chalkidiki peninsula has been studied in detail by Syrides (1990). Several sedimentary formations have been recognised. All the fossiliferous sites with mollusc faunas (TRF, KR1-1, KR1-2, SHL, LKM, NGN-1, KRN, FOK, SANI) have been located in Trilophos Fm (Fig. 2 a). Trilophos Fm consists of fossiliferous clays, sands and sandy limestones, it overlies the continental red beds of Triglia Fm (late Vallesian-lowermost Turolian) and is overlain by the sands, clays and marls of Gonia Fm (Pliocene). In the lower levels of Gonia Fm a fossiliferous locality (SLT) (Fig. 2 a) with micromammals of Lower Pliocene age was found (Syrides, 1990).

In Serres-Strymon-Strymonikos gulf basin, the stratigraphy is not well clarified. Several workers (Gramann & Kockel, 1969; Armour-Brown et al., 1977; Karystineos, 1984) present lithostratigraphic data and columns but not for the entire basin. In the SW margins of Strymon basin Gramann & Kockel (1969) recognise the marine "Dafni beds" overlain by the brachish "Houmniko beds" with *Limnocardiids* and *Dreissenids*. All our fossiliferous sites (Akti Kerdylion-1, Kato Kerdylion-1, Aidonochori-1, Skini-1, Houmniko-1, Perdikari-1) can be placed in "Houmniko beds". The fossiliferous localities are not well exposed and the stratigraphic succession is obscured. But in the hill of Skini a succession of fossiliferous sed-

iments and underlying continental red beds is clear (Fig. 2 b). This allows a first lithostratigraphic correlation with Chalkidiki similar deposits.

In the basin of Xanthi-Komotini the stratigraphic data are very poor and the absence of outcrops is considerable. In Vistonis lake along the east coast (Glykoneri) as well as along the west coast (Nea Kessani) there are some small outcrops. The fossiliferous limestones in Glykoneri-1 overlie red beds and are overlain by terrestrial sands and clays (Fig. 2 c). A micromammalian fauna found in these clays (locality "Kessani-1" KES) suggests an early Ruscinian age (Syrides et al., in prep.). Although the stratigraphic succession between Glykoneri and Nea Kessani is not visible, a first correlation with Strymon basin, as well as with Chalkidiki is possible (Fig. 2).

Chronology

The fossiliferous localities of the Chalkidiki peninsula contain a Paratethyan fauna of Pontian age. Trilophos Fm has been dated to late Miocene (Pontian) because it overlies late Vallesian-lowermost Turolian red beds and is overlain by the Pliocene (Ruscinian) sediments of Gonia Fm (Syrides, 1990). In the Serres-Strymon basin the mollusc faunas are also dated to Pontian and the fossiliferous sediments overlie red beds. In the Xanthi-Komotini basin the Paratethyan faunas also have a Pontian age. The fossiliferous sediments overlie red beds and are overlain by sediments containing a micromammalian fauna of early Ruscinian.

All the above-mentioned data coming from the study of the Paratethyan faunas of N Greece as well as lithostratigraphic and chronological data indicate the following:

- Common and constant faunal composition in the various localities
- Pontian (Upper Miocene) age of the faunas
- Similar lithostratigraphic succession (fossiliferous sediments overlie red beds)
- Similar lithology of fossiliferous beds in all basins.

Considering all these data it is clear that during the Upper Miocene (Pontian) Paratethys invaded the three (at least) basins of Northern Greece (Fig. 3).

In Chalkidiki the "framing" of the fossiliferous sediments of Trilophos Fm between the underlying red beds of Triglia Fm (upper Vallesian-lowermost Turolian) and the overlying Pliocene (locality SLT is early Ruscinian) sediments of Gonia Fm confirm the Pontian age of the Paratethyan invasion.

An analogous "framing" of the fossiliferous sediments between red beds (Miocene?) and Pliocene (KES) sediments is possible in Glykoneri-1 (Xanthi-Komotini basin) (Fig. 2 c).



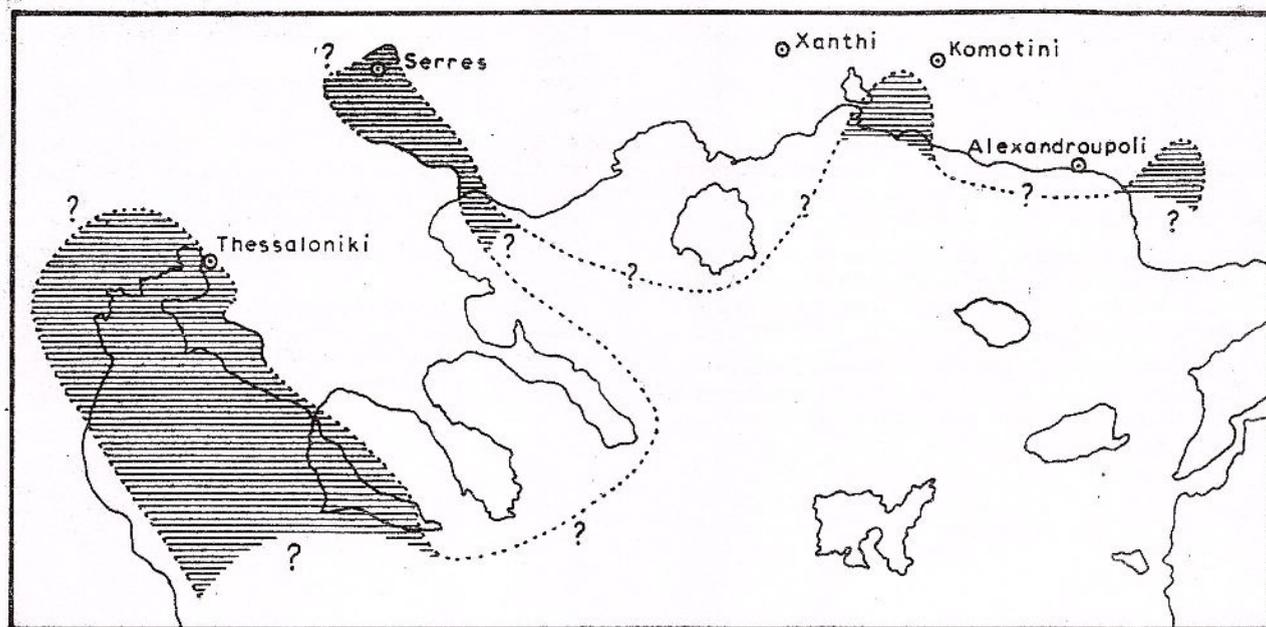


Fig. 3 - Palaeogeographic sketch map indicating the late Miocene (Pontian) Paratethyan invasion in Macedonia and Thrace, N Greece.

The distribution of Paratethyan molluscs localities in northern Greece and the discovery of new localities in Xanthi-Komotini basin in combination with the existence of Paratethyan faunas in East Thrace and Dardanelles (Gillet et al., 1978) support the hypothesis of a Paratethyan invasion into the Aegean region through East Thrace and Dardanelles.

Besides the abundant indications and evidences for a Paratethyan invasion during late Miocene in N Greece, a more extensive study is necessary in order to prove in details this invasion and to give a new stratigraphic reconsideration for the area.

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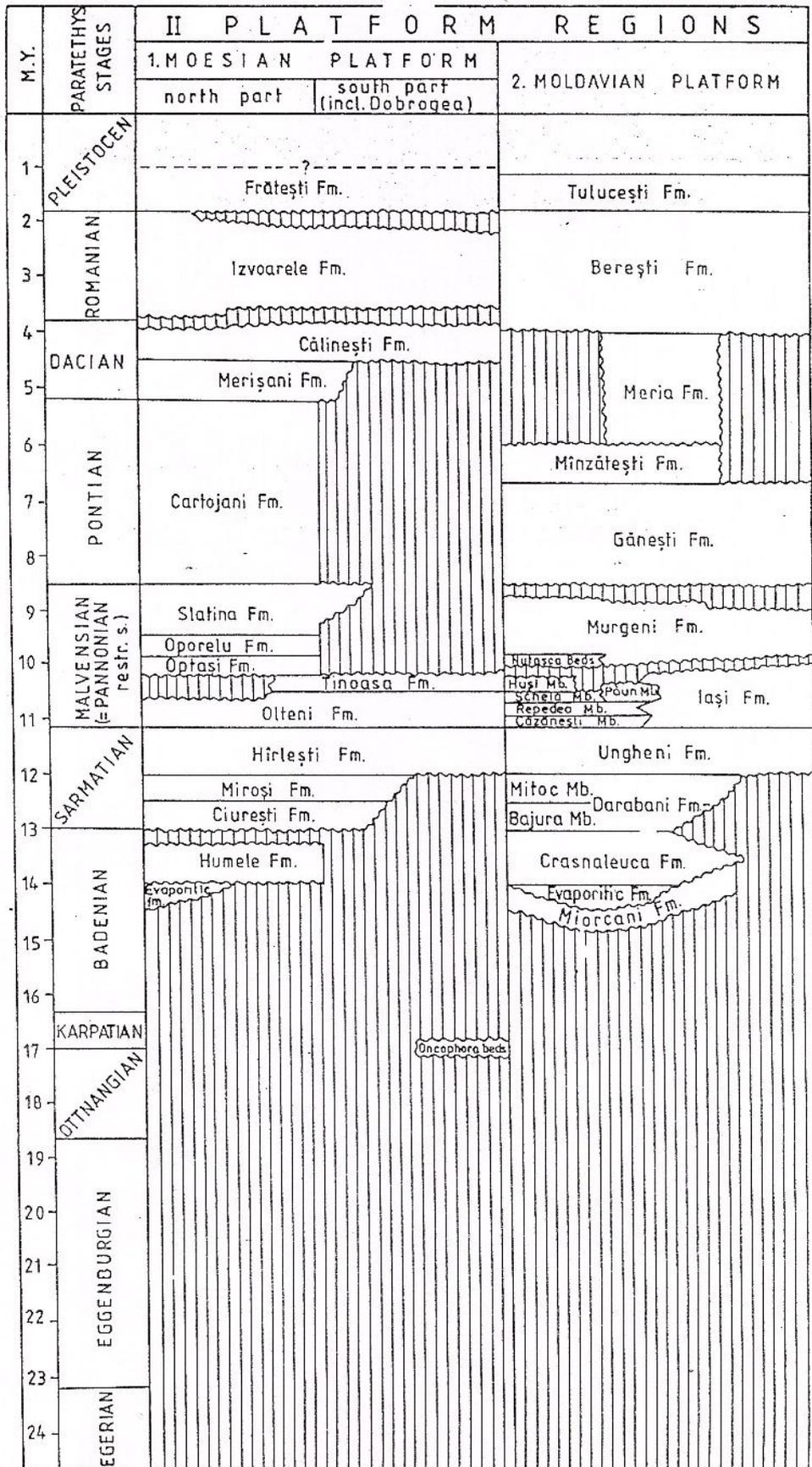
TABLES WITH THE CORRELATION OF THE NEOGENE DEPOSITS IN ROMANIA

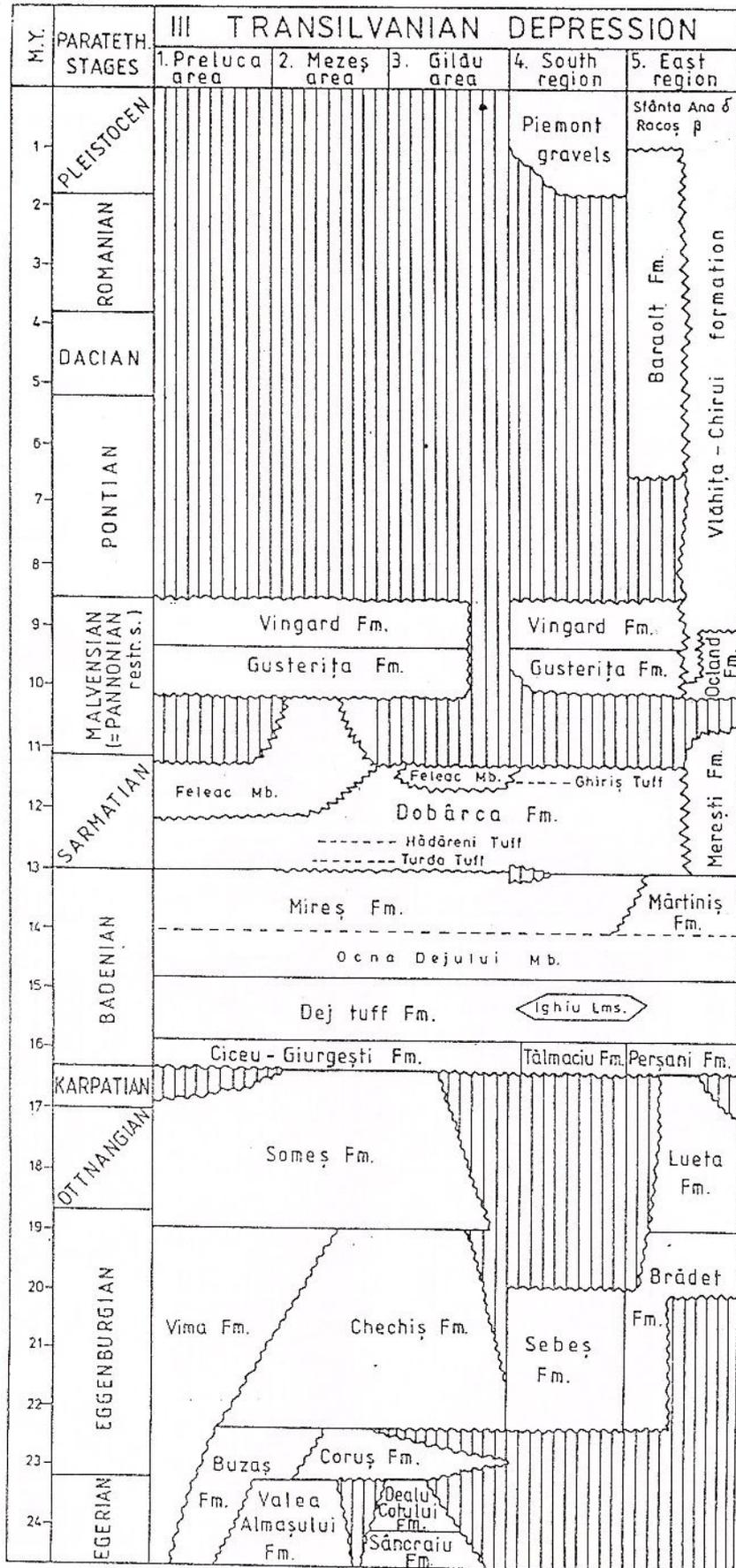
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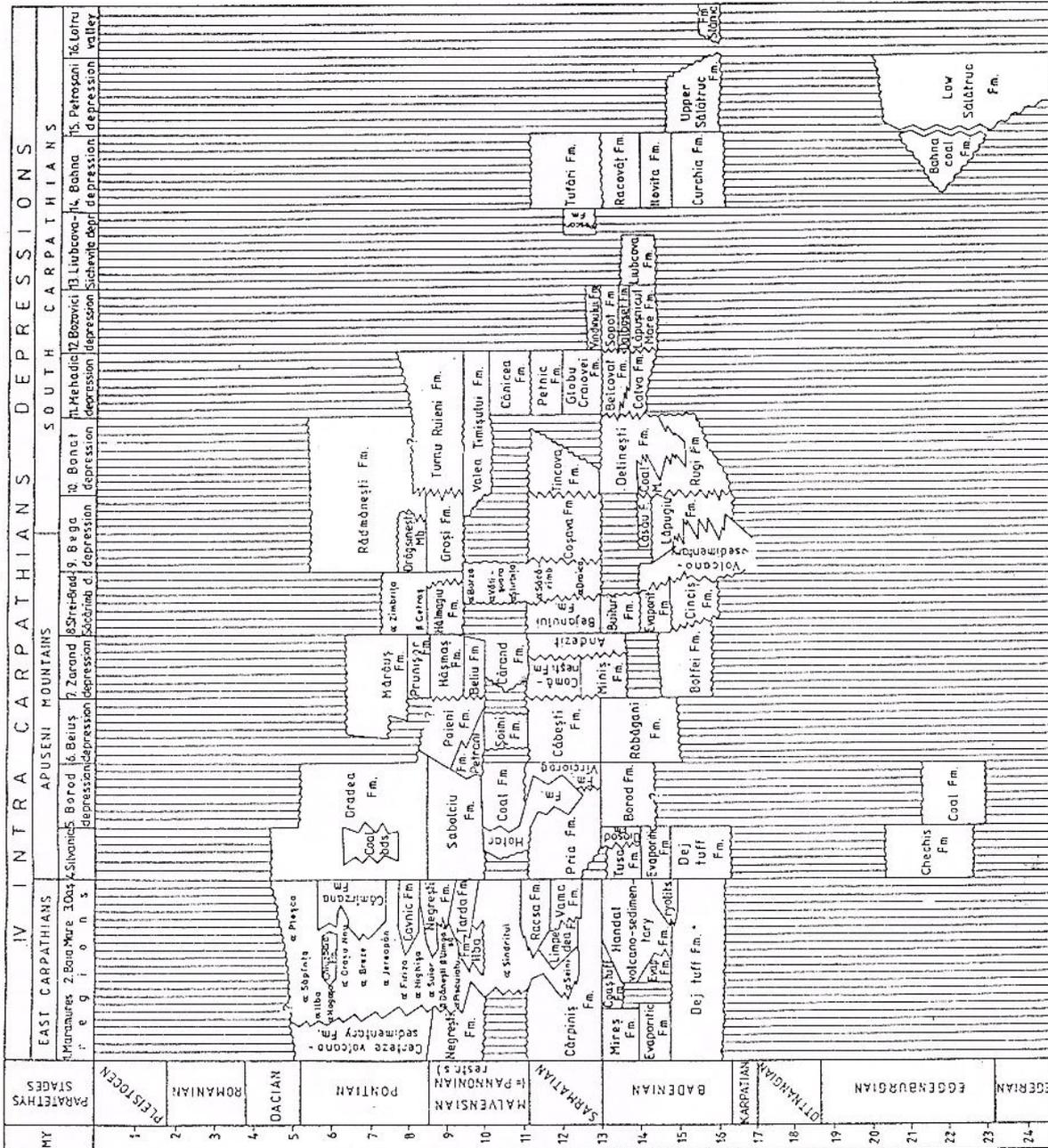
In the tables with the correlation of the Neogene deposits in Romania we try to present, in a normal succession, all the Miocene and Pliocene lithostratigraphic units described in the literature, denominated according to the international standard norms and ordered on major structural units.

For datings and correlations the most recent integrate biostratigraphic data and age determinations have been used.









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c) cărți:

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d) hărți:

Ionescu, I., Popescu, P., Georgescu, G. (1990) Geological Map of Romania, scale 1:50,000, sheet Cîmpulung. *Inst. Geol. Geofiz.*, București.

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b) special issues:

Strand, T. (1972) The Norwegian Caledonides. p. 1-20. In: Kulling, O., Strand, T. (eds.) *Scandinavian Caledonides*, 560 p., Interscience Publishers.

c) books:

Bălan, M. (1976) Zăcămintele manganifere de la Iacobeni. Ed. Acad. Rom., 132 p., București.

d) maps:

Ionescu, I., Popescu, P., Georgescu, G. (1990) Geological Map of Romania, scale 1:50,000, sheet Cîmpulung. *Inst. Geol. Geofiz.*, București.

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