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LOWER CRETACEOUS CARBONATE AND CARBONATE-EVAPORITE SEDIMENTATION IN THE EAST-WALLACHIAN SECTOR OF THE MOESIAN PLATFORM (EASTERN ROMANIAN PLAIN)[1]

BY

ANDREI DRĂGĂNESCU[2]

with an appendix by

S. RĂDAN

Abstract

A Lower Cretaceous autochthonous sedimentation of a carbonate-evaporite type is described in the eastern part of the Romanian Plain on the basis of petrologic material provided by drillings. The concerned sedimentary sequence may be subdivided in 3 complexes: I) the micritic limestone complex, including mainly micrites accommodating in places allochemical varieties of limestones; II) the facies-zoned complex, exhibiting a very complicate constitution, consisting of three, isochronous, laterally interfingering, petrogenetic systems: the tidal-flat system (of a carbonate-evaporite consistence), the carbonate-lagoonal system (mainly limestones of a bahamite type) and the reef system (skeletal and oölitic limestones); III) the reefal limestone complex consisting mainly of skeletal limestones of a coralgal, brialgal and algal type. The main petrogenetic processes pointed out by the studied sedimentary sequence are: syngenetic processes including deposition of calcareous sediments on account of inorganic or biogenic extraction of $CaCO_3$ from seawater as well as evaporite and dolomite formation in relation with sabkha cycles, and the diagenetic processual sequence — a new diagenetic line — involving dolomitization-dedolomitization-aggrading recrystallization responsible for most of the spar calcite in the analysed material. Particularly worth mentioning are 4 very interesting petrologic aspects discussed in the paper: 1) the microcoprolithic (fecal pellet) limestones, abundant in the tidal-flat system of the 2-nd sedimentary complex and consisting of very well-preserved fecal pellets ascribed to crustaceans, gastropods and polychaete annelids; 2) the anhydrite-after-gypsum pseudomorphs exhibiting two textural types; 3) the

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[2] Institutul de geologie și geofizică, str. Caransebeș 1, 7000 București.

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penecontemporaneous, submarine neomorphism of some micritic fabrics to fine microspar; 4) the replacement origin for almost all spar calcite in studied material by a new diagenetic line; dolomitization-dedolomitization-aggrading recrystallization.

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GENERAL STATEMENT

The area considered in the present paper is located in the east-Wallachian sector of the Moesian Platform [3], geographically lying in the eastern part of the Romanian Plain, bounded by the localities: Hîrșova, Brăila, Smeeni and Călărași. The work relies on the petrologic data provided by the sedimentary sequence stratigraphically interposing between the Upper Jurassic dolomitized reefal limestones and the Hauterivian variegated marly-calcareous series in the drillings performed in the following sectors: Bordeiu Verde (drilling no. 1), Ianca (drilling no. 2), Berlescu (drillings no. 3, 4 and 5), Zăvoaia (drilling no. 6), Smirna (dril-

[3] The Moesian Platform is the geological platform, generally epihercynian in age, placed between the Carpathian and Balkanian Orogens, geographically assembling the teritories of the Romanian Plain (= Wallachian Plain), Central and Southern Dobrudgea (all in southern Romania) and the pre-Balkanian Plateau (northern Bulgaria). The Romanian part of the Moesian Platform is usually divided in four sectors: the west-Wallachian sector, the central Wallachian sector, the east-Wallachian sector and the Dobrudgean sector.

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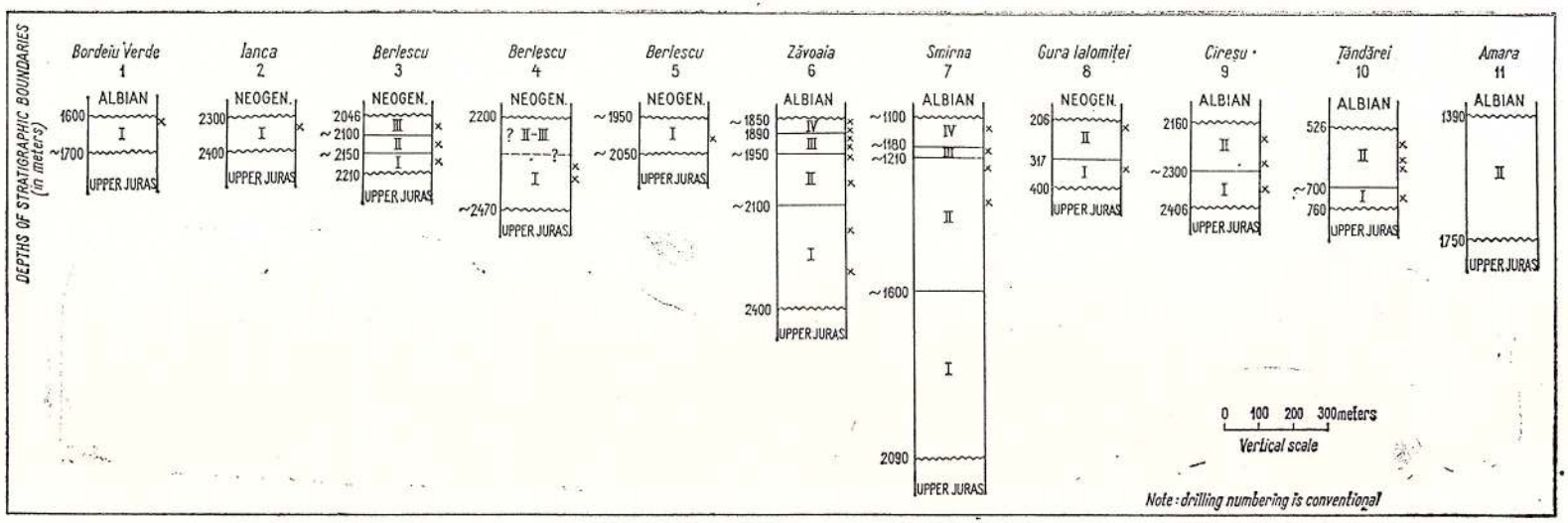


Fig. 1. — The Neocomian sections crossed by the drillings in study area. IV = the variegated marly-calcareous complex (out of discussion); III = the reefal limestone complex; II = the facies-zoned complex; I = the micritic limestone complex. x = position of the mechanical cores extracted from the Neocomian sections.

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ling no. 7), Gura Ialomiţei (drilling no. 8), Cireşu (drilling no. 9) and Ţăndărei (drilling no. 10) (the drilling numbering is conventional) (Fig. 1 and 2).

The concerned stratigraphic interval includes deposits resulted from an autochthonous sedimentation, consisting of limestones, dolo-

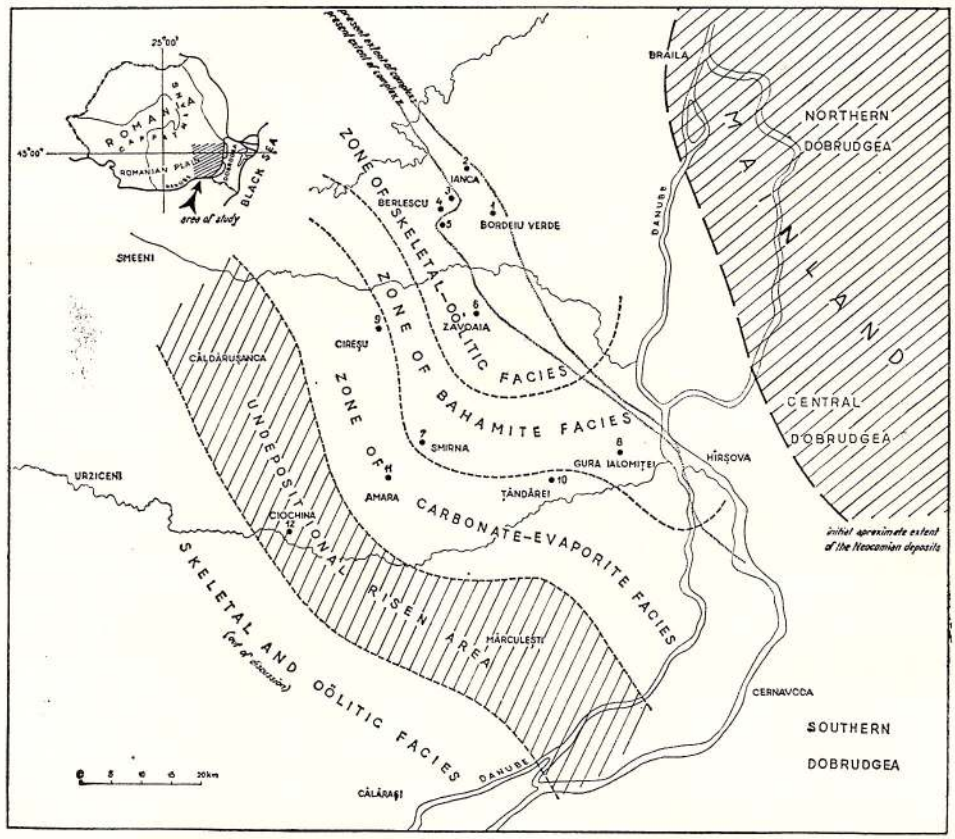


Fig. 2. — Facies map in the east-Wallachian sector of the Moesian Platform (eastern Romanian Plain) including the drilling locations, the ancient and present extents of the I and II Neocomian sedimentary complexes, and the sedimentary facies distribution during the deposition of II complex.

mites and evaporites; the terrigeneous supply is represented only by rare and weak influxes of silty quartz grains and argillaceous material.

The age of this sequence may be established by analising the supra- and subjacent formations as well as by interpreting the data provided by studied sequence itself. Thus, as the variegated marly-calcareous series, overlying the under-discussion sequence, is assigned to the

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Hauterivian stage (by comparison with southern Dobrudgea and northern Bulgaria; Patrulius and Tocorjescu, 1962; Drăgănescu, 1971 [4]), and the Upper Jurassic limestones and dolomites, standing for the basement of the concerned sequence, certainly include, besides the Oxfordian stage, the Kimmeridgian one (by comparison with central Dobrudgea as well as on faciological and paleontological grounds; Patrulius, unreported data; Drăgănescu, 1971 [5]), the discussed sequence results to be placed in the Tithonian-Valanginian time range; but, inasmuch as there are sufficient reasons to accept also the presence of the Lower Tithonian (Danubian) substage within the Upper Jurassic dolomitized limestone series (Patrulius, unreported data; Drăgănescu, 1971 [6]), the sequence under-discussion appears restricted to the Upper Tithonian (Ardescian)?-Berriasian-Valanginian range. This opinion is also supported by evidence provided by discussed sequence itself: the basal term, represented by micritic limestones bearing dolomite rhombohedra and considered as the first term of the Neocomian section in different areas of the Moesian Platform (Patrulius, 1964 a; Patrulius, unreported data), contains, in Zăvoaia sector (drilling no. 6, 2 295—2 296 m depth interval), an assemblage of calpionelids [*Crassicollaria massutiniana* (Colom), *C. parvula* Remane, *Tintinnopsella carpathica* (Murg. and Fil.), *Crassicollaria* cf. *intermedia* (Durand Delga)] corresponding to the *Crassicollaria* biozone and suggesting an Upper Tithonian age (calpionelids determination by Pop); the uppermost term is represented by reef limestones, whose facies is analogous to that of the Valanginian beds of southern Dobrudgea. Accordingly, the most plausible age for the entire under-discussion sequence, in present author's opinion, is the Berriasian-Valanginian one, possibly Upper Tithonian for its lowermost part. This assignation is in a good agreement with the more general point of view advanced by most of the previous researchers, according to which the discussed sequence belongs to the Neocomian stage.

A short bibliographic review reveals that the different studies bearing on the Neocomian sequences crossed by drillings in the east-Wallachian sector of the Moesian Platform pointed to the occurrence of some marine calcareous facies and of a peculiar one; this latter facies consists of coprolithic limestones, dolomites and evaporites and was thought the product of the temporary installation of some evaporite-lagoonal conditions in connexion with the emerged neighbouring territory of central and northern Dobrudgea. General remarks appear in the synthesis

[4] A. Drăgănescu. Studiul forajelor adînci Ţăndărei şi Gura Ialomiţei din platforma moesică (The Study of the Deep Drillings of Ţăndărei and Gura Ialomiţei in the Moesian Platform; unpubl. report). 1971. Arch. I.G.G., Bucharest.

[5] *Op. cit.* 4.

[6] *Op. cit.* 4.

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study made by Barbu et al. (1966)[7] and in the the paper of Popescu et al. (1967). Summary and strictly descriptive lithological-petrographical studies (bulletins of analysis and unpublished reports) were made also by some departments for petrographical analysis. Analysing the Neocomian series pierced by drillings in the southern central-Wallachian and east-Wallachian sectors, Patrulius (1964, and unpublished data) concludes that in some areas marine calcareous sequences have been accumulated, while in others — alternating marine limestones and coprolithic limestones bearing dolomite and sulfates have been deposited. This author estimates the latter lithologic assemblage as an expression of a marine facies including evaporite-lagoonal episodic reccurrences. In the lithofacies map of the Romanian territory for the Neocomian stage, elaborated in collaboration with Ştefănescu, Patrulius (1969) achieved the first areal figuration of the facies in the east-Wallachian sector; he distinguishes a mixed, marine-hypersaline lagoonal, carbonate-evaporite facies as a strip generally oriented NW-SE, bounded to the east by the mainland of central and northern Dobrudgea and to the west by normal marine, calcareous facies.

The researches that I undertook (Drăgănescu, 1969 a[8], 1969 b[9], and 1971[10]) on the drillings performed in the Moesian Platform revealed to me a lot of details giving much complication and somehow altering the previously accepted general image concerning the types and the distribution of the sedimentary facies during the Neocomian time interval in the east-Wallachian sector. This is the reason which determined the present author to resume this stratigraphic interval in a systematic study purposing to establish as detailed and exact as possible the lithostratigraphic sequences crossed by the mentioned drillings and then to correlate them in order to obtain, on the basis of a minutious petrological analysis, the picture of the sedimentary pattern for each lithostratigraphic unit over the discussed area. Finally comparing these sedimentary patterns with doubtless data supplied by recent carbonate and carbonate-evaporite marine sediments (and also by some ancient similar formations), the author attempted clearing up the genetic significance of these patterns. In addition, besides the approach of the depositional aspects, the main diagenetic processes are outlined.

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[7] C. Barbu, E. Vasilescu, A. Sterescu, T. Deicin, C. Paraschiv, O. Oprea, S. Dicescu and Gh. Dumitrescu. Studiul geologic complex al platformei moesice (The Complex Geological Study of the Moesian Platform; unpubl. report). 1966. Arch. I.C.P.G., Bucharest.

[8] A. Drăgănescu. Studiul stratigrafic al forajelor din sectorul nord-estic al platformei moesice (The Stratigraphical Study of the Drillings Performed in the NE of the Moesian Platform; unpubl. report). 1969 a. Arch. I.G.G., Bucharest.

[9] A. Drăgănescu. Studiul forajelor adînci Cireşu, Smirna şi Zăvoaia din platforma moesică (The Study of the Deep Drillings of Cireşu, Smirna and Zăvoaia in the Moesian Platform; unpubl. rep.). 1969 b. Arch. I.G.G., Bucharest.

[10] *Op. cit.* 4.

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Although the petrologic material examined was quantitatively not rich, the total number of the cores extracted from the Neocomian intervals reaching only a few tens [11], the results provided many interesting and conclusive data as regards the manner of development of the Neocomian sedimentation in the east-Wallachian sector.

In the Neocomian section of the east-Wallachian sector, present author distinguishes 4 lithostratigraphic units (from bottom to top) (Fig. 1) : I, the micritic limestone complex ; II, the facies-zoned complex; III, the reefal limestone complex ; IV, the variegated marly-calcareous complex. The purpose of present paper is to describe only the first three complexes, that is, the Upper Tithonian?-Berriasian-Valanginian interval, which displays an autochthonous sedimentation The variegated marly-calcareous complex, Hauterivian in age, is out of discussion.

The considered deposits are highly variable in thickness, decreasing radially from about 900 m in Smirna sector to about 100—300 m in Berlescu, Bordeiu Verde, Zăvoaia, Amara, Ţăndărei and Gura Ialomiţei sectors. Each of the component series (complexes I, II and III) displays variable thickness too. The variations in thickness were caused by temporal and spatial changes in the subsidence intensity as well as by a subsequent, partial or total, removal of Neocomian upper terms in some areas owing to the erosional stages installed in connexion with the Austroalpine and Early Austrian movements. The diastrophic and subsidential aspects in the east-Wallachian sector are discussed in detail by present author in the above-mentioned unpublished reports. Present paper approaches only the aspects of a petrologic-faciologic order. Worth specifying is that the Neocomian sedimentation represents a distinct cycle bounded by two depositional disconformities : the lower disconformity lies at, or slightly below the Jurassic/Cretaceous boundary, underlined in placed by an erosional gap lying in the Upper Jurassic series ; the upper disconformity, which is manifest in the Hauterivian-Aptian range, is partly a depositional gap and partly an erosional one.

I. THE MICRITIC LIMESTONE COMPLEX

This complex stands for the first term of the Neocomian sedimentation and overlies the erosional-sculptured surface of the Upper Jurassic reef limestones and dolomites. It was identified in the following sectors : Smirna (about 500 m in thickness), Cireşu (at least 60 m), Zăvoaia (about 300 m), Ţăndărei-Gura Ialomiţei (about 50—100 m) and Berlescu-Bordeiu Verde (50—200 m) (Fig. 1 and 2). Directly west of all these sectors, namely in Amara and Ciochina sectors, the drillings pointed to the lack of this complex (the drillings no. 11 and 12) ; here the Upper Jurassic dolomitized series underlies either the facies-zoned complex (drilling no. 11) or the Hauterivian marly-calcareous complex (drilling

[11] The lithologic control between the cored marchs was offered by the electric logs and sieve samples.

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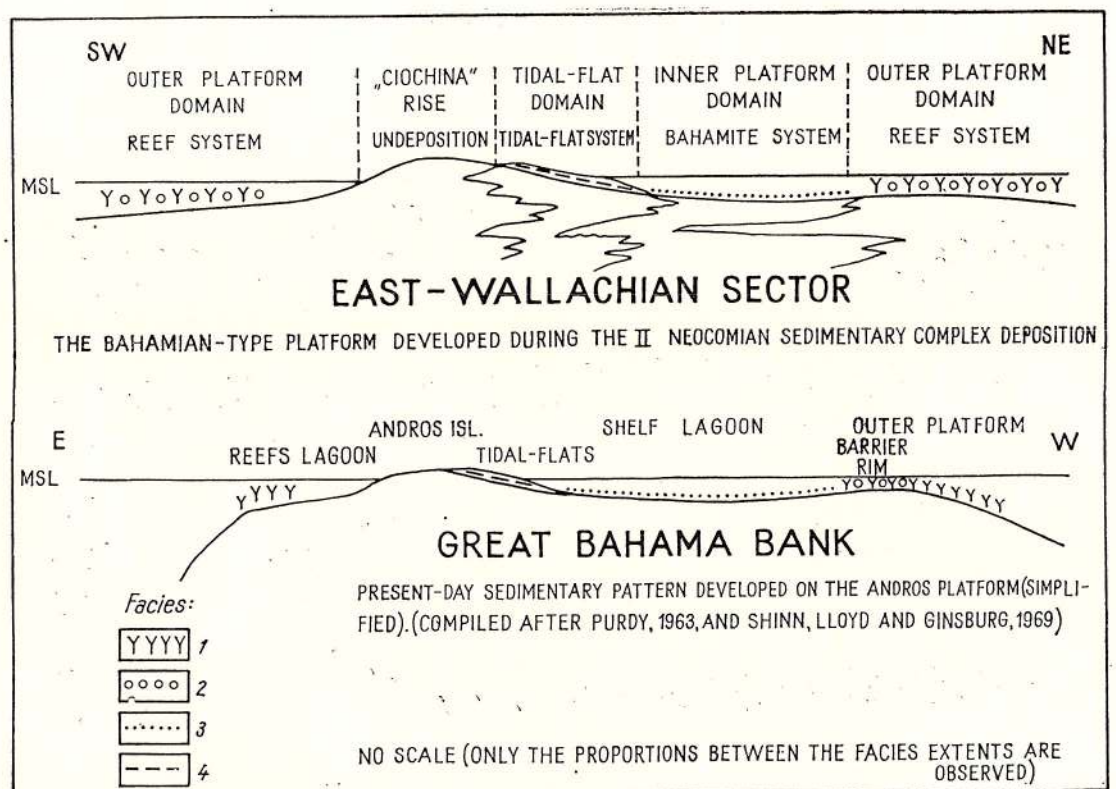


Fig. 3. — Comparison between the present-day carbonate sedimentation on the Great Bahama Bank and the Neocomian sedimentation in the east-Wallachian sector of the Moesian Platform during the deposition of II complex.

1. skeletal facies; 2. oölitic facies; 3. bahamite facies; 4. tidal-flat facies.

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no. 12). This finding claims accepting the existence of a threshold of the Upper Jurassic morphologic surface at the time of the micritic complex deposition, this threshold bordering the discussed depositional area to the west and SW. The threshold, oriented NW—SE, is thought to have been aligned lengthways the localities of Căldărușanca—Ciochina—Amara —Mărculești. This threshold behaved as an undepositional area, slightly higher elevated relatively to the surrounding bottoms, either in a submerged position or in an emerged one as a low, wide ridge. Inasmuch as central and northern Dobrudgea were emerged during the Neocomian time interval, according to the present general opinion, the discussed depositional area appears as a large bay clossed to the west, south and east, and widely opening towards the north.

The micritic limestone complex consists mainly of micrites, most of them more or less dolomitic, and subordinately of skeletal, micritic limestones, skeletal, micritic, pelletal limestones, micritic pelletal limestones, and other allochemical varieties in quite minor amounts (nomenclature Bissell and Chilingar, 1967) [12].

The micrites are usually homogeneous, seldom grumous, and consist of cryptocrystalline calcite particles (less than 4 micra in size). Dolomite rhombohedra frequently occur within the micrite matrix lending a slighter or a stronger dolomitic character to the limestone. The dolomite rhombohedra-bearing micrites represent the commonest petrographic type within complex I (Pl. III, Fig. 1—5; Pl. VI, Fig. 1—2; Pl. VIII, Fig. 1—2; Pl. XIII, Fig. 1—4). The dolomite rhombohedra are usually 30—125 micra in size. They occur in variable amounts within micrite, from scattered crystals to compact aggregates. The dolomite crystals are either apparently clear, free from impurities, or turbid, including calcareous dust dispersed in an approximately even distribution within crystals; a few micrite layers contain rhombohedra displaying only the cores occupied by rhomb-shaped patches of fine-grained lime dust; all these fine calcareous inclusions represent relict micritic material and suggest a replacement origin for the dolomite crystals.

The skeletal, micritic limestones are sporadically spread throughout the micritic limestone complex. They are mainly fossiliferous micrites and subsidiary sparse biomicrites (nomenclature Folk, 1962) (Pl. I, Fig. 1—2; Pl. IV, Fig. 1; Pl. VI, Fig. 3); both of these petrographic types show the skeletal grains floating within micrite matrix. Dolomite rhombohedra occur, in a variable proportion up to 50 volume percent, as scattered crystals or aggregated in irregular-shaped patches (often evolved to very fine spar calcite by dedolomitization). The biogenic material consists of ostracods, echinoderm plates, calcispheres (partly being calcitized radiolarians), fibrospheres (cadosinids and stomiosphae-

[12] In order to term the carbonate rock types as suggestive and exact as possible, I used in each situation the most adequate name, able to express at best the main features of the rock; the names proceed chiefly from Folk's (1959, 1962), Dunham's (1962) or Bissell and Chilingar's (1967) classifications.

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rids), calpionelids and *Aptychus debris*. The ostracods, the calcispheres and the echinoderm ossicles occur over the entire depositional area, while the *Aptychus debris*, the calpionelids and the fibrospheres have been encountered only in the northern part of the studied area, suggesting that the depositional environment gets more pelagic in facies northward (a fact confirming the interpretation on the northward opening of the depositional area). For the entire depositional area, the described skeletal material suggests a sedimentation in a relatively low turbulence (at least at the sediment/water interface) marine environment; the water depth is thought to have varied from several meters in the southern part up to at most 100 m in the northern part of the depositional area.

The pelletal limestone varieties were encountered only in the northern sector, where they occur either as the main constituent of the micritic complex section in drilling 3 Berlescu or as episodic reccurrences in the dolomitic micrites of the complex I in drilling 1 Bordeiu Verde. As a rule, they are represented by more or less fossiliferous pelmicrites (micritic, pelletal limestones, micritic-pelletal limestones, and skeletal, micritic, pelletal limestones in Bissell and Chilingar's 1967 nomenclature) consisting subordinately of skeletal grains (fibrospheres, calcispheres, and ostracods) and mainly of pellets of a cryptocrystalline grain type (probably originating from the hardening of some mud-aggregates — nomenclature Purdy, 1963), spheroidal or slightly ovoidal in shape, 130—150 micra in size and with a homogeneous micritic internal structure, embedded in a grumous micrite matrix; in places, the pellets are directly merged into one another (as small levels and lenses of pelletal limestones „cemented" by replacive spar calcite; see details in Pl .II, Fig. 4—5). Minor amounts of pellets are bound together in irregular-shaped lumps up to 500 micra in size.

Small patches of pelletal-intraclastic limestones (intramicrites and intrasparites) occur within the pelletal limestones, as co-associated microfacies, where the pellets and the lumps (if present) are accompanied to a variable extent by elongate, angular to moderately rounded, arenite or microrudite in size, intraclasts with an internal micritic structure (micrite containing scattered fibrospheres — Pl. I, Fig. 5—8). Associated to the „intraclastic phenomenon", the oölitic accretion phenomenon manifested, the end-product of this microfacies type being some microlenses of "sparry", oölitic, intraclastic-pelletal limestones; the oölitic grains vary from superficial oöliths to true oöliths, as nuclei serving pellets or finaly arenitic intraclasts (Pl. I, Fig. 6—8).

The spar calcite within the above-mentioned „sparry" allochemical varieties of limestone although apparently looks like a pore-filling calcite, actually stands for a replacement product formed by a diagenetic sequence involving dolomitization-dedolomitization-aggrading recrystallization (see details in the explanation to Pl. I—II). Worth emphasizeing is that this diagenetic sequence stands for a new postdepositional processual line and was noticed by present author in numerous limestone series of different ages; it results in dedolomitic microspar and dedolo-

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mitic pseudospar (new terms) which highly mimmic the neomorphic as well as the precipitative (orthospar) calcite varieties (D r ă g ă n e s c u, in preparation) [13]. Another conclusive example for this new diagenetic sequence, offered by the allochemical limestones of complex I (drilling no. 3 Berlescu), is represented by which seems to be at first sight a-„depositional perturbation facies", consisting of a gradual scale of petrographic terms from „disturbed" pelmicrites with irregular-angular spar calcite eyes („dismicrites") up to completely "broken" pelmicrites into equidimensional, angular or subangular, arenitic or ruditic in size, „intraclasts" embedded in spar calcite ("intrasparites"); this gradational suit may be watched in thin-sections along one or two centimeters, and actually it represents the product of the above-mentioned diagenetic sequence, where the dolomitized patches have undergone a dedolomitization (replacement of dolomite by microspar) followed by an aggrading recrystallization of dedolomitic microspar to larger pseudospar (Pl. II, Fig. 1—3). The micritic limestones, more or less free from allochems, were also affected in places by the dolomitization-dedolomitization sequence resulting in a microspar varying in participation and distribution from tiny patches irregularly scattered within micrite (leaving the appearence of a clotted texture) to a network enclosing some unaffected areas (pseudointraclasts) (Pl. I, Fig. 1—4). But, as already stated, most of the micritic limestones have undergone dolomitization only, without to be further submitted to dedolomitization.

As regards the micrite in this complex, it occurs as both independent rock (micrite = calcilutite = mudstone) and matrix in some allochemical limestones. It originated from the lithification of a former lime mud; this mud has most probably resulted from physico-chemical precipitation processes and seems to represent an ancient equivalent of the recent aragonite muds occuring in the inner, sheltered parts of the Bahama Banks, particularly of Great Bahama Bank. On the Bahama platforms the aragonite muds form in shallow, quiet, warm water environments by chemical precipitation, without any biogenic influence, on account of the slight increase in the seawater salinity over the banks in response to the restrictions in water circulation and to the strong evaporation caused by the warm and dry climate (C l o u d, 1962). This slightly increased salinity is enough to bring the seawater at the supersaturation level, with respect to its calcium carbonate content, required by $CaCO_3$ precipitation in the mineralogic form allowed by the seawater Mg/Ca ratio; calcium carbonate separates from the seawater mass as cryptocrystalline aragonite needles only (C l o u d, 1962; T a f t, 1967; W i n l a n d, 1969; etc.). Inasmuch as the climatic conditions prevailing over the Moesian Platform during the Neocomian time were of a tropical-

[13] A. D r ă g ă n e s c u. Dolomitization-dedolomitization-aggrading Recrystallization — a New Postdepositional Processual Sequence in the Shallow-water Marine Limestones and its Petrological Implications (in preparation).

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subtropical type (as revealed by the reefal faunas occurring on the Romanian territory in the Upper Jurassic-Aptian stratigraphical range) such depositional processes were quite possible. Furthermore, the semi-closed and epicontinental character of the depositional area discussed in present paper suggests a total similitude with regard to the depositional environment, conditions and processes and consequently to the mineralogic consistence of resultant sediments, between most of the discussed area and the inner parts of the Bahama Banks. Thus the micritic limestones of complex I are thought to represent ancient aragonite muds formed in a quiet, relatively shallow water environment and express a typical lime mud facies (sensu P u r d y, 1963). The skeletal material, by its composition and distribution, also suggests sheltered depositional conditions with basinal influences in the northern part of discussed area.

As regards the episodic pelletal facies and its associated microfacies, occuring in the micritic section of the northern part of the studied area, their depositional conditions seem to have been slightly different from those previously discussed. The tendency of micritic mud to agglutinate into pelletal grains, the occurrence of more or less washed pelletal and intraclastic episods, and the manifestation of the oölitic accretion phenomenon, all these facts suggest a higher energy level for the depositional environment comparatively to the (more or less skeletal) mud facies previously commented.

Mineralogically, the deposits of the complex I are thought to have been initially aragonitic in consistence, except for the skeletal grains (mainly calcitic). But as long as the aragonitic sediments are maintained under the seawater influence they preserve their rhombic, metastable mineralogic form; in this state the sediments may undergo dolomitization by penecontemporaneous or early diagenetic (at shallow depths of burial) replacement of aragonite by dolomite (I l l i n g et al., 1965; D e f f e y e s et al., 1964, 1965; S h i n n et al., 1965; L u c i a, 1968; M u r r a y, 1969; etc.). The dolomitization noted in the limestones of complex I seems, taking into account the large size of the dolomite crystals comparing with the extremmely small crystal size of present-day penecontemporaneous-replacement dolomites, to stand for a case of early diagenetic dolomitization brought about probably by a process of seepage refluxion (sensu A d a m s and R h o d e s, 1960) completed by a further increase of the Mg/Ca ratio in the seawater-derived, Mg-rich, dolomitizing, interstitial solutions percolated into the calcareous sediments. Later on, the dolomite was submitted in places to a late diagenetic sequence: dedolomitization-aggrading recrystallization (that is, replacement of dolomite by dedolomitic microspar and aggradation of this microspar to larger pseudospar).

Faciologically, the complex of micritic limestones expresses a marine sedimentation in a shallow, quiet water environment under a lime mud facies (sensu P u r d y, 1963), with reccurrences of episodic pelletal and intraclastic-oölitic facies suggesting a more agitated environment. The last two facies temporary replace in places the lime mud facies in the northern part only of the east-Wallachian sector, in other words exactly

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in that part where the depositional area widely opened up being largely linked to the rest of the Neocomian basin. Therefore, the sedimentation took place on an extensive inner platform domain (= inner shelf domain) under a prevalent lagoonal, lime mud facies, with some local tendencies of accumulation under outer platform facies in the northern part of the depositional area. The occurrence of calpionelids and stomiospherids in the northern part of the area suggests some basinal influences in the northern sectors of the carbonate platform; hence, the possibility for the inner platform domain to have been directly linked to the basinal one, the outer platform domain (outer shelf domain) having been slightly manifest only in places at the join of the two extreme domains.

II. THE FACIES-ZONED COMPLEX

The facies-zoned complex, laid down apparently in depositional conformity with the underlying, lower complex, displays a variable thickness: 100—200 m thick in Ţăndărei and Gura Ialomiţei sectors, about 400 m in Smirna sector, at least 300 m in Amara sector, about 150 m in Cireşu sector, at least 100 m in Zăvoaia sector and 50—100 m in Berlescu sector (Fig. 1). The Căldăruşanca-Ciochina-Mărculeşti threshold continues to behave as an undepositional risen area (the Upper Jurassic limestones and dolomites directly underlie the Hauterivian marly-calcareous complex along this threshold).

Unlike the lower complex, present complex exhibits a large petrologic variety, the sections crossed by the borings pointing to an interesting and evident facies zonation (Fig. 2): laterally interfingering facies zones succeeding as belts that parallel the Căldăruşanca—Ciochina—Mărculeşti threshold, each of them observing a semicircular alignment with the concavity facing NNE. Their drawing relies on the finding that from S and SW towards N and NE the following petrologic assemblages succeed: evaporites subordinately containing dolomite and micrite intercalations (Amara sector); evaporites, micrites, dolomites and microcoprolithic-microoncolitic calcarenites (Cireşu and Ţăndărei sectors); poorly skeletal micrites and microcoprolithic-lumpal, sometimes oölitic, calcarenites (Smirna sector); poorly skeletal, lumpal-pelletal calcarenites (Gura Ialomiţei sector); oölitic calcarenites, bioconstructed and corpuscular skeletal limestones (Zăvoaia and Berlescu sectors). This succession, displaying a notable gradual scale of substitution of the terms from an assemblage to the next one, argues for distinguishing 3 laterally interfingering facies zones, disposed from the "Ciochina" threshold (the margin of the discussed depositional area) towards the interior of the discussed depositional area (that is, towards N and NE) as follows: A) the zone of carbonate-evaporite facies; B) the zone of bahamite facies; C) the zone of skeletal-oölitic facies.

From genetic point of view these facies zones are thought to have the following significances:

A) The deposits included in the carbonate-evaporite facies zone represent a tidal-flat system being the product of a cyclic sedimentation

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under a dry, warm climate on the coastal plain extending on the north-eastern flank of the "Ciochina" threshold and submitted to the mean sea-level periodic fluctuation.

B) The bulk of the deposits of the bahamite facies zone makes up a carbonate-lagoonal system, resulted from a marine sedimentation on an inner platform domain (sheltered, shallow water environment).

C) The deposits of the skeletal-oölitic facies zone make up a reefal system, evolved from a marine sedimentation on an outer platform domain (agitated, shallow water environment).

Therefore, from the emerged "Ciochina" threshold towards NE, that is, towards the off-shore parts of the discussed depositional basin, 3 depositional domains extended : the shore zone was occupied by the tidal-flat domain ; this was followed seaward by the protected, inner domain of the continental shelf ; this latter domain was contiguous seaward with the unsheltered, outer domain of the shelf. All these three domains stretched on a shelf whose surface was near the mean sea-level in the south-western (peripheral) parts of the discussed depositional area, and slowly sank towards N and NE, but keeping at very shallow depths (depths more probably in the range of meters than of tens of meters).

The disposition of these 3 domains is similar to that noted in the present-day low latitude areas with a marine, shallow water, carbonate sedimentation. This similitude is striking especially when compared with the Andros Platform (Great Bahama Bank), whose facies pattern and asymmetric morphology (the flank west of Andros Island very extensive and exhibiting all of the 3 foregoing sedimentary domains, unlike the flank east of the island much shorter and displaying the reefal facies nearly joined to Andros Island) seem to be reproduced by the present case (Fig. 3). Thus, data provided by drillings point to the fact that directly west of "Ciochina" threshold the stratigraphic interval corresponding to the facies-zoned complex is occupied by a reefal, skeletal-oölitic facies (the borings of Călărași, Urziceni, etc.). Consequently, a cross section made SSW—NNE through the discussed area shows a carbonate platform with an oblong flank (the north-eastern one) exhibiting the complete carbonate depositional pattern developed on the flank west of Andros Island, and a short flank (the south-western one) confined to a reefal facies only, similarly to the flank west of Andros Island ; the equivalent of Andros Island seems to be the "Ciochina" threshold. From another angle, the facies distribution displays typical features of a carbonate platform of a bahamian type, as it was synthetically conceived by P u r d y (1963), the bahamite (nomenclature B e a l e s, 1958) facies zone representing the sheltered, inner platform facies placed between the two, outer platform, skeletal-oölitic facies (to these ones, the tidal-flat facies adding as the product of the interaction between a risen area occuring within the inner platform domain and the cyclic oscillation in the mean sea-level).

Further on, the petrologic data which have served as a basis for the afore exposed sedimentological interpretations are presented ; the 3

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facies zones will be successively approached in the following order : the carbonate-evaporite facies zone, the bahamite facies zone and the skeletal-oölitic facies zone.

A) The Zone of Carbonate-Evaporite Facies

The carbonate-evaporite facies, recorded in Țăndărei, Cireșu and Amara sectors, lies directly east of the undepositional area of the "Ciochina" threshold, parallel and closely joined to it. This facies includes limestones, dolomites and sulphates (mainly of calcium).

In Țăndărei sector (drilling no. 10) clayey micrites, microoncolitic-microcoprolithic calcarenites with *Favreina salevensis*, evaporites and dolomites occur (Pl. VII—XII and XVII—XIX).

In Cireșu sector (drilling no. 9) micrites, pelmicrites, microcoprolithic calcarenites with *Favreina salevensis*, and evaporites were intersected (Pl. XIII—XVI). The *Favreina salevensis*-bearing calcareous-evaporite facies of Cireșu was already briefly reported by P a t r u l i u s (1964 b) on the occasion of the study undertaken by this author on the Neocomian microcoprolithic limestones containing *Favreina salevensis* from Atîrnați drilling (the central-Wallachian sector of Moesian Platform) : the mentioned author used the association between the coprolith *Favreina salevensis* and evaporites at Cireșu as an argument supporting the interpretation that *Favreina salevensis* biofacies corresponds to a marine environment with an abnormal, slightly increased salinity, interpretation initially infered from the lack of typical marine organisms in the section of Atîrnați.

In Amara sector (drilling no. 11) a sequence consisting of evaporites with minor amounts of dolomite and micritic limestone was crossed.

For present work a petrologic study was accomplished only on the carbonate-evaporite series of Țăndărei and Cireșu sectors, the petrologic material from Amara sector having not been available to the author.

1. *I n Ț ă n d ă r e i s e c t o r*, the carbonate-evaporite sequence (crossed within the 526-about 700 m depth interval by the drilling no. 10) was mechanically cored in the depth intervals of 574—575 m, 620—621 m and 634—635 m.

a) At the depth of 574—575 m, a sequence of greenish, clayey, micritic limestones imbued with gypsum occurs ; it contains, at several levels, evaporite intercalations as irregular-shaped lenses or nodules consisting of gypsum, anhydrite and celestite.

The clayey micrites "invaded" by gypsum (Pl. VIII, Fig. 3—4) consist of cryptocrystalline calcite particles (in present case less than 3 micra in size) and clayey material, and contain rare ostracods. Gypsum occurs within host clayey micrite as colourless or whitish, flattened crystals varying from less than 1 mm to 1 cm in length ; the crystals occur either solitary or twinned or aggregated in small bodies whose angular outlines are given by crystallographic faces (crystallotopic structure of

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Maiklem et al., 1969 [14]). The gypsum crystals appear either pure, clear, or impurified with calcareous dust as disseminations. Some parts of the host micrite appear "broken up", "expanded", with the constitutive micritic particles slightly distanced from one another and poikilitically entrapped within gypsum cement optically representing single large crystals. Other parts of the micrite are obviously mechanically dissociated by the growth of gypsum crystals. In others, the micrite is completely invaded with gypsum crystals either disposed in random orientation or merged into irregular bodies; these networks and aggregates of gypsum crystals completely dismember the host micrite that appears partly enclosed as calcareous pieces or particles within large gypsum crystals and partly entrapped as calcareous tails and pieces (calcareous streaks in nomenclature of Maiklem et al., 1969) between merged gypsum crystals. All these observations suggest that: the clayey micrite was settled as clayey calcareous (probably aragonitic) mud containing scattered ostracods; the gypsum formation took place later, subsequently to the clayey lime mud deposition; the gypsum was laid down from interstitial superconcentrated solutions, within the still soft, clayey calcareous mud, either by carbonate replacement or, most frequently, by interstitial precipitation accompanied by mechanical displacement of the host mud, the gypsum crystals having grown by entrapping or by pushing aside the soft clayey calcareous mud.

Therefore, two depositional generations may be distinguished within a calcareous layer, belonging to two successive stages and representing two lithogenetic units: the calcareous unit (unit α), standing for the product of the first depositional stage, and the gypsum unit (unit β), representing the product of the second stage. Unit β is developed within unit α.

The lensoid or nodular evaporite intercalations, up to several cm thick, display either a nodular-mosaic to a mosaic internal structure or a massive one (nomenclature Maiklem et al., 1969), and consist of gypsum, anhydrite and, subordinately, celestite.

The gypsum occurs as flattened crystals 50—500 micra on the long side (therefore, much smaller than the gypsum crystals imbueing the micrites), aggregated into masses with a blocky texture (nomenclature Maiklem et al., 1969).

The anhydrite masses occur as irregular-shaped zones within the gypsum masses. Under the microscope they display a blocky-fibrous texture (new term) consisting of a mosaic of approximately rectangular or parallelogramic aggregates 50—500 micra long associated in a blocky texture, each aggregate exhibiting a parallel-fibrous internal texture made up of parallel lath-shaped crystals disposed perpendicularly to the long side of the aggregate. The anhydrite lath-shaped crystals are 30—100 micra in length by about 3—5 micra in diameter (Pl. VIII, Fig. 5). Between the anhydrite masses and the gypsum ones there are mineralo-

[14] In present paper the fabric nomenclature for anhydrite masses of Maiklem et al. (1969) was successfully used for gypsum masses too.

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gically mixed zones consisting of both gypsum and anhydrite. Under the microscope, these zones reveal interesting and eloquent facts concerning the anhydrite genesis : the gypsum crystals, approximately rectangular or parallelogramic in shape, are aggregated in a blocky texture and contain inside them variable amounts of lath-shaped anhydrite crystals; the anhydrite prisms accommodated within a gypsum crystal appear developed with their long axes approximately perpendicular to the large faces (010) of the host flattened gypsum crystal (that is perpendicular to the long sides of the gypsum crystal in thin-section); the anhydrite prisms within a gypsum crystal are neither exactly parallel to one another nor exactly perpendicular to the (010) faces of the gypsum crystal; a gradual scale of terms was noticed from gypsum crystals bearing a few anhydrite prisms through gypsum crystals "invaded" by anhydrite prisms up to anhydrite aggregates (with parallel-fibrous internal texture) similar in shape and size to the gypsum crystals. These observations support firstly that the gypsum-anhydrite association does not represent a syngenetic paragenesis (depositional concrescence) but an alteration assemblage caused by the replacement of gypsum by anhydrite, and secondly that the rectangular or parallelogramic aggregates making up the anhydrite masses represent anhydrite-after-gypsum pseudomorphs, each aggregate of fibrous anhydrite crystals corresponding to a single initial gypsum crystal. The turning of gypsum crystals into pseudomorphic anhydrite aggregates was already noticed or assumed in both present-day marginal-marine carbonate-evaporite deposits (Butler, 1969; Kinsman, 1969; Kendall, Skipwith, 1969; etc.) and ancient evaporite-bearing rocks of a different age (Stewart, 1953; Henderson, 1959; Kerr, Thomson, 1963; Murray, 1964; etc.). The present material supplies convincing arguments for the anhydrite-after-gypsum pseudomorphic replacement in ancient rocks and the same time it allowed establishing the exact manner which the replacement proceeded by to yield pseudomorphs of a textural type representing, to the author's knowledge, a new type of internal arrangement of the replacive anhydrite prisms within the replaced host gypsum crystal.

The gypsum masses locally contain abundant colourless or light-blue bipyramidate, short celestite prisms. The celestite prisms are 60—150 micra long by 30—60 micra in diameter, and appear developed in or between the gypsum crystals, suggesting a genesis quite similar to the anhydrite, i.e. by gypsum replacement (Pl. VIII, Fig. 6—7).

Therefore, the lensoid or nodular evaporite intercalations consist of primary gypsum, of a direct precipitation, and secondary anhydrite and celestite, proceeding from gypsum replacement, all these evaporites suggesting a deposition from hyperconcentrated solutions. The complex evaporite intercalations of each lithologic level make up an apart lithogenetic unit, formed during a stage subsequent to those responsible for the formation of the gypsum-bearing clayey calcareous mud accommodating the respective level.

The occurrence of several levels bearing complex evaporite intercalations (gypsum-anhydrite-celestite nodules or lenses) within the cal-

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careous-clayey-gypsum section suggests a cyclic sedimentation expressed by a succession of cyclothems. A complete cyclothem is assumed to be represented by a calcareous-clayey-gypsum layer enclosed between the tops of two successive levels bearing complex evaporite intercalations; consequently, a cyclothem would consist of a calcareous-clayey-gypsum layer displaying at the top of a level accommodating complex evaporite intercalations; a such cyclothem appears consisting of 3 lithogenetic units, corresponding to 3 successive depositional stages (this genetic interpretation keeps valid even if the accolades combining the lithologic elements into cyclothems within the discussed mechanical core are otherwise drawn) : α) the (more or less clayey) micrite unit; β) the gypsum unit, developed within unit α; γ) the gypso-anhydrite unit, deposited within and/or over unit α.

Within a cyclothem, a consistent upward increase in the concentration of the depositional solutions (that is, an increase in the restriction of the depositional conditions) is markedly evident.

The interpretation of the foregoing data themselves as well as the comparison with the present-day marginal-marine carbonate-evaporite sediments of Persian Gulf, Madre Lagoon (Texas), Baja California (Mexico), Bonaire Island (Netherlands Antilles) (Curtis et al., 1963; Shearman, 1963; Evans, Shearman, 1964; Illing et al., 1965; Butler, 1964, 1969; Kinsman, 1969; Kendall, Skipwith, 1969; Masson, 1955; Fisk, 1959; Phleger, 1969; Lucia, 1968) lead to the conclusion that the formation of cyclothems may be explained by two ways : either by a permanent supratidal sedimentation with periodical influxes of clayey-calcareous mud from the subtidal marine shallows owing to the flooding of the supratidal surfaces by seawater during the highest spring tides or storms, thus the requirement of the cyclic fluctuation in the salinity of depositing solutions being satisfied (every seawater influx meaning a dilution and a supply of lime mud, and the time interval elapsed between two consecutive floodings corresponding to a stage of drastic increase in salinity and evaporite deposition), or by a cyclic sedimentation of a tidal-flat type due to periodic oscillation in the mean sea-level. I accepted the latter genetic pattern as the most natural and plausible in present case. In accordance with the embraced genetic model, the lithogenetic units of a cyclothem formed as follows :

α. The deposition of the calcareous unit consisting of lime mud containing rare ostracods and clayey (most probably wind-blown) material, took place in a shallow-subtidal, protected, marine environment with a relatively normal-marine salinity.

β. The gypsum unit, precipitated within the soft mud of unit α from superconcentrated interstitial solutions, suggests intertidal genetic conditions, possibly upper intertidal. This genetic assigning relies on several facts. Firstly, in the present-day marginal-marine carbonate-evaporite sedimentation, gypsum precipitation as the first evaporite term from interstitial supersaline marine-derived solutions starts in the upper intertidal zones, going on upward, that is landward. Secondly, both the existence of a calcareous unit previously formed and that of a complex evaporite

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unit laid down subsequently and under more restrictive saline conditions, restrict the genetic environment to (upper) intertidal conditions. Thirdly, the large size of gypsum crystals supports salinities less drastic than those specific to a supratidal environment, pleading for an intertidal origin too.

γ. The gypso-anhydrite (complex evaporite) unit represents a typical supratidal product, laid down from seawater-derived hyperconcentrated brines accumulated over or within the clayey calcareous sediments of unit α (brought onto a supratidal position). The much more drastic values in concentration of depositing solutions under supratidal conditions caused the formation of complex evaporites, where, besides gypsum, the anhydrite and celestite occur. The initial mineral, directly precipitated from the hyperconcentrated solutions, was gypsum, which, owing to the high values of solution superconcentration, formed as tiny crystals, much smaller than those of the unit β. The anhydrite and celestite formed at the expense (by penecontemporaneous replacement) of gypsum. The anhydrite formation by gypsum replacement, as the genetic manner invoked in present case, represents a process ascertained on a large scale in the recent carbonate-evaporite sediments of the supratidal marginal-marine areas. The celestite formation relates, in the present-day carbonate-evaporite marginal-marine sedimentation around the Persian Gulf (E v a n s, S h e a r m a n, 1964; K i n s m a n, 1969), to the apparition within the supratidal carbonate-evaporite sediments of some seawater-derived pore-filling brines strongly enriched in strontium; this enrichement is related to some processes such as replacement of aragonite (7 000—9 000 ppm Sr) by dolomite (600—700 ppm Sr) or gypsum, and inversion of aragonite to calcite (less than 1 000 ppm Sr). The considerable amounts of strontium released into the hypersaline interstitial fluid by these processes, in the presence of $SO_4^{=}$ ions, generate celestite, either by direct precipitation from the solution or by replacement of previously formed gypsum; the gypsum substitution by celestite is favoured by the lower value of the celestite solubility product as against that of gypsum. In the present case the occurrence of the celestite as prisms disseminated within gypsum masses suggests a replacement process of gypsum by celestite under supratidal conditions, on account of the reaction of gypsum with seawater-derived hyperconcentrated and strontium-enriched interstitial solutions (enrichment probably resulted in the present case from replacement of aragonite of the host calcareous sediments by dolomite and/or gypsum, processes quite liable to have occurred taking into account the general depositional context).

b) *At the depth of 620—621 m* in Țăndărei sector, a sequence of thick (1—5 dm) and thin (0.2—1 dm) gypsum beds spaced by interbedded limestone beds 0.1—1.5 dm thick invaded by gypsum was transected.

The thick gypsum beds (1—5 dm thick) (Pl. VIII, Fig. 8—9; Pl. XVII, Fig. 4—6) consist of gypsum masses often associated with or embedded in greenish clay (the volumetric participation of the clayey material is usually highly subordinated to that of the evaporites). These gypsum

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beds display several structures (nomenclature M a i k l e m et al., 1969): massive structure, mosaic structure, nodular-mosaic structure and nodular structure. While the massive structure is nonnodular and free from clayey material, the last 3 structures belong to the "nodular" type and consist of gypsum nodules up to a few cm in size and clayey material filling the internodular spaces (these 3 "nodular" structures are defined in terms of the spacing of gypsum nodules within clayey mud). Texturally, these gypsum masses consist of flattened, small crystals 100—300 micra on the long side by 40—80 micra in thickness, aggregated in blocky, felted or aligned-felted textures (as they appear under the microscope) (nomenclature M a i k l e m et al., 1969). These thick gypsum beds are analogous to the gypso-anhydrite levels of the previously described core, and suggest deposition from seawater derived superconcentrated solutions on the surface of a supratidal area; the clayey material was probably derived from an eolian or coluvial redistribution of some possible residual clayey deposits born on the threshold of Căldărușanca-Ciochina-Mărculești. The masses of nodular, nodular-mosaic, or mosaic gypsum (Pl. XVII, Fig. 4—5) seem to have formed by precipitation as irregular-shaped nodules within the clay mud, roughly simultaneously to the deposition of this latter one; growing, the nodules reached finally different degrees of spacing or merging. The masses of massive gypsum (Pl. XVII, Fig. 6) suggest to have been laid down on the supratidal surfaces where or whenever the clay influxes were missing; consequently, they formed as pure evaporite soft sediments set down on the depositional surface and not as hard bodies grown within a sedimentary framework (clay muds) as is the case of the "nodular" types.

The thin gypsum beds (2—10 cm thick) (Pl. XVII, Fig. 1—2; Pl. XVIII, Fig. 4—7) are free from clayey material, display only nodular or nodular-mosaic structures and contain, between the gypsum nodules, calcareous fragments (sensu M a i k l e m et al., 1969) of the same petrographic type with the supra- and subjacent calcareous beds; the gypsum crystals are small and clear, devoid of calcareous material. These observations suggest that these thin gypsum beds (displaying "nodular" structures, lacking clay material and containing calcareous fragments) have formed under supratidal conditions similar to the thick gypsum beds, but unlike these latter ones they were born within previously formed quasiconsolidated calcareous sediments and not above them (as is the case of the thick gypsum beds that lack calcareous fragments, containing clay material in the "nodular" varieties instead). Accordingly, the thin gypsum beds represent intercalations within a limestone layer, even though now they appear as evaporite beds developed between limestone beds. Therefore, two or three limestone beds (0.1—1.5 dm thick) separated from one another by thin nodular nodular-mosaic gypsum beds actually (genetically) represent a single calcareous layer, subsequently dissociated by growth of supratidal thin evaporite beds inside it.

The limestone beds (subsequently invaded by gypsum of another generation than the foregoing gypsum beds) (Pl. IX—XII; Pl. XVII, Fig. 1—3; Pl. XVIII) are represented by calcarenites consisting of two

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types of allochems : microoncolites and fecal pellets (microcoprolithes). These calcareous grains contribute in variable proportions to the limestone bulk, leading to different petrographic rock types : pelletal calcarenites, pelletal microoncolitic calcarenites and microoncolitic calcarenites. Both of these grain types, initially of an aragonitic consistence (comparing to the similar recent products), display a micritic internal structure 1—2 micra in crystal size ; some grains show the constituent micrite partly or totally aggraded to a fine microspar (5—6 micra in crystal size).

The microoncolites (Pl. IX), thought as algal products of schyzophycean (blue-green) algae, appear either monoenveloped or polienveloped, with the nuclei, when preserved, represented by approximately spheroidal pellets. The microoncolites are 150—400 micra in diameter, usually lying in the 300—400 micra size range. The concentric micrite laminae making up the envelopes doubtless stand for algal products. As regards the origin of pelletal nuclei, I feel they cannot represent fecal pellets owing to the following reasons : the nuclei, more or less spherical, are completely different in shape from the fecal pellet types occurring in the limestone ; the oncolite-producing algal activity was concomitant to the production of pellets, inasmuch as the latter do not bear the hallmark of the algal activity, appearing devoid of micritic envelopes or other peculiar traces suggestive of an algal intervention. The pelletal nuclei probably originated from processes of algal precipitation-agglutination and seem to represent the embryo-colonies of the blue-green algae which subsequently yielded the oncolite-type deposition. The nuclei may be defined as algal mud-aggregates. The nucleus often is missing, its place being occupied by an area, apparently monocrystalline, of gypsum ; the gypsum deposition have been accomplished by replacement of the nucleus or by infilling of the void resulted from a previous leaching of the nucleus. In many microoncolites, the micrite laminae of the envelopes are distanced from one another on account of the development of concentric or wedge-like gypsum strips between them; in a more advanced stage the envelopes appear disrupted, radially cracked, the voids between the resultant fragments being filled up with gypsum ; further on, in extreme cases, the microoccolites, completely dissociated, dismembered within a gypsum mass, appear as phantomatic dusty cvasicircular bodies consisting of spaced calcareous particles. In all above-described situations, the gypsum areas developed within microoncolites display single optical orientations over large surfaces in thin sections.

The fecal pellets (Pl. X and XI) belong to 3 types : *a*) large, keg-shaped fecal pellets, 600—800 micra long by about 300 micra in diameter, exhibiting internal longitudinal tubes and belonging to the species *Favreina salevensis* (P a r é j à s), considered to represent coprolithes of decapode crustacean (P a r é j à s, 1935, 1948 ; B r o n i m a n, 1955) ; *b*) small ovoidal fecal pellets probably produced by worms (polychaete annelids) ; *c*) small, elongate-cylindrical fecal pellets probably standing for dejections of gastropods (the last two fecal pellet types were genetically assigned by comparison with the fecal pellets described in the recent carbonate sedi-

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ments of the Great Bahama Bank; Korniker, Purdy, 1957; Cloud, 1962).

The binder of the calcarenites is represented by both micrite matrix and spar gypsum cement (instead of spar calcite cement).

The micrite matrix fills about 30—40 volume percent of the intercorpuscular space and display, when unaggraded, a crystallinity equal or slightly larger than that of calcareous grains (1—3 micra).

Much of the micrite matrix as well as some of the allochems are evolved (Pl. XII) to a fine microspar (5—6 micra in crystal size). A short parenthesis is required for discussing briefly this microspar, which stands for a worth analysing case as it might contribute, by its particular features, to a better understanding of both the manner and the moment of development of one of most important and still rather obscure processes: the micrite-to-microspar neomorphism. Present microspar patches do not stand for a depositional material but represent the product of a postdepositional process, because they occur developed within both micrite matrix and calcareous grains, often transecting the depositional fabric, i.e. lying at the same time on several close grains or on grains and the matrix between them; the preferentially replaced material seems to have been the micrite matrix. On the other hand, the above mentioned gypsum cement has obviously formed subsequently to the enlargement of the micrite fabrics to microspar, because now the sulfate cement dissociates at different degrees the microspar patches from pieces with torn outlines to crystal-by-crystal dismembrations with the tiny microspar crystals floating within gypsum cement, this latter often consisting of a single large crystal. Accordingly, at least the recrystallization (without inversion of aragonite to calcite) of the micrite particles to microspar crystals proceeded before the sediment cementation by gypsum has taken place, that is certainly before the calcareous sediment be taken out of the seawater influence. But as long as the calcareous sediment was kept under seawater influence, it preserved its aragonitic consistence and soft state. Therefore, the microspar resulted from an extremely early (penecontemporaneous?) aggrading recrystallization of the cryptocrystalline aragonite particles composing the calcareous mud and grains. This process must logically resulted in a soft, aragonitic microspar. Probably this early aggrading recrystallization in soft, aragonitic state was followed much later, during late diagenetic stages, by the aragonite-to-calcite inversion. Consequently, this microspar raises several problems difficult to explain such as: the aragonite micrite-to-aragonite microspar recrystallization in soft state and within a soft sediment kept under the seawater influence; the aragonite-to-calcite inversion in isolated microspar crystals already entrapped within crystalline media such as the large gypsum crystals in present case (Pl. XII).

Coming back to the micrite matrix, this one occurs in uneven distribution within a limestone bed (Pl. XVII, Fig. 1), resulting in an irregular alternation of calcarenites (grainstones), devoid of micrite matrix, and lutaceous calcarenites (packstones and wackstones), displaying the intercorpuscular spaces filled up with micrite. The micrite matrix shows

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no preferential association to one or other of the two calcareous grain types — the pellets or the microoncolites. The vertical variation in the micrite matrix participation seems to mirror rather the variation in the pelleting rate of the mud by the pellet-makers and the variation in the supersaturation level of the seawater with respect to its $CaCO_3$ content, than the variation in the energy level of the depositional environment. The micrite matrix is analogous to the marine aragonite muds forming now, mainly by physico-chemical precipitation, in protected shallow water environment (the carbonate-lagoonal environments in Bahamas, Persian Gulf, etc.).

Some calcarenite levels contain large, elongate and flattened intraclasts (1—2 cm long by 3—5 mm thick) consisting of micrite bearing scattered fecal pellets and microoncolites, therefore displaying features characteristic of flat-pebbles (Pl. XVIII, Fig. 1—3). They probably resulted from the penecontemporaneous reworking of some of the calcareous sediments that, accumulated on the intertidal flats, have undergone a slight consolidation by compaction and, further, by deshydration, have broken down into flattened pieces; the merely semiindurated state of the pebbles at the moment of redeposition is demonstrated by the observation that the calcareous grains of the host calcarenite have imprinted themselves on the pebbles surface, slightly penetrating them.

The core from the depth of 620—621 m shows also, at the bottom of a limestone bed (Pl. XVII, Fig. 1; Pl. XVIII, Fig. 3—6), the occurrence of a level 2—3 cm thick consisting of lutaceous calcarenites displaying a slightly waved, laminar structure with thin, elongate laminoid fenestrae developed in places between the laminae (laminoid fenestral fabric, type A?; Tebbut et al., 1965); each of the laminae is made up of the same, above-described fecal pellets and microoncolites entrapped within a dense (when unaggraded) micrite matrix. This level, displaying features of an algal stromatolite, points to the existence, in places, within the sedimentary succession, of the deposit usually typical of the tidal-flat domain namely the stromatolitic algal mats. The laminoid fenestrae are disposed almost horizontally or slightly waved, upward converging only in the small protuberances of the algal crust (Pl. XVII, Fig. 1 and Pl. XVIII, Fig. 3, just below the calcarenite/stromatolite contact).

The limestone beds express a carbonate-lagoonal sedimentation of a pellet mud facies on a sheltered, inner platform domain in shallow subtidal and lower intertidal environments with salinities slightly exceeding the normal marine one (by comparison with the present-day depositional areas displaying similar sediments; Illing, 1954; Cloud, 1962; Purdy, 1963; Illing et al., 1965; Kendall, Skipwith,1969; etc.).

Most of the binder in the limestone beds (about 60 — 70 volume percent of the intercorpusclar space) is represented by a sparry gypsum cement (Pl. IX—XI). This gypsum, developed inside (and not between) the calcarenitic beds, occurs as cement, small lenses and thin bands, and consists of flattened, large gypsum crystals (0.1—1 cm on the long

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side) which appear juxtaposited, approximately parallel to each other, consequently exhibiting approximately the same optical orientation (under the microscope). The large size as well as the approximately single optical orientation of the gypsum crystals account for the behaviour of the gypsum cements, lenses and bands as monocrystal on large surfaces in thin sections. In accordance with the microscopic evidence, the development of gypsum within the host calcarenite took place as follows : gypsum passive precipitation as spar cement of a single optical orientation within available intercorpusclar spaces, between the oncolite laminae, into the molds left by leaching of oncolite nuclei and into the empty internal tubes of the *Favreina salevensis* fecal pellets ; displacive and/or replacive further growth of gypsum crystals and/or patches at digfferent intensities resulting in local consumption of calcareous grains to different extents, detachment of the calcareous grains from one another, dissociation of the oncolites or pellets and/or replacement of some oncolite nuclei or envelopes by gypsum ; stil further, local growth of gypsum patches yielding compact bodies of gypsum as small lenses and/or thin bands both of them less than 1 cm thick and fibrous in texture, consisting of juxtaposited flattened crystals of the same optical orientation and identical to that of gypsum crystals belonging to the interstitial cement. These small compact bodies of gypsum display a streaky massive structure (nomenclature M a i k l e m et al., 1969), containing disseminations, patches and transversal bridges of micritic or corpusclar calcareous material (more or less dismembered into the constitutive micrite particles) (Pl. XIII, Fig. 1 ; Pl. XVIII, Fig. 1) ; the gypsum crystals are not clear, but include torn calcareous material (Pl. IX, Fig. 6—7) ; the gypsum body/host limestone boundary is not as sharp as is the case of the already-described thin gypsum beds with nodular or nodular-mosaic structures ; in the case of present small bodies of gypsum with massive structure the passage from host limestone to gypsum body is gradual, the limestone undergoing a gradual disintegration from a limestone with gypsum cement up to gypsum body enclosing calcareous material floating within or between the gypsum crystals. Therefore, this generation of gypsum formed within unconsolidated calcarenitic sediment, evolving from the cementation of host sediment to the strong dissociation of the latter by a local excessive growth of gypsum patches to form massive structures of small gypsum bodies; it suggets, considering its large crystal-size, the soft state of the host sediment, as well as the data reported on present-day marginal-marine sedimentation, a deposition under upper intertidal conditions from seawater-derived brines.

A second gypsum generation noticed within the calcarenitic beds and crossing the above-described first gypsum generation is represented by quasivertical gypsum veinlets (0.1—1 cm thick) of a fibrous structure, consisting of crystals disposed perpendicular to the veinlet walls (Pl. XVII, Fig. 3). The gypsum crystals are devoid of calcareous impurities. In places angular fragments of limestone are enclosed. These veinlets seem to have formed by the infilling of some dessication cracks(or other types of cracks) born in the consolidated gypso-calcareous sediment during its exposing

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to subaerial conditions on supratidal areas. The veinlets may affect one or more calcarenitic beds (and the interstratified gypsum beds).

Concluding on the 620—621 m depth interval in Ţăndărei sector, worth underlining is the cyclic pattern of the sedimentation, suggesting a deposition of a tidal-flat type. A complete cyclothem seems including 3 lithogenetic units: α) the calcareous unit; β) the early gypsum unit; γ) the late gypsum unit. The whole above-discussion suggests the following genetic pattern for the 3 units:

α. The calcareous unit represents the shallow subtidal-lower intertidal unit, suggesting a calcareous sedimentation of a protected, inner shelf type in a pellet mud facies with local formation of flat-pebbles and algal stromatolites on the intertidal flats.

β. The early gypsum unit represents the upper intertidal unit, including the first gypsum generation (deposited inside the still unconsolidated unit α, that is, the gypsum cement and the small gypsum bodies of a streaky massive structure developed within the calcarenitic beds.

γ. The late gypsum unit is the equivalent of the gypso-anhydrite unit described in the cyclothem type section of the previous core, represents the supratidal unit and consists of the thick gypsum beds (accumulated over the unit α, the thin gypsum beds of a mosaic or nodular-mosaic structure (developed inside unit α, quasiconsolidated) and, if correctly interpreted, the gypsum veinlets.

The supratidal unit clearly prevails over the other units within the lithologic section of the core.

c) *At the depth of 634—635 m* in Ţăndărei sector, the drilling no. 10 intercepted dolomites only (idiotopic mosaic texture and 40—90 micra in crystal-size), displaying a laminar structure consisting of irregularly waved laminae separated from each other by rows of elongate laminoid fenestrae (laminoid fenestral fabric = LFF, type A; T e b b u t t et al., 1965) (Pl. XI, Fig. 8; Pl. XIX). The colour of the laminae varies from yelow to blackish green, the dark tinges being due to a blackish matter, thought to be of an organic nature, finely disseminated within the dolomite crystals. The darker laminae (corresponding to those consisting of organic matter-rich dolomite rhombohedra) exhibit a finer crystal-size than the light coloured laminae. Therefore, the dolomites display features of an algal stromatolitic deposit, formed during an intertidal stage and dolomitized during the next, supratidal stage. The petrographic arguments for an early dolomitization are: the preservation of the depositional fabrics (LFF, A); the presence of the organic matter and its uneven distribution from a lamina to the other; the reverse dependance of the dolomite crystal-size upon its content in organic matter. The causes responsible for the supratidal penecontemporaneous dolomitization in Ţăndărei sector are, I think, those generally involved in such situations (I l l i n g et al., 1965; K i n s m a n, 1969; etc.): evaporative superconcentration of the interstitial seawater-derived solution, implicitely bringing about a strong increase in the absolute concentration in magnesium ions; the impoverishment of interstitial fluid in calcium ions, therefore the increase in Mg/Ca

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ratio of the pore waters, on account of aragonite and gypsum precipitation ; the propitious mineralogic constitution of the calcareous sediments (aragonitic alone) in Ţăndărei sector, the aragonite exhibiting a remarkable afinity to dolomitization, as the present-day littoral areas with marine carbonate-evaporite sedimentation argue. But, inasmuch as, on the one hand, the modern penecontemporaneous dolomitization always yields extremely fine dolomite and, on the other hand, the discussed core shows both well-preserved laminoid fenestral fabrics and a reverse correlation between the crystal-size and the amount of organic matter accommodated within the crystals, it is to conclude that the dolomitization under discussion proceeded by two steps : the first step, previously discussed and represented by penecontemporaneous supratidal dolomitization, gave rise to an extremely fine dolomite preserving the depositional fabrics and largely retaining the initial, algal, organic matter ; later, during the second step (probably early diagenetic) this dolomicrite underwent an aggrading recrystallization, partly inhibited in places by the presence of the organic matter, evolving to the present 40—90 micra crystal-sized dolomite.

This core has a double importance. On the one hand, it confirms the occurrence of the deposit indicative for the intertidal areas, namely the stromatolitic algal mats, in Ţăndărei sector, supporting the carbonate-evaporite sequence in this sector as a tidal-flat deposit. On the other hand, this core directly demonstrated that during the supratidal stages, besides and in close relation to the other main processes as gypsum precipitation, anhydrite and celestite formation, the process of penecontemporaneous dolomitization of the shallow subtidal-lower intertidal calcareous deposits brought onto a supratidal position was manifest. This process intensely takes place today on the western coast of Quatar Peninsula (I l l i n g et al., 1965), where the lagoonal aragonitic pelletal muds and the intertidal algal stromatolitic deposits, brought onto supratidal position on the coastal flats, react with the interstitial seawater-derived magnesium-rich brines, getting gradually replaced by dolomite landward.

The electric log and sieve samples (chips) certify that the same carbonate-evaporite lithology maintains over the entire section of the facies-zoned complex in Ţăndărei sector.

Synthesizing all the core data, an ideal complete cyclothem in Ţăndărei sector appears to consist of :

α. calcareous, shallow subtidal — lower intertidal unit ;
β. gypsum, upper intertidal unit ;
γ. complex evaporite-dolomite, supratidal unit.

Inasmuch as both the gypsum unit and a part of the complex dolomite-evaporite unit are accommodated inside the calcareous unit, the lithologic column of a complete cyclothem never exhibits a sequence of 3 lithologic levels corresponding to the 3 lithogenetic units, but at the best it consists of two lithologic levels: a lower, calcareous level bearing evaporites and/or dolomite, and an upper, evaporite level (representing just a part of the third, supratidal unit, namely the supratidal evaporite de-

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posit accumulated upon the mass of the calcareous unit and not inside it). I estimate the mean thickness of a cyclothem averaging within several dm—1 m range.

2. *In Cireşu sector*, the carbonate-evaporite sequence (traversed by drilling no. 9 within 2 160-about 2 300 m depth interval) was mecanically cored within 2 208—2 209 m and 2 279—2 280.80 m depth intervals.

a) At the depth of 2 208—2 209 m, a sequence of micrites (containing pelmicrites and dismicrites as local microfacies) and pelletal (microcoprolithic) calcarenites occur, accommodating evaporites (gypsum and anhydrite) as both intercalations (lenses and irregular-shaped strips) and disseminations (scattered crystals and microaggregates).

The micrites (Pl. XIII, Fig. 5) display a homogeneous or clotted (grumous) texture and contain scarce ostracods and dassycladacean algae (genus *Actinoporella*; determination P a t r u l i u s, 1965). Gypsum occurs disseminated within micrites as tiny crystals (crystallotopic structure of M a i k l e m et al., 1969) or irregular-shaped microaggregates consisting of crystals of identical optical orientation (the nodular structure of M a i k l e m et al., 1969). The gypsum microaggregates and crystals are developed only between the clots within those microscopic fields displaying a clotted texture, suggesting to have formed subsequently to the host micrite by interstitial precipitative or displacive growth (accompanied or not by a replacement of calcareous material) within the most porous parts of the lime mud. It is to be assumed that the clotted texture in present micrites is a primary fabric, derived from a depositional floccular texture, so that the superconcentrated solutions had to circulate within original lime mud only through the more porous, interfloccular spaces (between the present clots), laying down gypsum within these spaces only. To be more exact, the clotted texture represents, I feel, a depositional fabric subsequently accented by diagenetic neomorphic processes (which led to a larger aggradation of the interfloccular mud in comparison with that of the micrite particles composing the floccules). The subsequent deposition of gypsum as against the time of calcareous sediment formation is also suggested by the observation that the cavities of the alga *Actinoporella* are filled with micrite, while the gypsum appears developed only within the skeleton wall of the alga (by replacement, microscale solution-precipitation and/or mechanical displacement? of the test).

Within some clotted-textured areas of the micrites, the clots, more sharply outlined, appear as spheroidal pellets of a mud-aggregate type (nomenclature P u r d y, 1963, alias friable aggregate of I l l i n g, 1954). The microvolumes made up of such pellets embedded in a micrite matrix displaying a crystallinity slightly larger than that of pellets (but both of them less than 4 micra in crystal size) stand for the mentioned pelmicrite lenses accommodated within the micritic limestones.

The dismicrites (Pl. XIII, Fig. 6) occur also as microlevels or microlenses irregularly dispersed within micritic limestones. They consist of dense micrite with vaguely clotted texture and extremely fine crystal-

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linity (features of algal calcilutite), containing ocular or vermicular birds' eyes. Therefore the dismicrites display a pronounced character of an algal deposit. Some eyes are filled with equant spar calcite cement, others — with gypsum, and others — with both of these minerals (it is difficult to establish the genetic sequence; apparently, calcite is the first generation and gypsum — the second, inasmuch as the calcite shows corrosions at the contact with the euhedral gypsum crystals).

The pelletal (microcoprolithic) calcarenites (Pl. XIV; XV; XVI, Fig. 1—4) consist of the 3 types of fecal pellets already described in Țăndărei sector (*Favreina salevensis* large, keg-shaped fecal pellets with internal longitudinal tubes; small, elongate-cylindrical pellets ascribed to gastropods; and small, ovoidal pellets probably excreted by worms) embedded in a spar calcite usually 8—20 micra in crystal size. The pellets are composed of micrite particles less than 1 micron in size; the outer parts of the pellets often display a coarser micrite texture 2—3 micra in crystal size, which, in contrast with the extremely fine micrite making up the central parts of the pellets, leaves the appearence of microspar rims. These rims probably proceeded from the aggradation of the original extremely fine micrite composing the pellets. As regards the ,,intercorpusclar'' spar calcite, the petrographic evidence supports a replacement origin by dolomitization developed mainly along the grain boundaries followed by dedolomitization and local aggrading recrystallization ofdedolo mitic microspar to larger pseudospar (the new diagenetic sequence already mentioned in section I; for details see explanations to Pl. XIV – XVI). Other fabrics (contact solutions, grain deformations) suggest that the original pelletal sands were submitted to rather strong pressures before and during the development of the first step of the above-mentioned diagenetic sequence (the dolomitization) (Pl. XIV—XV). Consequently these pelletal limestones represent grainstones resulted from some pelletal sands assumed to have undergone the following postdepositional processes: early pore-compaction (penecontemporaneous process) accounting for a good merging of the pellets; enlargement of the extremely fine micrite composing the pellets to coarser micrite (penecontemporaneous process developed inside the pellets, from their surface inwards) on account of the interstitial circulation of the seawater-derived solutions; more or less strong grain-compaction as a result of the continuation of the compaction process (early diagenetic process) illustrated by the mentioned compression fabrics, and concurrently to this latter process the network-like dolomitization in response to the apparition of some seawater-derived, magnesium-rich solutions (solutions also responsible, together with the active external pressures, for the contact solution phenomena noticed in these limestones); dedolomitization followed by local aggrading recrystallization (both of them presumed as late diagenetic processes) probably at the moment when the seawater-derived, magnesium-rich solutions were replaced by magnesium-poor, connate waters (for arguments and details see explanations to Pl. XIV—XV). The apparently total lack of the micrite matrix in these grainstones may be ascribed to the high activity of pellet-makers.

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All these calcareous sediments characterize the present-day shallow subtidal and lower intertidal areas of the marginal-marine domains with quiet water, carbonate sedimentation.

The limestone beds are crossed by veinlets (Pl. XIII, Fig. 7) composed of calcite, anhydrite and, subordinately, gypsum. The calcite represents the first precipitated generation, appearing largely crystallized, either in a parietal position or desaggregated and enclosed as angular fragments within sulfate mass. The sulfates stand for the second precipitated generation, displaying a felted texture and occupying the central parts of the veinlets. The veinlets are supposed to represent originally hard fillings (as against the soft ones in a subsequently described core) of some dessication cracks formed within the hardened calcareous sediment brought onto a supratidal position. In these cracks the marine-derived superconcentrated solutions laid down salts in the order of the increase of their solubility: primarily — calcium carbonate, and then — the sulfates. The anhydrite obviously prevails among the sulfates; it occurs as small lath-shaped crystals in random orientation, lacking any relationship with gypsum and offering no evidence for a genesis by gypsum replacement, suggesting a direct precipitation from superconcentrated solutions. The large participation of the anyhdrite comparatively to gypsum and its primar character (direct precipitation) involve extremely high values (at supergeneous scale) of concentration and temperature (at least 35—40°C) for the precipitating solution (as the experimental data and the theoretical estimates suggest; Posnjac, 1938, 1942; MacDonald, 1953; Murray, 1964).

The sulfate intercalations incorporated into the calcareous sequence occur as lenses or irregular-shaped strips some centimeters thick and consist of anhydrite alone, or of anhydrite (at least 80 volume percent) and, subordinately, gypsum. They display either a streaky nodular-mosaic structure, or a massive one, or a contorted bedded one (nomenclature Maiklem et al., 1969); the crystallinity varies from 30 micra to 1—2 mm.

The evaporite intercalations of a streaky nodular-mosaic structure contain calcareous (micritic in consistence) tails and pieces with torn outlines (calcareous streaks) floating between sulfate nodules; the nodules, more or less merged, are composed of anhydrite and subordinately gypsum, and exhibit a blocky-fibrous texture (idem Pl. VIII, Fig. 5). The anhydrite displays all the features already remarked in Țăndărei sector (core from 574—575 m depth) for the type of gypsum replacement: pseudomorphs of anhydrite after gypsum, as rectangular or parallelogramic aggregates (in thin section), each aggregate consisting of parallel lath-shaped small anhydrite prisms; some gypsum crystals appear only partly replaced by the tiny anhydrite prisms. Therefore these evaporite intercalations were initially composed of gypsum only, subsequently undergoing anhydritization. They suggest to have formed within unconsolidated or semiconsolidated calcareous sediment by gypsum precipitation (from interstitial supraconcentrated solutions) as nodules, whose growth led

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to their partial merging and to the dissociation of host sediment to calcareous streaks.

The evaporite intercalations of a massive structure show a lath-shaped texture (nomenclature Maiklem et al., 1969) and consist exclusively of large, lath-shaped anhydrite prisms 1—2 mm long in random orientation (idem Pl. XVI, Fig. 8). These intercalations are free from calcareous streaks. The anhydrite of these intercalations seems to be a primar sulfate, laid down by direct precipitation from marine-derived hyperconcentrated solutions of high temperatures (over 34°C after MacDonald, 1953; over 42°C after Conley, Bundy, 1958). The problem concerning the direct precipitation of the anhydrite in the natural environment is still an unsolved question, despite the affirmative answer suggested by the positive results of the laboratory experiments and by the theoretical computations. So, although the physico-chemical requirements are accomplished in some present-day marginal-marine supratidal areas with carbonate-evaporite sedimentation no conclusive evidence was obtained for the direct precipitation of the anhydrite versus its replacement origin at the expense of gypsum. Thus, Kinsman (1969) shows that the anhydrite formation on the supratidal flats of the Trucial Coast (Persian Gulf) initially takes place by the relatively penecontemporaneous replacement of gypsum crystals in the upper 4—5 cm of the sabkha „giving rise to well-formed pseudomorphs". But Kinsman goes on stating that „ the pseudomorphs begin to lose their shape with time, and primary diagenetic anhydrite (anhydrite with no gypsum precursor) is precipitated upon and between the older pseudomorphs, producing as an end product the larger ovoid and variously shaped anhydrite nodules" (Kinsman, 1969, p. 834). Unlike Kinsman, Butler (1969), studying the same supratidal deposits of the Trucial Coast, assumes that the anhydrite is entirely of a secondary origin, derived from gypsum, and that " there is no conclusive evidence to suggest that any anhydrite has formed by direct precipitation from concentrated sea water" (Butler, 1969 p. 81). In the present case the petrologic data support a direct precipitation origin for anhydrite.

The evaporite intercalations of a contorted bedded structure exhibit an aligned-felted texture (nomenclature Maiklem et al., 1969) and consist of crystals of anhydrite and subordinately gypsum, 30—300 micra long (Pl. XIII, Fig. 8). These evaporite intercalations are devoid of calcareous streaks too, but contain angular calcareous fragments (sensu Maiklem et al., 1969). The anhydrite within these intercalations suggests, by its micrographic features, both genetic possibilities : it seems to have partly formed by gypsum replacement and partly — by direct precipitation concurrently with gypsum from superconcentrated solutions.

The last two types of evaporite intercalations, both of them lacking calcareous streaks, reveal, by contrast to each other, two different genetic lines. The anhydrite intercalations of a massive structure suggest to have originated from gravitational accumulation of free crystals in an open space, while the gypso-anhydrite intercalations of a contorted bedded structure seem to have grown by deposition and replacement (gypsum-to-

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anhydrite conversion) as originally hard bodies in a limited, close space (the formation in close space is probably responsible for the distortion of evaporite body structure).

All of these structural-textural, mineralogic and genetic findings guide to the conclusion that the evaporite intercalations devoid of calcareous streaks originated under supratidal conditions, namely : the massive, exclusively anhydrite intercalations settled, as soft sediments consisting of free crystals, on the floor of some standing bodies of hypersaline water, while the contorted gypso-anhydrite intercalations were laid down from interstitial solutions as originally hard, clean lenses or strips within quasi-consolidated host sediment. Thus, the hyperconcentrated solutions accumulated on supratidal surfaces have exclusively precipitated anhydrite, owing to the extremely increased values in their salinity and temperature, while the interstitial solutions, exhibiting relatively lower concentrations and temperatures, have yielded both gypsum and anhydrite (by both direct precipitation and gypsum-by-anhydrite replacement).

The calcareous streaks-bearing gypso-anhydrite intercalations, whose anhydrite appears to be exclusively of a gypsum replacement type, have formed in an original mineralogic configuration somehow different of that one exhibited now (initially consisting only of gypsum) and probably during a stage preceeding the stage of clean evaporite intercalations formation (namely concurrently with the deposition of interstitially disseminated gypsum described within micrites and dismicrites). These intercalations, initially consisting of gypsum only, have subsequently undergone an anhydritization, probably during the stage of clean evaporite intercalations formation, that is during a supratidal stage. I interpret the stage of formation for the interstitial-disseminated gypsum and the calcareous streaks-bearing evaporite intercalations, both these deposit types accumulated (as already stated) within still unconsolidated calcareous sediment, as representing an upper intertidal stage.

Synthesizing the presented material, the following genetical image concerning the sequence from the 2 208—2 209 m depth may be infered : the sedimentation develops a cyclic pattern of a tidal-flat type ; a complete cyclothem appears to consist of 3 lithogenetic units :

α. The calcareous, shallow subtidal-lower intertidal unit deposited in a marine environment of a carbonate-lagoonal facies with relatively normal-marine salinity, includes ostracods-bearing micrites, algal dismicrites, pelmicrites containing mud-aggregate-like pellets, and fecal pellet sands.

β. The gypsum unit was precipitated under upper intertidal conditions from marine-derived superconcentrated interstitial solutions that imbued the unconsolidated calcareous sediment (= unit α ; some processes took place (within the calcareous sediment) : precipitation of calcium carbonate and gypsum as disseminations within the calcareous sediment, and formation of gypsum hard bodies of a streaky nodular-mosaic structure.

γ. The gypso-anhydrite unit points to its having originated under supratidal conditions, reflecting a drastic increase in concentration values

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for the marine-derived solutions, for both those interstitial and those accumulated at the surface of the supratidal area; the following processes took place: depositions of anhydrite soft bodies at the surface of the supratidal areas (above the bulk of the calcareous unit), depositions of hard, gypso-anhydrite lenses and strips inside the quasiconsolidated calcareous unit, replacements of gypsum by anhydrite in both supratidal-deposited gypso-anhydrite bodies and upper-intertidal-formed gypsum bodies, formation of calcite-gypsum-anhydrite veinlets as fillings of some dessication cracks within calcareous sediment.

b) At the depth of 2 279—2 280.80 m, micritic limestones occur, containing anhydrite intercalations (each of them some cm thick).

The micritic limestones contain variable amounts of ostracods, varying from fossilifereous micrites to biomicrites; angular silty quartz grains (probably wind-blown material) occur in a low proportion giving concentrations at some microlevels (Pl. XVI, Fig. 5). The limestones include micronodules of sulfate and are crossed by sulfate veinlets.

The sulfate micronodules (Pl. XVI, Fig. 6) are rectangular or parallelogramic in shape, 200 micra to 2 mm long, and consist of anhydrite prisms arranged in felted texture (nomenclature M a i k l e m et al., 1969); in places the nodules contain small gypsum patches. These micronodules strongly suggest initial large gypsum crystals partly or totally replaced by aggregates of tiny anhydrite prisms, a repetition of the anhydrite-after-gypsum pseudomorphs already described within previously examined cores (but, unlike these previously described ones displaying an internal parallel-fibrous texture, in the present case the pseudomorphs exhibit an internal felted texture).

The veinlets (Pl. XVI, Fig. 7), 50—500 micra wide, consist of lath-shaped crystals 150—300 micra long of anhydrite and, subordinately, gypsum; the veinlets display a relatively aligned felted texture (nomenclature M a i k l e m et al., 1969) oriented either parallel to veinlet walls or perpendicular to them; in each of these two fabric types (types liable to replace one another even within the same veinlet) the sulfate crystals are neither perfectly parallel to one another, nor do they indicate growths from the wall towards the interior of veinlet; they suggest a gravitational accumulation, initially unconsolidated, of free, anhydrite and gypsum crystals inside some dessication cracks. The sulfate crystals are thought to have precipitated from the superconcentrated solutions which were infilling the dessication cracks. The precipitated crystals settled down in the dessication cracks in a position depending on the ratio between the crystal size and the crack width: the large crystals, with the long side exceeding the crack width, settled in a relatively parallel position to the crack walls (vertical in the case of flattened gypsum crystals, and indifferent within the plan paralleling the crack walls in the case of anhydrite prisms), while the crystals equal to or smaller than the crack width settled in a quasihorizontal position (with the long side in an indifferent orientation relatively to the walls position). These two accumulative types yielded the two textural types of sulfate veinlet: the first accumulative

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type gave rise to the felted texture quasiparallel to veinlet walls, while the second accumulative type originated in the felted texture quasiperpendicular to veinlet walls. After deposition, the evaporite crystals often kept on with growing, as demonstrated by some anhydrite prisms in the perpendicular textural type which protrude from the veinlet space penetrating, often rather deep, the walls of host micrite. These observations also suggest that during veinlet formation, the host calcareous sediment was merely semiindurated permitting the in-situ-growing sulfate crystals to penetrate it.

The evaporite intercalations accommodated within the micritic sequence display a bedded massive structure (nomenclature M a i k l e m et al., 1969) and consist of 1—2 mm long anhydrite prisms only, disposed in a lath-shaped texture (nomenclature M a i k l e m et al., 1969), with the lath-shaped crystals either in random orientation or in a slightly aligned quasihorizontal position (Pl. XVI, Fig. 8). These intercalations are similar to those exclusively anhydritic described in the previous core, and suggest open-space accumulations of originally soft sediment on supratidal surfaces by direct precipitation of free anhydrite crystals from standing bodies of marine-derived hypersaline solutions.

Therefore, the 2 279—2 280.80 m depth interval provides a sedimentary sequence displaying a cyclic pattern of a tidal-flat type; a complete cyclothem includes:

α. The calcareous unit consisting of ostracod-bearing lagoonal micrites, and corresponding to the shallow subtidal (-lower intertidal?) stage;

β. The gypsum unit, represented by large gypsum crystals (subsequently evolved to anhydrite) accommodated within the micritic limestones of unit α, formed probably under upper intertidal conditions (upper intertidal stage);

γ. The gypso-anhydrite unit composed of the anhydrite intercalations (laid down upon the calcareous unit, and not within it), gypso-anhydrite veinlets and pseudomorphs of anhydrite after intertidal gypsum, and suggesting a supratidal depositional environment (supratidal stage).

Concluding on Cireşu sector worth underlining is the strong similitude in facies with Ţăndărei sector. The sedimentation, cyclically developed under a tidal-flat regime, consists of a sequence of relatively identical cyclothems, each of them containing, when complete, three lithogenetic units: α) the calcareous, shallow subtidal-lower intertidal unit; β) the gypsum, upper intertidal unit; γ) the gypso-anhydrite, supratidal unit.

The three lithogenetic units of a cyclothem appear strongly interpenetrated spatially, the calcareous unit accommodating both the gypsum unit and a large part of the gypso-anhydrite unit; this fact makes a complete cyclothem to exhibit two lithologic levels only: a lower, calcareous level containing gypsum disseminations (as single crystals — some of them anhydritized —, or microaggregates) and gypso-anhydrite intercalations (as lenses and irregular-shaped strips); an upper, exclusively anhydritic level representing a part of the supratidal unit, namely the

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sulfates (primary anhydrite) settled down upon, not within, the calcareous unit.

3. *In Amara sector* (drilling no. 11), according to the data mentioned in different unpublished reports, the sequence is almost entirely made up of anhydrite and gypsum, suggesting that the evaporite stages were plainly dominant; the carbonate material, utterly subordinate, consists of micritic limestones and dolomites. Unfortunately, the petrologic material from this sector was not available to the author for study.

The foregoing petrologic data suggests the following conclusions concerning the development of the sedimentation in the carbonate-evaporite facies zone:

The sedimentary succession in the carbonate-evaporite facies zone was born by the superposition of a great number of cyclothems; a complete cyclothem is lithogenetically made up of a calcareous unit and two evaporite units, one gypsic and another complex evaporitic (anhydrito-gypso-dolomitic). As already mentioned, the conditions and processes pointed out by this cyclothem type suggest a sedimentation of a tidal-flat type. This conclusion is also supported by the regional sedimentary pattern: the areal configuration of this facies zone and the vicinity of an emerged rise. Therefore, the carbonate-evaporite facies zone represents an ancient tidal-flat domain.

A complete depositional time-cycle consisted of a shallow subtidal-lower intertidal period corresponding to the stage of the calcareous unit sedimentation, and of an upper intertidal-supratidal period corresponding to the two stages responsible for the deposition of the evaporite units (gypsum unit and anhydrite-gypsum-dolomite unit).

During the shallow subtidal-lower intertidal period, calcareous deposits have formed under sheltered, inner platform conditions, belonging to lime mud facies, pellet mud facies and skeletal mud facies (sensu Imbrie, Purdy, 1962; Purdy, 1963), with local participation of intertidal algal mats.

During the upper intertidal-supratidal period the entire area gradually became a belt of coastal salt flats, equivalent to present-day coastal sabkhas developed around the Persian Gulf. Exactly as on the modern sabkhas, on these fossil sabkhas sedimentation proceeded under the direct influence of marine-derived brines. So, on the one hand, the sea-derived waters imbued the marginal-marine calcareous deposit lifted in upper intertidal or supratidal position, their renewal having probably been accomplished either by flood recharge (sensu Butler, 1969) or by landward interstitial capilar migration; on the other hand, the seawater accumulated on the supratidal surfaces on account of capilar ascension of marine-derived interstitial waters or by direct flooding of these surfaces by sea waters during storms or high spring tides. The interstitial waters are assumed to have got superconcentrated by the process of capilar concentration (sensu Müller, 1960; Friedman, Sanders, 1967). The marine waters accumulated on the supratidal surfaces were also submitted

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to a process of high concentration by strong evaporation. Thus, all the marine waters migrated into upper intertidal and supratidal areas evolved to brines, whose concentration was increasing landward, so that in the upper intertidal flat space they were precipitating gypsum, while in the supratidal flat space they were laying down gypsum and anhydrite; in places, simultaneously to these last products the calcareous sediments have undergone dolomitization (not to confuse this penecontemporaneous dolomitization with the early diagenetic dolomitization ascertained in many limestones in present paper and displaying no relation to sabkha cycles); all these processes were sometimes accompanied by celestite formation as a supratidal by-product of the calcium carbonate replacement by sulfate (gypsum) and/or dolomite. In the upper intertidal zones the evaporite deposition took place, from the interstitial brines, within the previously formed and still soft calcareous (sometimes more or less clayey) sediments. In the supratidal zones, evaporite depositions occurred both within the previously formed and already hardened calcareous (more or less clayey) sediments, and over them (in this latter case the evaporite deposition being sometimes accompanied by clayey mud accumulation); within the calcareous sediments, evaporite and dolomite formation took place from the interstitial brines; over the calcareous sediments, evaporite settled down as soft sediments from the standing bodies of supra-saline water accumulated on the supratidal flats (brine pools and brine pans). The evaporite veinlets are assumed to have formed also during the supratidal stages by infilling with salts the dessication cracks opened within the quasiconsolidated calcareous sediments.

As regards the evaporite minerals genesis with relation to the sabkha cycles in the studied area, it is to conclude:

— gypsum appears always as the first mineral laid down from the marine-derived solutions (usually by direct precipitation, but sometimes by replacement of calcareous material); it is an upper intertidal and supratidal product;

— anhydrite is mainly a secondary mineral, formed by replacement of gypsum, only a few situations suggesting a primary origin by direct precipitation from brines; the anhydrite seems to be only a supratidal product;

— celestite appears as a secondary, replacement mineral formed under supratidal conditions at the expense of gypsum;

— dolomite is also a secondary mineral, supratidally formed by calcium carbonate (probably aragonite) replacement.

The lack of halite within the studied area suggests that the salinity of precipitative brines has not exceeded values of 250—300 ‰, inasmuch as the halite precipitation requires concentrations (salinities) at least as low as 300 ‰ (S c h m a l t z , 1966).

On theoretical grounds, it is to be expected that the sedimentary sequence across the carbonate-evaporite facies zone should display a substantial horizontal variation in lithology, from an almost completely evaporitic composition landward to a mainly calcareous one seaward,

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in accordance with the presumption that the upper intertidal-supratidal sedimentation prevails landward and the shallow subtidal-lower intertidal one — seaward. This theoretical assumption is consistent with the field evidence; so, watching the variation in sedimentation across the carbonate-evaporite facies zone (using the data provided by the 3 drillings), one may ascertain that generally from SW towards NE (that is from „Ciochina" land towards the sea) the participation of evaporites decreases and that of limestones increases.

On the whole, the carbonate-evaporite deposits of this facies zone appear outlined, in the vertical section of the east-Wallachian sector, as an well-individualized sedimentary system: the tidal-flat system.

B) The Zone of Bahamite Facies

This facies zone parallels the previous one, lying directly NW of it. The bahamite facies is represented by skeletal-lumpal-pelletal, muddy calcareous facies in Smirna and Gura Ialomiței sectors.

1. *In Smirna sector* (drilling no. 7), the calcareous sequence contains poorly skeletal-pelletal, micritic limestones (1 239—1 240.30 m depth) and lumpal-oölitic limestones (1 338.50—1 339 m depth).

The poorly skeletal-pelletal, micritic limestones (fossiliferous, pellet-bearing micrites; in Folk's 1959 nomenclature) consist of scarce skeletal grains and fecal pellets floating in homogeneous or grumous micrite (Pl. VI, Fig. 4). The skeletal grains are represented by ostracods and spiral or biseried foraminiferids; the fecal pellets belong to both the *Favreina salevensis* type (crustacean fecal pellets) and the elongate-cylindrical, small type (probably molluscan fecal pellets). The limestones contain angular, silty, quartz grains in minor amounts (about 10 volume percent). Therefore, this limestone type resembles some of the limestones of the carbonate-evaporite facies zone, expressing the same quiet, shallow marine environment illustrative of the protected, inner platform domain. The foraminiferids exhibit strong tendencies of micritization (grain-diminution of Orme, Brown, 1963; Wolf, 1965 a) which may be ascribed to the algal boring-inorganic infilling mechanism described from modern sediments (Bathurst, 1966; Kendall, Skipwith, 1969; Alexandersson, 1972), the former ones appearing more or less turned into cryptocrystalline grains (nomenclature Purdy, 1963). The terrigeneous quartzose material probably resulted from an eolian transport.

The lumpal-oölitic limestones (Pl. VII) consist of oölitic grains (= oöids=oöliths) and pseudoölitic grains (= pseudoöids = pseudoöliths) embraced by a finely or very finely crystalline spar calcite often displaying features of a basal binder. The microscopic examination suggests that this spar calcite is a diagenetic pseudospar resulted from replacement of calcareous grains and possibly some micrite matrix; it seems to represent the product of the new diagenetic sequence (already mentioned in section I) involving dolomiztization-dedolomitization-aggrading recrystallization (for details see explanations Plate VII).

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The pseudoölitic grains (= the nonoölitized grains accompanying the oölitic grains, the same time representing the source of nuclei for these latter) are represented by pellets and grapestone lumps.

The pellets, 100 — 500 micra in size, belong to the fecal pellet type as well as to the cryptocrystalline grain type. There are two kinds of fecal pellets: fecal pellets of the *Favreina salevensis* type (crustacean fecal pellets) and small ovoidal fecal pellets (probably produced by worms). The spheroidal to irregular-ovoidal pellets seem to be of a cryptocrystalline grain type, probably proceeded from the hardening of some mud aggregates (nomenclature Purdy, 1963).

The grapestone lumps (nomenclature Illing, 1954; Purdy, 1963) are composed of pellets, smaller grapestone lumps (consisting of pellets too) and/or oöliths. The lumps reach up to 2 mm in size and are made up either of one (monogeneous) or of more (poligeneous) of the enumerated component types. The coexistence of the pellets or some small lumps and the oöliths within the large grapestone lumps proves that the processes of pelletal and lumpal accretion took place concurrently with the oölitic accretion processes.

The oölitic grains, representing more than 90 volume percent of the allochem bulk, mainly belong to the superficial oölith type, each of them usually possessing a large nucleus (0.5—1.5 mm in diameter) and a single thin circumlamina of oölitic accretion; in minor amounts true oöliths occur, whose small nuclei (100—150 micra in diameter) bear thick oölitic envelopes consisting of numerous laminae (some of them partly or completely micritized probably by the algal boring-inorganic infilling mechanism described by Bathurst (1966) and Alexandersson (1972) in modern shallow-marine sediments). The nuclei are represented by pellets and lumps. The pellets have yielded, by oölitization, both superficial oöliths and true oöliths. The lumps, owing to their large size, underwent a superficial oölitic accretion only; they appear as an ancient equivalent of present-day botryoidal lumps (Illing, 1954; Purdy, 1963; Kendall, Skipwith, 1969). The oöliths with pelletal nuclei were frequently aggregated, along with or without other grain types, in grapestone lumps (consisting of 2—4 aggregated grains), which, in their turn, were submitted to a new oölitic accretion, finally resulting in superficial oöliths bearing complex nuclei (these oöliths are also equivalent to modern botryoidal lumps). This fact supports the opinion that the processes of pelletal-lumpal accretion and oölitic accretion developed concurrently. The oöliths are 1—1.5 mm in diameter, therefore rather uniform in size comparatively to the large range of nucleus size; the present data confirm the fact, already described by different authors, that the oölitic accretion tends to homogenize the grain size by a higher rate of oölitic accretion on the smaller pseudoölitic grains.

In the last analysis one may assert that the processes of oölitic accretion concurred with the processes of pelletal-lumpal accretion, the grains bearing, at the moment of burial, the print of this concurrence and

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falling into the oölith class or pseudoölith class in accordance with the last process undergone : cölitic accretion or lumpal type accretion. The simultaneous development of the lumpal-pelletal and oölitic accretions suggests that the sedimentation took place in no case within an oölitic facies area, but within a grapestone facies area (sensu P u r d y , 1963) submitted to the influences of a neighbouring oölitic facies.

Therefore, Smirna sector belonged to an inner platform domain, where the deposition developed under skeletal-pellet mud facies and grapestone facies, that is, under a calcareous bahamite facies (sensu B e a l e s , 1958 : the ancient analogous of all present-day facies distinguished within the sheltered, inner areas of the carbonate platforms). Along the same line, worth mentioning is that bahamite facies and bahamite rock should be regarded as two distinct notions which must not be confused : the term of bahamite facies encompasses the lime mud facies, skeletal/pellet mud facies and grapestone facies, that is the facies prevailing in the sheltered, inner domains of the carbonate platforms and bordering seaward upon the skeletal-oölitic facies specific to the agitated, open environment of the outer domains of the carbonate platforms, whereas the bahamite rocks stand for the petrologic group represented by limestones varying from calcilutites to calcirudites and consisting of calcareous mud and/or allochems of a micritic consistence suggesting to have originated from physico-chemical precipitation-accretion-aggregation processes only.

The transient manifestation of some oölitic facies influences (outer platform facies) in Smirna sector suggests the existence of a lateral shifting in facies perpendicular to the direction of the facies zones development, probably in correlation with the lateral movements within tidal-flat domain lying directly S and SW.

2. *I n G u r a I a l o m i ţ e i s e c t o r* (drilling no. 8 : 220—226 m depth core) medium and coarse lumpal-pelletal calcarenites were identified (Pl. III, Fig. 6—8). The allochems are chiefly represented by grapestone lumps (consisting of pellets) and pellets (of a cryptocrystalline grain type, possibly mainly proceeding from mud-aggregates); coated grains subordinately occur, bearing micrite envelopes and varying from monoenveloped to polienveloped (microoncolitic) grains ; in minor amounts finely arenitic skeletal grains occur (ostracods, echinoderm ossicles, miliolids). The allochems are embedded within a microspar matrix, thought as derived from a former micrite matrix. These limestones express a bahamite facies (sensu B e a l e s , 1958) of a grapestone facies type (sensu P u r d y , 1963).

The presented bahamite facies fully corresponds to a shallow marine calcareous sedimentation developed within an inner platform domain under its specific, carbonate-lagoonal facies analogous to the present-day lime mud facies, pellet mud facies, skeletal mud facies and grapestone facies (initially outlined on Great Bahama Bank, Andros lobe ; I m b r i e, P u r d y , 1962; P u r d y , 1963); sporadic influences of the N and NE outer platform facies were felt in places.

The bulk of bahamite deposits of this facies zone at present makes up the carbonate-lagoonal sedimentary system.

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C) The Zone of Skeletal-Oölitic Facies

This facies zone stretches parallel to the previous one, lying directly NE and N of it. Inasmuch as the facies-zoned complex is missing NE of the line joining the localities of Hîrșova and Berlescu, being removed by erosion, a little part only of the skeletal-oölitic facies zone preserved until today. Thus, calcareous skeletal and/or oölitic facies (both of them sensu Purdy, 1963) were encountered in Zăvoaia and Berlescu sectors only.

1. *In Zăvoaia sector* (drilling no. 6) oölitic calcarenites (2 033—2 033.40 m depth core) and skeletal limestones (1 956—1 956.50 m depth core) occur.

The oölitic calcarenites (oosparites) consist of true oöliths (several oölitic laminae) whose nuclei are represented by pellets of a cryptocrystalline grain type; minor percents of pseudoöliths (pellets of a cryptocrystalline grain type) occur. The rock is a medium-grained calcarenite (note: the thin sections were not available for a detailed study and for taking photographs). These limestones express a typical oölitic facies of open, outer platform environment.

The skeletal limestones are extremely complex, consisting of biolithite masses and allochthonous skeletal masses (Pl. IV, Fig. 2—3). The biolithite masses include an assemblage of coraligeneous structures, algal structures and encrusting foraminiferic structures of a „nubecularian" type (sensu Elliot, 1966); the algal structures are prevalent and belong to the pseudostromata type (sensu Wolf, 1965 b, 1965 c), consisting of extremely dense and various-textured (from homogeneous to pelletoid) algal calcilutite (schyzophycean calcilutite); the foraminiferic structures occur as nodules within the algal structures. The allochthonous skeletal masses (sensu Carozzi, 1960) include biomicrites and biopelmicrites; they consist of a micrite matrix containing numerous skeletal grains (fine shells of pelecypods, ostracods and rare calpionelids), algal pellets about 30 micra in size made up of extremely dense calcilutite (algal mud-aggregates), and microoncolitic algal nodules centrated or not by a skeletal grain. These skeletal limestones express a skeletal facies of a coralgal type (sensu Purdy, 1963) typical of an unprotected, outer platform environment. There is no doubt that they stand for an ancient reef complex.

2. *In Berlescu sector* (drilling no. 3), pelletal, skeletal calcarenites (2 128.90—2 129.20 m depth core) (biosparites to biopelsparites — Pl. II, Fig. 6—7) occur. The skeletal grains are the main contributor to the limestone bulk and consist of echinoderm plates. The pellets appear ovoidal or spheroidal in shape, 60—150 micra in size and belong to the cryptocrystalline grain type. Irregular-shaped lumps, made up of pellets, subordinately occur. These limestones express a skeletal facies (sensu Purdy, 1963). Both the pellets and the lumps seem to have formed by algally (schyzophycean) induced precipitation-aglutination-aggregation processes, inasmuch as on the one hand it is very difficult to accept the occurence of some pure physico-chemical

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processes of this kind, without any organic intervention, under a skeletal depositional facies, and on the other hand the constitutive micrite approaches the algal calcilutite (as it was defined by W o o d, 1941; C a r o z z i, 1960; W o l f, 1965 b, 1965 c). The spar calcite „cement" exhibits clear features of a pseudospar resulted from the diagenetic sequence including dolomitization-dedolomitization-aggrading recrystallization (arguments: scattered or grouped dedolomitized rhombohedra, angular outlines of micrite grains, partial or total consumption of grains, rhombohedra obviously penetrating the grains of a micritic consistence). The calcarenites of Berlescu sector may be ascribed to a skeletal facies of an echinalgal type (skeletal facies including limestones consisting of echinoderm pieces and corpuscular algal products of blue-green algae, the latter also termed algal grains of the Oncolithi Group; sensu W o l f, 1965 c).

Therefore, the skeletal-oölitic calcareous facies corresponds to a reefal sedimentation within an outer platform domain under coralgal, echinalgal and oölitic facies. As a whole, the deposits belonging to this facies zone make up, at present, the reef sedimentary system.

The facies-zoned complex, on the whole, represents the sedimentary product of a highly facies-differentiated, carbonate platform. The depositional area within discussed perimeter was sketched as a large bay widely opening northward and closed on the other three sides by the mainland of Dobrudgea and „Ciochina"rise[15]. Three sedimentary domains, semicircular in shape, stretched within this area from S to N, that is, from a near-shore position towards an off-shore position: the marginal-marine domain, the inner shelf domain and the outer shelf domain. Three depositional facies were corresponding to these three domains: the carbonate-evaporite facies, the calcareous bahamite facies and the calcareous skeletal-oölitic facies. The simultaneous development of these three facies led to the accumulation of a sedimentary body — the facies-zoned complex — consisting of three lateral-interfingering petrogenetic systems: the tidal-flat system, the carbonate-lagoonal system and the reef system. These three systems succeed now in the above-mentioned order from S and SW towards N and NE.

III. THE REEFAL LIMESTONE COMPLEX

This complex was recognized only in the central part of the studied area, in Berlescu (about 50 m thick), Zăvoaia (50—60 m) and Smirna (20—30 m) sectors, overlying apparently in depositional conformity the

[15] The carbonate-evaporite facies zone eventually has its continuation towards SE, in south Dobrudgea, where prof. B ă n c i l ă recently described („On the presence of a Purbekian — Wealdian gypsiferous formation in the Fetești-Constanța region" — in Romanian; in press) a thick evaporite sequence stratigraphically interposing between the upper Jurassic limestones and the Lower Cretaceous ones, sequence dated by mentioned author as belonging to the uppermost part of the Upper Jurassic.

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facies-zoned complex. The absence of this complex in all the other sectors seems to be connected with the very active postneocomian-prealbian erosion. Only in Ciochina sector it seems to be a depositional gap and not an erosional one (because here the Hauterivian complex does exist, directly overlying the Upper Jurassic limestones and dolostones).

1. *In Berlescu sector* (drilling no. 3: 2 078—2 079 m depth core) biodolomicrites occur, consisting of a dolomicrite matrix containing many, arenite-sized, skeletal grains: pieces of echinoderms, tests of dassycladacean algae (genus *Actinoporella*), briozoan, pelecypod and brachiopod debris, and scarce foraminiferids. The dolomicrite matrix consists of a micrite matrix of extremely fine crystallinity (algal calcilutite?) invaded by dolomite microrhombohedra (at most 8 micra in size) (possibly of a penecontemporaneous replacement of the calcareous mud under submarine conditions). I consider these limestones as representing a skeletal facies of a brialgal type (skeletal facies including limestones consisting of algal material and zoogeneous material characterized by the presence of frequent briozoan remnants and the lack of coral debris).

2. *In Zăvoaia sector* (drilling no. 6: 1 906—1 906.50 m depth core), gresous, pelletal, lumpal calcarenites and microcalcirudites occur, consisting of lumps, pellets and angular quartz grains (the last ones in amounts of 10—20 volume percent and 60—90 micra in size) (Pl. IV, Fig. 4—6; Pl. V). The lumps, made up of pellets and/or smaller lumps, are of a grapestone lump type (nomenclature Illing, 1954) and of a micritic lump type (new term: lump displaying either a homogeneous-micritic consistence, in present case on account of a perfect merging of the component allochems also micritic in consistence, or a slightly nonhomogeneous-micritic consistence, in present case owing to the partial merging only of the component grains). The pellets, up to 125 micra in size, ovoidal or spheroidal in shape, and of a homogeneous-micritic consistence, display features of cryptocrystalline grains (nomenclature Purdy, 1963). Within „intercorpuscular" spaces, diagenetic equant spar calcite occurs, suggesting to represent, by its fabrics, an allochem-replacive calcite, resulted from the already-mentioned diagenetic sequence involving dolomitization-dedolomitization-aggrading recrystallization (for details see explanations to Pl. V). Therefore, this spar calcite stands for former patches of grouped dolomite rhombohedra; scattered dedolomite rhombohedra occur disseminated within the micrite of the calcareous grains too. The micrite composing the allochems exhibits features of an algal calcilutite (extremely fine crystallinity less than 1 micron, extremely dense and compact texture) corresponding to the descriptions given by Wood (1941) and Wolf (1965 b, 1965 c). The calcareous grains display strong tendencies of merging to each other, often welding to yield pseudodismicritic microvolumes (see explanations to Pl. V). These observations suggest an algal origin for the calcareous grains; the author considers them as algal pellets and algal lumps (calcareous grains of a micritic consistence formed by algally (blue-green algae) produced or induced processes of precipitation-flocculation-agglutination-

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aggregation). Therefore, in present case the pelletal-lumpal limestones do not express a bahamite facies (sensu B e a l e s, 1958) of a grapestone type (sensu P u r d y, 1963): pure physico-chemical accretion-aggregation; they suggest a skeletal facies (sensu P u r d y, 1963) of an algal type, represented by allochthonous limestones (sensu C a r o z z i, 1960) made up of algal calcilutite organized in corpuscular productions (= algal grains).

3. *In Smirna sector* (drilling no. 7 : 1 183.50—1 184.50 m depth core), gresous biomicrites occur consisting of abundant skeletal grains and angular silty quartz grains (in amounts of 10—20 volume percent) embedded within a micrite matrix (Pl. VI, Fig. 5—7). The micrite matrix often displayes enlargement to microspar. The skeletal material is represented by: foraminiferids, fragments of briozoans, corals, dassycladaceans, pelecypods and echinoderms; all the skeletal grains exhibit micrite envelopes representing supposingly the product of the algal boring-inorganic infilling mechanism reported from recent sediments by B a t h u r s t (1966) and A l e x a n d e r s s o n (1972); the foraminiferids display different stages of grain-diminution to an internal micritic texture (this process is thought to have proceeded by the same boring-infilling mechanism above-refered to). The limestones of Smirna sector indicate a skeletal facies (sensu P u r d y, 1963) of a coralgal type.

As a general conclusion, the calcareous reefal complex, as much as it is preserved, suggests a strong homogenization of the depositional conditions in comparison with the previous complex, the sedimentation taking place under a general skeletal facies, the whole area displaying features of an outer platform domain; within this general reefal facies, some differentiations may be noted depending on the biocoenosis type which yielded the calcareous deposits: coralgal facies, brialgal facies and algal facies. Taking into account that in Cernavoda sector (western-southern Dobrudgea) the same skeletal facies develops overlying the carbonate-evaporite facies of the facies-zoned complex (as the data presented by B ă n c i l ă suggest [16]), one may conclude that the sedimentation of the calcareous reefal complex proceeded under a skeletal facies of outer shelf domain over the entire eastern part of the Romanian Plain (that is, also in those sectors where this complex now appears eroded). Probably the rise of Ciochina-Mărculeşti was lowered to a submerged position, having lost its role of threshold during the deposition of this third complex, even though some of its sectors, such as Ciochina sector, possibly behaved as undepositional areas.

GENERAL CONCLUSIONS

The petrologic study of the Neocomian sequences crossed by drillings in the eastern Romanian Plain emphasizes the following general conclusions:

[16] *Op. cit.* 15

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During the Upper Tithonian?-Berriasian-Valanginian time interval, an autochthonous, carbonate and carbonate-evaporite sedimentation developed, originating in 3 sedimentary complexes: the lower, micritic limestone complex; the middle, facies-zoned complex; and the upper, reefal limestone complex.

The complex of micritic limestones generally expresses a marine sedimentation in a lime mud facies within a sheltered, inner platform environment.

The facies-zoned complex represents the product of a sedimentation on a carbonate platform with strong facies differentiations; this complex includes 3 isochronous petrogenetic systems which laterally interfinger succeeding from S towards N as folows: the tidal-flat system, the carbonate-lagoonal system and the reef system; these 3 systems represent the products of the 3 sedimentary facies — the carbonate-evaporite facies, the bahamite facies and the skeletal-oölitic facies — typical of the 3 depositional domains of any marine carbonate platform under warm and dry climate: the marginal-marine domain (or the tidal-flat domain), the inner shelf domain and the outer shelf domain.

The complex of reefal limestones on the whole represents, as its name reveals, a reefal system formed on an extensive outer shelf domain under coralgal, brialgal and algal facies; this complex suggests the reccurrence of the uniformity of the depositional conditions over the entire studied area, as is the case with the first complex.

The general sequence of the main petrogenetic processes noticed in the whole studied material appears as follows:

The syngenetic processes consisted in: direct deposition of calcareous (mainly aragonitic) sediments by physico-chemical or organogenous extraction of $CaCO_3$ from seawater; direct precipitation of gypsum and possibly anhydrite; anhydrite-after-gypsum replacements, sometimes of pseudomorphic types (two textural types of pseudomorphic replacement are described, in both of them each gypsum crystal being replaced by an aggregate of tiny anhydrite prisms); celestite formation probably also by gypsum replacement; penecontemporaneous, extremely finely crystalline dolomitization. All these processes, except the calcareous sediment formation, took place with relation to sabkha cycles.

The early diagenetic stages (diagenesis after some burial, but still under seawater influence) include the main bulk of dolomitization, developed as finely or medium crystalline rhombohedra.

The late diagenetic stages (diagenesis after rather deep burial and developed under the influence of connate waters) were responsible for the dedolomitization-aggrading recrystallization processual sequence ascertained in many cores. A worth mentioning fact is the total lack of the pore-filling calcite (=orthospar=calcite cement) in the examined petrologic material; so, all the spar calcite, apparently calcite cement in many instances, actually represents microspar or pseudospar usually of a dedolomitic type (proceeding from the diagenetic sequence of dolomitization-dedolomitization-aggrading recrystallization) and seldom of a neomorphic

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type (direct aggrading neomorphism of some depositional micritic fabrics to larger calcite).

As regards the age of each sedimentary complex, accepting that these complexes conformably overlie each other, that the calpionelid assemblage of micritic limestone complex in Zăvoaia sector suggests an Upper Tithonian age and that the complex of reefal limestones is the equivalent of the Valanginian reefal limestones of southern Dobrudgea (Cernavodă sector), present author put forward the following presumptive chronostratigraphic assignation :

I. The micritic limestone complex is Upper Tithonian?-Lower Berriasian in age;

II. The facies-zoned complex belongs to the Upper Berriasian-Lower Valanginian time-interval;

III. The reefal limestone complex stands for the Upper Valanginian term.

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QUESTIONS

M. Săndulescu: How does the author explain the lack of some facies zones especially that of carbonate-evaporitic facies in the area stretching towards the north-Dobrudgean mainland?

Answer: The mainland probably included both the northern and central Dobrudgea (according to the annexed facies map). The present-day lack of facies zones towards the Dobrudgean mainland is accounted for both diverse erosional phases installed subsequently to the Neocomian time, and probable original disposition under an intricated pattern of the facies zones at the contact with the Dobrudgean shore.

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DISCUSSIONS

M. Săndulescu. When emphasizing the scientifical and practical quality of the paper drawn up at an up-to-date level, and presented by our colleague A. Drăgănescu, I should like to remark that the explanation of the asymmetrical character of facies zones, and hence their lack towards the north-Dobrudgean mainland may be found if taken into account the occurrence of some tectonic accidents of a transcurrent fault type (strike slip faults) which would have cut the primary depositional zones, shifting towards north-west the north-Dobrudgean compartment (faults trending NW—SE). However, such strike slips have been already noticed in the Pecenega-Camena fault zone by our late colleague O. Mirăuță. This structural control of the present pattern of the facies zones is underlined by the perfect parallelism between the Ciochina threshold and the Fierbinți fault.

APPENDIX

CLAY-MINERAL STUDY OF SOME NEOCOMIAN CORES IN ȚĂNDĂREI DRILLING

BY

SILVIU RĂDAN[1]

Method

Two samples from the cores at 574—575 m depth (sample *A*) and 620—621 m depth (sample *B*) were prepared for X-ray diffraction analysis aiming at the knowledge of their clay-mineral composition. With a view to satisfy this purpose, each sample was treated with HCl 0,1 N in order to remove the carbonate fraction, then washed and shaked in distiled water. The suspension containing the fraction less than 2 microns was collected and two slides of oriented aggregates were prepared. Three X-ray diffractograms were obtained from each slide: untreated, exposed to a saturated atmosphere of ethylene-glycol and heated at 550°C. Calculation of the relative amounts of clay minerals in each of the samples, on the basis of these diffractograms was, achieved using Biscaye's method (1965).

Results

TABLE

Clay-mineral composition of the samples

Sample	Depth	Clay-mineral composition		
		Illite (%)	Chlorite (%)	Mixed-layer structures
A	574—575m	90	10	—
B	620—621m	85	15	10—14 M, 14 C—14 M

[1] Institutul de geologie și geofizică, str. Caransebeș 1, 7000 București.

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The illite displayes a good crystallinity in both samples, and I (002)/I (001) ratio shows rather high values (0.46 in sample *A* and 0.58 in sample *B*), typical of Al-rich illites.

The chlorite of sample *A* is of a poor swelling, Fe—Mg-rich type, while the chlorite of sample *B* is of a Mg-rich type and belongs to the C_M type, according to Lucas classification (1962).

The random mixed-layer structures belong to 10—14 M and 14C—14M types (according to Lucas classification, 1962) and were identified only in sample *B*. They are believed to have resulted from degradation of illite (10—14 M) and chlorite (14C—14 M).

Discussion

The mineralogic composition of clay fraction in both samples is characteristic, as a rule, of the deposits laid down under dry climate, where the physical wethering prevails. The lack of mixed-layer structures in sample *A* suggests aggrading action, rather drastic insolation and cation abundance, as the main features of the original depositional environment. The sample *B*, in oposition to the former, expresses an environment submitted to a larger circulation of the water or to a stronger influence of the shore proximity, where degraded minerals were not able to get back to their original structure. The clays of sample *A* suggest a subtidal depositional environment, while those of sample *B* seem to stand for an intertidal or supratidal product.

Some processes specific to the deep-burial stage of diagenesis are thought to have been superposed to the transformations developed during the above-commented, marine, pre-burial stage of diagenesis. So, this opinion is supported by the fact that the illite, though Al-rich in both samples, is however sligthly Fe-enriched in sample *A*, and the chlorite is Mg-rich in sample *B* and slightly Fe -enriched in sample *A*. The gradual enhance in aluminium content of the illite and magnesium content of the chlorite with increasing burial depth of sediment in present case are in a very good aggrement with Dunoyer's data (1969).

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PLATE I (a, b)

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PLATE I (a, b)

Berlescu sector. Complex I.

Thin sections in the core from 2 182.15—2 182.90 m depth in drilling no. 3

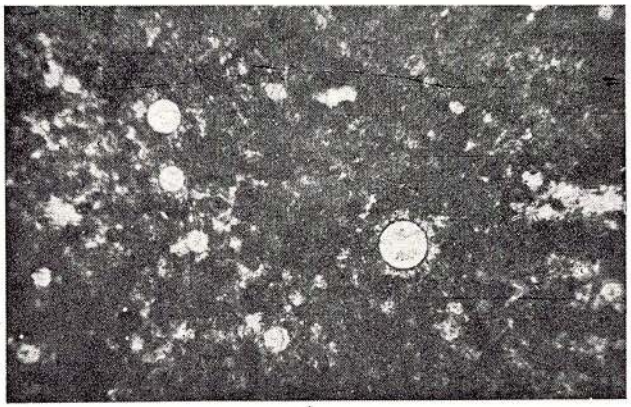
Fig. 1—2. — Fossiliferous clotted micrites bearing frequent fibrospheres (cadosinids and stomiospherids) and calcitized radiolarians. The clotted texture seems to be due to an irregular, slight dolomitization followed by dedolomitization resulting in a microspar network enclosing micrite clots.

Fig. 3—4. — Fossiliferous (fibrospheres) micrite or pelmicrite with pseudointraclastic microfacies resulted from an irregular dolomitization-dedolomitization leaving intraclast-like patches of relict micrite embedded within a dedolomitic microspar network.

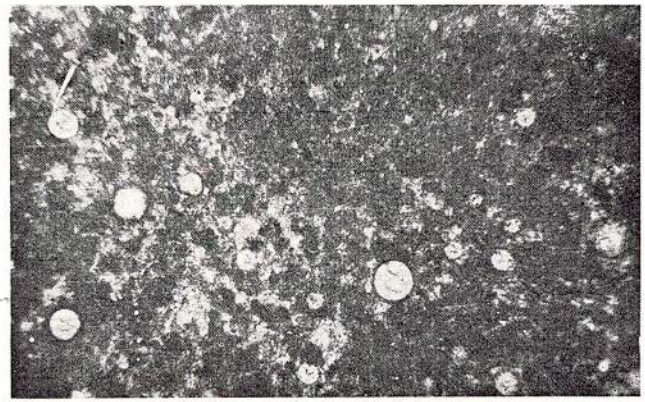
Fig. 5. — Intramicrite as microfacies within micritic limestones, consisting of a fibrosphere-bearing pelmicritic groundmass containing micritic intraclasts.

Fig. 6—8. — Oölitic, intraclastic microfacies within micritic limestones, represented by patch-like, ,,sparry" or micritic, pelletal, oölitic, intraclastic calcarenites (i = large intraclasts consisting of fibrosphere-bearing pelmicrite; $o_{1,2}$ = oöliths). The intercorpuscular spar calcite seems to be a pseudospar resulting from a network-like dolomitization (on eventually interestitial micrite, tiny grains and margins of larger grains), followed by dedolomitization (replacement of dolomite by dedolomitic microspar) and more or less aggrading recrystallization (aggradation of dedolomitic microspar to larger pseudospar) (arguments: grains with poligonal outlines, abnormally loose packing, spar ,,cement" with angular outlines consisting of straight and rather long segments linked at different angles while its component calcite crystals are anhedral in shape and very small in size).
All photographs under plane polarized light.

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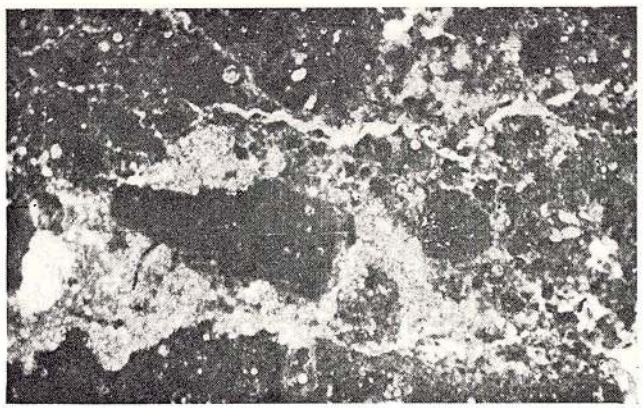


1



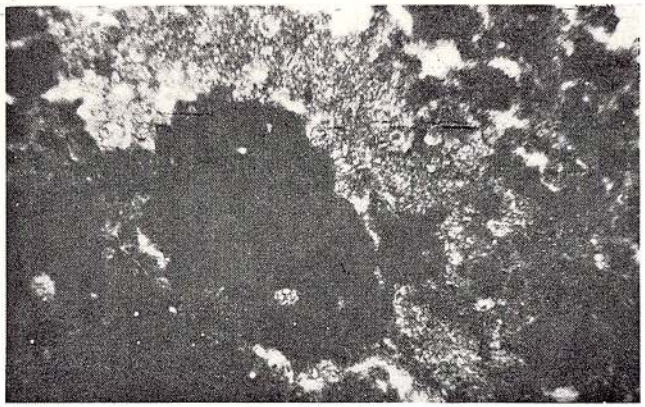
2

Fig. 1, 2, 4 0 125 250 μ



3

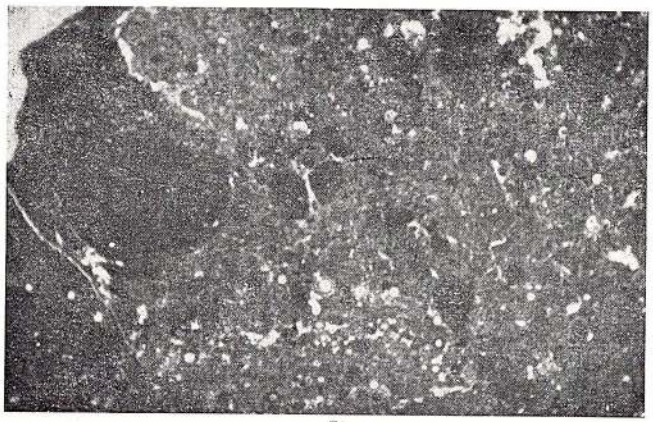
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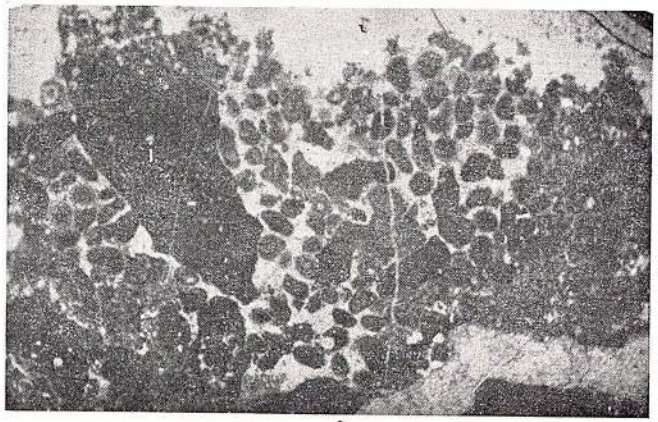
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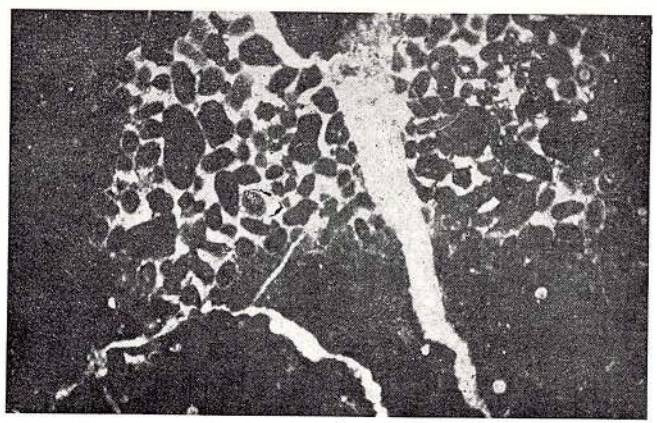
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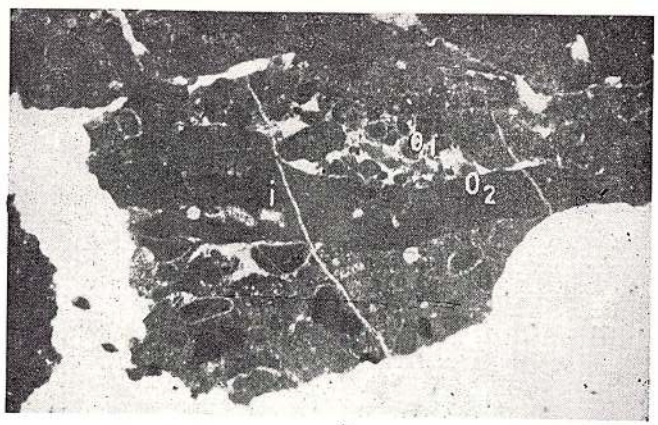
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6



7



8

0 0,25 0,5 1 2mm

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PLATE II (a, b)

Berlescu sector

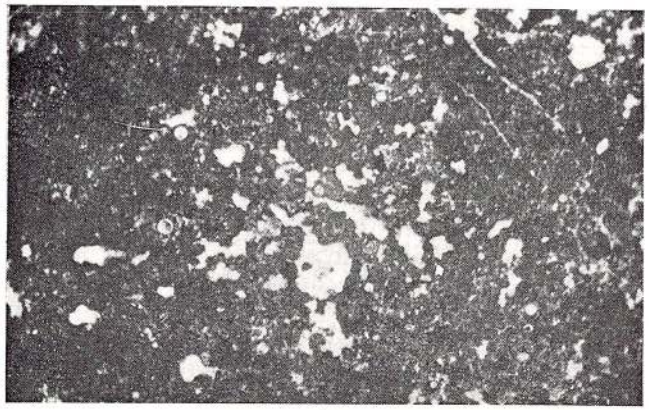
Thin sections in the cores from drilling no. 3

Fig. 1—3. — Complex I (2 182.15 — 2 182.90 m depth). The pseudo- "depositional perturbation facies" consisting of pelmicrites, ,,disturbed" pelmicrites, ,,dismicrites" and ,,intrasparites". The photo 3 shows a vertical passage from fossiliferous pelmicrites (top) to ,,dismicrites" (bottom). All spar calcite eyes and patches, responsible for such an interpretation at first sight owing to their pore-filling appearence, actually represent a diagenetic product of the dolomitization-dedolomitization-aggrading recrystallization sequence. Arguments: all sparry areas, although ,,filled" with tiny anhedral calcite crystals, are large-angular in shape strongly suggesting former dolomite patches; many spar calcite areas display angular-serrate outlines whose component segments, grouped by twos, threes or fours, close more or less completely in small rhomb-shaped spaces the marginal zones of these sparry areas; within spar calcite areas, micrite relicts occur exhibiting diffuse, angular outlines. Consequently, this ,,depositional facies" actually stands for a diagenetically acquired fabric.

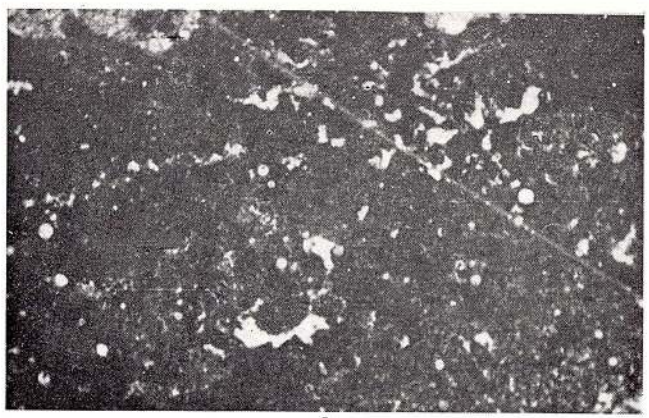
Fig. 4—5. — Complex I (1 182.15—2 182.90 m depth). Fine alternation of fossiliferous pelmicrites and pelsparites, containing in places (Fig. 5, lower half) micritic intraclasts (some of them displaying sparry coatings: oölitic envelopes or diagenetic fine spar calcite rim?). The pelsparites are correctly termed with relation to their probably perfect washing, although their calcite ,,cement" exhibit strong features of the type described from Fig. 1—3, present plate (probably an early pore-compaction of the original grainstone followed by the diagenetic sequence of dolomitization-dedolomitization-aggrading recrystallization developed mainly along the grain boundaries).

Fig. 6—7. — Complex II (2 128.90—2 129.20 m depth). Pelletal, crinoidal calcarenites (biosparites and biopelsparites), whose spar calcite "cement" displayes clear features of a replacement spar calcite by the diagenetic sequence of dolomitization-dedolomitization-aggrading recrystallization (arguments: many scattered or grouped dedolomized rhombohedra, angular outlines of the grains).
All photographs under plane polarized light.

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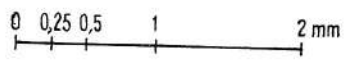
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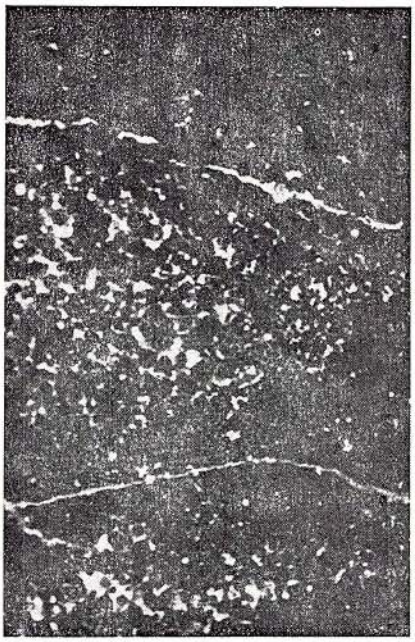


3



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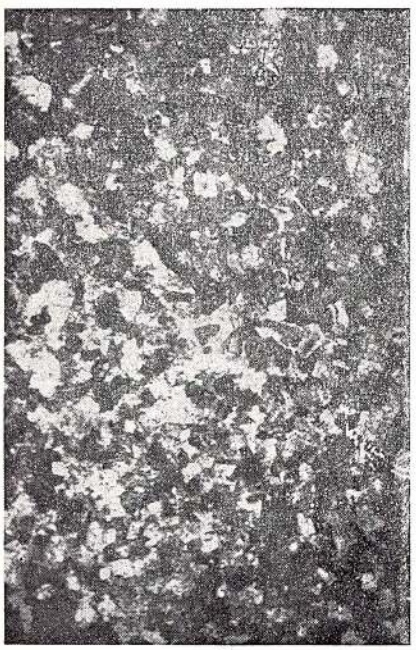
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0 0,25 0,5 1 2mm

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PLATE III (a, b)

Berlescu and Gura Ialomiței sectors
Thin sections in the cores from drillings no. 4 and 8

Berlescu sector (drilling no. 4)

Fig. 1—2. — Complex I (2 372—2 373.20 m depth). Dolomite-bearing micrite. The dolomite rhombohedra exhibit rhomb-shaped nuclei of relict micrite.

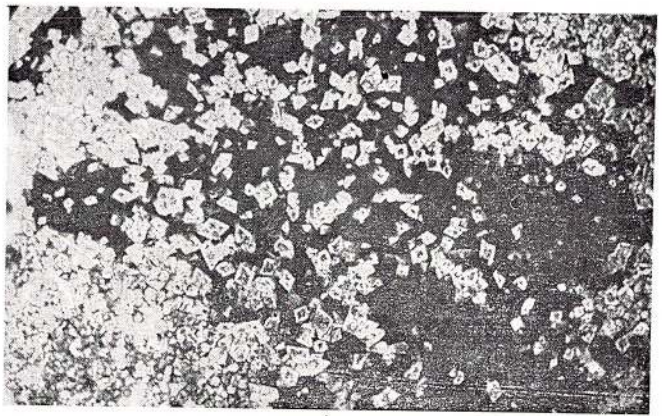
Gura Ialomiței sector (drilling no. 8)

Fig. 3—5. — Complex I (343—343.50 m depth). Micrite invaded by dolomite rhombohedra each of them displaying rhombic nuclei of relict micrite.

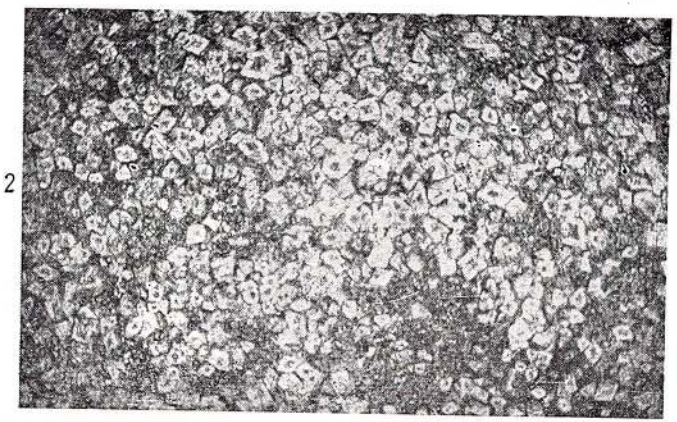
Fig. 6—8. — Complex II (220—226 m depth). Lumpal-pelletal calcarenites consisting of grapestone lumps (composed of pellets), pellets and rare skeletal grains embedded within a microspar matrix, locally aggraded to larger pseudospar (probably dolomitization-dedolomitization-aggrading recrystallization on the micrite matrix and the smallest grains).

All photographs under plane polarized light.

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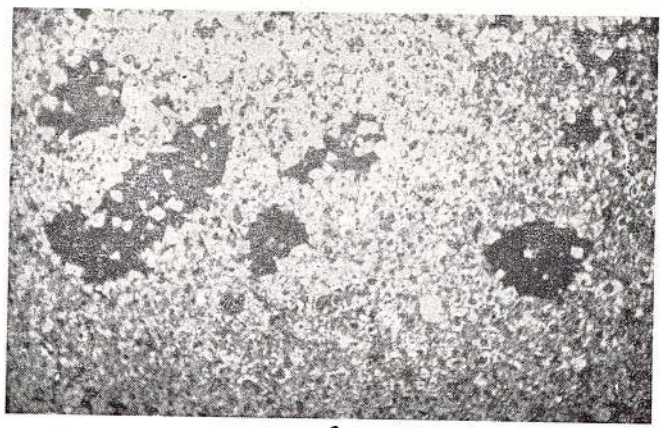


1

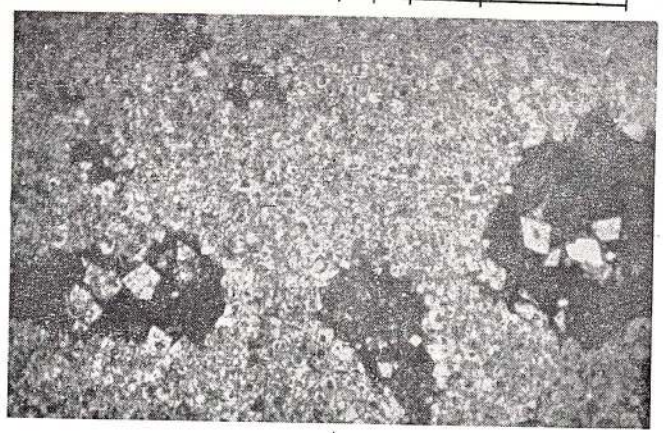


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Fig.1,2,3 : 0 0,25 0,5 1 2mm



3

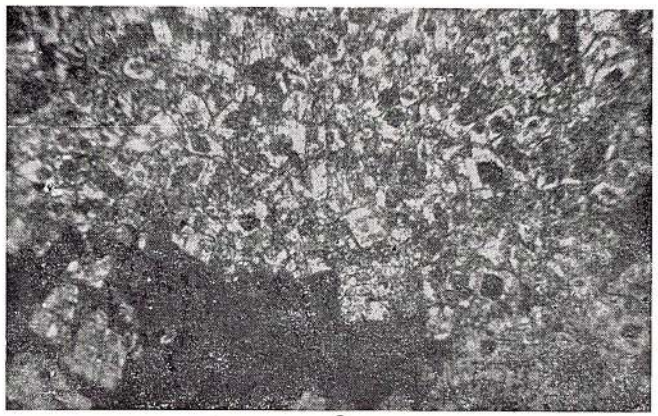


4

0 125 250 μ

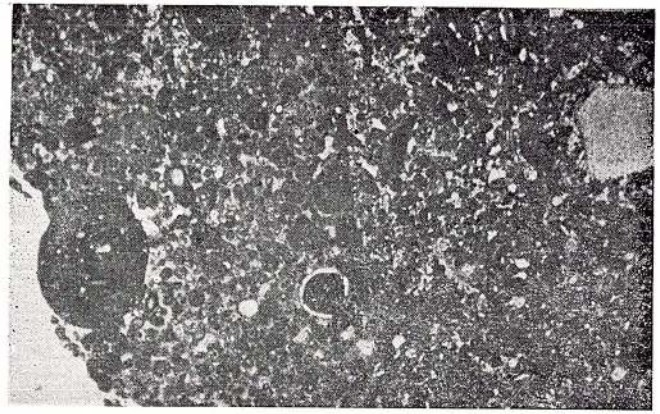
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0 125 250 μ



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7



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Fig. 6,7,8: 0 0,25 0,5 1 2 mm

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PLATE IV (a, b)

Zăvoaia sector

Thin sections in the cores from drilling no. 6

Fig. 1. — Complex I (2 295—2 296 m depth). Fossiliferous micrite containing scattered dolomite rhombohedra. The skeletal material consists of echinoderm pieces, tiny ostracods and (not visible in the photo) calpionelids.

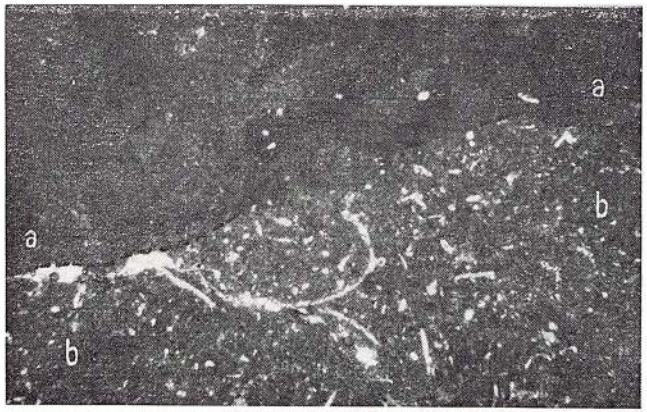
Fig. 2—3. — Complex II (1 956—1 956.50 m depth). Skeletal limestones including biolithite masses (coraligenous structures, algal structures and foraminiferic structures) (Fig. 2:a) and interbiolithitic sediments (biomicrites, and biopelmicrites, Fig. 3).

Fig. 4—8. — Complex III (1 906—1 906.50 m depth). Algal calcarenites and calcirudites varying from pelletal, lumpal limestones (grainstones) (Fig. 4—6), some of them exhibiting the algal grains partly merged (Fig. 7), to pseudodismicrites of a complete or almost complete merging of the grains (Fig. 8) (see also the next plate). All photographs under plane polarized light.

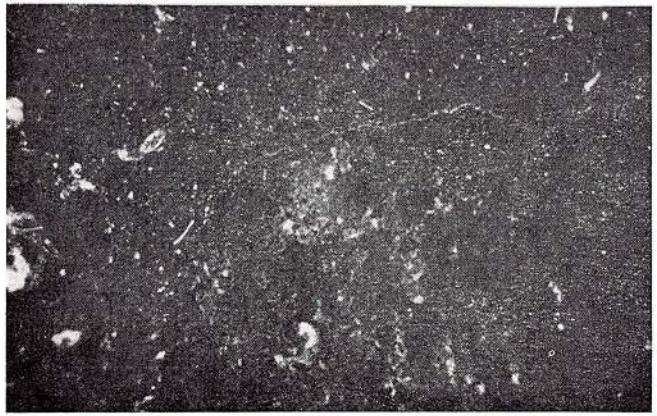
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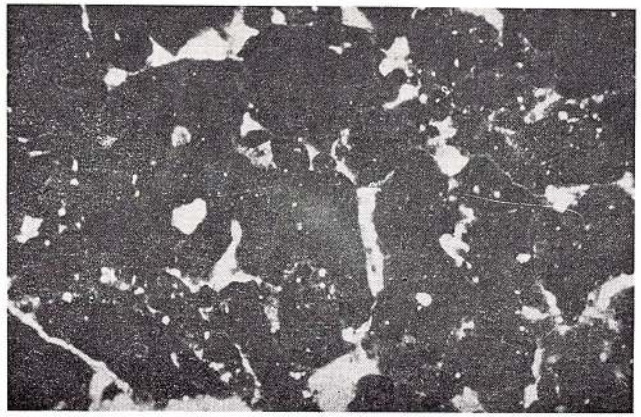
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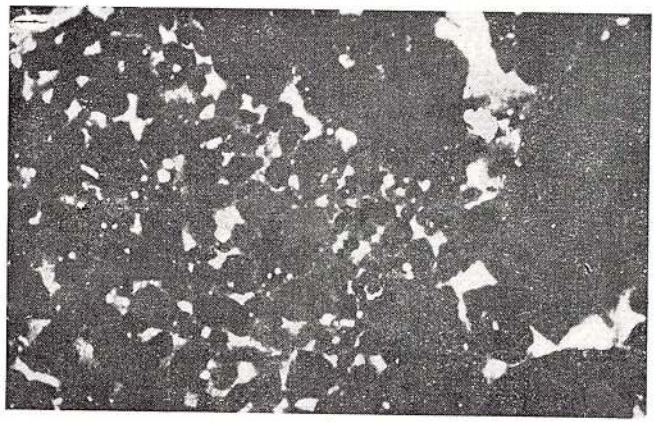


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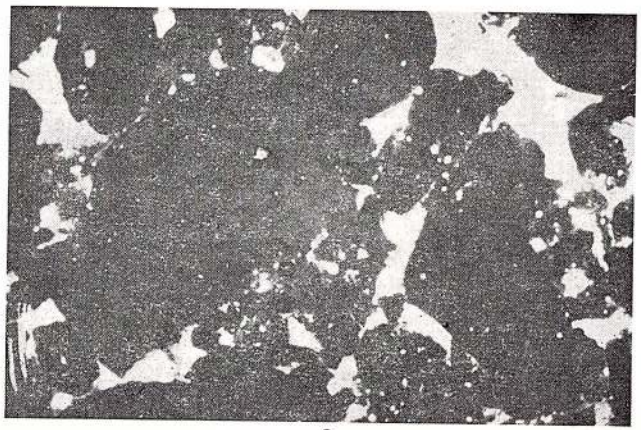
0 0,25 0,5 1 2 mm

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0 0,25 0,5 1 2 mm

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PLATE V (a, b)

Zăvoaia sector

Thin sections in the core from 1 906—1 906.50 m depth (Complex III) in drilling no. 6

Fig. 1—1a. — Pelletal, lumpal algal calcarenites (idem Pl. IV, Fig. 4—6). Fig. 1a represents an enlargement in Fig. 1.

Fig. 2—2a. — Algal pseudodismicrite resulted from an almost complete merging of the algal nodules, pellets and lumps. Fig. 2a stands for an enlarged part of the Fig. 2.

The spar calcite ,,cement'' (this Plate and Fig. 4—8, Pl. IV) displays evident features of a replacive type formed at the expense of depositional petrographic constituents, probably by dolomitization-dedolomitization-aggrading recrystallization developed mainly on the tiny grains and along the margins of larger grains (arguments: the rhombohedral-shaped margins of the spar areas, relict angular patches with micritic internal structure floating within spar-areas, angular outlines of algal grains, scattered dedolomite rhombohedra consisting of dedolomitic microspar).

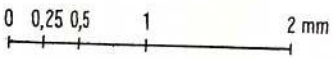
These algal limestones (this Plate and Fig. 4—8, Pl. IV), originally algal sands, suggest to have undergone an early merging of the calcareous grains owing to a penecontemporaneous algal aggregation followed by a very early diagenetic pore-compaction; consequently, in a first stage, grainstones exhibiting an apparently perfect welding of the component grains resulted; subsequently, by a dolomitization-dedolomitization-aggrading recrystallization developed mainly along the grain boundaries, the welded grainstones evolved to the present ,,loosely packed'' grainstones (Pl. IV., Fig. 4—6; Pl. V, Fig. 1—1a) or, if this diagenetic sequence resulted in a patch-like spar calcite, to the present pseudodismicrites (Pl. IV, Fig. 7—8; Pl. V, Fig. 2—2a).

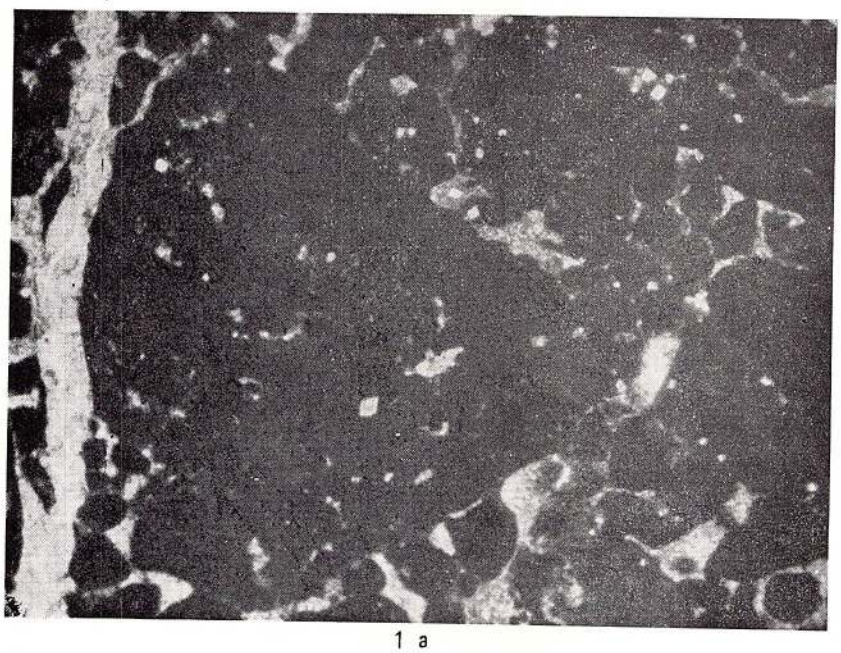
All photographs under plane polarized light.

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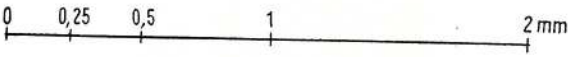


1



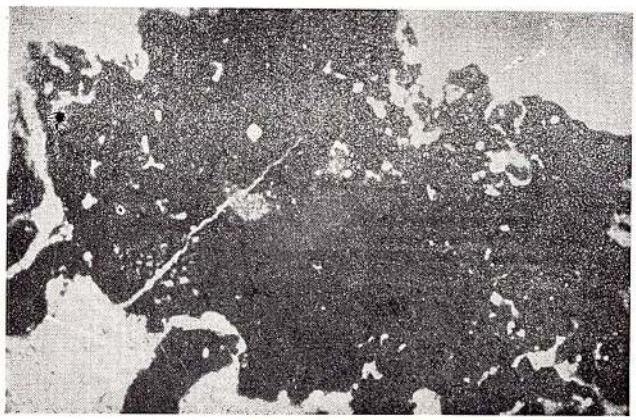


1 a

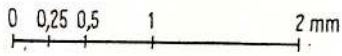


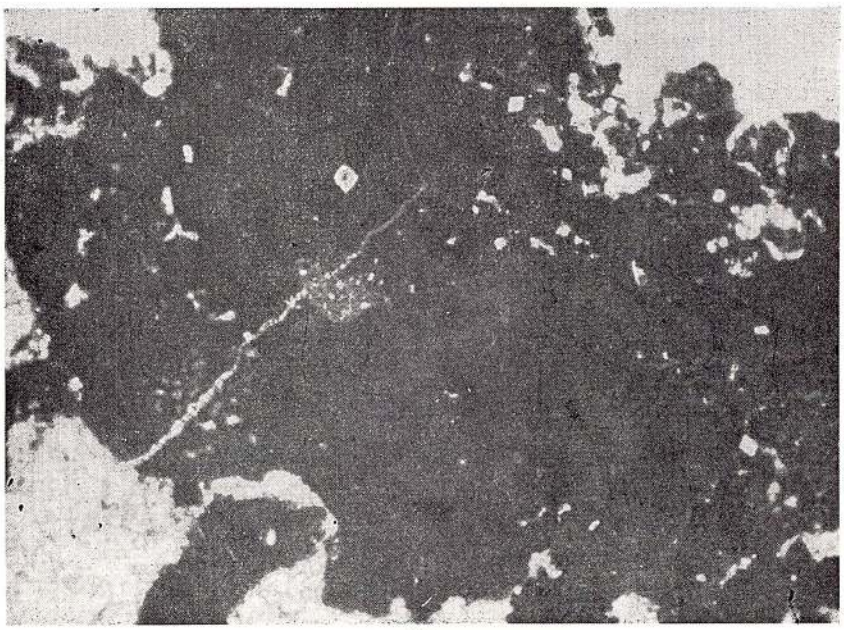
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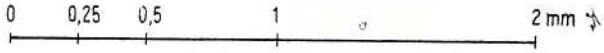


2





2 a



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PLATE VI (a, b)

Smirna sector

Thin sections in the cores from drilling no. 7

Fig. 1—2. — Complex I (1 983.50 — 1 987.50 m depth). Dolomite rhombohedra-bearing micrites.

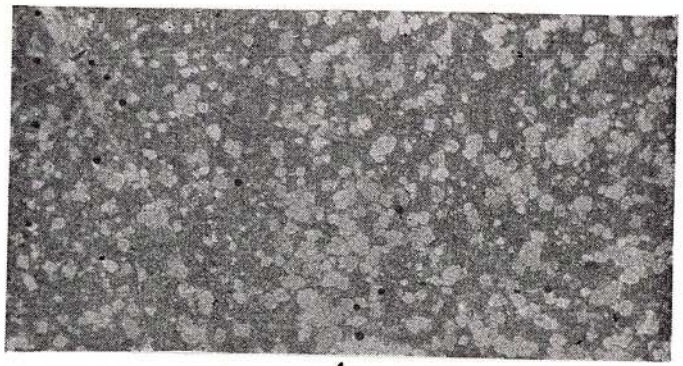
Fig. 3. — Complex I (1 856—1 856.50 m depth). Micrite containing calcispheres, ostracods and other small skeletal debris.

Fig. 4. — Complex II (1 239.50—1 240.50 m depth). Micrite containing foraminiferids (most of them submitted to a strong micritization), fecal pellets (*Favreina salevensis*, noted „f" in the photomicrograph) and quartz grains.

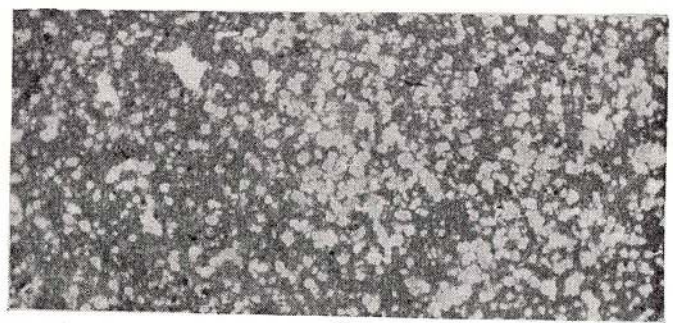
Fig. 5—7. — Complex III (1 183.50—1 184.50 m depth). Greseous biomicrite containing abundant skeletal grains (foraminiferids, fragments of corals, briozoans, echinoderms, etc.) submitted to a strong algal action (micritic envelopes and micritization). In the photomicrograph 7, some large limeclasts (intraclasts or lumps?) occur, including tiny pellets, micritized foraminiferids, quartz grains, etc. Most of the micrite matrix is evolved to fine microspar.

All photographs under plane polarized light.

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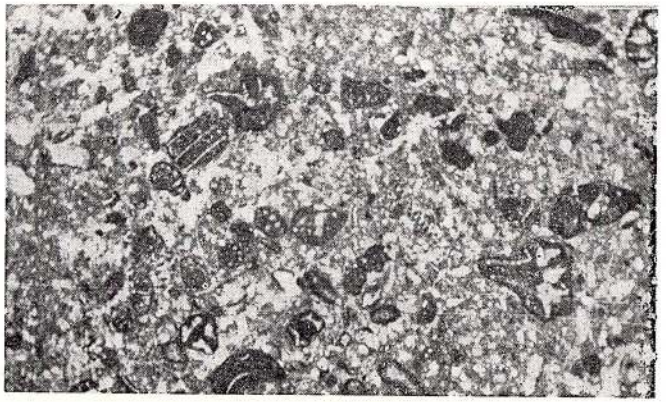
0 0,25 0,5 1 2 mm

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0 0,25 0,5 1 2 mm

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PLATE VII (a, b)

Smirna sector

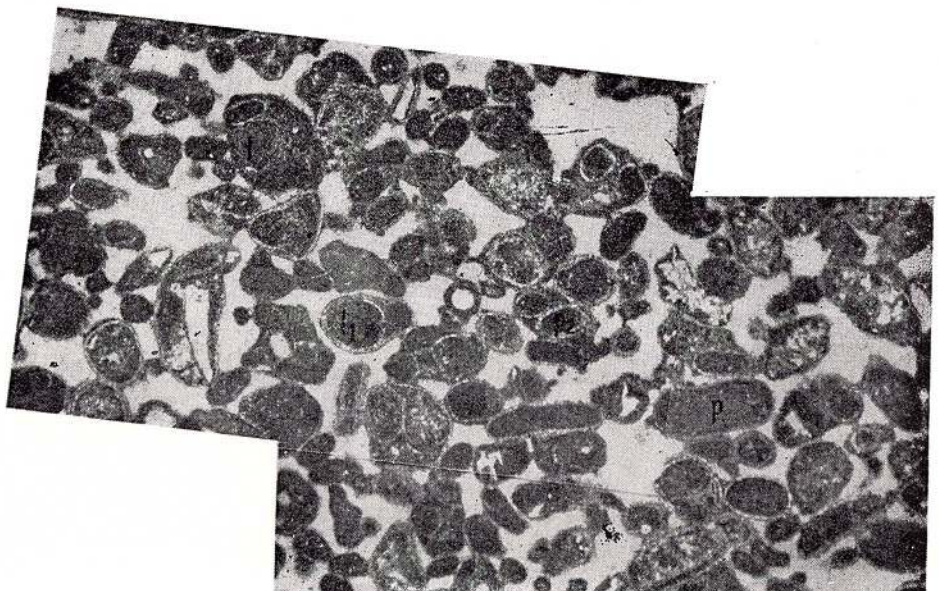
Thin sections in the core from 1 338.50—1 339 m depth (Complex II) in drilling no. 7

Fig. 1—3. — Lumpal-oölitic limestones consisting of: grapestone lumps (l, l_2), many of them displaying superficial oölitic envelopes (bothryoidal lumps: l_1), oöliths (o), pellets of a fecal pellet type (*Favreina salevensis* = p, p_1, and ovoid fecal pellets = p_2) and of a cryptocrystalline grain type. The pellets and the oöliths are most frequently bound together within composite grains = grapestone lumps (l_3).

The calcite „cement" exhibits strong features of a replacive, diagenetic spar calcite (pseudospar): abnormally loose packing, the „cement" patches cut, transect and partly or totally assimilate the allochems. In addition, other features (the „cement" patches show frequent angular outlines but a xenotopic structure, in places penetrating the preserved grains as rhombohedral-shaped protrusions; the small size of the pseudospar crystals aligned along the margins of the pseudospar areas as against the large length of the straight segments composing the angular outlines of the pseudospar areas) suggest that the pseudospar proceeded from a network-like dolomitization (of some allochems and possibly some former micrite matrix) followed by dedolomitization and, in most instances, recrystallization of dedolomitic microspar to larger pseudospar.

All photographs under plane polarized light.

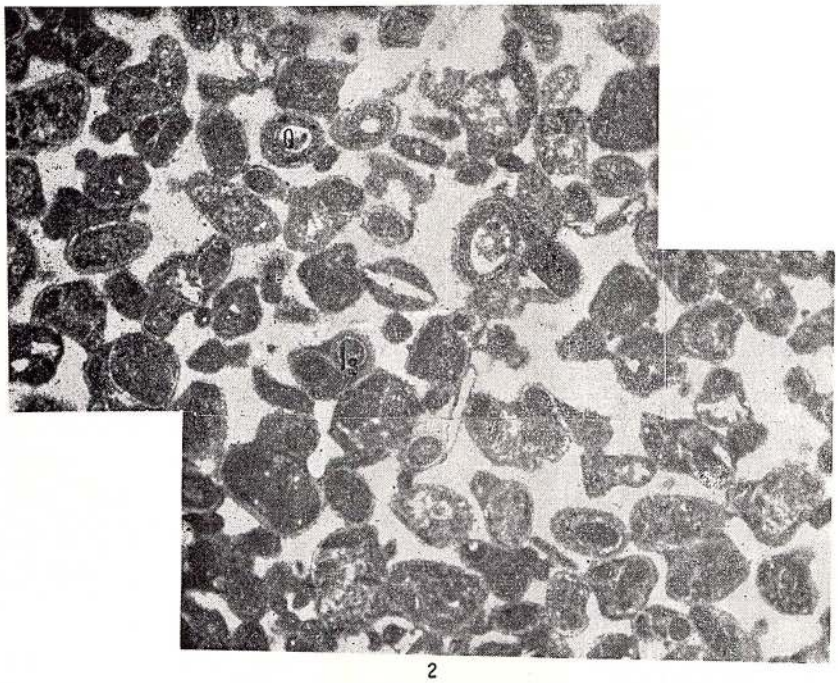
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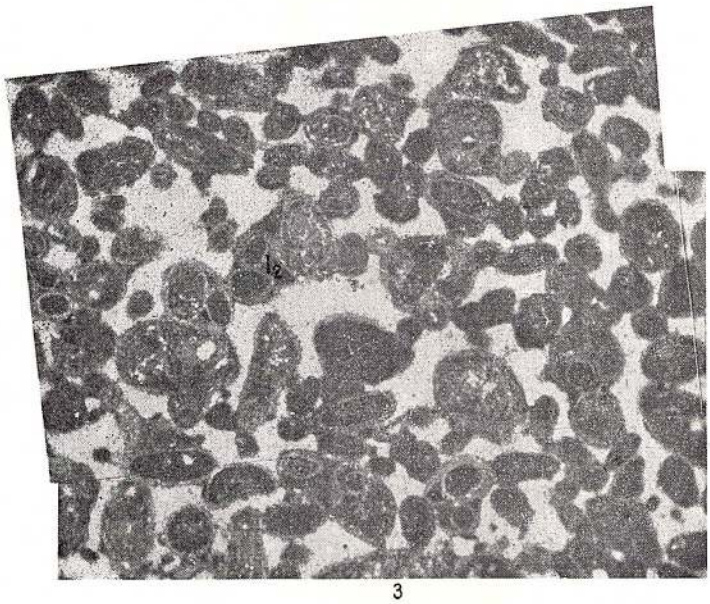
0 0,25 0,5 1 2 mm

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2



3

0 0,25 0,5 1 2 mm

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PLATE VIII (a, b)

Țăndărei sector

Thin sections in the cores from drilling no. 10

Fig. 1--2. — Complex I (728—729 m depth). Dolomite-bearing micrite. Plane polarized light.

Fig. 3—4. — Complex II (574—575 m depth). Gypsum mash-bearing clayey micrite. The gypsum grew as large euhedral crystals (the white areas) dissociating the soft host mud (the grey areas) in pieces and particles entrapped within or between the sulfate crystals. The large size of these gypsum crystals belonging to lithogenetic unit β strongly contrasts with the small size of gypsum crystals of lithogenetic unit γ (Fig. 5—7, this Plate). Origin of the thin section: clayey micrites imbued with gypsum. N +.

Fig. 5. — Complex II (574—575 m depth). Anhydrite with blocky-fibrous texture. Each of the anhydrite aggregates (some of them outlined with dashed line) consists of quasiparallel, neddle-like, anhydrite prisms and represents a single initial gypsum crystal subsequently replaced by anhydrite. Origin: lensoid evaporite intercalation with mosaic structure. N +.

Fig. 6—7. — Complex II (574—575 m depth). Celestite prisms (c) scattered within a finely crystalline gypsum mass blocky to microcrystalline in texture. Origin: lensoid evaporite intercalation with mosaic structure. Plane polarized light.

Fig. 8—9. — Complex II (620—621 m depth). Gypsum displaying a texture varying from blocky to felted or aligned-felted. Origin: thick gypsum bed with mosaic structure. N +.

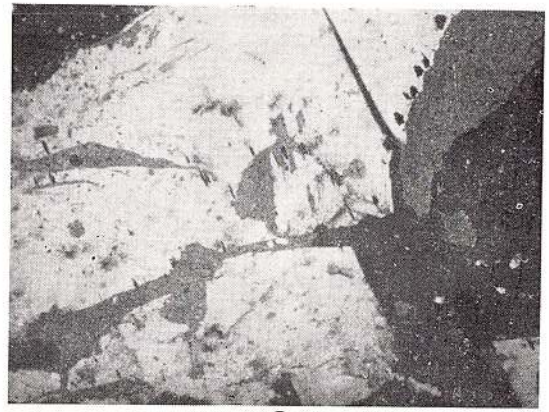
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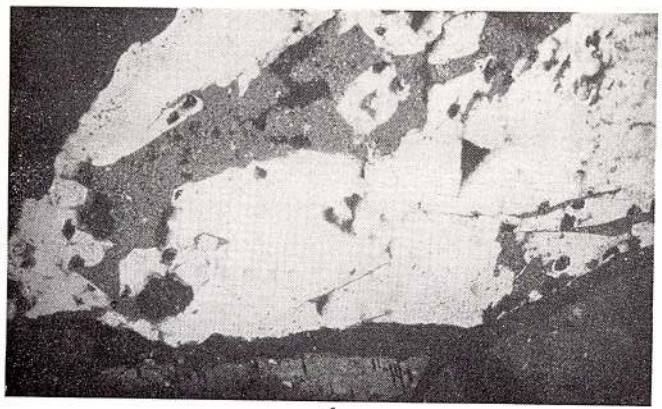
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Fig. 1, 3, 4: 0 0,25 0,5 1 2 mm

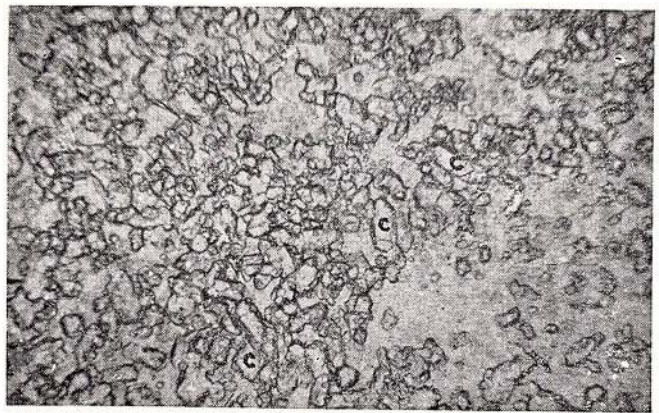


5

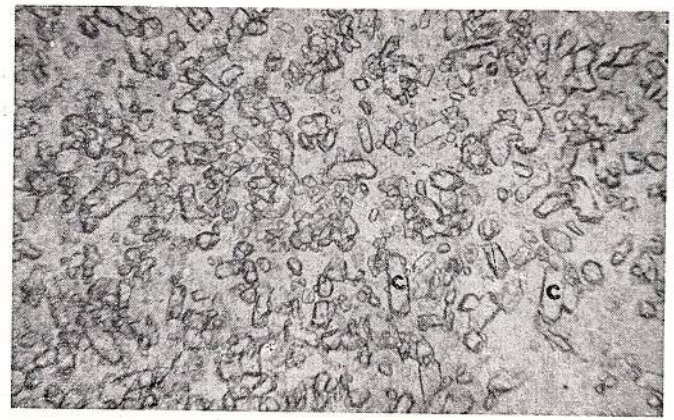
Fig. 2, 5: 0 125 250 μ

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6



7

0 125 250 μ



8



9

0 0,25 0,5 1 2 mm

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PLATE IX (a, b)

Țăndărei sector

Thin section in the core from 620—621 m depth (Complex II) in drilling no. 10

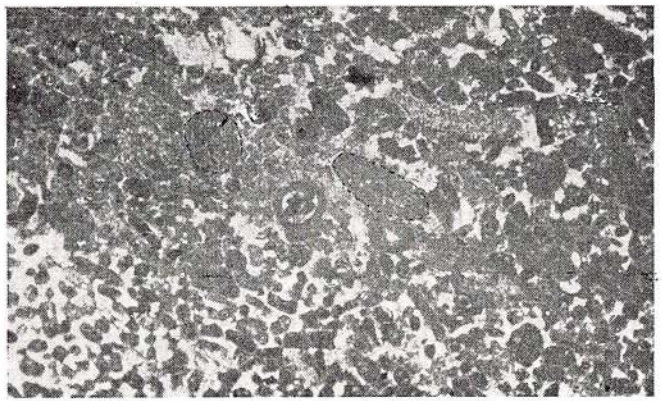
Fig. 1—8. — Pelletal (microcoprolithic) and/or microoncolitic calcarenites (grainstones) or lutaceous calcarenites (packstones and wackestones) displaying various fabrics and textural relationships between their petrographic components.

The calcareous material consists of microoncolites (abundant in Fig. 3,4,5), fecal pellets (abundant in Fig. 1,2,6—8) and micrite matrix (most of the micrite matrix and many of the grains exhibit their constituent micrite evolved to microspar as in Fig. 1,6,8). All the light spaces in the photographs represent gypsum patches as void-or pore-filling cement or as replacive or displacive growths within or between the grains (the latter case led finally to thin gypsum lenses or strips (g) with massive structure, as in Fig. 6,7). Most of the microoncolites lack the original calcareous nucleus, exhibiting a central patch of gypsum instead; some of them show gypsum growths between the micritic laminae of their envelopes (Fig. 3, 4). Some well-visible fecal pellets of the *Favreina salevensis* type, either well-preserved or cracked to brocken by gypsum growth, occur in the photos 1, 3, 6 and 8 (some of them outlined with dashed line). The photo 1 also shows very clear, small, cylindrical fecal pellets ascribed to gastropods.

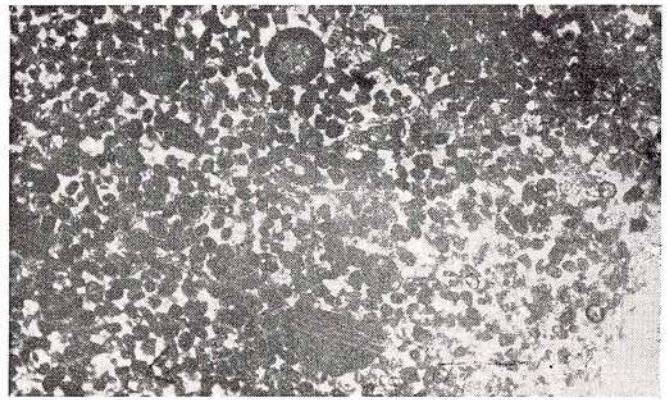
Some details concerning the infilling of pore spaces by gypsum followed by its displacive or replacive growth between or within the calcareous grains are seen in Fig. 5 (enlargement in the left of Fig. 4): the pores are filled up with gypsum; some microoncolites (m) show gypsum patches instead of nuclei; the calcareous grains (both microoncolites, m, and *Favreina salevensis* fecal pellets, f) „imbued" with gypsum display various degrees of dismembration.

All photographs under plane polarized light.

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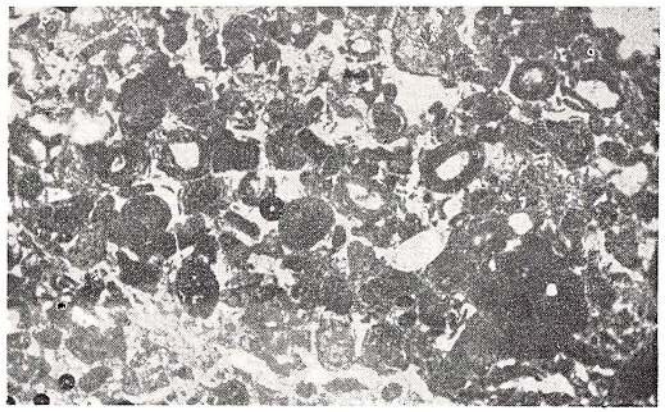
1



2



3

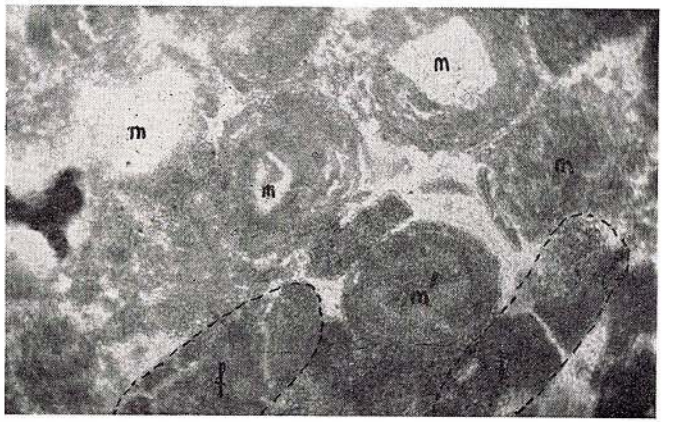


4

0 0,25 0,5 1 2 mm

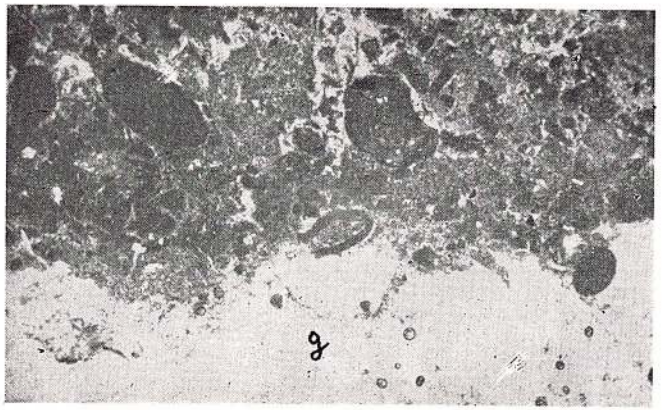
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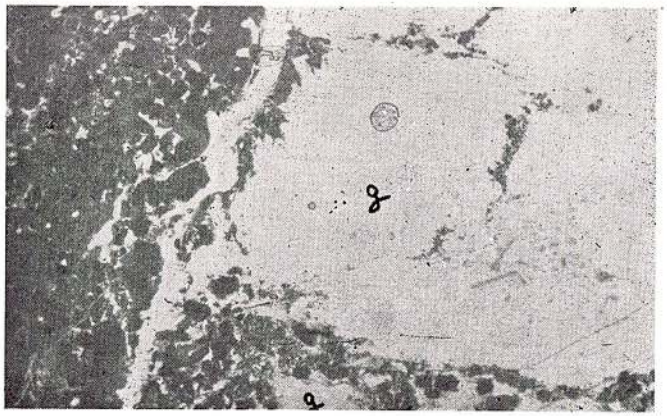


5

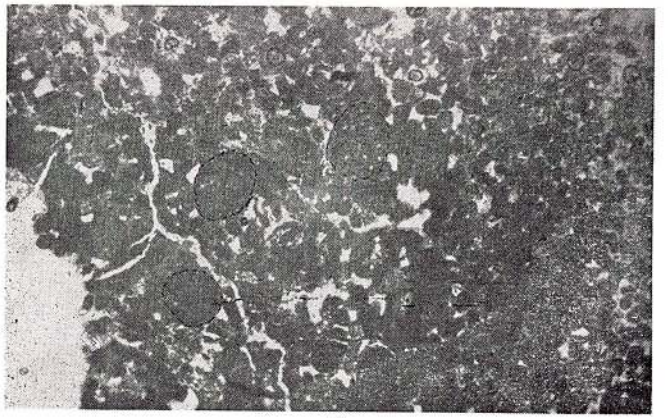
0 100 200 m



6



7



8

Fig. 6-8: 0 0,25 0,5 1 2 mm

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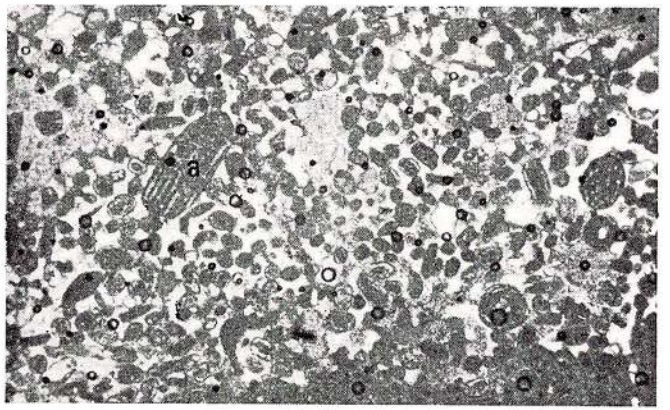
PLATE X (a, b)

Țăndărei sector

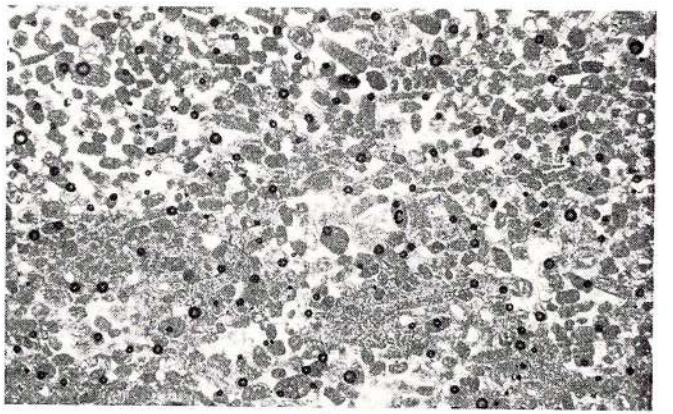
Thin sections in the core from 620—621 m depth (Complex II) in drilling no. 10

Fig. 1—8. — Various fabrics and textures in the same pelletal (microcoprolithic) and/or micro-oncolitic calcarenites or lutaceous calcarenites presented in Pl. IX. The intercorpuscular spaces filled with micrite matrix (mostly evolved to microspar) or gypsum cement. The pellets belong to 3 types : a) large, keg-shaped fecal pellets of crustaceans (*Favreina salevensis*) (clearly visible in Fig. 1, 5, 6, 8); b) small, elongate-cylindrical fecal pellets probably excreted by gastropods (typical in Fig. 7); c) small, ovoidal fecal pellets ascribed to polychaete annelids (abundant all over the photomicrographs). Note: for other petrographic aspects see explanations to Pl. IX, XI and XII.
All photographs under plane polarized light.

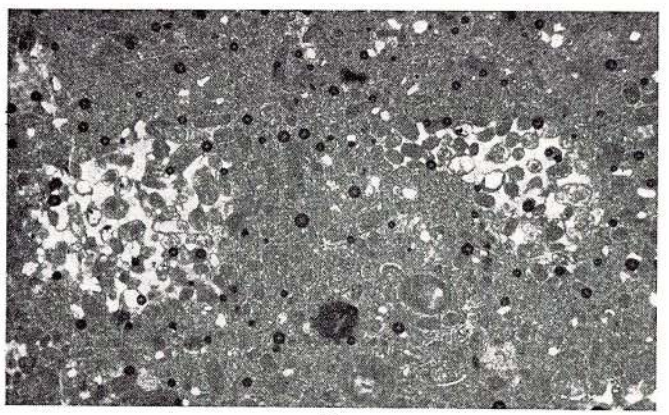
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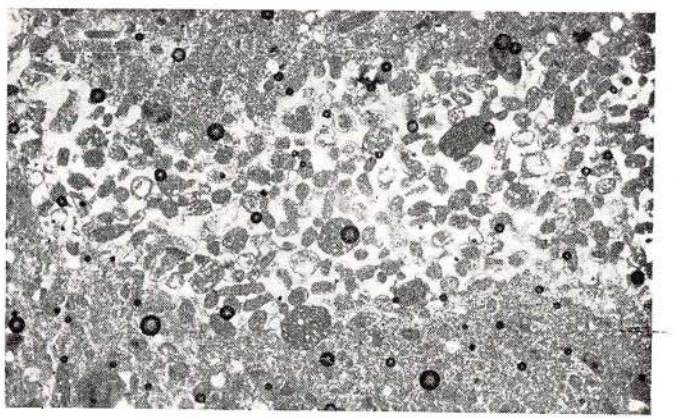
1



2



3

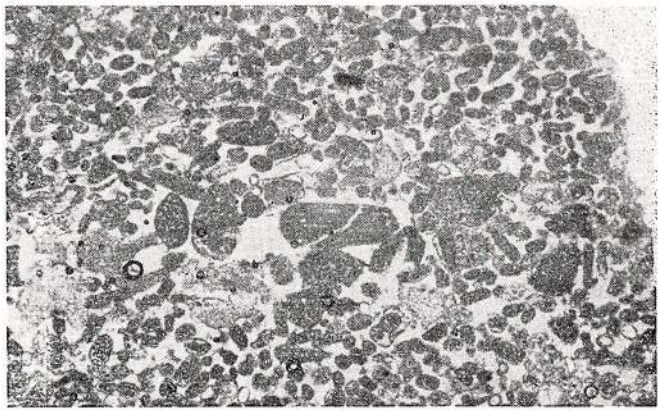


4

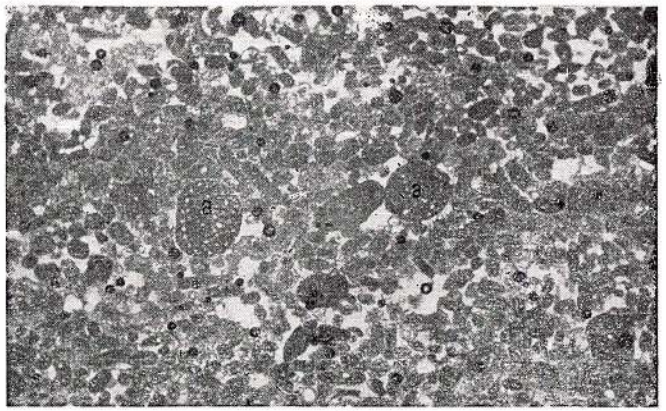
0 0,25 0,5 1 mm

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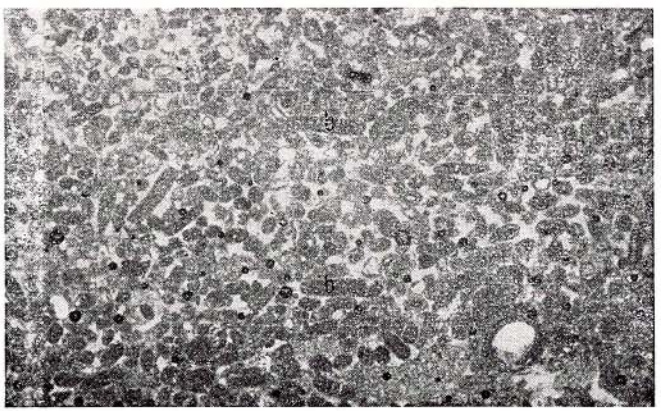
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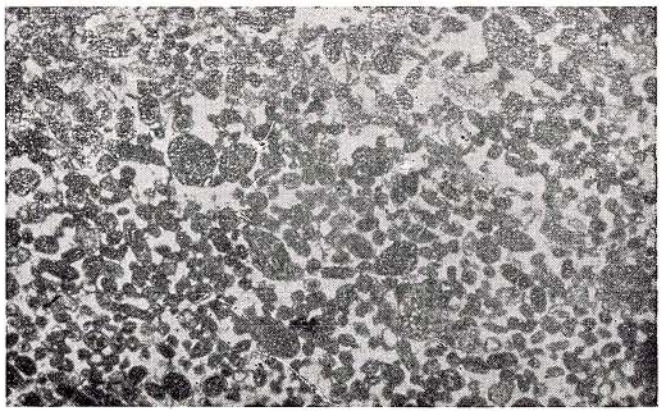
5



6



7



8

0 0,25 0,5 1 mm

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PLATE XI (a, b)

Țăndărei sector

Thin sections in the cores from drilling no. 10

Fig. 1—7. — Complex II (620—621 m depth). Details in the pelletal-microoncolitic limestones presented in Pl. IX—X, purposing to illustrate especially the dismembration of the original lime sands or muddy sands by gypsum displacive or replacive growth:

Fig. 1 — (detail in Fig. 4, Pl. X): Pelletal lutaceous calcarenite overlain by pelletal calcarenite cemented by gypsum. The pelletal lutaceous calcarenite consists mainly of small, ovoidal fecal pellets (worm dejections) embedded in an originally micritic matrix; almost all micrite matrix and most of the pellets aggraded to fine microspar 5—6 micra in crystal size; the unaggraded pellets surrounded by microspar rims. The pelletal calcarenite is cemented by gypsum displaying single optical orientation (the same orientation of the cleavage over the entire gypsum network); the fecal pellets belong to the already-mentioned *Favreina* type, small elongate-cylindrical type and small ovoidal type; the displacive growth of gypsum accounts for the local abnormally loose packing; many pellets are more or less replaced or dissociated by gypsum.

Fig. 2. — (detail in Fig. 8, Pl. X) Pelletal calcarenite consisting of the 3 above-mentioned fecal pellet types and minor amounts of microoncolites cemented by gypsum (of a single optical orientation). Gypsum assimilates partly or totally many pellets; many microoncolites display subepidermal circular narrow gypsum rims.

Fig. 3. — idem Fig. 2. Note the marked tendency of gypsum to replace, disintegrate, or break the calcareous grains. The preferentially disintegrated grains are those evolved to microspar.

Fig. 4—5. — (Fig. 5 shows a detail in Fig. 6, Pl. X) Pelletal lutaceous calcarenites containing patch-like pelletal calcarenites cemented by gypsum. Frequent, large *Favreina* fecal pellets occur more or less cross-sectioned. Most of micrite matrix and many grains display aggradations to microspar 5—6 micra in crystal size. A gradual passage is noted between microspar areas and gypsum patches, the former undergoing a particle-by-particle progressive disintegration within gypsum.

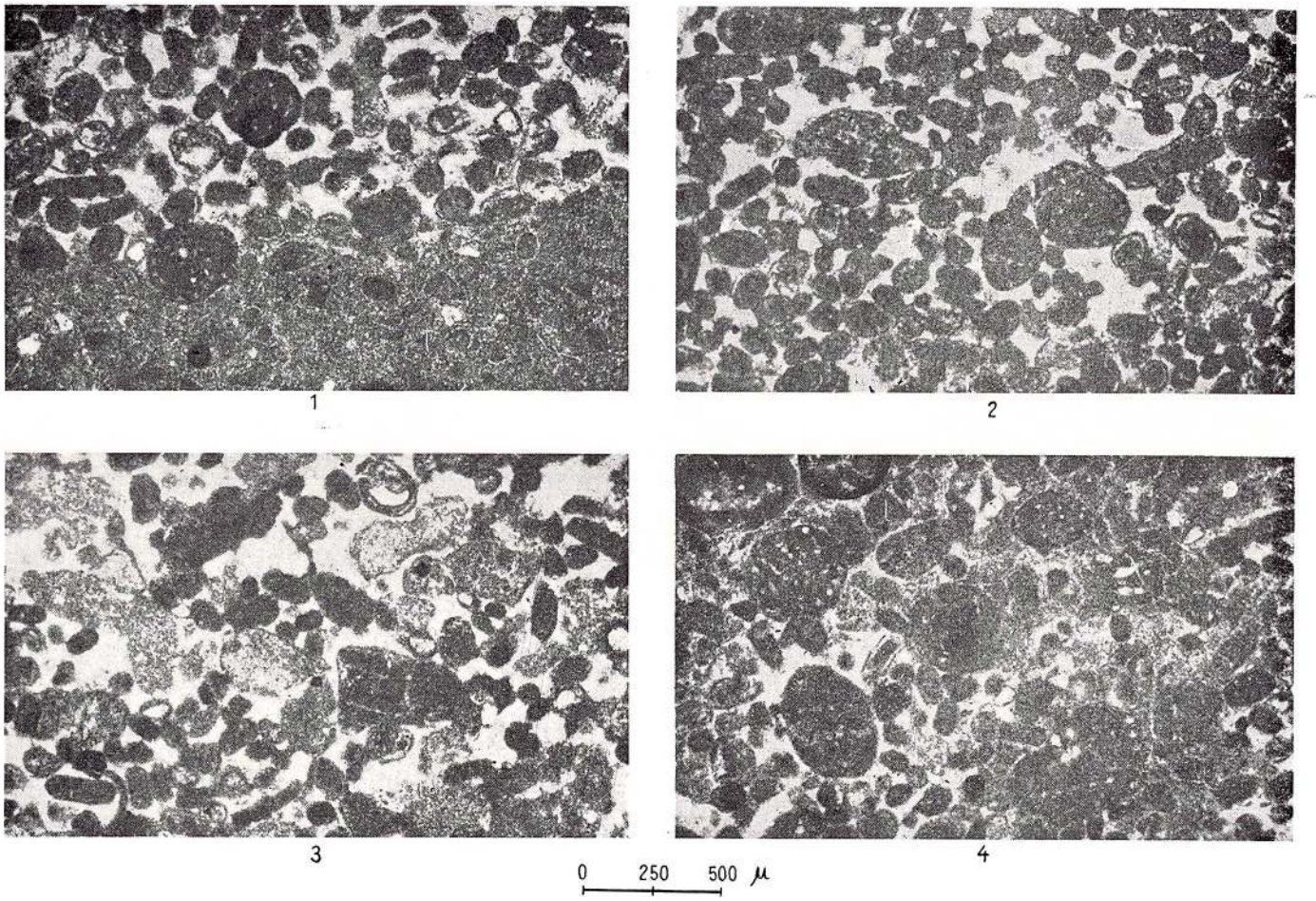
Fig. 6. — (detail in Fig. 7, Pl. X) idem to Fig. 4—5. Typical, lengthwise-sectioned, elongate-cylindrical fecal pellets (ascribed to gastropods) occur.

Fig. 7. — Microoncolitic limestone strongly affected by subsequent energic development of gypsum. Almost all calcareous material shows the component micrite particles enlarged to microspar. The gypsum highly dismembers the limestone by vigurous displacive and replacive growth between and within the calcareous grains. At present, remnants of allochems as well as shapeless microspar patches float within a gypsum mass.

Fig. 8. — Complex II (634—635 m depth). Dolomite consisting of euhedral rhombohedra more or less stained by dark organic matter. The more the organic matter within the crystals, the finer the crystallinity.

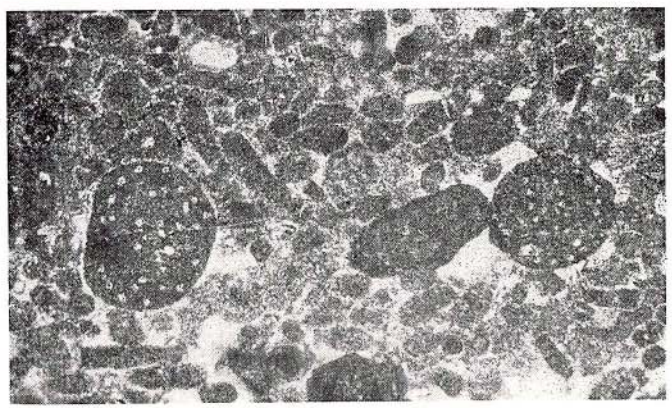
All photographs under plane polarized light.

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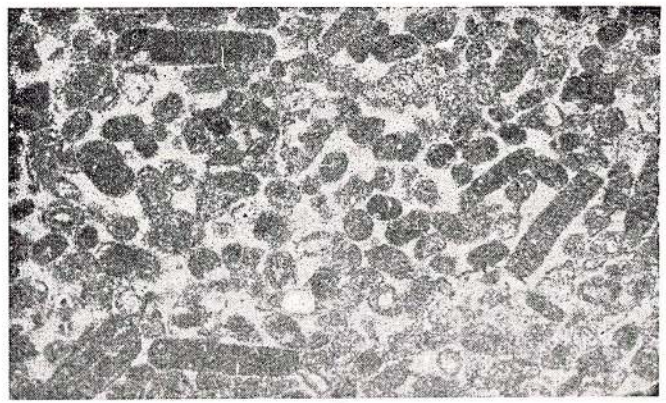


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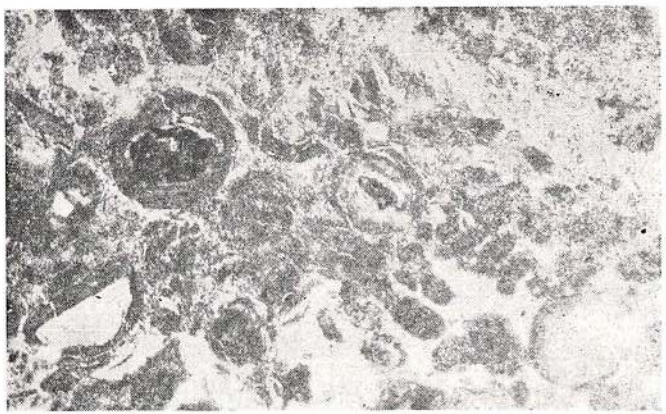
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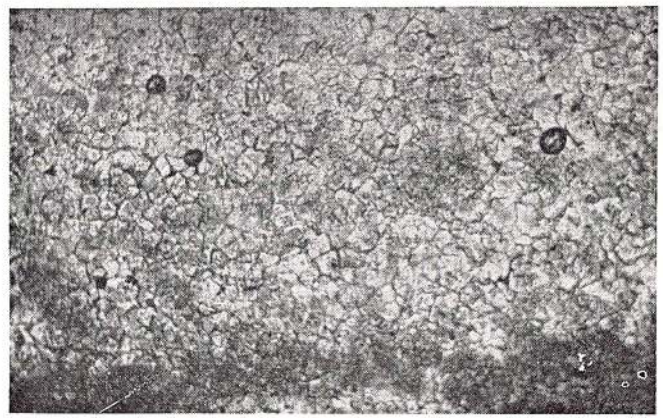
5



6



7



8

0 250 500 μ

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PLATE XII (a, b)

Țăndărei sector

Thin sections in the core from 620—621 m depth (Complex II) in drilling 10 — Details at high magnitude, under plane polarized light, in the limestones presented in Pl. IX—XI, illustrating the microspar problem—

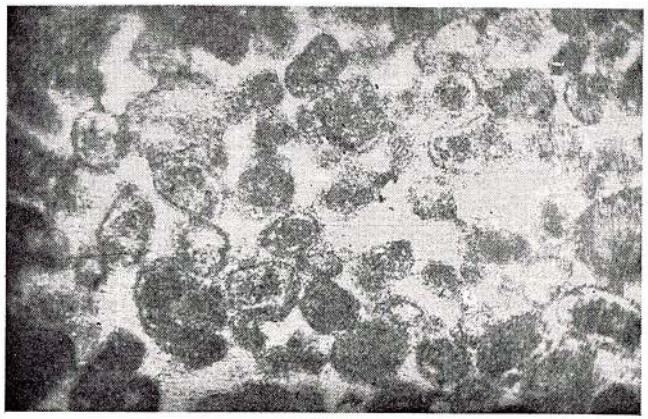
Fig. 1—2. — Pelletal calcarenites consisting of fecal pellets and rare microoncolites embedded within a gypsum mass. Most of the pellets are replaced by microspar (as irregular-shaped patches and fringing rims) which in its turn appears more or less dissociated within gypsum mass as individual microspar crystals or shapeless microspar patches. The microspar rims surrounding the calcareous grains are often separated from the respective grains by narrow gypsum rings. The pellets show disintegrations within or replacements by gypsum, they appearing sometimes as phantomatic or torn patches. Note the single orientation of the cleavage (single optical orientation) of gypsum in each photograph.

Fig. 3,5 and 7. — Calcarenite showing microspar 5—6 micra in crystal-size developed as large patches covering numerous calcareous grains; the preserved calcareous grains (mainly fecal pellets)) exhibit an internal micritic structure partly replaced by microspar as peripheral rims or small patches. The microspar areas and the preserved grains float within a gypsum mass of a single optical orientation in each photograph. The gypsum penetrates the microspar areas, gradually dismembering them from microspar fields containing irregular-shaped gypsum patches to microspar blebs with torn outlines accommodated within gypsum mass and still further to individual microspar crystals scattered within gypsum fields.

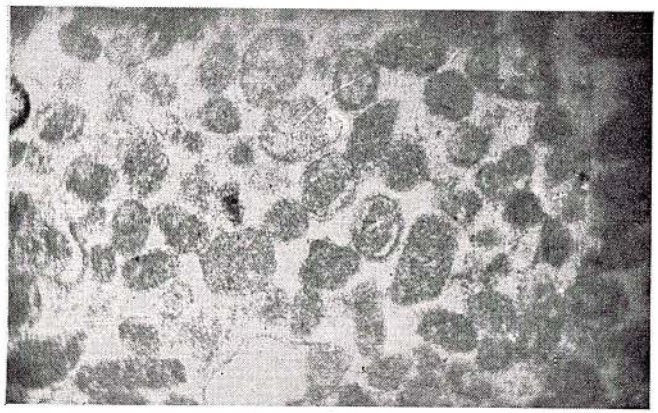
Fig. 4, 6 and 8 — Lutaceous calcarenite consisting of a former micrite matrix (now almost entirely aggraded to microspar 5—6 micra in crystal size) and calcareous grains mainly of the fecal pellet type (most of them partly or totally evolved to microspar). At present, the aggraded micrite matrix and pellets make up a microspar groundmass containing unaffected pellets (displaying a micritic consistence). These relict grains appear surrounded by microspar rims probably formed at the expense of former micrite matrix.

All these facts suggest that the microspar resulted from an approximately penecontemporaneous aggradation of micrite fabrics (matrix and grains), before gypsum cementation took place. By all means, this microspar must be considered as formed at a moment when the calcareous sediments were still soft and under the seawater influence. Therefore, its formation proceeded in soft state and under aragonitic form (considering the sediments as originally aragonitici by comparison with their present-day analogous).

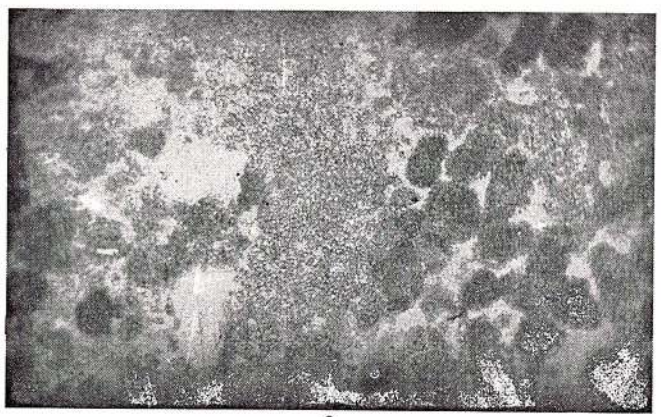
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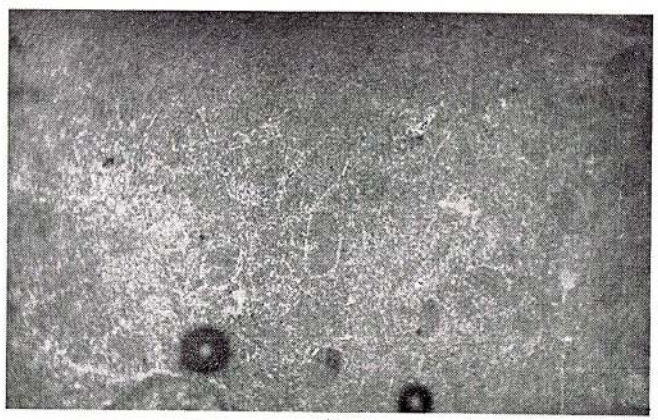
1



2



3

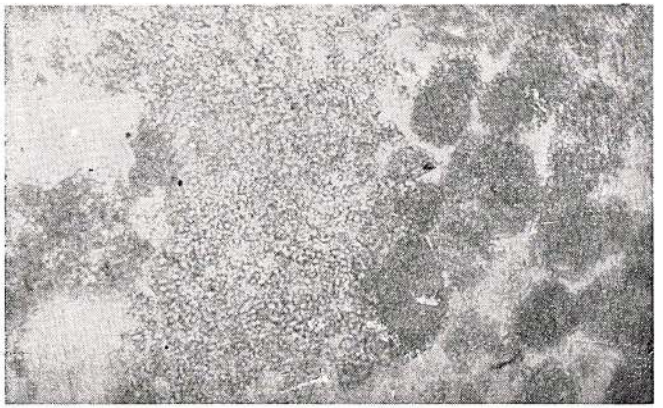


4

0 100 200 μ

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5



6



7



8

0 25 250 μ

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PLATE XIII (a, b)

Cireşu sector

Thin sections in the cores from drilling no. 9

Fig. 1--4. — Complex I (2 351—2 353 m depth). Micrite invaded by dolomite rhombohedra as disseminations, crowds and thin strips. Plane polarized light.

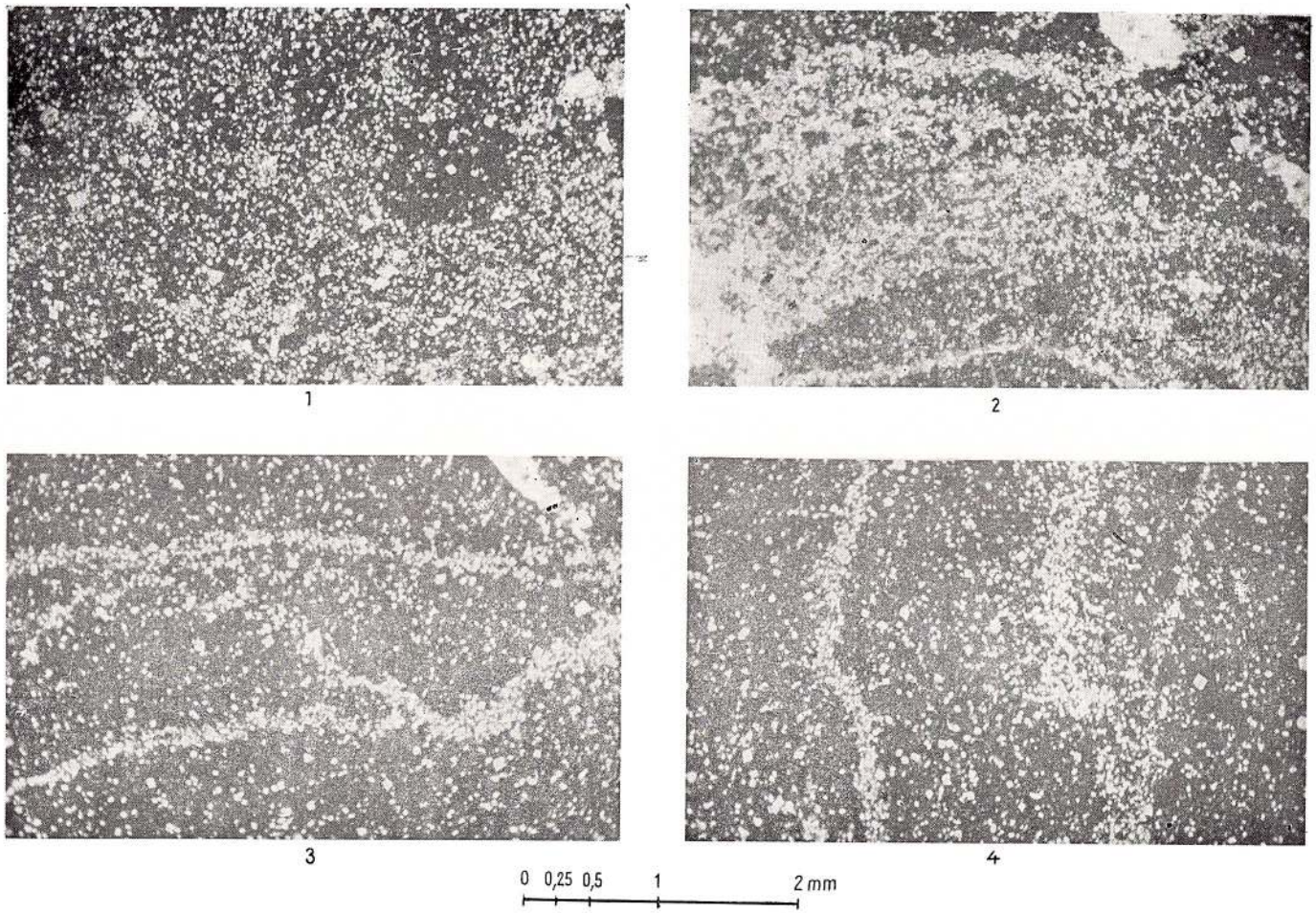
Fig. 5. — Complex II (2 208—2 209 m depth). Clotted micrite containing rare ostracods, *Actinoporella* tests (A) and tiny gypsum microaggregates and crystals (the white spots inside the micrite). The *Actinoporella* test consists of gypsum (of a replacive type). Plane polarized light.

Fig. 6. — Complex II (2 208—2 209 m depth). The same clotted micrite containing dismicrite textures (exhibiting strong features of an algal product) (d). The sparry eyes in dismicrite patches are filled with gypsum and/or calcite. Plane polarized light.

Fig. 7. — Complex II (2 208—2 209 m depth). Photomicrograph showing the contact between a calcite-evaporite veinlet (upper right half of photo) and the host micrite (lower left half of photo). The veinlet consists of anhydrite (a) in a central position and calcite (c) in a parietal position including some gypsum patches (g). The host micrite contains tiny gypsum micro-aggregates (the white spots) and is pierced by a fine calcite veinlet. N +.

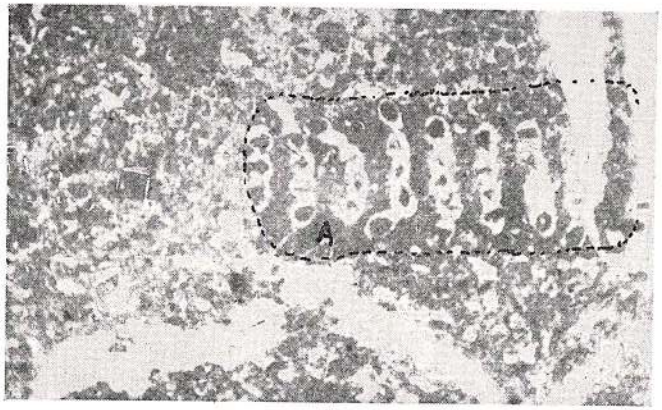
Fig. 8. — Complex II (2 208—2 209 m depth). Evaporite intercalation with contorted bedded structure and aligned-felted texture, consisting mainly of anhydrite and subordinately of gypsum. N +.

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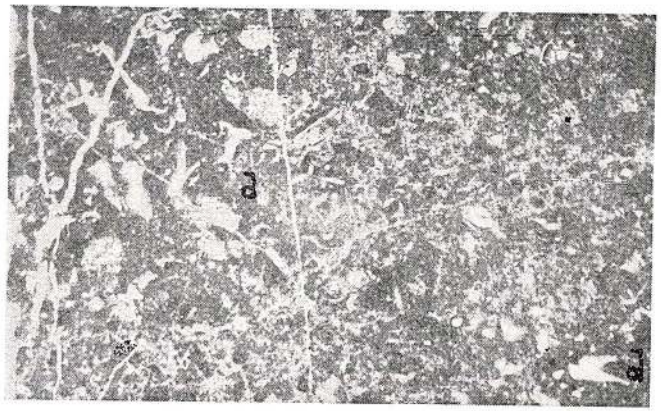


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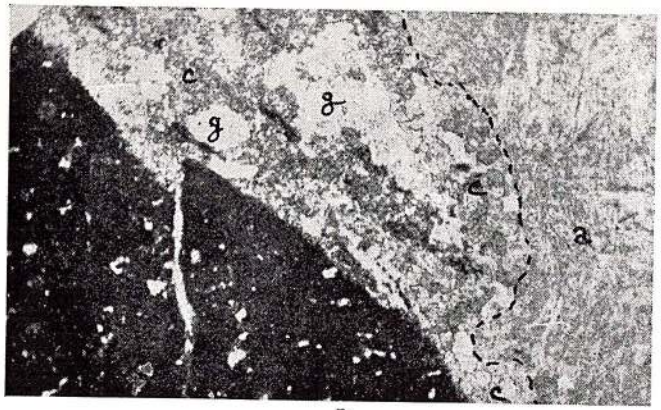
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5



6



7



8

0 0,25 0,5 1 2 mm

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PLATE XIV (a, b)

Cireşu sector

Thin sections in the core from 2 208—2 209 m depth (Complex II) in drilling no. 9

Fig. 1, 3, 5 and 7. — Pelletal (microcoprolithic) calcarenites. Textural and structural general features. Plane polarized light.

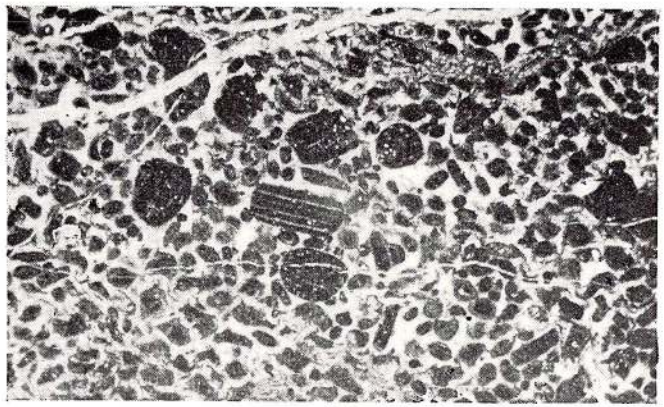
Fig. 2, 4, 6 and 8. — details in the pelletal (microcoprolithic) calcarenites presented in Fig. 1, 3, 5, 7. Plane polarized light.

The pelletal calcarenites consist of fecal pellets (belonging to the same 3 types described in Ţăndărei drilling, Pl. IX—XI: crustacean, large, keg-shaped, *Favreina salevensis* fecal pellets; small, elongate-cylindrical fecal pellets ascribed to gastropods; and small, ovoidal fecal pellets probably excreted by worms) embedded in a very finely or finely crystalline spar calcite. The fecal pellets consist of micrite particles less than 1 micron in size; the outer parts of the pellets appear often aggraded to a coarser micrite 2—3 micra in crystal size; in some pellets this larger micrite replaces the whole fine micrite.

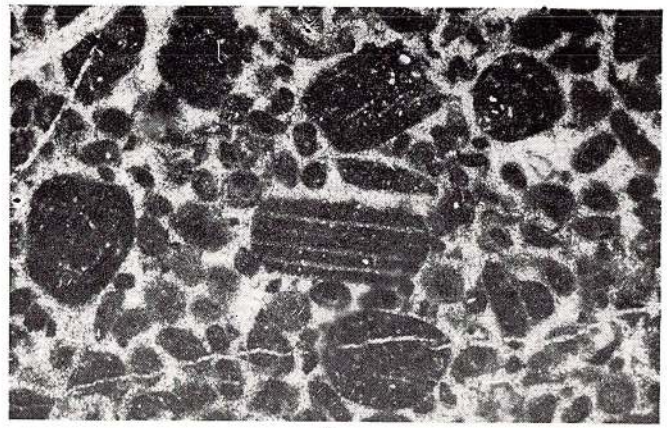
The limestones often show an abnormally loose packing. The ,,intercorpuscular'' spar calcite is 8—20 micra in crystal size in places enlarged to 30—60 micra crystal size; the spar calcite displayes a granular texture, seldom showing slight tendencies towards a drusy texture; usually, the larger the spar patch, the larger its crystallinity; the contact between the spar calcite ,,cement'' and the calcareous grains (either finely micritic or coarser micritic in consistence) is sharp, no gradual passage existing from the pellet crystallinity to the spar calcite one; the spar ,,cement'' cuts and transects the pellets and their internal fabrics, often lending them new and angular outlines; sometimes, the pellets appear more or less ,,disolved'', assimilated within the spar ,,cement''; a careful observation of the spar calcite network shows that most of its outlines are angular-serrate consisting of short, straight segments associated in poligonal forms strongly suggesting rhombohedra margins. In many instances the pellets show the destruction of their outer parts along the contact between them; in these cases the contact between them is usually masked by a thin strip of spar calcite (as in Fig. 6, lower left). All hese observations support the opinion concerning the replacement of the ,,intercorpuscular'' spar calcite by the diagenetic sequence of dolomitization-dedolomitization-local aggrading recrystallization: network-like dolomitization develope on the small grains and margins of larger grains; dedolomitization leading to dedolomitic microspar 8—20 in crystal size; local aggradation of this microspar to 30—60 micra pseudospar. The contact solution phenomena suggested by the destruction of the pellets along their contacts involve a rather strong compaction before and during the circulation of the solutions responsible for the „intercorpuscular'' dolomitization.

As regards the aggradation of the original extremely fine micrite composing the outer parts of the pellets to larger micrite, this process developed from pellet surface inwards and is thought to have taken place before the development of the above-mentioned diagenetic sequence, that is, it represents a penecontemporaneous process.

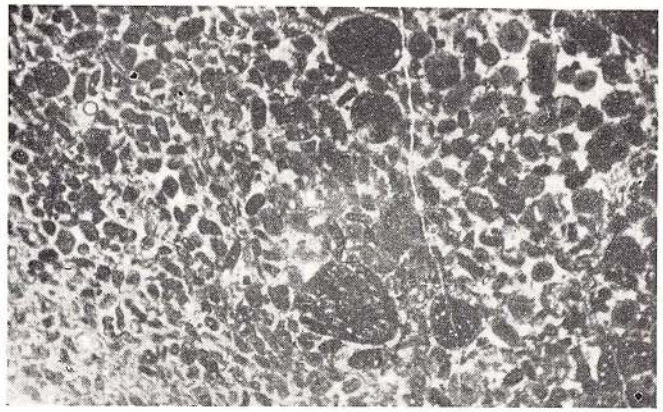
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1



2



3

Fig. 1,3: 0 0,25 0,5 1 2 mm

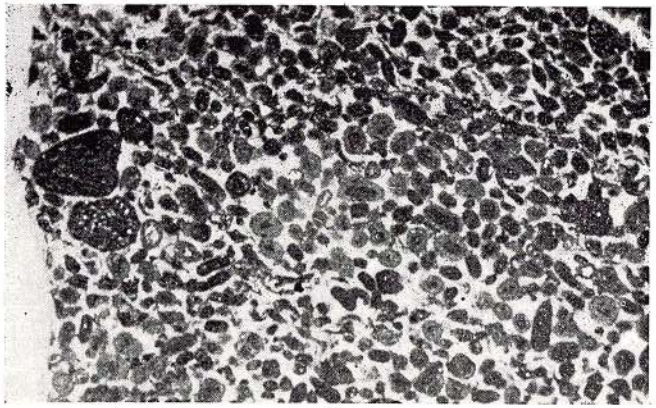


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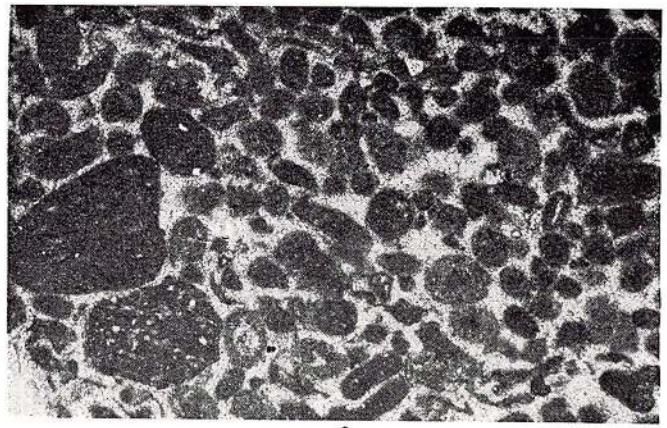
Fig. 2,4: 0 250 500 μ

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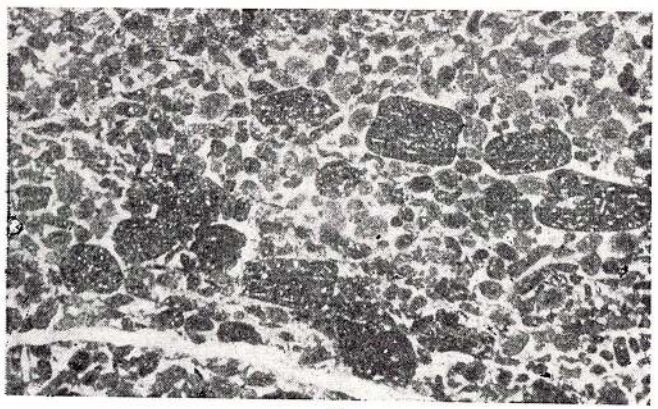
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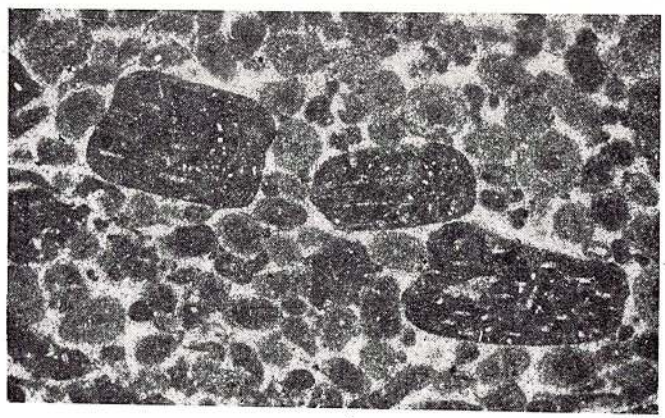


6



7

Fig.5,7: 0 0,25 0,5 1 2 mm



8

Fig.6,8: 0 250 500 μ

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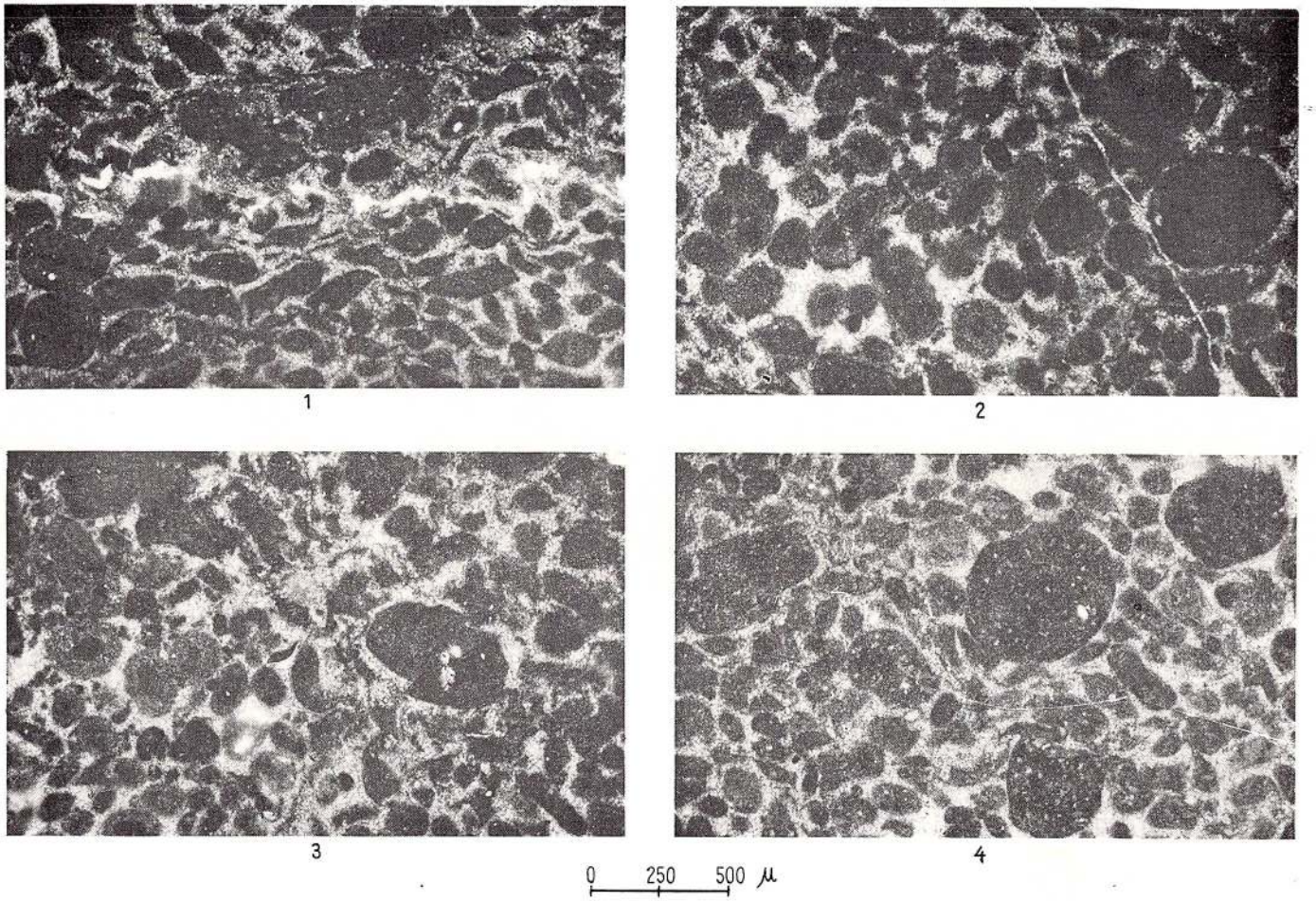
PLATE XV (a, b)

Cireşu sector

Thin sections in the core from 2 208—2 209 m depth (Complex II) in drilling no. 9

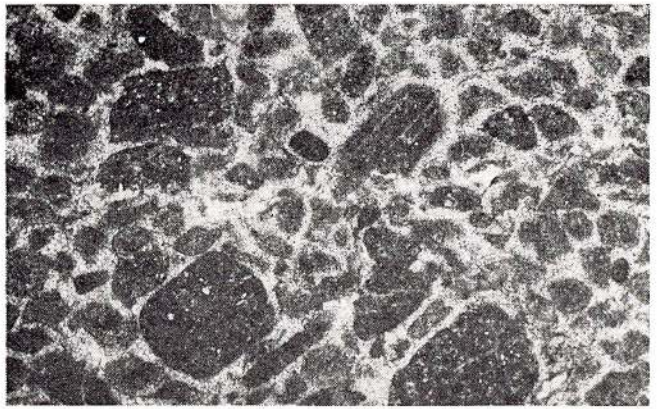
Fig. 1—8. — Details in the same pelletal grainstones presented in the Pl. XIV. The replacement origin of the spar „cement" by the diagenetic sequence of dolomitization-dedolomitization-local recrystallization is more obvious than in the previous plate: the pellets are reoutlined, cut, more or less assimilated or invaded by the „intercorpuscular" spar calcite; the edges of the spar calcite areas display angularities suggesting margins of former dolomite rhombohedra. Many large pellets (Fig. 3, 4, 8) show marked deformations by fracturing followed by slipping of the resultant fragments along the fracture planes, as well as partial „melting"(obvious in Fig. 1, 3, 5) on account of contact solution phenomena, proving that the pelletal grainstones were submitted to rather strong pressures before and during the development of the network-like dolomitization. The photomicrographs, following from Fig. 1 through Fig. 8, show an increasing in the intensity of the diagenetic sequence, exhibiting the resultant network-like spar calcite inscribing in a gradual scale from thin strips along the grain boundaries (Fig. 1) to a spar groundmass accommodating more or less destroyed fecal pellets (Fig. 8). The cluster of pellets in the left of Fig. 2 represents a relict, undismembered, small part of the original pelletal sand as it looked after the pore-compaction and the aggradation of the outer parts of the pellets to coarser micrite have taken place and before the „intercorpuscular" dolomitization has started.
All photographs under plane polarized light.

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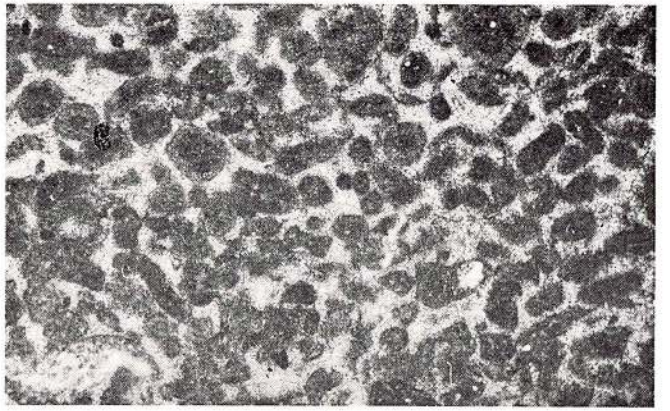


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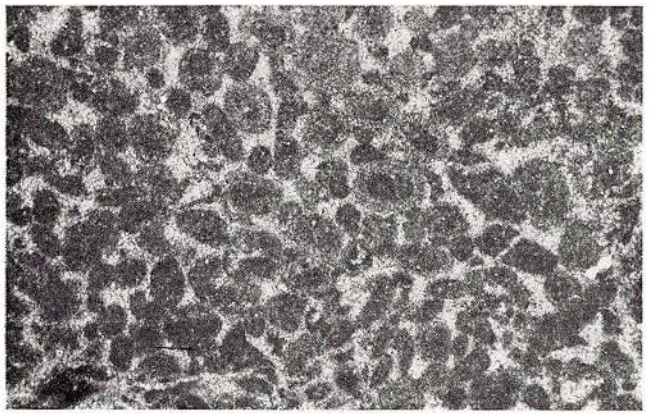
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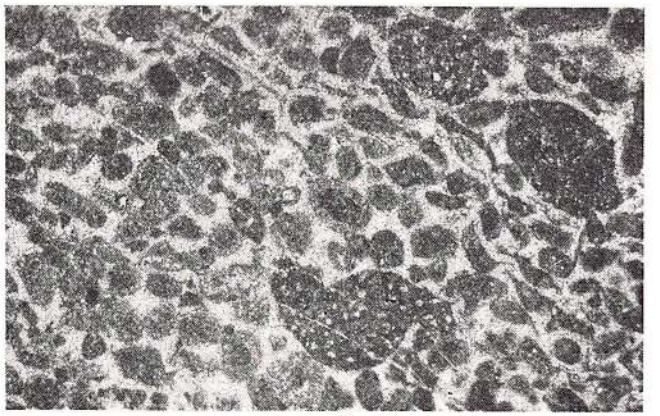
5



6



7



8

0 250 500 μ

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PLATE XVI (a, b)

Cireşu sector

Thin sections in the cores from drilling no. 9

Fig. 1—4. — Complex II (2 208—2 209 m depth). The same pelletal calcarenites presented in Pl. XIV—XV. The limestones appear strongly "recrystallized" with pelletal relics floating in a very finely or finely crystalline calcite; they stand for the extreme case ascertained within these pelletal calcarenites, where the diagenetic sequence of dolomitization-dedolomitization-recrystallization reached the highest intensity, deeply penetrating the large pellets and almost completely consuming the smaller ones. Plane polarized light.

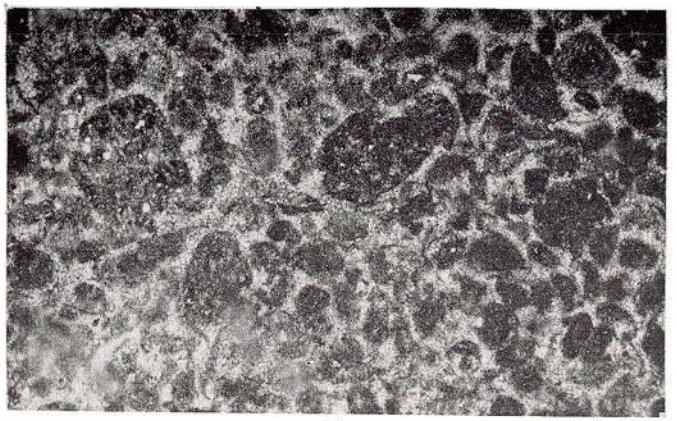
Fig. 5. — Complex II (2 279—2 280.80 m depth). Micrite containing frequent ostracods and numerous quartz grains. Plane polarized light.

Fig. 6. — Complex II (2 279—2 280.80 m depth). Micrite containing large anhydrite micronodules (a) with felted texture. A large micronodule appears in the middle left of the photomicrographs, and other two micronodules are partly visible in the left and lower left of the photomicrograph. Plane polarized light.

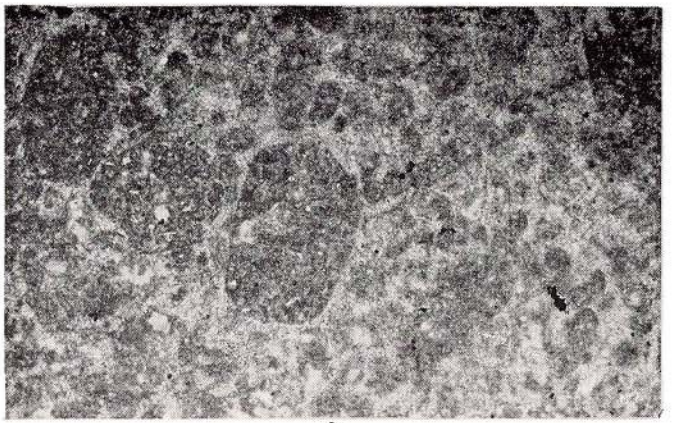
Fig. 7. — Complex II (2 279—2 280.80 m depth). Micrite (containing tiny quartz grains) crossed by a thin evaporite veinlet (v) consisting of anhydrite and gypsum crystals disposed quasiperpendicular to the veinlet walls; the evaporite crystals are not parallel to one another; some of them pierce the veinlet wall, penetrating the host micrite. These observations support a gravitational accumulation of free evaporite crystals within a dessication crack, followed by in place growth of some of the evaporite crystals, those of an excessive growth managing to exceed in length the veinlet width and to penetrate the host micrite. N+.

Fig. 8. — Complex II (2 279— 2 280.80 m depth). Anhydrite exhibiting lath-shaped texture. Origin : anhydrite intercalation of a bedded massive structure. N +.

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1



2



3



4

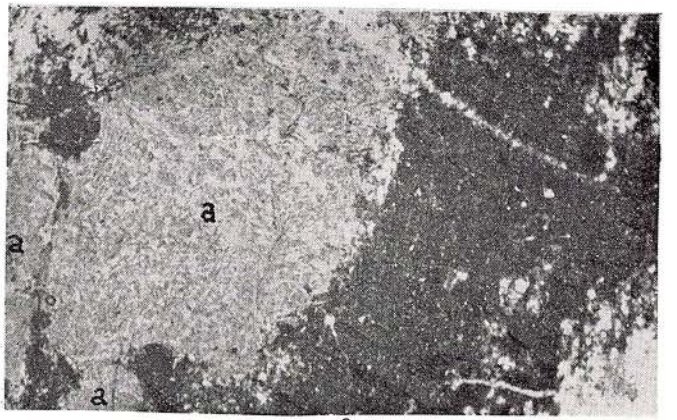
0 250 500 μ

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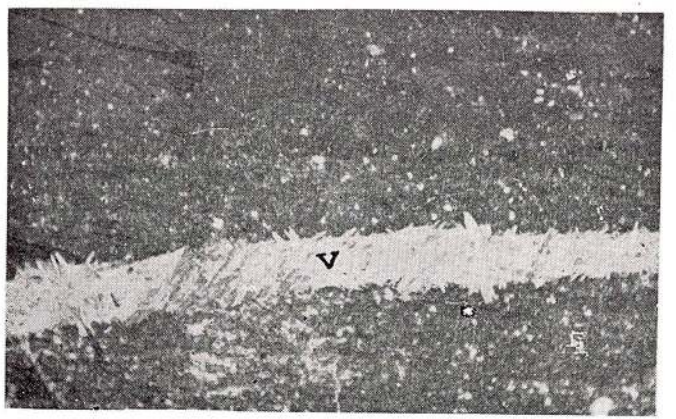
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5



6



7



8

0 0,25 0,5 1 2 mm

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PLATE XVII (a, b)

Țăndărei sector

Core specimens from 620—621 m depth (Complex II) in drilling no. 10

Fig. 1. — Core specimen showing a calcareous sequence containing three thin gypsum intercalations: the upper one (g_1) with monocrystal behaviour displays streaky massive structure, while the two lower intercalations (g_2 and g_3) are obviously polycrystalline and exhibit nodular to nodular-mosaic structure. The limestone consists of a lower, stromatolitic level (s), and an upper, calcarenitic sequence (c) composed of microoncolitic-microcoprolithic (fecal pellets) calcarenites with micrite matrix or gypsum cement.

Fig. 2. — Core specimen consisting of two thin gypsum beds with nodular-mosaic structure developed within the microoncolitic-microcoprolithic calcarenite (this last one preserved as a thin strip in the middle of the core specimen, as remnants at the top and the bottom of the core specimen, and as calcareous fragments entrapped within gypsum beds among the gypsum nodules).

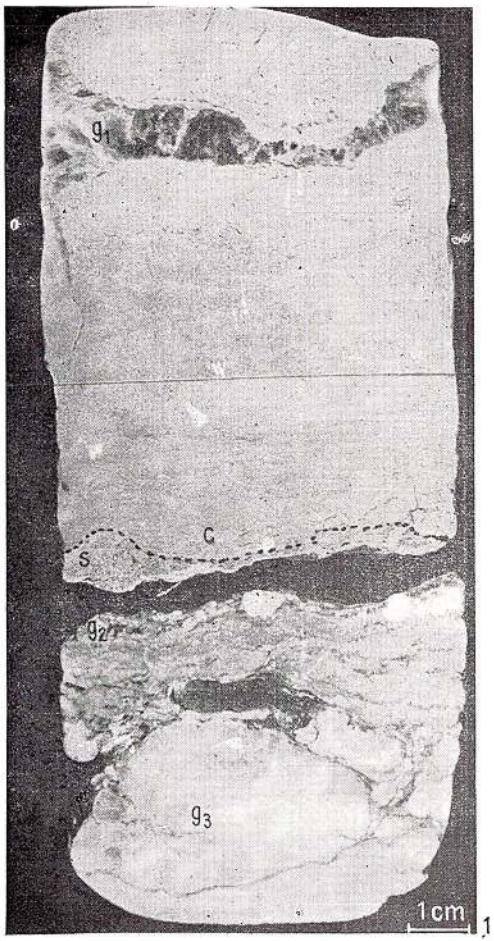
Fig. 3. — The oposite side of the core specimen (upper half) presented in Fig. 1, showing the microoncolitic-microcoprolithic limestone and the thin gypsum intercalation with streaky massive structure (g_1) cut across by a quasivertical, slightly waved gypsum veinlet (v) containing rare angular calcareous fragments.

Fig. 4. — Core specimen provided by a thick gypsum bed with nodular-mosaic to mosaic structure. The internodular spaces are filled up with greenish clay material.

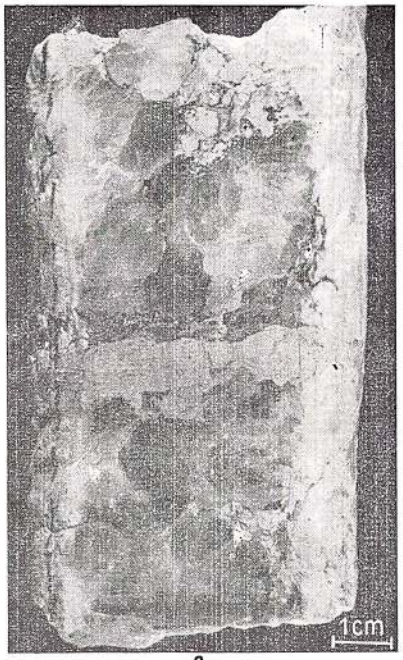
Fig. 5. — The oposite side of the upper half of the core specimen in Fig. 4, showing a well-visible gypsum nodule (n).

Fig. 6. — Core specimen proceeding from a thick gypsum bed with massive structure. This gypsum bed type is free from clayey material.

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1



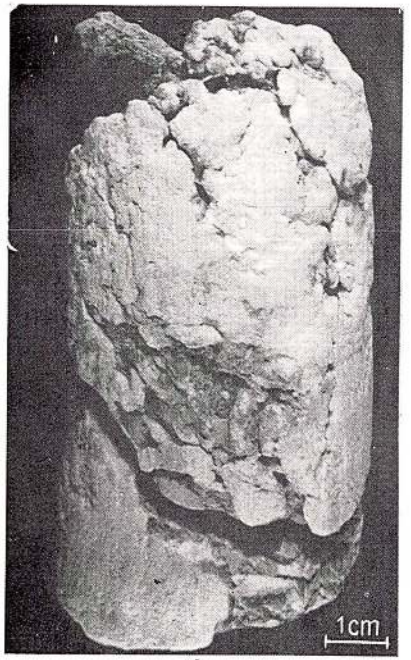
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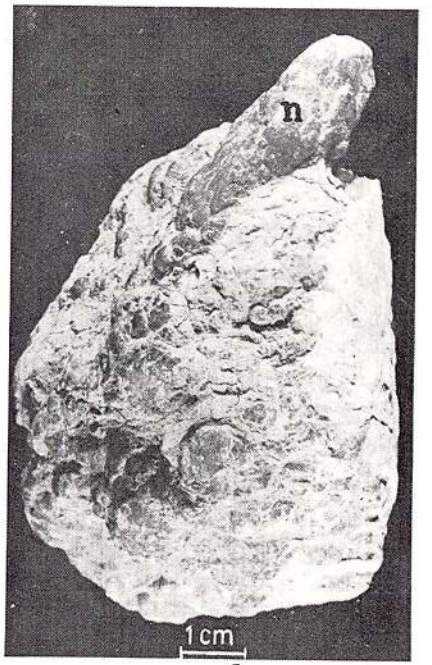
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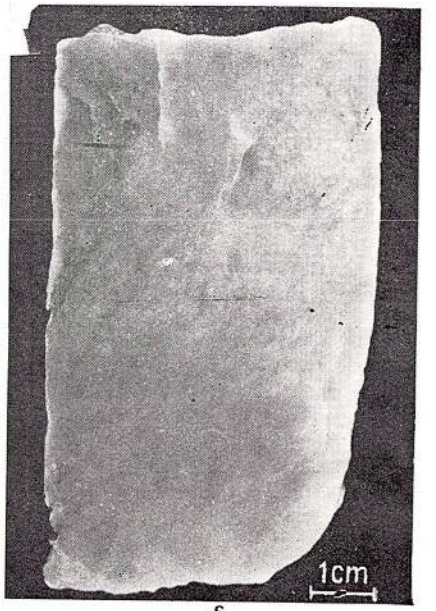
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4



5



6

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PLATE XVIII (a, b)

Țăndărei sector

Details in the cores specimens presented in Pl. XVII, Fig. 1—2. Complex II, 620—621 m depth

Fig. 1. — Detail in the upper middle of the core specimen presented in Pl. XVII, Fig. 1. The microoncolitic calcarenite, bearing a flat-pebble type, large intraclast (i), contains gypsum as cement and as a narrow band (g_1) of a streaky massive structure (resulted from the local excessive growth of the gypsum cement).

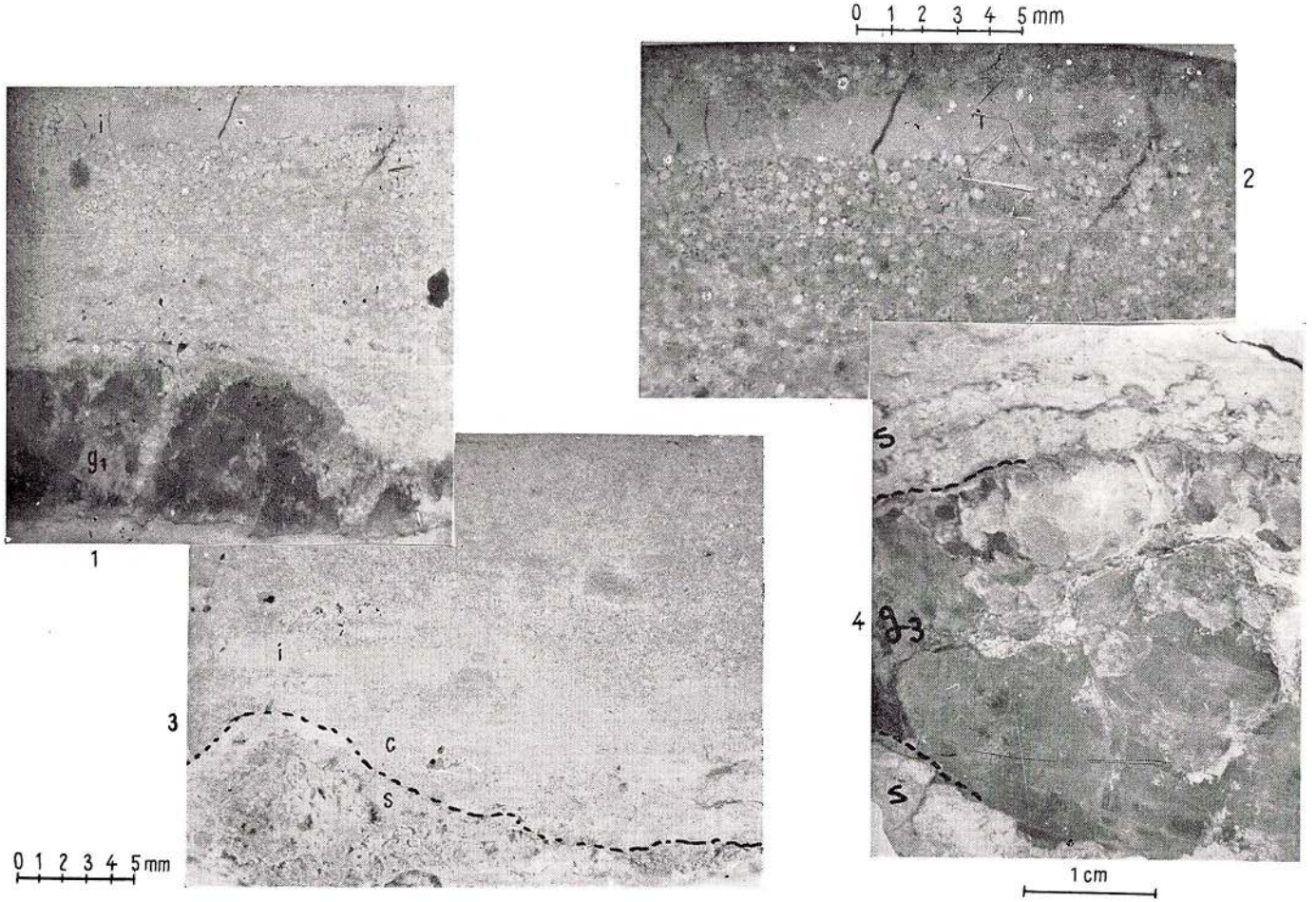
Fig. 2. — Detail in the upper part of Fig. 1, this Plate: microoncolitic calcarenite bearing a large intraclast (i) whose surface appears moulding the grains of the host sediment. The imprinting of the grains of the host calcarenite on intraclast surface proves the merely semiindurated state of the intraclast at the moment of its reworking and burial within microoncolitic sand.

Fig. 3. — Detail in the middle of the core specimen in Pl. XVII, Fig. 1, showing the stromatolite (s)/calcarenite (c) contact. The calcarenitic sediment contains large intraclasts of a flat-pebble type (i).

Fig. 4—6. — Details in the lower part of the core specimen in Pl. XVII, Fig. 1: stromatolitic laminae (s) enclosing a thin gypsum bed of a nodular-mosaic structure (g_3). Fig. 5—6 are enlargements in Fig. 4.

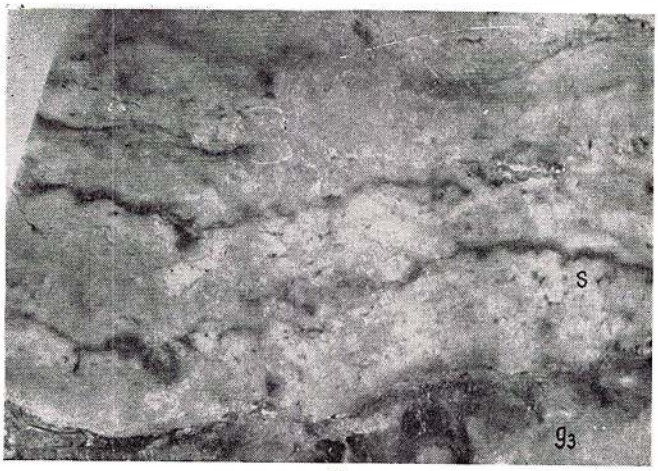
Fig. 7. — Detail in the middle of the core specimen in Pl. XVII, Fig. 2: two thin gypsum beds of a nodular-mosaic structure (g) separated by a thin calcarenite level (c) and containing calcareous fragments resulted from the disruption of the host calcarenite.

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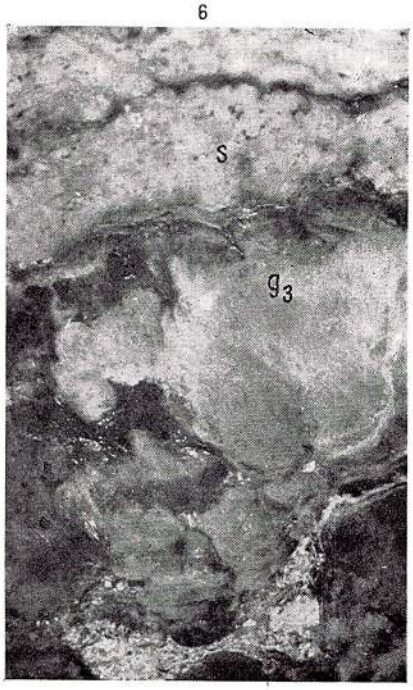
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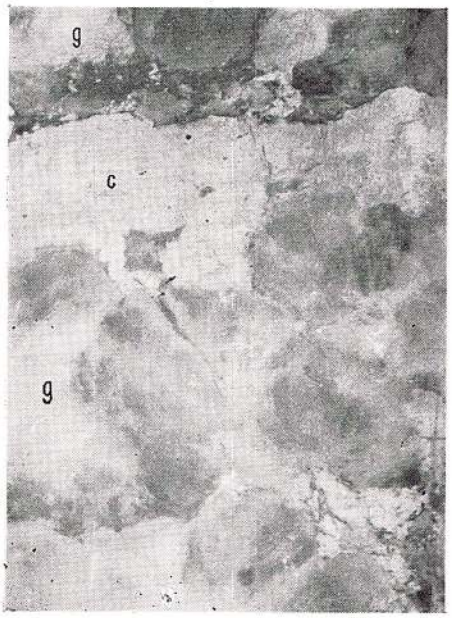
5

Fig.5,6: 0 1 2 3 4 5 mm

6



7



1 cm

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PLATE XIX (a, b)

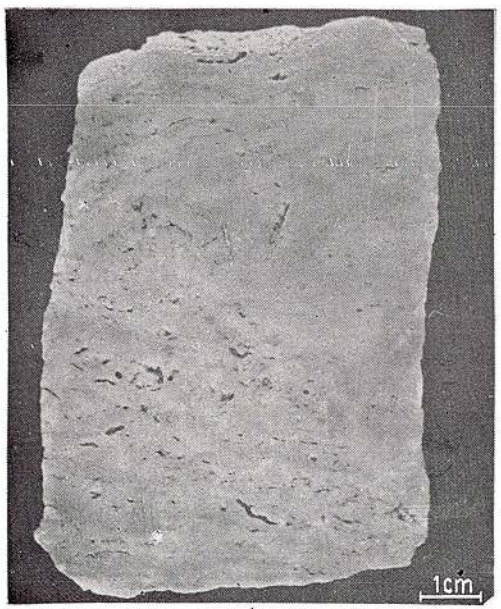
Țăndărei sector

Core specimens from 634—635 m depth (Complex II) in drilling no. 10

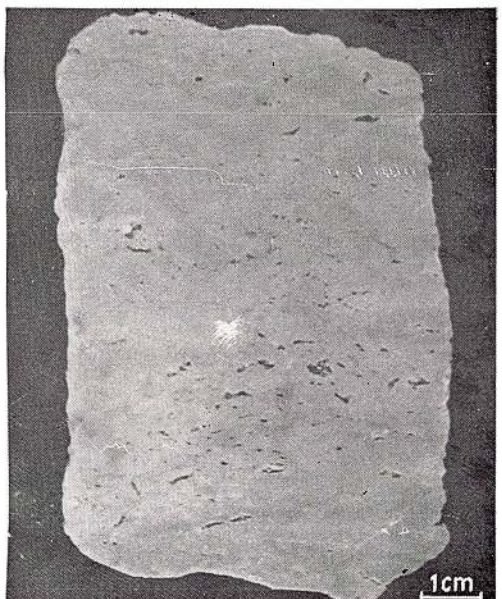
Fig. 1—2. — Core specimens of dolomite exhibiting waved laminar structure, the laminae being separated by rows of laminoid fenestrae. The slabs suggest former calcareous stromatolitic deposits subsequently dolomitized.

Fig. 3—5. — Details in the dolomitized stromatolitic sequence from 634—635 m depth. The laminoid fenestral fabric is conspicuous in photos 3—4. The alternation of dark and light levels, depending on the amount of organic matter entrapped within dolomite crystals, is clearly demonstrated by photo 5.

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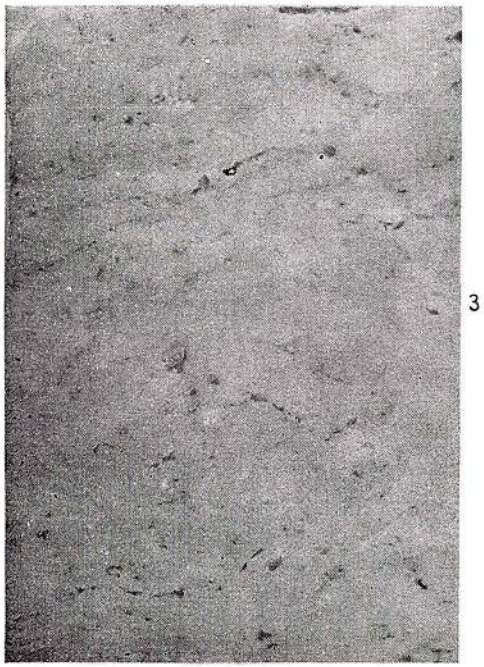
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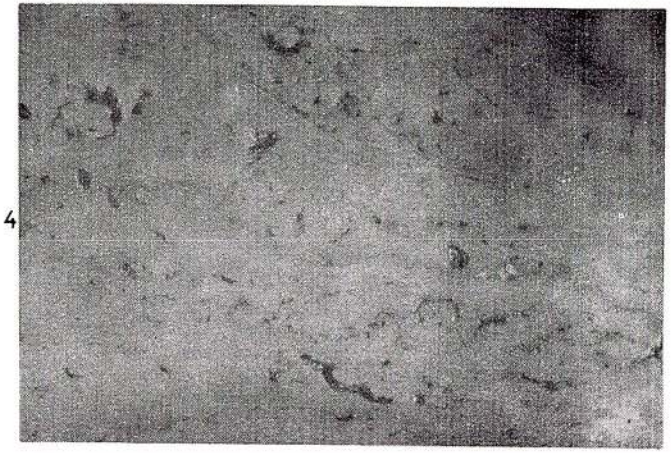
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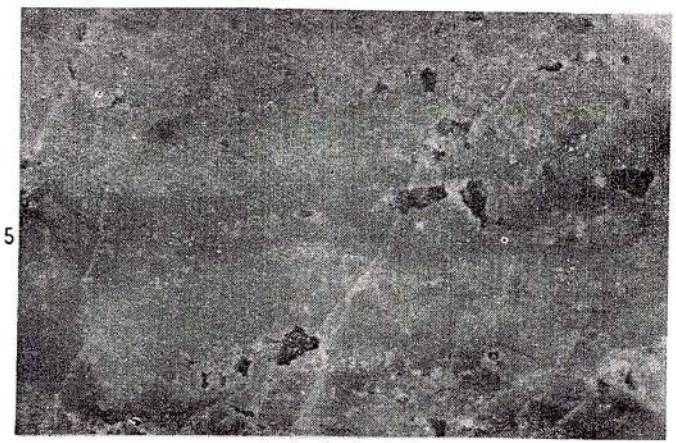
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3



4



5

Fig.3,4: 1 cm

0 1 2 3 4 5 mm

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ORIGIN OF SOME MESOZOIC BASINAL LIMESTONES FROM THE REŞIŢA ZONE (SOUTH CARPATHIANS)[1]

BY

GRIGORE POP[2]

Abstract

Within the Reşiţa Zone the typical basinal carbonate sequence is represented by the Upper Oxfordian — Berriasian Valea Aninei, Brădet and Marila Limestones. Valea Aninei Limestones are differing from the other carbonate units due to the high frequency of allodapic limestones as well as banded and nodular cherts. Brădet Limestones occur in the marly nodular limestone facies, and the Marila Limestones are essentially made up of pelagic micrite containing calpionellids. In all these limestones micrites of pelagic origin are predominant, and their share increases from older to younger limestones and from east to west. The deposition of carbonate sediments took place in the outer part of the Reşiţa Trough, within its basinal margin and slope, bellow aragonite compensation depth and over that of calcite, through accumulation of pelagic particles as well as redeposition of some sediments, initially formed in shallow marine environments of the adjacent threshold (Semenic). Redeposition of non-consolidated sediments and reworking of the semi-consolidated ones occured through various submarine ,,flows'' (slidings, semi-fluid flows, turbidity currents); owing to the submarine transport, sediments of diverse origin were mixed in various proportions and modes. Lithification of basinal sediments, selectively displayed as a function of their fabrics, has taken place through the compaction, neomorphism, partial replacement by silica of biogenous origin, cementation, and locally dolomitization processes. Compaction, partially neomorphism and cementation of sediments as well as their replacement by silica had occurred during a very early diagenetic stage.

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[1] Received on 23rd July 1975 and accepted for publication on July 30th 1975.

[2] Institutul de geologie şi geofizică, str. Caransebeş 1, 7000 Bucureşti.

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INTRODUCTION

Within the South Carpathians the Jurassic and Cretaceous carbonate formations occur in some areas known as „sedimentary zones", and generally represent the Middle Jurassic to Aptian interval. In each of these main zones carbonate formations have a certain lithologic constitution and some lateral and vertical facies variations involving important petrological implications.

The Reşiţa Zone is the innermost sedimentary zone within the South Carpathians and is located in the inner part of the Getic Realm thus bordering to the west upon the Supragetic Unit. Within this zone the Upper Paleozoic and Mesozoic formations (Triassic, Jurassic and Lower Cretaceous) make up several frequently asymmetric and faulted longitudinal anticlines and synclines (Fig. 1).

As a whole the Jurassic and Cretaceous sedimentary sequence consists of Lower Jurassic lacustrine conglomerates, sandstones and bituminous shales (< 500 m thick), Middle Jurassic to Hauterivian predominant basinal limestones and marls (600—1000 m thick), Barremian-Aptian shallow water limestones (especially algal boundstones and grainstones) (300—600 m thick) and Albian and Cenomanian? glauconitic and calcareous sandstones (< 200 m thick) (see : Năstăseanu, 1964).

A few years ago the author began petrological investigations of the Mesozoic basinal and shallow water carbonate formations from the Reşiţa Zone; the results of this study, partially recorded in two unpublished reports (Pop and Jeana Ionescu, 1972 [3]; Pop, 1973[4]),

[3] Gr. Pop and Jeana Ionescu. Petrological Study of the Mesozoic Carbonate Deposits from the Central Part of the Reşiţa Zone. 1972. (in Romanian) Arch. Inst. Geol. Geophys., Bucharest.

[4] Gr. Pop. Petrological Study of Mesozoic Carbonate Deposits from the Northern Part of the Reşiţa Zone. 1973. Arch. Inst. Geol. Geophys., Bucharest (in Romanian).

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are going to be presented in several papers. The first paper will include the data concerning the sedimentary environments and some diagenetic aspects of the Upper Oxfordian-Berriasian basinal limestones.

BASINAL LIMESTONES

Lithostratigraphic Units

The typical basinal carbonate deposits from the Reşiţa Zone are successively represented by the Valea Aninei Limestones (Upper Oxfordian-Kimmeridgian), the Brădet Limestones (Lower Tithonian), and the Marila Limestones (Topmost Lower Tithonian-Upper Berriasian pro parte)

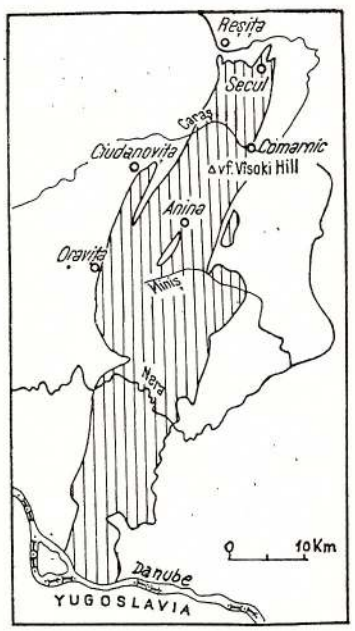


Fig. 1. — General distribution of the Mesozoic sedimentary formations in the Reşiţa Zone (South Carpathians).

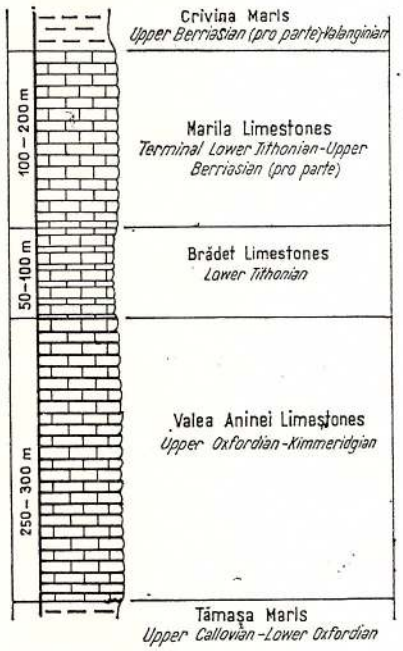


Fig. 2. — Basinal carbonate sequence from the Reşiţa Zone.

(Răileanu et al., 1957, 1964; Mutihac, 1959; Năstăseanu, 1964; Aurelia Bădăluţă-Năstăseanu, 1965[5]; Pop, 1974) (Fig. 2; Pl., Fig. 1).

[5] Aurelia Bădăluţă-Năstăseanu. Geology of the Anina Area with Particular View on Jurassic (Reşiţa Zone, Banat). 1965. Summary of the Dissertation Thesis (Unpublished, in Romanian).

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Between these limestones there is a gradual passage. The basinal carbonate sequence is passing in the same manner both to the older Tămașa Marls (Upper Callovian-Lower Oxfordian) and to the younger Crivina Marls (Upper Berriasian pro parte-Valanginian). Only in the eastern part of the Reșița Zone the Valea Aninei Limestones transgressively and unconformably overlie older sedimentary formations as well as the metamorphic basement.

The thickne.s of basinal limestones di play longitudinal and, particularly, transversal variations. Thus in the central part of the Reșița Zone their thickness attains its maximum while in the marginal areas of this zone (eastern and western) it is decreasing.

The three basinal carbonate units may be easily recognized owing to the higher or lower frequency of certain limestone types and also of cherts.

Valea Aninei Limestones. This lithostratigraphic unit lithologically [6] is the most varied being made up of different pelagic and allodapic limestones [7] as well as transitional fabrics between the two main group of limestones and frequently bedded and nodular cherts (Pl. I, Fig. 2; Pl. II, III, IV, Fig. 1; Pl. VII—IX, X, Fig. 1—4).

Pelagic limestones having the highest share especially consist of micrite and biomicrite containing corresponding skeletal grains (calcitized radiolaria and sponge spicules and scarce thin-shelled bivalves, cadosinids, *Globochaete* and belemnite guards), micrite nodules (intraclasts) and pelletoid particles, and subordinately nodular limestones (intramicrudites or nodular packstones) which also include certain fraction of some of the above mentioned allochems.

Allodapic limestones representing almost half of the bulk of the Valea Aninei Limestones are composed of skeletal, intraclast, pelletoid and pelletal grainstones-packstones whose allochems consist of more or less sorted and micritized crinoids, pelecypods, benthonic foraminifera, algal crust and structure grains, shallow water and basinal intraclasts (s.l. F o l k, 1962), pelletoids and pellets. These allochems and also the pelagic fraction entail various mixtures thus resulting different homogeneous to non-homogeneous carbonate fabrics. In the eastern part of the Reșița Zone the crinoid biosparites are very frequent. Generally the frequence of allochemical limestones decreases from east to west. Locally in these deposits early diagenetic dolomitized limestones are encountered (Vîsoki Hill, Caraș Valley) (Fig. 3).

All these basinal limestones usually occur as clear-cut strata lying in direct contact or being separated by a shale pellicle and/or by banded cherts. Frequently the contact between two successive strata is of stylolite nature; such contacts are more frequently encountered among

[6] In this paper is essentially used the F o l k's (1962) and D u n h a m's (1962) carbonate rock classification terminology.

[7] Allodapic limestones (M e i s c h n e r, 1964) here designate all the limestones resulting from carbonate sediments formed in shallow water, and then transported and resedimented in basinal areas.

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allodapic limestones or between these limestones and the pelagic micrites. Allodapic limestones often display parallel and oblique laminae and occasionally graded bedding.

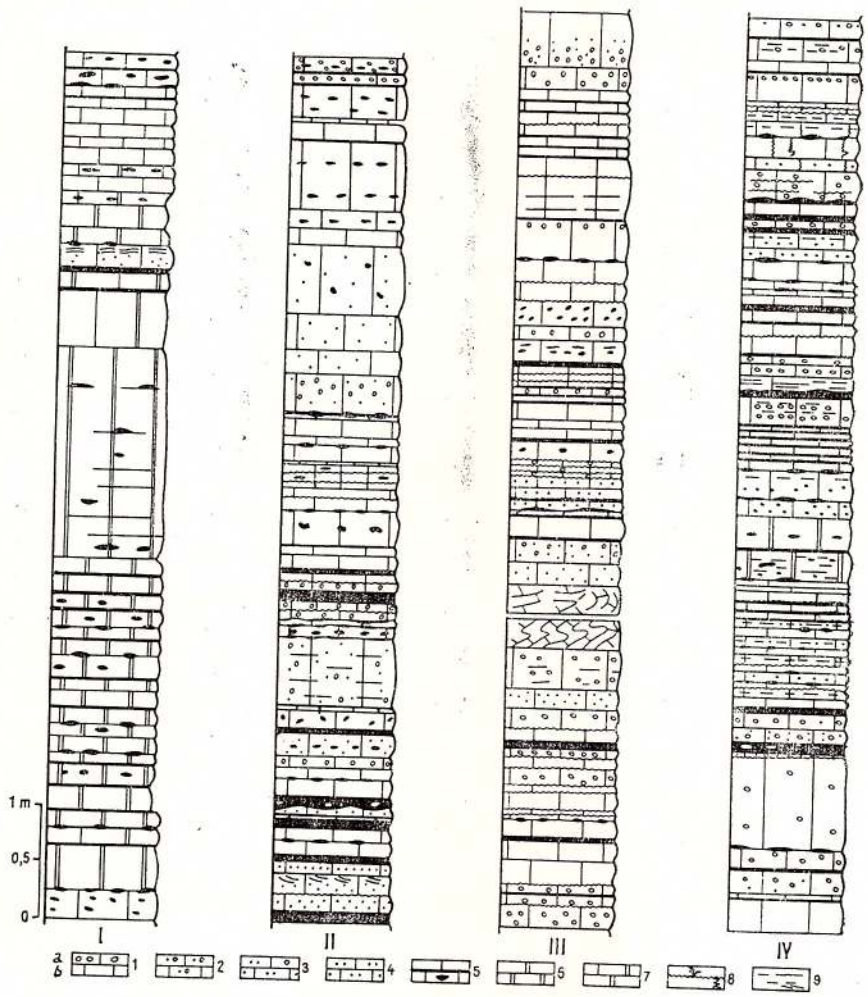


Fig. 3. — Partial sequences in the Valea Aninei Limestones from the Visoki Hill (lower (I) and middle (II) parts) and from north of Anina (basal (III) and middle (IV) parts).

1, pelagic micrite and biomicrite sometimes containing pelagic micrite nodules (intraclasts) (a) and nodular micrite (b); 2, allodapic packstones-grainstones including pelagic micrite intraclasts; 4, pelletal and pelletoid packstones-grainstones; 5, banded and nodular cherts; 6, dolomitized allodapic and pelagic limestones; 7, moderately dolomitized limestones; 8, stylolites; 9, parallel and oblique laminae.

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The thickness of strata shows a certain tendency depending on the limestone fabrics; in this way the highest frequency is marked by strata whose thickness ranges from 16 to 32 cm, and which are predominantly built up of micrites; as the thickness of beds decreases they are to an ever larger extent represented by micrites of pelagic origin. Beds whose

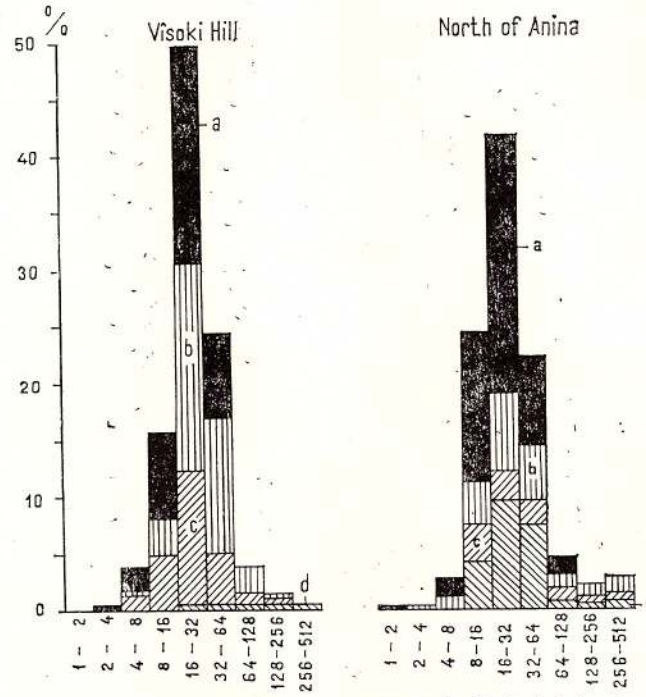


Fig. 4. — Thickness frequency of beds and their fabrics in the Valea Aninei Limestones.

a, pelagic sometimes nodular micrites and biomicrites; b, allodapic packstones-grainstones containing pelagic micrite intraclasts; c, allodapic pelletoid, pelletal and skeletal grainstones-packstones including pelagic micrite intraclasts; d, dolomitized allodapic and pelagic limestones.

thickness exceeds 32 cm are rather made up of allochthonous allochems (Fig. 4). In the Valea Aninei Limestones there are two or three massive limestone intercalations which are not reef limestones (Năstăseanu, 1964) but represent basinal heterogeneous mixtures of various shallow water and pelagic carbonate grains, wherein the latter often constitute a large fraction (Pl. I, Fig. 2; Pl. III, Fig. 3).

Valea Aninei Limestones often exhibit synsedimentary simple to intricate slump-folds and gravitational slidings; slump-fold aspects and the westward dip of slide planes in relation to bed surfaces point out

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almost always a corresponding dipping of the sea floor, inclusively in the innermost part of the Reșița Zone (Ciudanovița).

B r ă d e t L i m e s t o n e s. This lithostratigraphic unit also known under the name of „nodular limestones" is made up to a great extent of

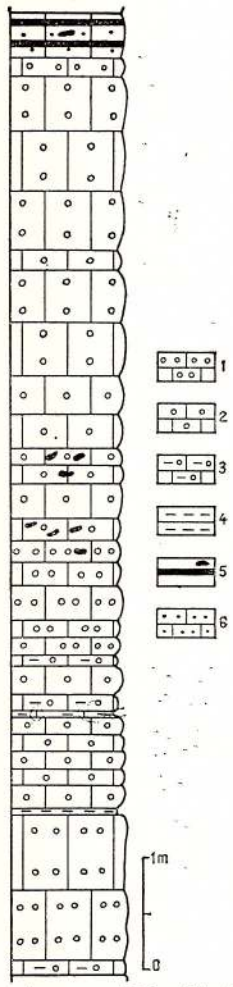


Fig. 5. — Partial sequence in the Brădet Limestones from the Miniș Valley (south-west of Anina).
1, nodular limestones; 2, nodular micrites and biomicrites; 3, marly nodular limestones; 4, marls; 5, banded and nodular cherts; 6, biopelsparites.

pelagic limestones and only scarcely of allodapic ones (Fig. 5). The pelagic limestones are represented prevalently by typical grey and greenish nodular limestones and subordinately by nodular micrites and biomicrites. These limestones are more or less argillaceous and the matrix of the nodules is occasionally weakly to moderately dolomitized. Besides micrite and biomicrite nodules, the allochems of these limestones consist of skeletal grains and rare pellets. The skeletal particles are represented by common to abundant *Saccocoma* and *Globochaete*, rare thin-shelled bivalves, calcitized radiolaria and accidentally some benthonic foraminifera. These

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limestones locally contain ammonite moulds and opercules (*Aptychus*), belemnite guards and brachiopods (*Pygope* and others).

Allodapic intercalations are built up of generally well sorted and in some cases poorly washed pelsparites and crinoid biopelsparites. These

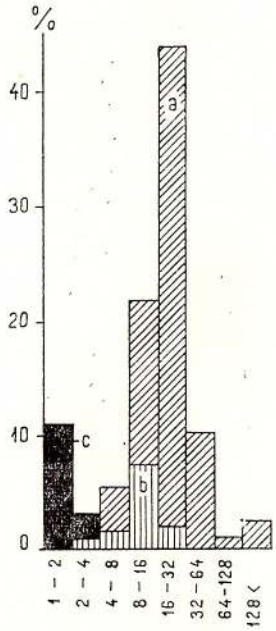


Fig. 6. — Thickness frequency of beds in the Brădet Limestones.
a, pure and marly nodular limestones; b, nodular micrites and biomicrites; c, marls.

limestones are relatively scarce or are missing, only in the eastern part of the Reşiţa Zone they are more frequently encountered.

The Brădet Limestones include thin marl and silty marl intercalations with frequent irregular micrite and biomicrite nodules. Cherts are much less frequent than in the Valea Aninei Limestones and occur as nodules and lenses and only rarely as thin beds (Fig. 5).

Although apparently the Brădet Limestones consist of a relatively small number of carbonate fabrics the transitional structures among them, the different content of the limestones in argillaceous matter and the presence of the marl intercalations amplifies their rock variety (Pl. IV, Fig. 2; Pl. V, VII; Pl. X, Fig. 5, 6; Pl. XI, Fig. 1—4).

Generally the Brădet Limestones occur as well bedded limestones whose average thickness is ranging from 16 to 32 cm (Fig. 6); the more conspicuous their bedding the smaller their content in argillaceous matter. The upper part of some limestone beds is richer in argillaceous matter and commonly contains micrite nodules; in other cases in the same upper part of a bed the carbonate fraction shows the tendency to form an incipient nodular structure. In such situations either the upper surface of beds is less distinct or the limestones display a transition towards the covering marl intercalations which usually show subdecimetric thicknesses (Pl. IV, Fig. 2).

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The Brădet Limestones present various synsedimentary gravitational deformations (slump-folds, sliding planes) which constantly indicate slidings towards the western part of the Reşiţa Zone (Pl. V, Fig. 2).

Marila Limestones. These limestones are the most homogeneous basinal carbonate deposits and are prevailingly formed of micrites of pelagic origin bearing calpionellids. Subordinately, especially in their lower part, there are some intercalations of nodular micrites and almost typical nodular limestones, occasionally weakly or moderately dolomitized; this lithostratigraphic unit also includes allodapic well sorted pelsparite and crinoid biopelsparite intercalations, sometimes intraclastic, and in their upper part — thin marls and shales. The Marila Limestones scarcely contain nodular and lens-like cherts (Pl. XI, Fig. 5, 6; Pl XII).

The petrographic monotony of the Marila Limestones is doubled by their stratonomic uniformity, the average thickness of the limestone beds being close to 20—40 cm. Locally these limestones display at some stratigraphic levels westward oriented synsedimentary slump-folds.

Constituents of Basinal Limestones

The major carbonate and non-carbonate constituents of rocks display both common and peculiar aspects in the three above described lithostratigraphic units. These constituents are microcrystalline calcite, allochems, sparry calcite, dolomite crystals, siliceous minerals and terrigenous particles.

Microcrystalline calcite. The very fine-grained calcite is the principal constituent of micrite and of certain allochems and their matrix, forming the basinal allochemical limestones.

In pelagic limestones such calcite occurs under the form of a non-homogeneous microcrystalline mosaic wherein the calcite crystals display a varied size-range in the limits corresponding to this end member of carbonate rocks (F o l k, 1959, 1962). Occasionally in micrites there are encountered irregular patches bearing no relation to the bedding in which calcite has been converted entirely into microspar ($< 10\mu$). The size variations of the microcrystalline calcite are more pronounced in biomicrites.

In micrite limestones containing argillaceous matter from the Brădet Limestones and from the upper part of Marila Limestones such calcite particles are more fine-grained and more uniform in size. These relationships seem to be general in the case of studied basinal micrite limestones, so one may admit that as the content of argillaceous matter decreases the microcrystalline calcite displays ever more varied shapes and sizes.

In contrast with calcite from pelagic micrite, the microcrystalline calcite from allochthonous allochems is more fine-grained and more uniform in size; this is the result of micritization processes of allochems in original shallow water environments.

Allochemical grains. In basinal limestones these grains include nodules, intraclasts, bioclasts, pelletoids, pellets and rarely ooliths.

Allochems here named „nodules" are always made up of micrite and biomicrite of pelagic origin and are the major constituent of the so-

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called „nodular limestones"; likewise these allochems are often encountered as subordinate constituents in other basinal limestones of pelagic and allodapic origin. In many cases the nodules may be distinguished from their micrite and biomicrite matrix through their more finely grain-sized microcrystalline calcite, the differences in density and orientation of skeletal particles and their slightly different colour due to a certain content of argillaceous matter, organic substance and ferromanganese impregnations. Their contours are either sharp, sometimes partially or entirely outlined by neomorphic microsparite pellicles, or diffuse; the former display varied shapes (irregular, angular to rounded), and are poorly sorted although in each case taken separately most of them are comprised among certain dimensional limits.

In some cases the nodules present an internal compound nodular structure indicating intraformational rework of such cohesive sediments, and other times they are surrounded by a more or less developed micrite coat of oncolithic type which is more dense and clouded and sometimes includes calcitized radiolaria and other planktonic calcite bioclasts; some nodules do also contain ammonite moulds and tabular bioclasts intersected by their edges, a fact proving their detrital origin. When the micrite matrix of the nodular limestones has been more or less dolomitized the nodules did undergo dolomitization only to a very insignificant extent. In allodapic limestones (especially in the Valea Aninei Limestones) nodular grains are associated in varied proportions with shallow water allochems from which they may be easily distinguished through their pelagic micrite fabric and planktonic microorganism content; at the contact with other allochems, nodular grains present varied mechanical deformations. In nodular limestones the internodular contacts often are stylolitic.

Nodular grains frequently include the same planktonic microfauna as their micrite and biomicrite matrix; this fact is more conspicuous in the Brădet Limestones and in the lower part of the Marila Limestones where both these grains and their matrix contain the same *Saccocoma* and, respectively, calpionellid assemblages (Pl. V, Fig. 1; Pl. VI, Fig. 1; Pl. VII, Fig. 3—6; Pl. VIII, Fig. 1; Pl. IX, Fig. 3,5; Pl. X, Fig. 6; Pl. XI, Fig. 1—4,6; Pl. XII, Fig. 1—5).

Basinal limestones contain both nectono-planktonic skeletal particles, some of them specific to certain stratigraphic intervals and allochthonous bioclasts initially formed in shallow marine waters.

The first group of skeletal grains includes more or less frequent calcitized radiolaria and sponge spicules, pelagic crinoids of *Saccocoma* type, *Globochaete*, calpionellids of the *Crassicollaria*, *Calpionella* and *Calpionellopsis* Zones, fine bivalves, cadosinids, ammonite moulds and opercules (*Aptychus*), brachiopods, belemnite guards, very rarely benthonic foraminifera and in addition other skeletal fragments of unknown origin. These bioclasts occur only in the pelagic limestones or fractions.

The second group forming allodapic limestones consists of crinoid monocrystalline calcite associated with polycrystalline molluscan fragments, benthonic foraminifera (textularids, miliolids), algal micrite

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crusts and structures, corals and rarely bryozoans. All these bioclasts are more or less micritized some of them to an extent that does not allow to recognize their particular aspects. Among skeletal grains from allodapic limestones only crinoid particles reach corresponding abundance to form peculiar fabrics such as the crinoid biosparites, which are very frequent in the Valea Aninei Limestones from the eastern part of the Reşiţa Zone. Occasionally the tabular molluscan bioclasts underwent some intrastratal fragmentations during the compaction of sediments.

Allochthonous allochems are also built up of micrite, pelsparite and biopelsparite intraclasts commonly poorly sorted and variably shaped. These intraclasts represent fragments of algal crusts and of very early lithified carbonate sediments and eventually „lump", „grapestone" and „micrite aggregate" type structure, originated in shallow marine waters.

Pelletoids and pellets are often encountered in different allodapic limestones and are the major constituent of pelmicrites, pelsparites, and biopelsparites scarcely occuring in the three lithostratigraphic units. These grains are of varied origin (see : B e a ls, 1965), however, most of them represent micritized allochthonous bioclasts. In the Brădet and Marila Limestones there are some pelletal limestones wherein these grains are of smaller size (30—40 μ) and better sorted.

S p a r r y c a l c i t e. Calcite from basinal limestones occurs under the form of microspar and coarse spar.

Microspar is encountered as patches in micrite matrix of some allochemical limestones, within some allochthonous grains and as liant in intergranular and intraskeletal spaces. In the last case the microspar is made up of relatively uniform-sized crystals or forms finely bladed and fibrous crusts covering allochems and filling intraskeletal voids. These crusts are limited outside by equant microspar or coarse spar and by the micrite matrix (Pl. VIII, Fig. 6 ; Pl. IX, Fig. 1—4).

The coarse-sparry calcite mainly occurs in intergranular spaces of better sorted allochemical limestones where it forms varied-sized crystals and overgrowths in lattice continuity, especially on the crinoid monocrystalline grains. Such calcite crystals either entirely occupy the intergranular spaces or are limited outside by equant or radial-fibrous microspar. In some cases one may observe two or three generations of overgrowths and their „zig-zag" boundaries (Pl. VIII ; Pl. IX, Fig. 1, 4).

D o l o m i t e s. In the lower and middle part of the Valea Aninei Limestones there are dolomite rocks which form either compact intervals of variable thickness, occasionally reaching 50—70 m (Vîsoki Hill), or subordinate intercalations. These rocks are built up of equant anhedral to euhedral dolomite mosaic containing some relics of initial sediments (finely disseminated clouded argillaceous and (or) organic matter, calcitized radiolaria and rarely benthonic foraminifera). Dolomite crystals (100—500 μ) do often have zoned rhombohedral shapes. In some strata the distribution of relics is variable and imitates the arrangement of parallel and oblique laminae of original carbonate sediments.

Intercrystalline spaces are filled with calcite as shown by Alizarin Red S staining. In dolomite rocks one may observe some joints where

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intersected dolomite crystals are regenerated; regenerated parts of these crystals are lacking relics and are bordered by calcite filling the joints.

Dolomite rocks show the same bedding as limestones and contain banded and rarely nodular cherts. Sometimes the dolomite intervals have parallel limits in relation to the bedding of the limestones, but in other cases these limits seem to intersect some limestone beds. In dolomite intervals slump-folds or other synsedimentary gravitational deformations have not been observed (Pl. IV, Fig. 1).

In some nodular limestones from the Brădet and Marila (lower part) Limestones the matrix usually contains up to 30 per cent of rhombohedral fine crystalline dolomite. Accidentally such crystals do also occur in the nodular grains. In some cases dolomite crystal spaces are occuppied by „dedolomite microspar"; likewise, dolomite rhombohedra are very seldom replaced by microcrystalline calcedony (Pl. XI, Fig. 2,3; Pl. XII, Fig. 3—5).

Cherts. Siliceous rocks occur as bedded, lens-like and nodular cherts.

Bedded cherts appear as thin bands between limestone beds and within some micrite ones. Lenticular and nodular cherts display a similar distribution; nevertheless these cherts seem to be chaotically distributed in micrites and more orderly distributed in allodapic limestones where they are situated on the planes of some parallel and oblique laminae (Fig. 3; Pl. II, Fig. 1; Pl. III, Fig. 1,2).

Cherts are composed of microscrystalline or fibrous quartz and probably cristobalite. All the cherts are diagenetic products and resulted from silicification of carbonate sediments (Pl. X, Fig. 2).

Bedded cherts are frequently encountered in slump-folds; likewise the presence of graded nodular cherts in some allodapic limestones from the Anina Valley (Valea Aninei Limestones) is to be noticed. In some biosparites the crinoid grains are occasionally substituted by silica.

Non-carbonate terrigenous particles. These particles are present in a small amount in the basinal limestones under the form of sand — to silt-sized grains and clayey minerals.

The terrigeneous grains are represented by quartz, feldspars (sometimes secondarily regenerated), muscovite and very seldom amphiboles (hornblende). They are met only at certain stratigraphic levels and are commonly associated with allochthonous carbonate grains (Pl. X, Fig. 1).

Generally, in the eastern part of the Reşiţa Zone the content of the limestones in terrigenous grains is higher and at more stratigraphic levels than in the western one; this rise in terrigenous content from west to east is accompanied by an increase in grain-size. Thus the crinoid biosparites from the lower part of the Valea Aninei Limestones occuring in the eastern part of the Reşiţa Zone contain up to 30 per cent of sand-sized non-carbonate terrigenous grains.

Clay minerals are widespread in almost all basinal rocks, especially in the marl intercalations and limestones from the Brădet and Marila (upper part) Limestones.

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Taking into account the manner under which the major constituents of the basinal limestones from the Reşiţa Zone occur, one may notice that within these limestones fine-grained calcite particles prevail, especially representing nannoplankton skeletons; these particles are relatively non-uniform sized in chemically purer limestones and more uniform-sized and finer grained in the limestones containing more than about 2% argillaceous matter. As a whole, the amount of microcrystalline calcite from the basinal carbonate units increases from the older limestones to the younger ones.

Allochthonous allochemical grains are generally poorly sorted and associated in different manner with pelagic microcrystalline calcite. As the sorting of these grains increases their liant is made up of coarser calcite and the micrite intraclasts show ever more pellet shapes; also as the amount of such allochthonous grains and their sorting increase, the allodapic limestone beds tend to become thinner and present more frequently parallel and oblique laminae.

Dolomite crystals from the Brădet and Marila Limestones, usually euhedral and finely crystalline are found more frequently in the micrite matrix of nodular grains, which has often a small content of argillaceous matter and silty grains.

In basinal limestones one may notice a positive relation between the frequency of cherts and the amount of calcitized radiolaria. The same positive relation is found between the frequency of allochthonous allochems and those of silt and sand terrigenous grains.

As regards fabric and stratonomical aspects in time one may state in the basinal carbonate sequence a decrease in the amount of allochthonous carbonate grains accompanied by an increase of their sorting degree, gradual uniformity of pelagic composition and bedding of the limestones, the diminution of chert frequency and calcitized radiolaria and a decrease of the number of synsedimentary gravitational deformations. At the boundary between the Valea Aninei and the Brădet Limestones a rather sudden diminution of the allodapic frequency takes place.

In space starting from the outer part (east) towards the inner part (west) of the Reşiţa Zone one may observe a lowering of the number of allodapic intercalations concomitantly with the increase of grain sorting diminution of bed thickness, also involving the massive intercalations from the Valea Aninei Limestones, and the decrease in frequency of the synsedimentary gravitational deformations (distal aspects).

Thus, the Upper Oxfordian-Berriasian basinal limestones generally display a slope facies in the outer and central parts of the Reşiţa Zone and a basin margin facies in the inner part of this zone (see also Wilson, 1970).

GENESIS OF BASINAL SEDIMENTS

Petrological aspects of basinal limestones from the Reşiţa Zone are indicative of the source of the sediments, their original mineralogical setting and the sedimentary environments.

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Main Types of Sediments and their Origin

Following up the share of various carbonate structures in basinal limestones it may be readily stated that their sediments were made up of carbonate ooze, nodular ooze and of allochthonous allochemical sediments (Fig. 7).

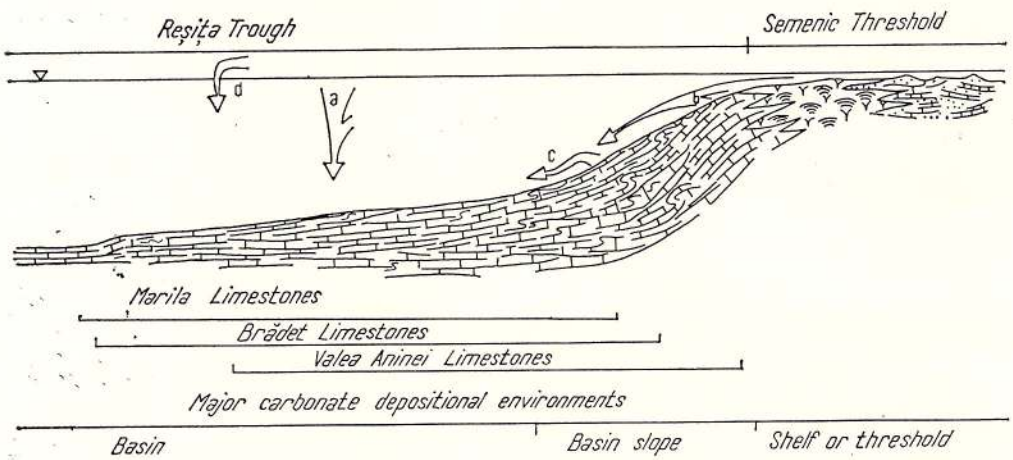


Fig. 7. — Deposition environments and the main sources of the basinal sediments from the Reşiţa Zone.

a, pelagic particles; b, shallow marine sediments redeposited as a result of submarine transport; c, reworking of the basinal sediments; d, detrital particles originated in some littoral and landmass areas and subsequently deposited in the basinal areas by the eolian transport.

Carbonate oozes. The formation of such oozes characterized the basinal sedimentation; they are usually proceeding from the accumulation of planktonic microorganism skeletons, espacially of coccoliths. Nowadays it is known that the so-called „pelagic limestones" do, to a great extent, consist of nannoplakton skeletons (see: Garrison, 1967; Garrison and Fischer, 1969; Flügel et al., 1968; Farinaci, 1969; Jenkyns, 1970 a, b, 1971, 1972; Bernoulli, 1971, 1972; Bernoulli and Jenkyns, 1970; and others). On account of this fact, one may assume that the basinal fine-grained carbonate sediments from the Reşiţa Zone chiefly consist of nannoplankton skeletons.

A certain amount of carbonate oozes probably originated in the adjacent shallow marine areas where they initially formed through generally known processes (Mathews, 1967; Stockman et al., 1967). It is presumable that at least a part of the dolomite crystals from the Brădet and Marila Limestones have detrital cores which previously formed in the same areas.

The non-uniform fine-grained character of carbonate oozes has been, to a great extent, determined by the initial sizes of skeletal parti-

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cles of pelagic origin to which have also contributed repeated resedimentation phenomena, through which these particles were partially desintegrated in their component parts, eventually their partial solution at the water/sediment interface and subsequent neomorphism processes.

Nodular sediments. As a whole the micrite nodules from the nodular limestones are considered to be of an early diagenetic nature. Undoubtedly, the observations on nodular limestones from the Reșița Zone pointed out that in the genesis of nodular structures an essential role was played by the early lithification of carbonate ooze. Howeve, the study of nodular limestones and of other types of limestones including micrite nodules supplies us with evidence that a large part of these nodules are intraclasts and to a lesser extent the products in situ of a specific diagenetic process affecting the carbonate oozes.

The intraclastic nature of nodular structure is illustrated by following aspects :

— compound structure of some nodules and their occurrence along with homogeneous micrite nodules in the same bed ;

— association of nodules with allochthonous allochems in allodapic limestone beds with current laminae and graded bedding ;

— a usually chaotic mixture of nodules with other types of sediments within the massive intercalations of the Valea Aninei Limestones where portions of some slump-folds are also recognized ;

— different density of bioclasts and the chaotically variation of colours of nodules in the same bed, aspects which also distinguish the nodules from their micrite matrix ; commonly the density of bioclats is less pronounced in nodules than in their matrix ;

— occurrence in some nodules of skeletal elements truncated by the contour of the nodules ;

— the frequently finer-grained character of the nodules in comparison with their micrite matrix and their local presence only at some levels or within the whole bed of micrite or biomicrite, practically pure from the chemical viewpoint, and with no signs of diagenetic segregation.

The contacts among the nodules or between these and their matrix, the shape and size of nodules are rather varied ; they are controlled both by the different coherence degree of sediments in course of consolidation, reworked as nodules, and by the subsequent lithification of nodular sediments. For this reason, the partial or whole transitional contacts between nodules and maxtrix or between two nodules, for instance, do not exclusively prove the formation in situ of nodules but also the possibility that the latter should represent reworked grains which have been diagenetically welded with the surrounding mass of sediments.

Consequently, the intrabasinal reworking of some pelagic more or less consolidated carbonate sediments led to the formation of nodular sediments with a fine-grained matrix. The micrite nodules probably arose through reworking of some nodules and nodular sediments already diagenetically formed or in course of formation and through the fragmentation of some crusts or beds of more or less lithified micrite sediments, with subsequent transport and deposition of the carbonate clasts. The

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last case is particularly evidenced by the existence of some intercalation of micrite breccias in the Brădet Limestones also containing some almost rounded micrite grains (Pl. VI, Fig. 2). Such grains at least should be at the origin of some nodular structures. Moreover in many cases the well bedded carbonate sediments, including the nodular ones, represent the effect of resedimentation.

Allochthonous shallow water sediments. The genesis of pelagic sediments was accompanied by the accumulation of some allochemical sediments previously formed in adjacent shallow marine environments. Their redeposition has been very frequent during the deposition of the sediments of Valea Aninei Limestones, and progressively more seldom, during the accumulation of those of Brădet and Marila Limestones.

The allochthonous sediments were to a large extent composed of marine shallow water bioclasts and compound grains (intraclasts s.l., F o l k, 1959, 1962), pellets and pelletoids and rarely coated grains (ooliths, oncoliths).

These allochemical sediments have been almost always mixed in various proportions with the pelagic ones through reworking and redeposition of the latter as matrix or intraclasts. The share of the diverse allochems was different determining the formation of sediments with frequently non-homogeneous fabrics. As a whole, however, during the Upper Oxfordian and the Kimmeridgian the supply in intraclastic (s.l.) and skeletal sediments (particularly crinoidal ones), prevailed especially in the eastern part of the Reşiţa Zone, whereas during the Tithonian and Berriasian in the basinal areas chiefly pelletal and pelletoid sediments have been redeposited.

Non-carbonate sediments. The accumulation of both pelagic and shallow water sediments was accompanied by the deposition of planktonic siliceous skeletons and of some terrigenous clayey, silty and sand-sized particles.

The siliceous particles accumulated as radiolaria skeletons and sponge spicules and they sometimes were an important constituent of carbonate oozes. According to the frequency of calcitized radiolaria and of cherts in basinal limestones, the climax in prolificity of the siliceous organisms (particularly of radiolaria) coincided with the formation of the Valea Aninei Limestones, subsequently their amount noticeably decreasing.

The terrigenous sediments have originated in some land and littoral areas from which they were transported in basinal environments, some of them together with shallow water carbonate sediments. The deposition of argillaceous particles has been probably more intense during the Lower Tithonian and since the Upper Berriasian.

Mineralogical Nature of Carbonate Sediments

The mineralogical composition of initial carbonate sediments may be generally estimated relying on major constituents of basinal limestones. Thus the fine-grained sediments of pelagic origin were composed of

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low-magnesian calcite which formed the skeletons of the nannoplankton and other planktonic microorganisms (probably calpionellids, *Saccocoma*, cadosinids and others). Belemnite guards, ammonite opercules and brachiopod shells, sporadically encountered in these limestones, are made up of the same mineral species.

The high-magnesian calcite constituted a reduced fraction in the pelagic sediments, being included in the tests of rare benthonic foraminifera met with especially in the Brădet Limestones; the aragonite with a similar content was present in the ammonite shells which have been, however, dissolved under submarine or intrastratal conditions very soon after their deposition.

The shallow marine carbonate sediments redeposited in the basinal area were probably formed to a great extent from aragonite and high-magnesian calcite.

The initial pelagic sediments comprised likewise numerous opaline skeletons (especially radiolaria), which in a certain time interval (Upper Oxfordian — Kimmeridigian) have been an important constituent of carbonate oozes.

Deposition and Redeposition of Sediments

The structures, textures and bedding of basinal limestones are suggesting that the accumulation of sediments represent a single effect or in many cases a combined one of several „mechanisms" such as : particle by particle deposition, resedimentation and submarine slidings of semiconsolidated sediments.

The planktonic skeletal particles have accumulated through particle by particle deposition; in the same way a part of the argillaceous and other terrigenous particles has been deposited as a result of eolian transport from land areas in marine ones. Owing to the slow and probably discontinuous deposition the carbonate and non-carbonate particles have accumulated as relatively thin strata of homogeneous composition.

In many cases, however, the particles deposited through the general processes of pelagic sedimentation were resedimented by several types of submarine „flows"; the redeposition has been particularly controlled by the submarine slope and its relief, by the subsidence regime (conjunction with certain fractures probably situated between trough and threshold) as well as by the physical properties of the sediments. Under the influence of these factors the non-consolidated fine-grained sediments were frequently drawn into submarine semifluid flows often evolving into turbidity currents; the more or less consolidated sediments have been submitted in many cases to gravitational slidings accompanied by various deformations. The redeposited non-consolidated sediments have been systematically retextured, and the submarine slidings gave rise to varied styles of simple to intricate, plastic to ruptural deformations; in the case of weekly consolidated sediments these deformations have sometimes caused the destruction of their initial fabrics and the

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formation of news ones, aspects already known in such limestones (B e r n o u l l i, 1971).

One of the more striking effects of the resedimentation has been the transport of sediments formed in shallow marine areas and their redeposition in basinal environments. Their transport in deeper marine areas has been chiefly carried out by turbidity currents which led to the formation of allodapic limestones frequently displaying gradded bedding and parallel and oblique laminae. The fabrics of these limestones also point out the fact that the transport of shallow water sediments also involved a various amount of non-consolidated and semi-consolidated sediments of pelagic origin from the higher parts of the basin slope towards the deeper ones. Subsequently these pelagic sediments have been redeposited as matrix of shallow water grains, as intraclasts with varied contours depending on the competence of reworked sediments, and as particles in the fine-grained intervals from the upper part of turbidite beds.

Accordingly, under control of the basin slope and of the movements to which it was submitted, the sediment bodies, exceeding a certain critical volume, have undergone slidings; these sediment slidings being at the origin of many resedimentation processes have often evolved into semifluid and turbulent flows determining the formation of fluxoturbidite and, respectively, turbidite limestones. The various ways of submarine reworking of sediments were probably interrelated, so that the display of one of them could have induced a „chain reaction" owing to which the sediments were affected by several types of submarine flows.

Some of these combined submarine flows were recognized in two or three massive intercalations from the Valea Aninei Limestones which are in fact basinal heterogeneous limestones, a chaotical mixture of pelagic and shallow water sediments displaying slump-folds, formed by wide-scale redeposition of such sediments.

The frequency of effects caused by gravitational slidings and resedimentation is differing from an area to another and also from one stratigraphic interval to another suggesting certain variations of the submarine relief. Generally, however, one may notice their decrease from east to west and also from the bottom to the top of the sequence; likewise the allodapic limestones have in the same sense the tendency to show a distal aspect. These facts reflect a corresponding attenuation of submarine slope and progressively a more distant redeposition of sediments from their source.

Depth Deposition of Sediments

The interpretation of depth deposition of basinal carbonate sediments is so far a most delicate probleme since this depth implies not only paleogeographic aspects and their tectonic control but also its influence on the sedimentation of previously deposited sediments.

Upper Oxfordian-Berriasian limestones from the Reşiţa Zone were commonly considered of bathyal origin taking into account the classical criteria, chiefly the presence of ammonite moulds and of planktonic

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microfauna (see : R ă i l e a n u et al., 1957). Actually, one may notice the pelagic nature of these limestones as well as the absence of features pointing to deposition in shallow marine environments ; in these limestones one may observe lack or scarceness of some aragonite shells (which could have been replaced by calcite) owing either to their solution on the sea floor or in intrastratal conditions under a certain depth, scarceness of benthonic organisms, especially as a result of lack of food, presence of chert and allodapic limestones and, in general, the lateral persistence of beds. All these elements evidence the bathyal depth of formation of basinal limestones.

However, these criteria have a relative character and hence the above mentioned aspects of limestones are likewise reflecting relative depths (H a l l a m, 1971). For this reason basinal limestones with similar facies as a whole were considered to have formed at depths of some hundred to about a thousand meters (B e r n o u l l i and J e n k y n s, 1970 ; B e r n o u l l i, 1971, 1972 ; W i l s o n, 1969) or even exceeding 4 000 meters ((G a r r i s o n, 1967 ; G a r r i s o n and F i s c h e r, 1969).

When appreciating the deposition depth of ancient carbonate sediments, in most cases data relating to the depth accumulation of recent pelagic carbonate oozes have been considered. When estimating the ionic activity products it has been stated that the superficial marine waters are saturated and supersaturated with respect to all $CaCO_3$ phases and are undersaturated in this respect with increase of their depth owing to increase of total pressure and CO_2 pressure and decrease of temperature (B e r n e r, 1965 ; P y t k o w i c z, 1965), facts subsequently verified in the waters of Central Pacific (P e t e r s o n, 1966 ; B e r g e r, 1967). According to H u d s o n (1967), if the solubility product of aragonite changes like that of calcite as a function of pressure and temperature, aragonite becomes thermodynamically soluble in ocean waters at a shallower depth than calcite, a fact later proved by *L i* et al. (1969), and assigned to the different amounts of CO_2 dissolved in these waters.

This differences in saturation of marine waters with respect to $CaCO_3$ as well as of solubility of aragonite and calcite would have accounted for the scarceness of aragonite organisms, preserved only as moulds, and the preservation of calcite skeletal grains (ammonite opercules, belemnite guards, brachiopod shells, crinoid pieces and others), which only very seldom present irregular traces of corrosion. If we take into account the fact that in warm seas, as those of the Jurassic, the aragonite and calcite solution rate might have been higher than nowadays, and that these minerals were submitted to an essential solution at comparatively shallower depths, we have grounds to admit depths of probably some hundred to about a thousand meters, for the accumulation of the Upper Oxfordian-Berriasian carbonate sediments. Besides, as noticed by H u d s o n (1967) and J e n k y n s (1970), it seems that traces of submarine solution of carbonates from ancient fossils and sediments are not necessarily indicative of great depth of sea waters.

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The mere fact of the increase of depth does not implicitely mean a corresponding solution of carbonates in sea water, the formation of carbonate particles and their accumulation showing a complex character. In near-surface sea waters (photic zone), where organic production of pelagic particles is taking place, the $CaCO_3$ is in a supersaturation state and, consequently of desequilibrium; this state has been assigned to the inhibition effect of Mg in solution on the precipitation of carbonates (P y t k o w i c z, 1965), to the complexing of Ca and CO_3 ions by the dissolved phosphorus and organic compounds as well to the inactivation of the $CaCO_3$ nuclei by the dissolved organic matter which is faster adsorbed from sea water onto their surface than the precipitation of $CaCO_3$ onto the same surface (C h a v e, 1965; C h a v e and S u e s s, 1967, 1970). The carbonate particles protected by their organic coatings may reach and accumulate at great depths; the adsorbed organic molecules prevent or inhibit the reactions between carbonate particles and the undersaturated deeper sea waters (S m i t h et al., 1968). The solution of such particles, probably displayed selectively (B e r g e r, 1967, 1968), may take place through the removal of their organic coating as a result of repeated resedimentation and bacterial activity.

Hence, if one admits the existence of such factors in the Upper Oxfordian-Berriasian environmental sedimentation within the Reşiţa Trough, it results that carbonate sediments of the Valea Aninei, Brădet and Marila Limestones have deposited about or under the aragonite compensation depth and over the calcite compensation depth, the two limits being probably situated at shallower depth than at present. Under these conditions, as also assumed by J e n k y n s (1970 b), if large aragonite and, occasionally, calcite skeletons have been dissolved, smaller particles with greater reaction surfaces have also to be dissolved; it is probable that metastable shallow water carbonate sediments, particularly the aragonite ones, redeposited in basinal environments, have undergone essential solutions intrastratally and on the sea bottom. Solution of these carbonate particles probably led to the increase of the concentration in Ca and CO_3 ions of interstitial fluids and subsequently—through their reprecipitation — to the early lithification of basinal sediments.

The high frequency of bedded and nodular cherts in the Valea Aninei Limestones does not indicate essential depths; the abundance of diagenetic cherts is in connection with the initial amount of biogenous silica which had resulted from prolific spreading of opaline organisms (especially radiolaria), an almost general phenomenon in the alpine peri-Mediterranean areas.

Paleogeographic and Paleotectonic Aspects

At the beginning of the Jurassic the South Carpathians area like the central Mediterranean zone (see: B e r n o u l l i, 1971, 1972) had been predominantly submitted to longitudinal block-faultings, probably as a result of spreading of the central part of the Tethys. In connection

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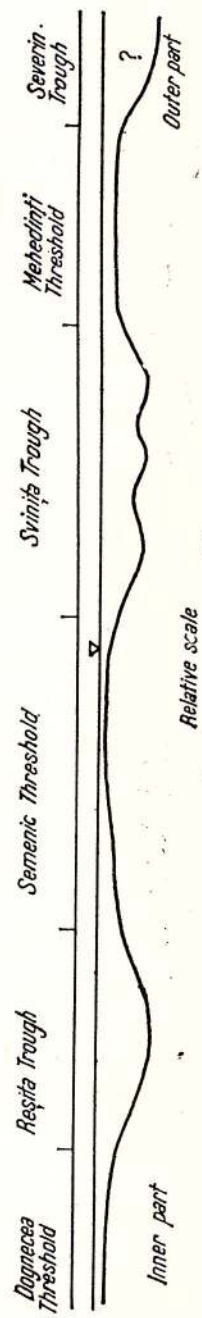


Fig. 8. — Major depositional environments from the South Carpathians during the Upper Tithonian-Berriasian.

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with these fractures there began a gradual separation of some elongated basins and swells under the shape of troughs (Reşiţa, Sviniţa, Severin, possibly Coşuştea) and thresholds (Dognecea, Semenic, Mehedinţi) (C o d a r c e a and P o p , 1970 ; P o p , 1973) which have afterwards variously persisted up to the end of the Lower Cretaceous (P o p , 1967) (Fig. 8)[8].

A long time, the Reşiţa Zone was interpreted as a „sedimentation zone", notion also implying basinal marine conditions relying on the presence of ammonite-bearing formations (Middle Jurassic-Neocomian) classicaly considered of bathyal origin. The areal significance of this zone was evidenced later by C o d a r c e a (1940, 1964) who pointed out that is was bordered by thresholds. Subsequent stratigraphic investigations have revealed lateral facies variations of the Mesozoic sedimentary formations, thus distinguishing central deeper water facies and marginal shallow water (locally littoral) ones (B o l d u r and B o l d u r, 1962; N ă s t ă s e a n u and D i n c ă, 1962; N ă s t ă s e a n u, 1963, 1964; R ă i l e a n u et al., 1964).

According to this interpretation whose starting point has been the analysis of facies of Middle Jurassic formations (B o l d u r and B o l d u r, 1962), the maximum depth of the basin corresponded to the central part of the present-day Reşiţa Zone, where the sedimentary formations build up a continuous thick sequence in fine-grained facies; the borders of the basin were located in the marginal parts (eastern and western) of this zone, where the Mesozoic sequences frequently display neritic facies, smaller thicknesses and at certain levels stratigraphic gaps and a transgressive position (N ă s t ă s e a n u, 1964; R ă i l e a n u et

[8] These block-faulting structures, some of them so far under discussion (Coşuştea Trough) or with a uncertain paleogeographic position (Severin Trough), had each a specific geologic evolution and a certain morphology, which determined the formation of some sediments with corresponding facies and stratigraphic relationships, as well as with some common aspects. Thus the troughs and thresholds (considered under the form of those from Fig. 8) progressively separated ones in relation to others and within each of them certain morphological variations did exist which subsequently have gradually attenuated; these aspects are reflected by lateral facies variations — more frequent in the Lower and Middle Jurassic formations and less frequent in those of the Upper Jurassic and Lower Cretaceous. Nevertheless in the Sviniţa Trough, for instance, some intrabasinal swells were preserved which led to the formation of some condensed sequences and to frequent redeposition of sediments. During the first stage of the separation of troughs and thresholds (Lower and partially Middle Jurassic) the sedimentation within them displayed a terrigenous character — firstly lacustrine and then marine, and from the Middle Jurassic onwards a generalization of carbonate sedimentation with varied intensity had started. The facies of sedimentary formations are suggesting, as a whole, a progressive differentiated sinking of trough accompanied by a higher or lower deposition rate of sediments, phenomenon which has also episodically and partially affected the swells especially in the course of the Neocomian. Thus, for instance, the Mehedinţi Threshold which during the Middle and Upper Jurassic presented typical shallow marine environments of Bahamas carbonate platform type, displayed in the Neocomian a tendency to sinking, as shown by deposition of micrites with scarce calpionellids. Such sinking movements also gave rise to the Severin Trough wherein sediments of the Sinaia and Comarnic Beds have accumulated (P o p, 1973).

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al., 1964). A similar interpretation was given to the thickness variations, stratigraphic relations and to the facies of the Upper Oxfordian-Berriasian limestones.

Data regarding fabrics, bedding, thickness, facies variation, stratigraphic relations and the regional geological framework of Upper Oxfordian-Berriasian limestones grouped into the Valea Aninei, Brădet and Marila Limestones point out that their sediments accumulated in basinal environments of the Reșița Trough. The same data are also indicative of the fact that these sediments were restricted to the outer part of the Reșița Trough corresponding to the eastern basin margin and slope [9] (see also W i l s o n, 1970). The deposition of sediments in dipping basinal areas is firstly indicated by the slump-folds and sliding planes which in all encountered cases point to a slope with westwards dipping inclusively in the innermost outcrops of such limestones (Ciudanovița); the same submarine morphology is evidenced by the presence of allodapic limestones whose share in the basinal sequence is decreasing from east to west, as well as by the increase and afterwards decrease of thickness of the three lithostratigraphic units in the same direction. The more reduced thicknesses of the synchronous sequence from the western part of the Reșița Zone, previously considered as an element of the shallower facies, its pelagic more homogeneous composition and the slump-folds are indicative in fact of a greater depth than in central or outer parts of this zone.

Consequently, the Upper Oxfordian-Berriasian limestones generally illustrate the sedimentation in the outer part of the Reșița Trough, and those corresponding to its inner part are missing from the Reșița Zone; the latter have been probably entirely or partially removed through erosion subsequent to post-Neocomian diastrophisms or tectonically overlain by the Supragetic Unit presumably at the end of the Upper Cretaceous. On this account, the outline of the Reșița Trough is difficult to be rendered, as we can only admit that its outer part mainly represented a wide basin slope with certain variations of submarine relief which have influenced the sedimentation, redeposition as well as gravitational sliding of non-consolidated and semi-lithified sediments.

The Reșița Trough was bordered by the Dognecea and Semenic Thresholds (C o d a r c e a and P o p, 1970). Nowadays within the thresholds areas Mesozoic formations of certain ages (Middle Jurassic) occur only in a few very small, areas so that their paleogeographic aspects are mainly indirectly inferred from the facies analysis of basinal formations.

During the Lower Jurassic in the Reșița Trough there were lacustrine environments and subsequently basinal marine ones; at the beginning of the Upper Jurassic the depth of this trough increased, thus determining the formation of a typical basinal sequence successively corresponding to the Tămașa Marls, Valea Aninei, Brădet and Marila Limestones, the Crivina Marls and Hauterivian biomicrites (pro parte).

[9] Here there are accepted only two major carbonate depositional environments; basin and shelf (shallow water); the slope environments are included in the basinal ones.

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In the threshold areas shallow marine conditions were more widespread during the Oxfordian and the Hauterivian and more restricted during the Kimmeridgian-Valanginian—as a result of their partial sinking.

The Hauterivian biomicrites showing an essential reduction of basin depth, progressively grade upwards into algal boundstones and grainstones of Urgonian type (Barremian-Aptian), which are indicating the extension of shallow marine environments within the trough area. Thus the austroalpine diastrophism led to uniformization of shallow marine environments over the whole inner part (Supragetic Unit and Getic Realm) of the South Carpathians (C o d a r c e a and P o p, 1970).

In the basinal carbonate sequence the Valea Aninei Limestones with frequent allodapic intercalations and slump-folds illustrate an active subsidence regime connected with some longitudinal fractures, probably a marked depth difference between the trough and the adjacent Semenic Threshold, and a relatively varied and periodically regenerated submarine relief. The allochthonous carbonate material from these limestones originated in the shallow marine environments of thethreshold which presumably represented an open carbonate platform.

Crinoid grainstones, abundant in the eastern part of the Reșița Zone, previously interpreted as a neritic-littoral facies, represent a slope facies; these limestones with reworked pelagic sediments (chiefly as intraclasts) deposited in the upper parts of the basin slope but below the level at which pelagic carbonate oozes still accumulated.

Brădet Limestones do commonly occur in the „marly Ammonitico Rosso" facies of A u b o u i n (1964) and likewise evidence a basinal slope. A much lower number of allodapic intercalations as compared to the Valea Aninei Limestones is indicative of a diminishing of the morphological difference between trough and threshold and at the same time a removal of the source of shallow water carbonate sediments. This situation is probably due to the partial sinking of the Semenic Threshold. These nodular limestones are at all events evidencing greater depth than the condensed so-called „carbonate Ammonitico Rosso" which characterizes the intrabasinal swells or seamounts (A u b o u i n, 1964; J e n k y n s, 1971; B e r n o u l l i, 1971, 1972). In the South Carpathians such condensed nodular limestones are very well exposed in some areas from the Svinița Zone (inner part of the Danubian Realm).

Marila Limestones are generally occurring in the facies of the Majolica Limestone widely spread in the central and western Mediterranean provinces (see: B e r n o u l l i, 1971, 1972); lithologically they are more homogeneous than the older basinal limestones from the Reșița Zone, show less synsedimentary gravitational deformations, contain relatively scarce allodapic intercalations but still do present frequent evidences of intrabasinal redeposition. These limestones suggest more pronounced deepening movements and increased pelagic carbonate supply — probably rather due to the prolific development of nannoplankton during the Middle Tithonian-Berriasian than to the decrease of sediment deposition depth (N ö e l, 1965; H a l l a m, 1971).

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The sinking of the adjacent threshold areas led to attenuation of the morphological difference between basin and thresholds and probably to the extension of the pelagic sediment accumulation area.

During the late Berriasian and the Valanginian, the supply of pelagic carbonate sediments diminished concomitantly with a slightly increase of the share of argillaceous ones, which determined the formation of the Crivina Marls. This evolution of the basinal sedimentation took place without any important changes of the sediment deposition depth.

LITHIFICATION OF CARBONATE SEDIMENTS

The lithification of the Upper Oxfordian-Berriasian sediments from the Reşiţa Zone was especially achieved by neomorphism, cementation, silicification and dolomitization processes. These processes generally had a selective character, depending on the fabrics of sediments, and mainly operated at different burial depths of sediments under the water/sediment interface and sometimes in basinal submarine environments.

Lithification of Carbonate Oozes

At present, it is admitted that the transformation of carbonate oozes into monomineral (low-magnesian calcite) microcrystalline aggregates (micrites) is the result of several complex processes comprised in the „aggrading neomorphism" concept (F o l k, 1965). According to this concept, the lithification of the primary porous, wet, multimineralogic and highly reactive carbonate oozes, should take place by the solution of finer particles and the proeminences of the larger grains, the conversion of aragonite into calcite and the high-magnesian calcite recrystallization accompanied by the transfer of Mg^{2+} into interstitial fluids, the pressure-solution at grain contacts, the cementation of carbonate particles by $CaCO_3$ precipitation under the form of syntaxial overgrowths on their surfaces and the corresponding decrease in porosity and interstitial fluids, and eventually by the allochthonous influx of Ca^{2+} and CO_3^{2-} in these sediments (F o l k, 1965; B a t h u r s t, 1970).

The share of these processes in the lithification of carbonate oozes probably depended on many factors such as: mineralogical nature of carbonate particles, their shapes and sizes, sedimentation rate, burial depth of sediments and their content in argillaceous matter and organic substance, and others.

The Upper Oxfordian-Berriasian carbonate oozes of pelagic origin were to a great extent composed of low-magnesian calcite, their particles were relatively non-uniform sized; they accumulated under a low sedimentation rate and by frequent redeposition and some of them contained certain fractions of argillaceous matter, organic substance and opaline microfossil skeletons. For this reason, the aragonite conversion into calcite and high-magnesian calcite recrystallization were minor pro-

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cesses ; also aragonite solution from the autochthonous and allochthonous particles, under a low sedimentation rate, probably took place penecontemporaneously with the sediment accumulation due to the subsaturation state of the deep sea waters with respect to aragonite.

It results that the solution of finer supersoluble particles, pressure-solution, precipitation of calcite under the form of syntaxial overgrowth cement on the microfossil skeletal particles and eventually the calcite recrystallization constituted important factors for the lithification of basinal carbonate oozes of pelagic origin. Generally the achievement of such processes probably depended on burial depth and age of the sediments and their insoluble content, as a matter of fact these aspects being noticed in the lithification of some Upper Cretaceous and Cenozoic carbonate oozes from certain deep sea and oceanic areas (see : B e rg e r and von R a d, 1972 ; P a c k h a m and L i n g e n, 1973 ; M a t t e r, 1974). As a result, the compaction of the carbonate oozes (usually now imperceptible) under the influence of lithostatic pressure, even if it was considered as having a minor implication, could play a certain role in pressure-solution at the contact points of grains and their welding.

Syntaxial overgrowth carbonates resulted from the dissolving of finer particles and from the pressure-solution at the intergrain contacts (see also M a t t e r, 1974), from the solution of aragonite particles of shallow water origin and their conversion into calcite, through the formation of early diagenetic stylolites (B a t h u r s t, 1970), as well as from the replacement of carbonate sediments by silica.

Transformation of carbonate oozes into compact and rather uniform-sized monomineral aggregates has been recently explained by F o l k (1974) through Mg^{2+} adsorbtion from interstitial fluids on the surface of calcite particles which should prevent from their normal growing over the equilibrium sizes of 2—4 microns, representing the average length of the original particles ; this should be a chemically-induced limitation of the growing of calcite particles determined by a small amount of interstitial Mg^{2+} proceeding from sea water and fine high-magnesian calcite particles.

As it was shown in the previous chapters, the basinal micrites have their particles of smaller and more uniform sizes when they contain argillaceous matter tending to get the „minimicrite" fabric of F o l k (1974). These fabrics were previously explained by the fact that a content of argillaceous matter greater than about 2% in carbonate sediments should promote more intense compactions, at the same time inhibiting their early lithification by cementation (Z a n k l, 1969). More intense compaction and slower lithification of such sediments were also observed in the case of Cenozoic sediments from some deep sea areas (see : M a t t e r, 1974).

Basinal carbonate oozes from the Reşiţa Zone containing argillaceous matter do not seem to have had initially a higher content of high-magnesian calcite than those chemically purer, from which a greater amount of interstitial Mg^{2+} resulted. For this reason, the eventually greater amount of Mg^{2+} could be retained in the interstitial fluids pro-

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bably by the semipermeable membranes of clay particles which braked free moving of these fluids and their ions; Mg^{2+} concentrated in this way and adsorbed on the surface of calcite particles inhibited the formation of the overgrowth cement. The clay particles had probably a similar inhibiting role as they partially or integrally isolated the calcite particles from the interstitial fluids.

Although the lithification of basinal carbonate oozes was conditioned, among other factors, by the burial depth of sediments and their age, however, at least their local lithification in submarine environments or near water/sediment interface cannot be excluded. At present, several cases of submarine lithification of the recent carbonate oozes predominantly of pelagic origin are known (Milliman, 1966, 1969; Fischer and Garrison, 1967; Garrison and Fischer, 1969).

The frequent synsedimentary gravitational deformations of sediments and the presence of micrite intraclasts at numerous stratigraphic levels in the basinal carbonate sequence from the Reşiţa Zone prove the early lithification of carbonate oozes, in some cases possibly even in submarine conditions; almost unanimously, it is believed that the submarine lithification of such sediments was favoured by their long exposure on the sea bottom due to a very low sedimentation rate accompanied by solution and reprecipitation of carbonates. The greater frequency of the synsedimentary gravitational deformation in the Valea Aninei Limestones could be caused not only by a more marked and labile slope but also by an earlier lithification of the sediment as a result of more intense neomorphism of metastable carbonate particles of shallow water origin redeposited in the basinal area.

Genesis of Nodular Limestones

The origin of nodular limestones focused the attention of many petrologists and continues to be one of the most delicate problems of the carbonate rock petrology. Such limestones, known under the name of „Ammonitico Rosso", "Knollenkalke" or "calcaires noduleux" are frequent in the Tethyan Jurassic sequences and have many common features as to their fabrics, and also certain local petrographic aspects which have led to more or less different interpretations regarding their genesis[10].

[10] Cayeux (1935) considered some nodular limestones as an „old pudding breccia". Lucas (1955 a, b) admitted the formation of such limestones by the partial cementation of pelagic carbonate oozes having a high content of argillaceous matter, under the shape of nodules with clear or sometimes diffuse contour; the sediments surrounding these nodules underwent intense differential compaction and dissolution, leading to an important reduction of their thickness and to deformations of stylolite type of the nodules. The same author also admitted the possibility of nodule formation in submarine environments, where currents could have removed their matrix, and accidentally their subsequent reworking. Atanasov (1958) considered Callovian and Kimmeridgian nodular structures from the Balkans of early diagenetic nature being made up by the concentration of the carbonate matter around some bioclasts

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In their formation a main part should have been played by: early cementation and differential compaction of more or less argillaceous carbonate oozes (L u c a s, 1955 a, b), submarine solution known under the name of "subsolution" (H o l l m a n n, 1962, 1964), diagenetic segregation in a broad meaning (H a l l a m, 1967), subsolution accompanied by shrinkage-cracking (G a r r i s o n and F i s c h e r, 1969), „flow" of semi-consolidated sediments in some cases and burrows (B e r n o u l l i, 1972).

Consequently, almost in all cases one may notice a certain content of argillaceous matter in the nodular limestones and as result various effects of compaction of the initial sediments; under the influence of the lithostatic pressure the nodules and their matrix underwent plastic and pressure-solution deformations. Also, most of the authors admitted the formation of nodular structures by early diagenetic lithification of pelagic oozes, sometimes even in submarine conditions. The early lithification of these carbonate sediments do represent a condition for the formation of nodular structures.

In the nodular limestones from the Reşiţa Zone there are no obvious elements which can attest the nodule formation by differential solution and/or shrinkage-cracking of some carbonate crusts or semi-consolidated ooze strata or by burrows. In exchange, within marly intercalations as well as in limestones with higher content of argillaceous matter one may notice incipient nodular structures in which the nodules have diffuse contours and gradually pass to their matrix. These nodular structures seem to be formed in situ generally by diagenetic segregation; their

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under the influence of a certain argillaceous content; the compaction of their more argillaceou matrix was deduced from its moulding aspects around the nodules and from the fact that the ammonites from the matrix are flattened while those from nodules are undeformed. H o l l m a n n (1962, 1964) interpreted the nodular structures from the Ammonitico Rosso from the Southern Alps, which bear evidences of some intense submarine solution, as relics of some carbonate strata already formed and then distroyed through a process called „subsolution". According to A u b o u i n (1964) the nodular facies is of syngenetic origin and resulted from discontinuous and slow pelagic sedimentation, interrupted by periods of submarine dissolution; nodular limestones often stratigraphically condensed should characterize the deep swells and their slopes. H a l l a m (1967) assigned the nodule formation to an early diagenetic segregation; the same processes should also control the genesis of limestone-marl alternations from the Ammonitico Rosso deposits. Analyzing the origin of nodular limestones, especially the Adnet ones, G a r r i s o n and F i s c h e r (1969) admitted that they reflect the periodical deposition of carbonate oozes followed by solution stages in the same place, accompanied by lithification and shrinkage-cracking; by the combination of these factors, the carbonate ooze strata evolved to nodules, partly embedded within a red fossiliferous clay matrix. B e r n o u l l i (1972) has found that many nodules from the Ammonitico Rosso limestones were formed in some places by „diagenetic flow" of semiconsolidated sediments, and in other places the nodular aspects should have resulted from burrows; the nodular aspects were in many cases accentuated by postdepositional intrastratal solutions.

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consolidation took place probably similarly to the lithification of oozes containing argillaceous matter. Consequently, the nodule formation is the effect of neomorphic solution and differential cementation of the calcite particles, especially by syntaxial overgrowth on these particles. In some cases, the influence of initial random distribution of carbonate and argillaceous particles, as a result of sediment redeposition, on the nodule „nucleation" is not to be excluded. The nodule formation was favoured by long time exposure of these sediments on the sea floor as a result of very low rates of sediment accumulation.

Redeposition of the sediments, which through such an early diagenetic evolution contained more or less lithified nodules and an unconsolidated argillaceous matrix, probably led to the formation of nodular sediments in which the nodules were laid down as detrital grains. Consequently, such primary in situ formation of nodular structures and their frequent redeposition as intraclasts had probably a major role in the genesis of nodular limestones from the Reşiţa Zone. The fragmentation of semi-consolidated crusts or beds of carbonate oozes by submarine slidings and the reworking of their fragments had also subordinately concontributed to the formation of nodular limestones.

In many cases, the nodular structures were amplified by compaction and subsequent neomorphic lithification.

Lithification of Allochthonous Allochemical Sediments

Lithification of these sediments was also achieved by compaction, neomorphism and cementation.

Subsequently to their accumulation in the basinal area, the allochthonous carbonate sediments were more or less affected by submarine and intrastratal solutions (especially aragonite particles) and by compaction.

The sediment compaction had as important effects the welding of some allochthonous micrite allochems (mainly pellet and pelletoid grains) and mechanical deformations such as various intergrain contacts resulting from their different consistence; also the compaction produced deformations of muscovite lamella from crinoid biosparites and breakage of some tabular skeletal grains. These deformations were probably accompanied by pressure-solution and tended to the formation of some stable frameworks of the sediments, concomitantly with an increase in intensity of their neomorphism and cementation.

In these sediments, the micrite allochems and also their fine-grained matrix usually of pelagic origin, generally underwent neomorphic lithification specifical to the carbonate oozes, subsequently evolving to microspar or even to sparry calcite. Also of neomorphic nature are the radial-fibrous crusts and the conical crystals perpendicularly grown on some bioclasts, which developed by replacement of the surrounding fine-grained matrix and sometimes of some micrite allochems (see: F o l k, 1965) (Pl. VII, Fig. 2).

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Adjoining allochems generally with similar micrite fabrics are frequently met, but some of them have radial-fibrous crusts on their surface. Radial-fibrous crusts have neomorphically formed presumably only on allochems which displayed the same mineralogical nature. On the intraclasts made up of pelagic micrite such crusts have never been noticed; the low-magnesian calcite of these intraclasts prevented from the nucleation of carbonate crystals of the crusts which formed in another mineral phase (probably high-magnesian calcite) (Pl. VIII, Fig. 6; Pl. IX, Fig. 1—4).

Syntaxial neomorphic overgrowths on monocrystalline grains had an important role in the lithification of allochemical sediments since crinoidal grains often constituted a significant fraction of these sediments. These overgrowths were made up by replacement of the fine-grained carbonate matrix and partially of micrite allochems especially pelletal and pelletoid particles; there are cases when such grains were wholly replaced by the calcite overgrowths, being recognized only due to the distribution of organic and argillaceous inclusions previously contained by these grains. There are also cases when the syntaxial overgrowths partially replaced some micrite allochems after they underwent intergrain deformations. Often, the boundaries of syntaxial overgrowths are saw-toothed, and sometimes the distribution of insoluble inclusions within them points out two or three generations of calcite, and consequently the discontinuity of neomorphic overgrowth formation (see: E v a m y and S h e a r m a n, 1965, 1969). In a single case a crinoid grain was met presenting two generations of syntaxial overgrowths: a first generation with its external part micritized in shallow marine environments (see: B a t h u r s t, 1966; A l e x a n d e r s s o n, 1971) and a second one which arose after the redeposition of this grain in the basinal area.

In all cases, the relations between radial-fibrous crusts and neomorphic syntaxial overgrowths demonstrate that the former were made up before the latter.

The neomorphism also led to frequent replacement of the primary fabrics of bioclasts by more or less coarse sparry calcite.

The lithification of allochemical sediments, especially of the well sorted and washed crinoid ones, also took place by Ca^{2+} and CO_3^{2-} precipitation as cement from the interstitial fluids within intergrain and intraskeletal voids. Cementation first took place under the form of radial-fibrous crusts, more or less developed on the surface of some allochems, and then as sparry crystals whose sizes usually show a centripetal increase within intergrain spaces. Sparry calcite from cements seems to be clearer than those considered of neomorphic origin. The precipitation of radial-fibrous crusts had a selective character, their crystal nucleation occurring only on certain allochems, probably depending on their mineralogical nature.

The sparry calcite and radial-fibrous crusts were also formed within the voids resulted from the solution of some aragonite bioclasts and within the chambers of some bentonic foraminifera. In the latter case, it is possible that the radial-fibrous crusts formed, at least partially, in shallow marine waters where these skeletons initially originated.

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The cementation of these sediments took place frequently by precipitation of calcium carbonate as syntaxial overgrowths on the crinoid grains. In sediments with a greater content of such grains, the cementation by syntaxial overgrowth had a major role in their lithification.

In the early stage of diagenesis, the allochemical sediments underwent compactions and formed neomorphic and cement radial-fibrous crusts and at least a part of overgrowths on monocrystalline calcite grains. The radial-fibrous crusts and a part of the syntaxial overgrowth originated in this early diagenetic stage by $CaCO_3$ crystallization probably under the form of high-magnesian calcite. This has been presumably due to a favourable Mg/Ca ratio in the interstitial fluids (see: F o l k, 1974). The formation of the two morpholgical types of high-magnesian calcite took place probably after their incipient burial, and locally even in submarine environments.

In the subsequent diagenetic stages, the formation of neomorphic and cement syntaxial overgrowths in conditions of deeper burial of the basinal sediments did continue. Under these conditions probably arose a great part of microspar from the matrix and sparry calcite as orthospar and pseudospar. During the later diagenetic stages, the dominant mineral forming phase of calcium carbonate was probably the low-magnesian calcite.

Silicification of Sediments

This process had an important role as to the basinal carbonate sediment lithification from the Reșița Zone, especially of the Upper Oxfordian-Kimmeridgian ones.

The existent positive relation between the amount of calcitized radiolaria from limestones and the frequency of bedded and nodular cherts represents an incontestable argument that silica is of biogenous origin resulting from the solution of opaline skeletons of planktonic microfossils. Silica from interstitial fluids, having predominantly such an origin, has probably precipitated as cristobalite which replaced the carbonate particles and filled the intergrain voids of sediments (Pl. X, Fig. 2).

The cherts from the basinal carbonate sequence from the Reșița Zone are mainly made up of calcedonic and fine grained quartz and subordinately of cristobalite, thus they represent quartz-rich cherts, an end product of the reaction: biogenous opal → cristobalite → quartz chert (see also H e a t h and M o b e r l y, 1971; W i s e et al., 1972; W i s e and W e a v e r, 1974; B e r g e r and v o n R a d, 1972; and others). Dissolution of opaline microfossils and subsequent silicification of carbonate sediments took place probably in a very early stage of diagenesis; this fact results from the presence of bedded cherts in slump-folds and from graded redeposition of some nodular cherts. However, some cherts, especially the nodular ones, seem to have been also formed in the later diagenetic stages by silica remobilization.

Following up the relic fabrics of the silicified sediments, one may notice that the bedded cherts are more frequently formed through sili-

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cification of allochthonous carbonate sediments than of pelagic ones; it is probable that their metastable carbonate composition favoured the early and intense silicification of these sediments.

Silicification of carbonate sediments released a relatively great amount of Ca^{2+} and CO_3^{2-} which subsequently participated in their early lithification.

Dolomitization of Sediments

In the basinal carbonate sequence from the Reşiţa Zone, dolomites occur as tiny crystals in the nodule matrix of the Brădet and Marila (lower part) Limestones, and as medium to coarse crystals in the Valea Aninei Limestones.

The genesis of fine dolomite crystals from the pelagic limestones and Cenozoic to Recent deep sea carbonate sediments was assigned to different causes (see: Berger and von Rad, 1972; Packam and Lingen, 1973). In the Brădet and Marila Limestones such crystals occur in the matrix of nodules which has almost always a certain content of argillaceous matter and silty detrital quartz; this assemblage suggests that at least part of the dolomite crystals was formed by regeneration in basinal sediments of some fine grained detrital dolomite initially originated in inter- and supratidal environments. This fact does not exclude the possibility of in situ forming of a certain part of these crystals (see also Thompson et al., 1968; Scholle, 1971).

The coarser crystalline dolomites from the Valea Aninei Limestones are of diagenetic nature, and resulted from the dolomitization of basinal carbonate sediments subsequently to an important stage of their silicification under the form of bands.

These dolomites were formed by local dolomitization of different basinal carbonate sediments through processes of „seepage refluxion" type (see: Deffeyes et al., 1965).

CONCLUSIONS

The typical basinal carbonate sequence from the Reşiţa Zone is successively represented by the Valea Aninei, Brădet and Marila Limestones (Upper Oxfordian-Berriasian).

The Valea Aninei Limestones are made up of different types of pelagic limestones (micrite, biomicrite and sometimes nodular ones), allodapic limestones (packstones-grainstones), and of limestones with mixed fabrics, banded and nodular diagenetic cherts and rarely dolomites. The Brădet Limestones do predominantly consist of *Saccocoma* nodular limestones, occasionally with a dolomitized matrix, which includes scarce allodapic intercalations. Marila Limestones are essentially built up of calpionellid micrites including rare allodapic intercalations and, in their lower part, nodular micrites.

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Beginning with the lower part of this basinal sequence and tracing it upwards, one may generally state an increase in the pelagic limestone share concomitantly with a decrease in frequency of allodapic limestones, of cherts and calcitized radiolaria and also of synsedimentary gravitational deformations; in the same time it is noted an increase of the sorting degree of allochthonous allochems from allodapic limestones. From east to west of the Reșița Zone one may observe a lowering of allodapic intercalations and an increase of their grain sorting, diminution of bed thickness (inclusively of massive intercalations from the Valea Aninei Limestones) and the decreasing frequency of synsedimentary gravitational deformations of carbonate sediments.

Basinal limestones resulted from the lithification of three main types of sediments: carbonate oozes, nodular oozes and allochthonous allochemical sediments. Carbonate oozes proceed from the accumulation of planktonic microorganism skeletons (especially of nannoplankton), and subordinately from the redeposition of some oozes previously formed in shallow marine areas. Nodular oozes, which led to the formation of a great part of nodular limestones, have arisen due to the reworking of early diagenetic previously formed nodules or those in course of formation, and of some crust or bed semi-consolidated ooze fragments resulted from submarine erosion and slidings. Allochthonous sediments formed through the redeposition within basinal areas of some shallow water sediments, constituted of bioclasts (especially the crinoid ones), compound grains (intraclasts, s.l.), pelletoids and pellets and scarcely coated grains (ooliths, oncoliths).

Pelagic oozes were to the greatest extent composed of low-magnesian calcite, and subordinately of opal contained in the siliceous microorganism skeletons while the allochthonous sediments were predominantly made up of aragonite and high-magnesian calcite.

The accumulation of sediments involves a peculiar effect, and in many cases a combined one of the particle by particle deposition, redeposition of such sediments and of those originated in shallow marine areas, as well as of submarine slidings of basinal sediments. Redeposition of sediments which is especially determined by the exceeding of a certain critical volume of sediment bodies under the influence of tectonic movements, of submarine relief and of a certain sedimentary rate took place through submarine slidings, semifluid flows and turbidity currents. These reciprocally connected main transport modes led to the gravitational deformations of sediments, their retexturation, and to the formation of some turbidite and fluxoturbidite limestones.

Carbonate sediments accumulated under the aragonite compensation depth and over that of calcite probably situated at shallower depth than today, and at the depth of several hundred to about a thousand meters. In these conditions aragonite particles accumulated in basin areas, underwent important dissolution before to be buried at various depths under the water/sediment interface. The high frequency of cherts (all of diagenetic origin) within the Valea Aninei Limestones does not imply a greater depth of sea waters, but is rather reflecting a prolific widespread of opaline

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organisms; likewise essentially micritic and homogeneous Marila Limestones suggest an increase of the supply of carbonate particles which had been sooner determined by a prolific development of nannoplankton than by the decrease of sea water depth.

Petrological data are suggesting that sediments of the Upper Oxfordian-Berriasian basinal sequence from the Reşiţa Zone have accumulated under basinal conditions in the outer part of the Reşiţa Trough (especially in the basinal margin and slope areas). Sediments formed in the inner part of this trough have been subsequently eroded and/or tectonically overlain by the Supergetic Unit. The Reşiţa Trough was bordered by Semenic Thresholds outside and the Dognecea one inside that during the Tithonian and Berriasian had displayed a tendency to partial sinking, thus determining the rise of the accumulation area of pelagic sediments and the diminution of the supply of shallow water carbonate sediments.

The lithification of basinal sediments took place through compaction, neomorphism, cementation, silicification and, locally, dolomitization processes which had selectively operated (as a function of their structures) at various depths under water/sediment interface, and partially in submarine environments.

Carbonate oozes have been submitted to lithification by neomorphic processes within which dissolution of finer supersoluble particles, pressure-solution, precipitation of syntaxial calcitic overgrowths on skeletal particles, and eventually recrystallization of calcite constituted essential factors. Likewise the formation of micrites has been probably conditioned by Mg^{2+} adsorbtion on calcitic particles which limited their growth to the corresponding sizes of these rocks. The formation of overgrowths on calcitic particles from sediments including argillaceous matter has been more limited as a result of their isolation by surrounding clayey particles, and of Mg^{2+} adsorbtion on their surfaces; Mg^{2+} concentrated in interstitial fluids under the influence of the same clayey particles, which played the part of semi-impermeable membranes.

The differential neomorphism of such more or less homogeneous sediments determined the in situ formation of nodular structures under the conditions of a long-lasting exposure on sea bottom as a result of a very low sedimentation rate; these structures have been subsequently amplified owing to their compaction and neomorphic lithification.

The lithification of allochemical allochthonous sediments took place through compaction under whose influence their grains had undergone weldings and varied pressure-solution and mechanical deformations, neomorphic processes especially obvious under the form of thin radial-fibrous crusts, microspar and pseudospar, and cementation through equant and radial-fibrous microsparry calcite, sparry calcite (medium to coarse crystalline) as well as through syntaxial overgrowths.

Banded and nodular cherts did predominantly form during an early diagenesis phase through dissolution of opaline planktonic skeletons, and afterwards replacement of carbonate sediments; silica has probably precipitated as cristobalite which subsequently graded, to a great extent,

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into quartz. It would seem that lithification affected more intensely allochthonous carbonate sediments with a metastabile mineralogical composition.

The dolomitization of the matrix of nodular limestones occurred, at least partially, through regeneration of some detrital dolomite cores. The formation of dolomites from the Valea Aninei Limestones had probably taken place in a later diagenetic stage through local dolomitization of basinal sediments through „seepage refluxion" type processes.

In the early diagenetic stage, the lithification of basinal sediments occurred through compaction, neomorphism, silicification, partially cementation and dolomitization, whereas during a later diagenesis, cementation and locally dolomitization have played a more important role in their lithification.

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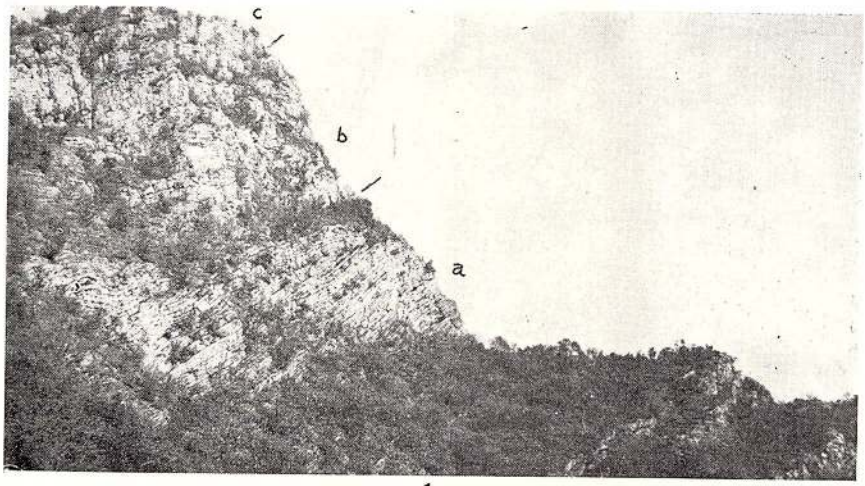
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PLATE I

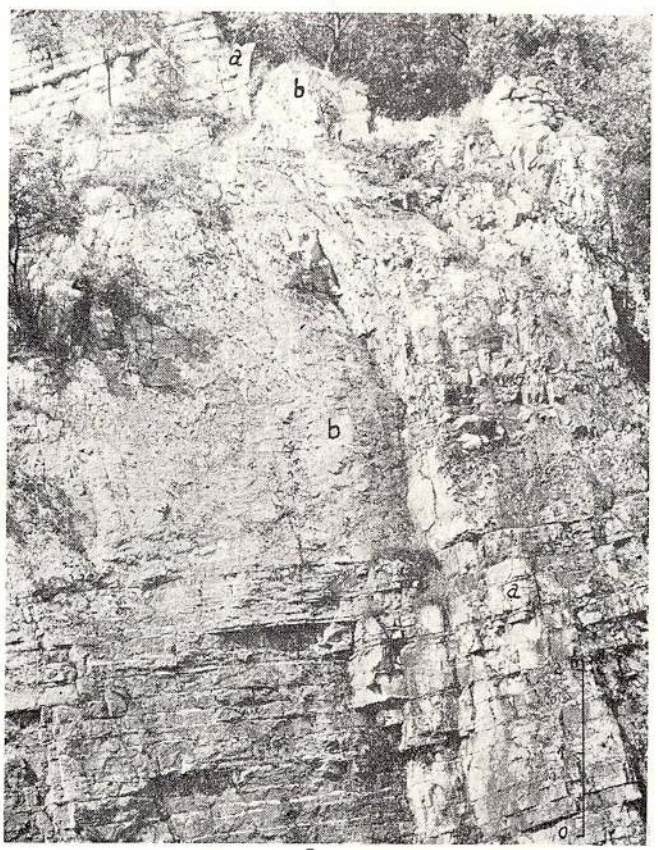
Fig. 1. — Basinal limestones from the Reşiţa Zone. a, Valea Aninei Limestones; b, Brădet Limestones; c, Marila Limestones. Northern slope of the Caraş Valley (about 3 km north-east of Caraşova).

Fig. 2. — Valea Aninei Limestones displaying relatively thin bedded pelagic and allodapic limestones and banded cherts (white ones) (a) and a thick intercalation of limestone (b) initially made up by the redeposition of a mixture of pelagic and shallow marine carbonate sediments. North of Anina (near the viaduct).

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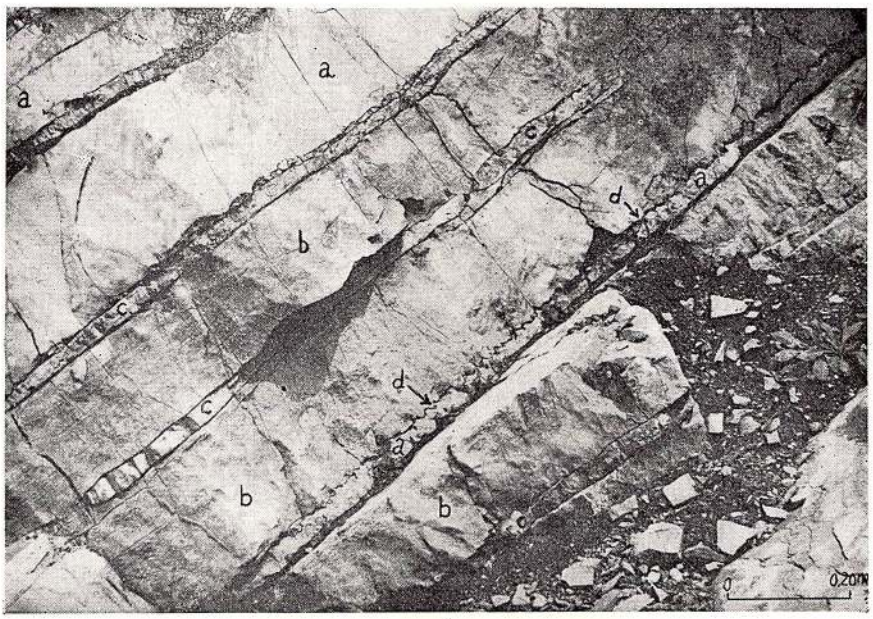
PLATE II

Fig. 1. — Detail of the Valea Aninei Limestones. a, pelagic micrites; b, laminated allodapic limestones; c, cherts; d, stylolites. Visoki Hill (on the Anina — Carașova road).

Fig. 2. — Two successive beds of coarser (a) and finer (b) grained allodapic limestones separated by a horizontal megastylolite surface (c). Valea Aninei Limestones. North of Anina (near the viaduct).

Fig. 3. — Crinoid biosparite layers showing multiple graded intervals; in their lower parts, these intervals are made up of sandy biosparites (cloudy parts). Lower part of the Valea Aninei Limestones east of Secul (on the Reșița — Văliug road).

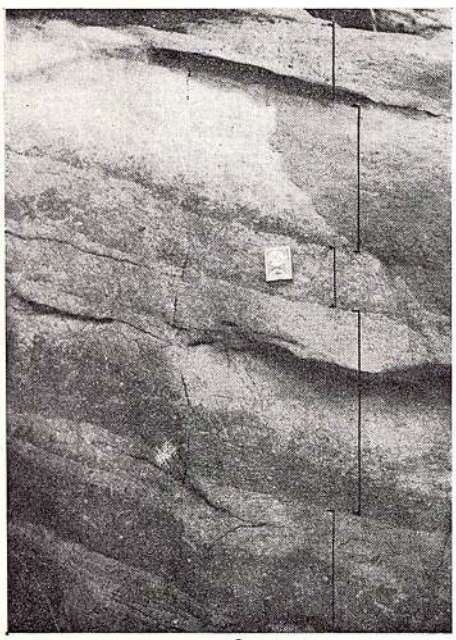
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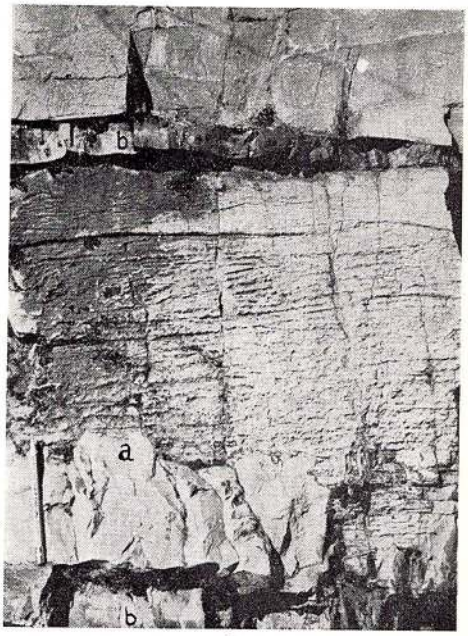
PLATE III

Fig. 1. — Parallel and oblique laminae in a fine grained allodapic limestone (a) and banded cherts (b). Middle part of the Valea Aninei Limestones. About 3 km north of Anina (on the railway).

Fig. 2. — Laminations in an allodapic limestone bed including conformably distributed nodular cherts (a); this limestone overlies a nodular limestone layer. Middle part of the Valea Aninei Limestones from the Visoki Hill (on the Anina — Caraşova road).

Fig. 3. — Thick basinal limestone intercalation pointing out some synsedimentary gravitational deformations (see arrows). Valea Aninei Limestones at about 4 km north of Anina.

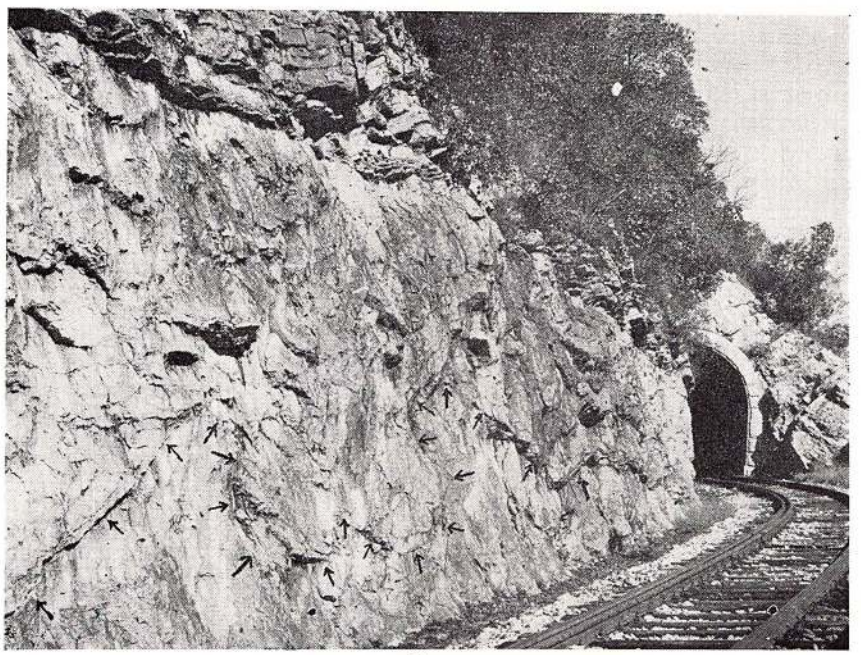
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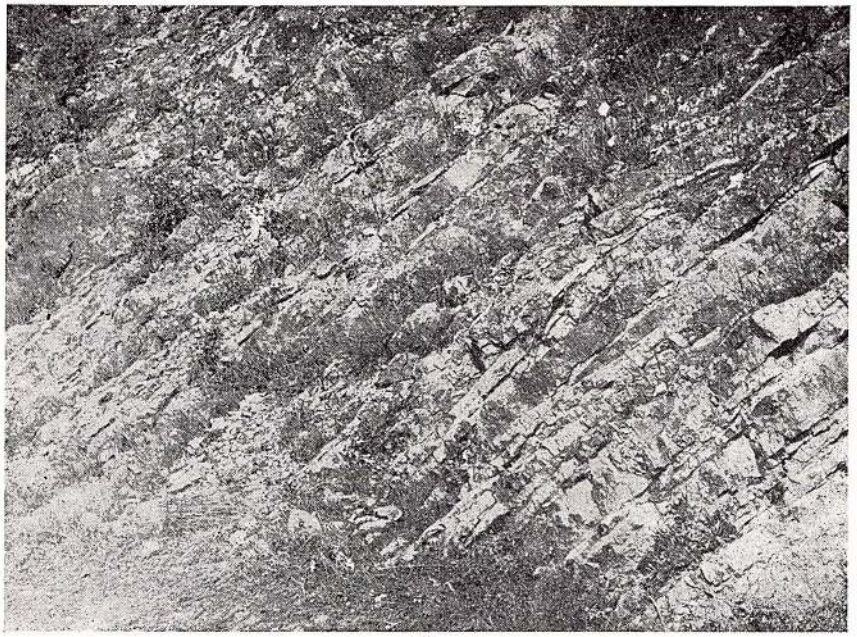
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PLATE IV

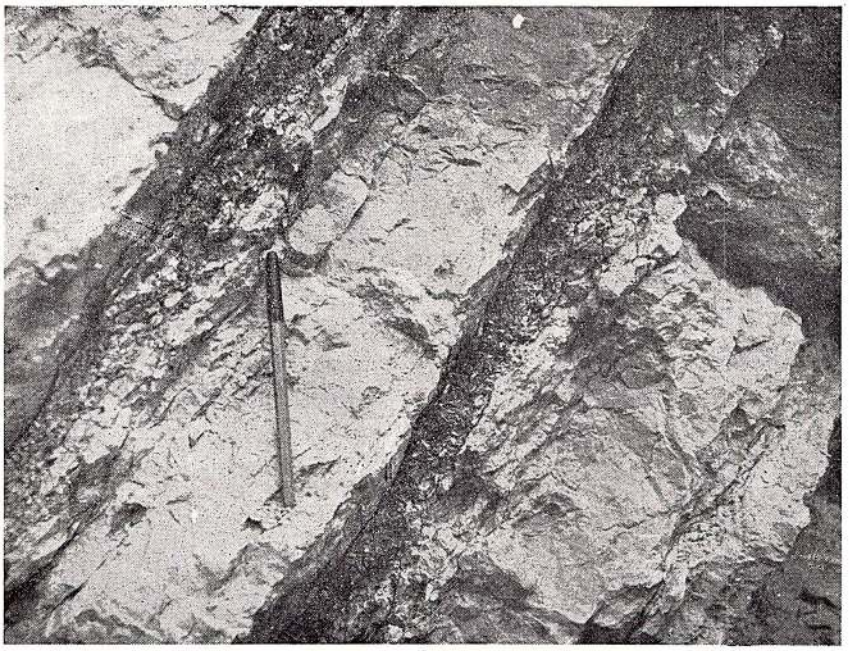
Fig. 1. — Dolomitized basinal limestones containing banded and rarely nodular cherts (whitish ones); the carbonate sediments were partially replaced by silica previously to their dolomitization. Lower part of the Valea Aninei Limestones from the Visoki Hill area (on the Anina — Caraşova road).

Fig. 2. — Nodular micrite beds upwards gradually passing into nodular marls. Brădet Limestones from the Miniş Valley (south-west of Anina).

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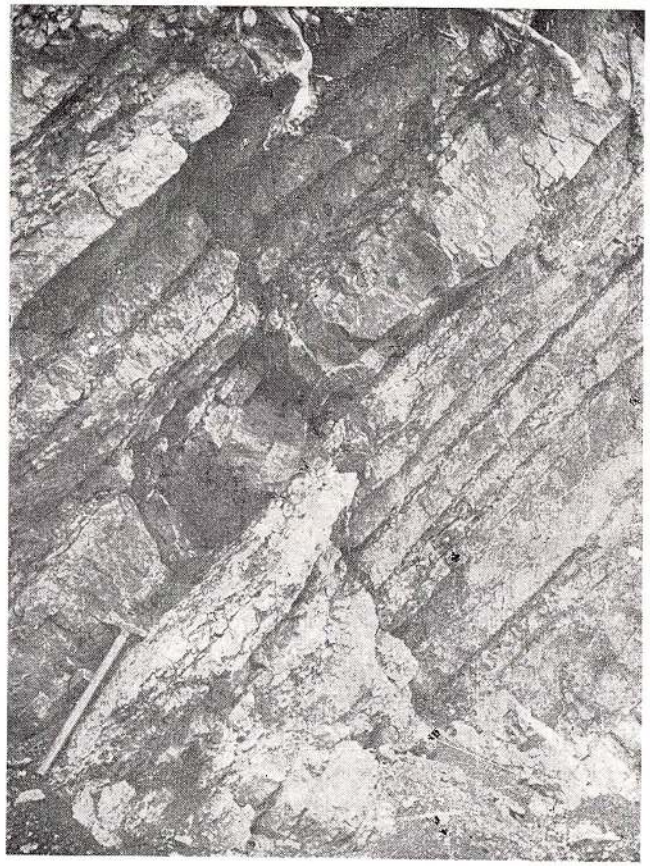
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PLATE V

Fig. 1. — Typically thin bedded marly nodular limestones. Brădet Limestones from the Miniș Valley (south-west of Anina).

Fig. 2. — Slump structure in the Brădet Limestones from the Miniș Valley (south-west of Anina).

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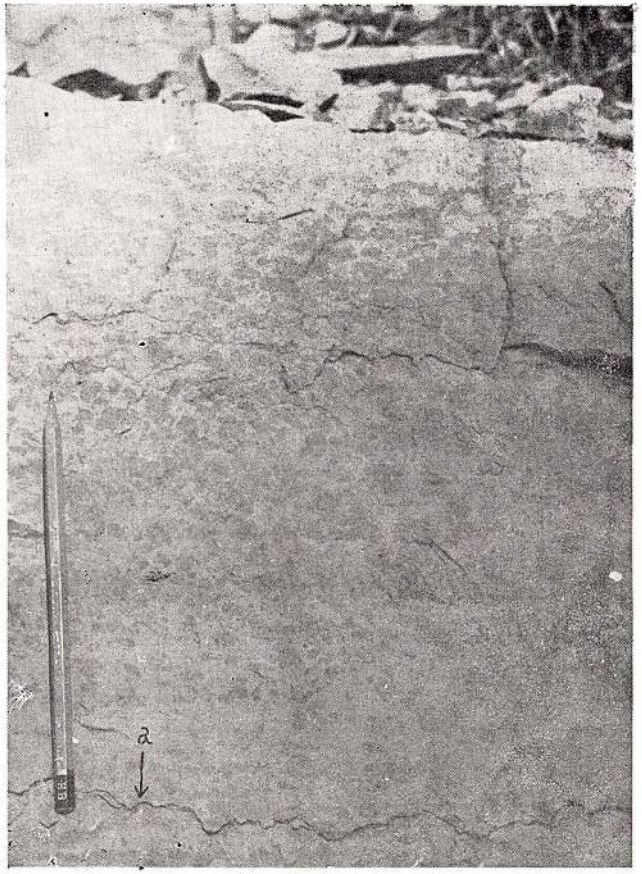
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PLATE VI

Fig. 1. — Nodular limestones wherein the nodules (clouded ones) are embedded in a more argillaceous and weakly to moderately dolomitized micrite matrix. Brădet Limestones from the Brădet plateau (Bibel quarry, west of Anina).

Fig. 2. — Intraformational carbonate breccia intercalation in the Brădet Limestones; the grains and their matrix are made up of *Saccocoma* biomicrite and micrite. Lindina Mare Valley (tributary of Nera River).

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PLATE VII

Fig. 1. — Micrite containing calcitized radiolaria and sponge spicules. Valea Aninei Limestones. North of Anina (near the viaduct).

Fig. 2. — Allodapic intraclastic packstones with a matrix mainly of pelagic origin. On the surface of elongated calcite bioclast perpendicularly grown neomorphic conical calcite crystals. Valea Aninei Limestones from the Visoki Hill area.

Fig. 3. — Allodapic intraclastic and skeletal packstone-grainstone including some rounded pelagic micrite intraclasts (a). Valea Aninei Limestones from the eastern border of the Reşiţa Zone (Comarnic).

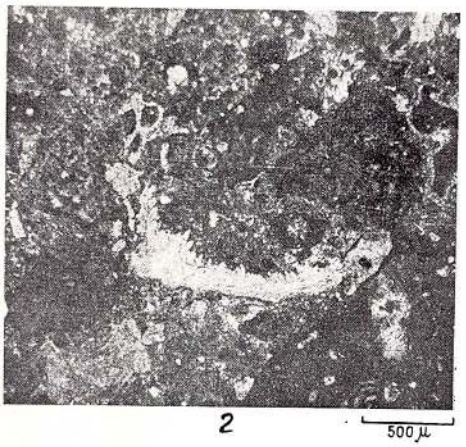
Fig. 4. — Allodapic intraclastic and skeletal grainstone bearing some irregularly edged pelagic micrite intraclasts (a); the liant of this limestone predominantly consists of neomorphic syntaxial overgrowths (on the crinoid grains) (b) and microspar (c). Valea Aninei Limestones. North of Anina (near the viaduct).

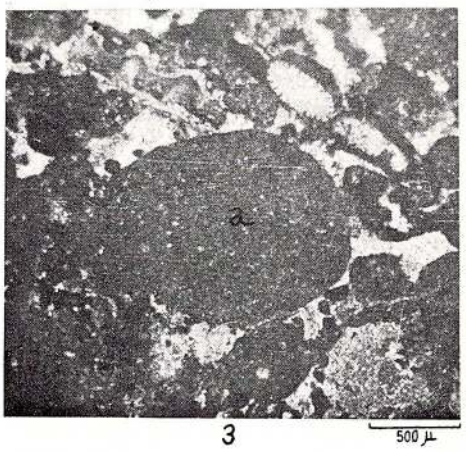
Fig. 5. — Allodapic intraclastic grainstone-packstone containing irregularly edged pelagic micrite intraclasts (a); between these intraclasts and other allochems there are differently shaped contacts formed under the influence of compaction. Valea Aninei Limestones. North of Anina (near the viaduct).

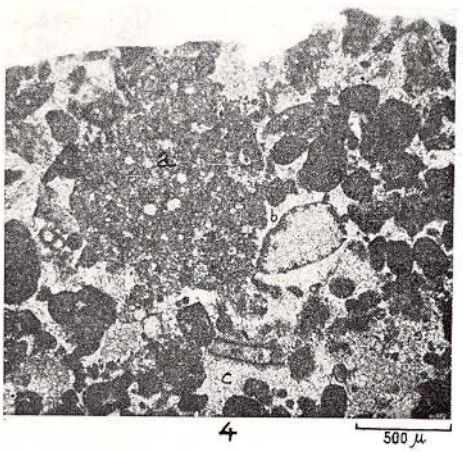
Fig. 6. — Intraclastic wackestone. Sometimes the intraclasts are made up of very fine-grained pelagic micrite and their borders are probably impregnated with ferromanganese oxides. Valea Aninei Limestones. Visoki Hill.

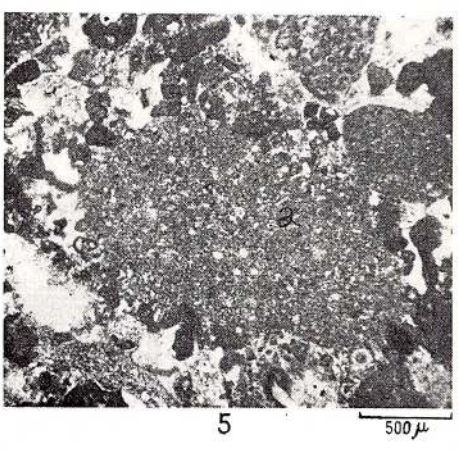
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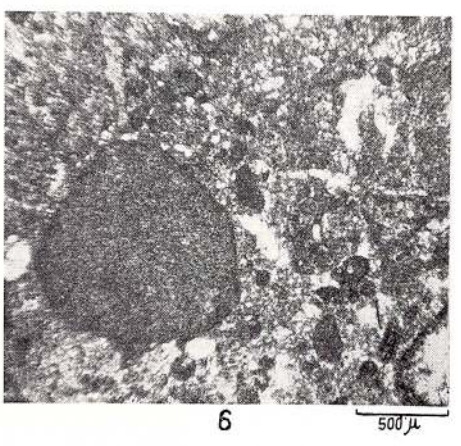












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PLATE VIII

Fig. 1. — Crinoid grainstone containing deformed pelagic micrite intraclasts (a) and intensely micritized crinoid grains (b) without syntaxial overgrowths; the grain liant is predominantly made up of neomorphic and partially cement syntaxial overgrowths. Valea Aninei Limestones from the easternmost part of the Reşiţa Zone (Comarnic).

Fig. 2. — Allodapic intraclastic and skeletal grainstone. The coarse crinoid grain displays discontinuous neomorphic syntaxial overgrowths probably under the form of three generations, marked by their insoluble inclusions. Also the pelletoid and micrite intraclastic grains are often welded similarly to those lithified in undisplaced shallow marine sediments. Valea Aninei Limestones from the Reşiţa-Văliug road (east of Secul).

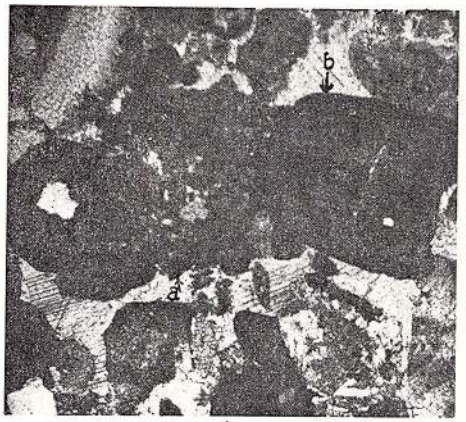
Fig. 3. — Allodapic intraclastic and skeletal grainstone. Syntaxial overgrowths on the crinoid grains formed by neomorphic replacement of some micrite allochems often show saw-toothed boundaries. Valea Aninei Limestones from the Comarnic area.

Fig. 4. — Pelletoid packstone-grainstone. Some crinoid grains (a) underwent corrosions and incrustations with ferromanganese oxides before to be redeposited in the original place of this sample. Valea Aninei Limestones. North of Anina (near the viaduct).

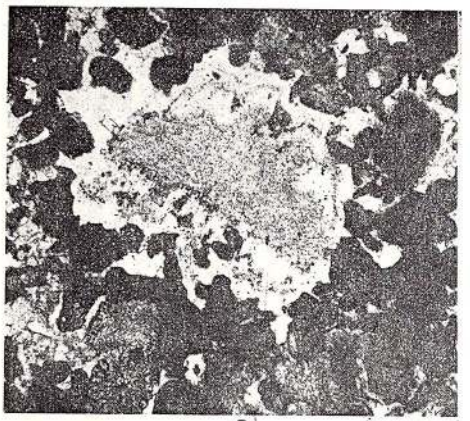
Fig. 5. — Allodapic intraclastic and skeletal grainstone. In this limestone the micrite allochems are often welded and the syntaxial overgrowths on the crinoid grains are very developed cementing many adjacent allochems and thus displaying a poikilotopic fabric. Valea Aninei Limestones. North of Anina (near the viaduct).

Fig. 6. — Allodapic intraclastic grainstone; the liant of their allochems is predominantly constituted of neomorphic syntaxial overgrowths on crinoid grains and some micrite grains present recrystallized thin radial-fibrous crusts. Valea Aninei Limestones from the Comarnic area.

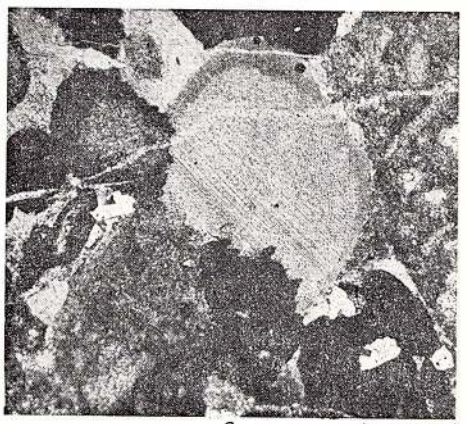
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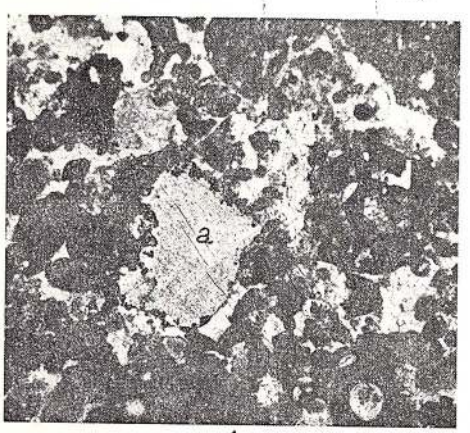
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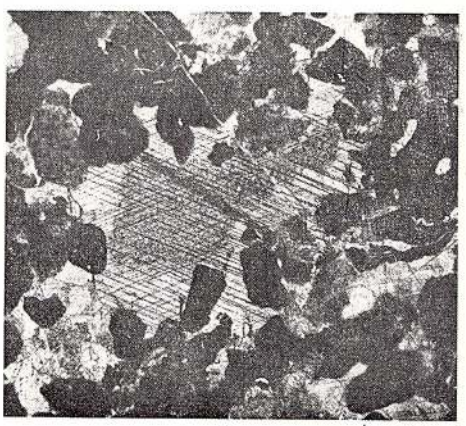
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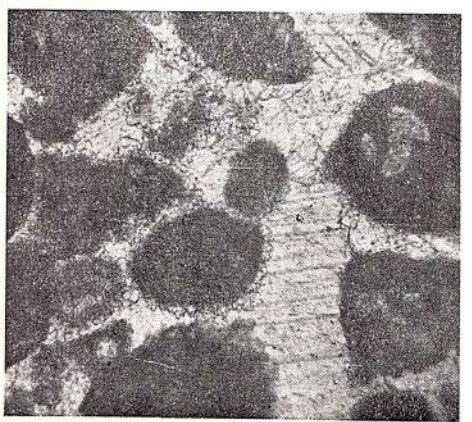
3 500 μ



4 500 μ



5 500 μ



6 100 μ

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PLATE IX

Fig. 1. — Allodapic intraclastic grainstone. The radial-fibrous crusts selectively formed on the certain micrite allochems, followed by the overgrowth neomorphism and cementation of the carbonate grains. Valea Aninei Limestones. East of Secul area.

Fig. 2. — Allodapic wackestone. Some micrite or micritized allochems of shallow marine origin present on their surface typically neomorphic radial-fibrous crusts. Valea Aninei Limestone. Visoki Hill area.

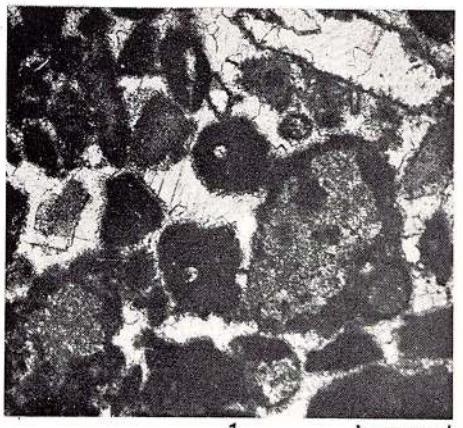
Fig. 3. — Allodapic intraclastic and skeletal packstone including deformed pelagic micrite intraclasts and some shallow water micrite allochems presenting thin radial-fibrous crusts on their surface. Valea Aninei Limestones. North of Anina.

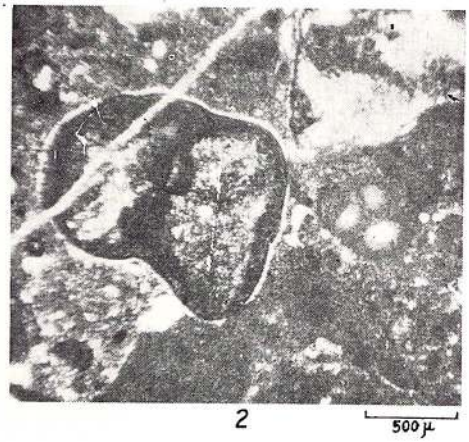
Fig. 4. — Algal boundstone fragment (a) containing internal micrite (b) and a crinoid grain (c), sedimented in their original shallow water environments; the internal sedimentation was followed by the neomorphic formation of a relatively thick radial-fibrous crust (d) and subsequently by the overgrowth of syntaxial calcite on the crinoid grains. This crust is comparatively more developed than those formed within basinal sediments. The boundstone fragment was encountered in a thick predominantly allodapic limestone intercalation from the Valea Aninei Limestones outcropping in the eastern border of the Reşiţa Zone (Comarnic).

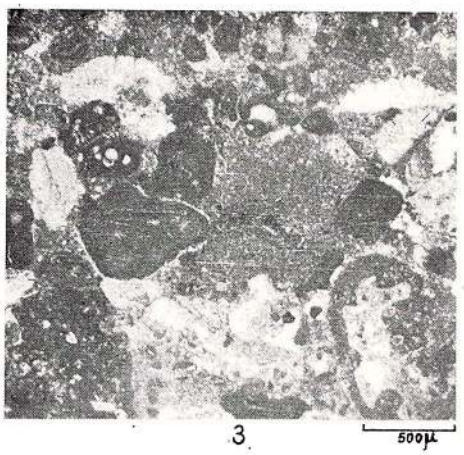
Fig. 5. — Pelletoid and skeletal (crinoid) grainstone or biopelsparite constituted of well-sorted shallow marine allochems and of some pelagic micrite intraclasts sometimes irregularly edged. Valea Aninei Limestones occurring at north of Anina (near the via duct).

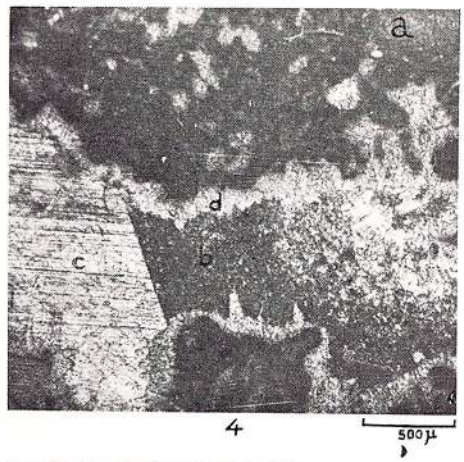
Fig. 6. — Well-sorted and finer-grained pelletoid and crinoid grainstone (biopelsparite) (distal facies) occurring as rare intercalations in the Valea Aninei Limestones from the western border of the Reşiţa Zone (Ciudanoviţa).

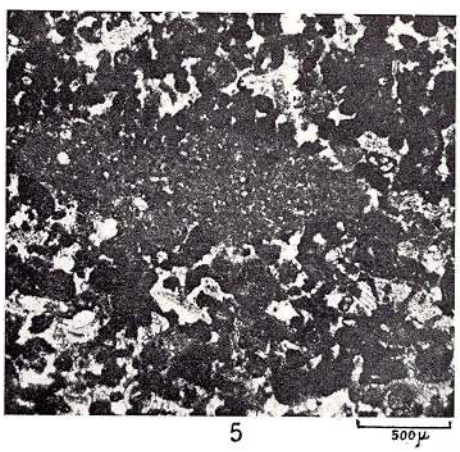
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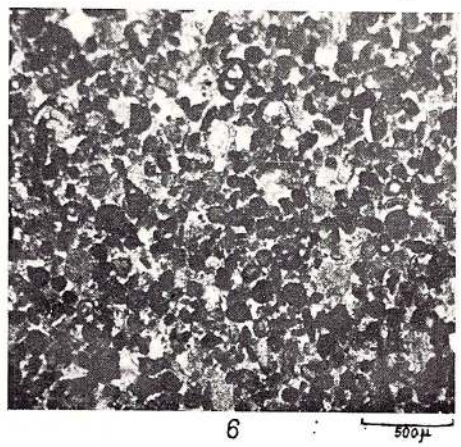












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PLATE X

Fig. 1. — Sandy grainstone made up of different shallow marine allochems and coated detrital quartz. Basal part of the Valea Aninei Limestones from the Reşiţa-Văliug road (east of Secul).

Fig. 2. — Partially silicified allodapic intraclastic packstone from the Valea Aninei Limestones. The silicified part (upper part of the picture) occurs as banded chert within the middle interval of this limestone layer. Visoki Hill area.

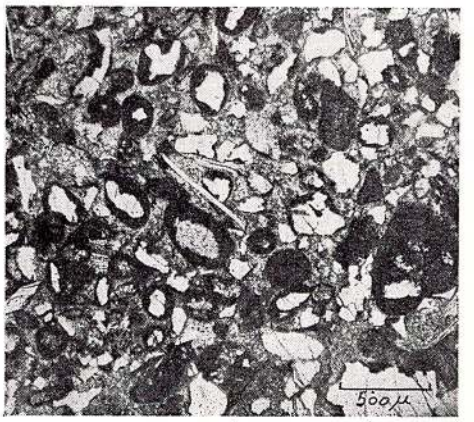
Fig. 3. — Medium to coarsely crystalline dolomite resulted from the dolomitization of some basinal carbonate sediments containing previously calcitized radiolaria (a). Valea Aninei Limestone occurring in the Visoki Hill area (on the Anina-Caraşova road).

Fig. 4. — Medium to coarsely crystalline dolomite. The dolomite crystals are rich in insoluble inclusions pre-existing in the basinal carbonate sediments subsequently diagenetically dolomitized. Intercrystalline voids were then filled by calcite. Valea Aninei Limestones from the Caraş Valley (about 3 km north-east of Caraşova).

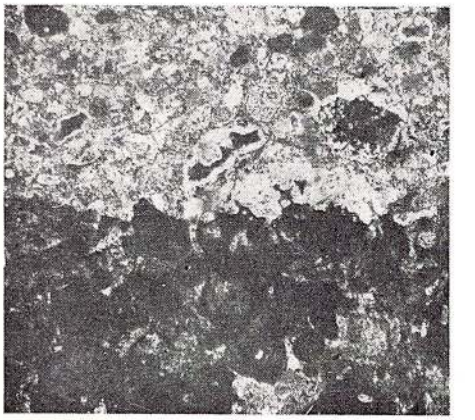
Fig. 5. — *Saccocoma* biomicrite in the Brădet Limestones from the Brădet plateau (Bibel quarry, west of Anina).

Fig. 6. — Nodular micrite bearing fine bivalves and some *Saccoma* pieces. The nodule (a) is made up of more fine-grained micrite than its matrix and only partially edged. It is probably of detrital origin. Brădet Limestones from the Miniş Valley (south-west of Anina).

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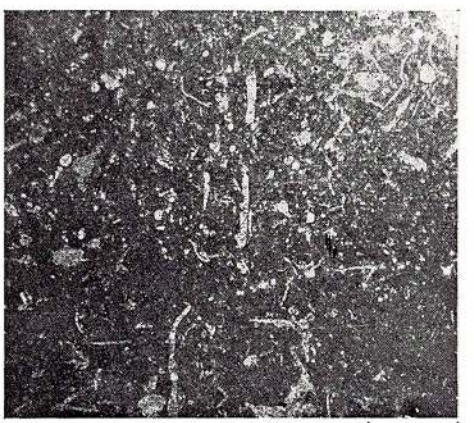
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PLATE XI

Fig. 1. — Nodular micrite containing calcitized radiolaria. Nodules are of detrital nature showing different fabrics in relation to their matrix. Brădet Limestones from the Bibel quarry (Brădet plateau).

Fig. 2. — Moderately dolomitized nodular micrite from the Brădet Limestones. Detritic micrite nodules sometimes containing *Saccocoma* are comprised in a partially dolomitized micrite matrix. East of Livadica (on the Danube).)

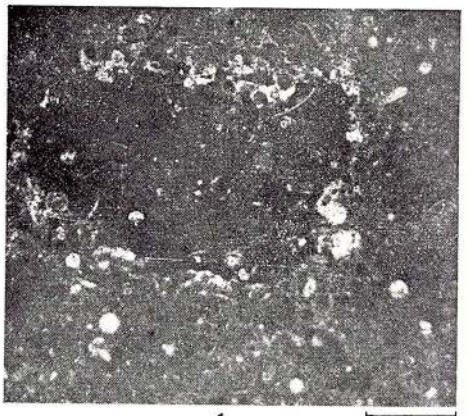
Fig. 3. — Partially dolomitized nodular micrite. The nodules are of detrital origin and display different fabrics among them and as against their moderately dolomitized micrite matrix. Brădet Limestones from the Miniş Valley (south-west of Anina).

Fig. 4. — Allodapic well-sorted pelletoid and skeletal (crinoid) grainstone (biopelsparite) forming a distinct bed in the Brădet Limestones from the Miniş Valley.

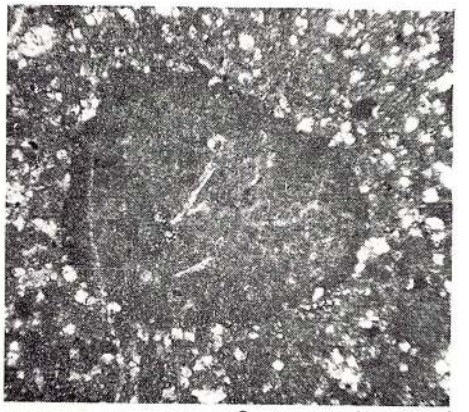
Fig. 5. — Relative homogeneous micrite containing calpionellids (here *Calpionella alpina*) from the *Calpionella alpina* subzone (late Upper Tithonian). Marila Limestones from the Mîndrişag Valley (tributary of Miniş River).

Fig. 6. — Nodular micrite from the lower part of the Marila Limestones including rare calcitized radiolaria, benthonic foraminifera and traces of burrows. In this case there is no connection between the detrital nodule and burrows. Marila Limestones from the Mîndrişag Valley.

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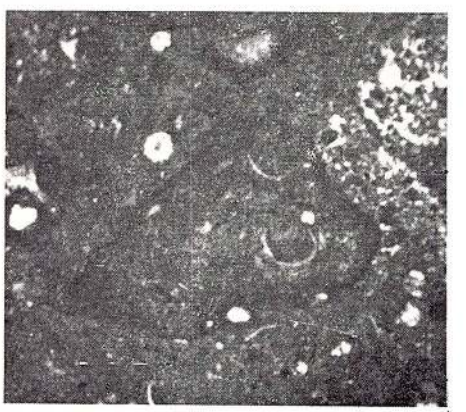
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PLATE XII

Fig. 1. — Nodular micrite containing some pelmicrite to pelsparite patches where the fabric difference between the nodules and their matrix is obvious proving the detrital nature of these nodules. Marila Limestones from the Livadica area (northern slope of Danube).

Fig. 2. — Nodular pelmicrite. In some well-rounded nodules the marginal micrite allochems are intersected by the nodule boundary, thus showing detrital origin of these nodules. Marila Limestones from the Livadica area.

Fig. 3. — Moderately dolomitized nodular micrite from the lower part of the Marila Limestones. The nodules are hardly distinguished from their undolomitized matrix. Mindrişag Valley (south-west of Anina).

Fig. 4. — More intensely dolomitized nodular micrite from the lower part of the Marila Limestones. Lindina Mare Valley (tributary of Nera River).

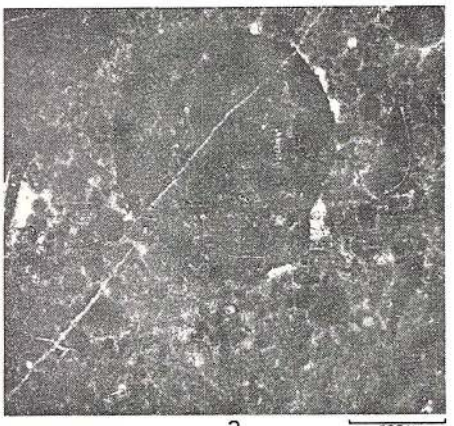
Fig. 5. — Moderately dolomitized nodular micrite from the lower part of the Marila Limestones. Some nodules have their margins finer-grained and more clouded probably as a result of their impregnation with insoluble matter; the nodules and their matrix contain the same calpionellid assemblage (*Crassicollaria* Zone). Mindrişag Valley.

Fig. 6. — Allodapic pelletoid grainstone-packstone from the Marila Limestones occurring in the Mindrişag Valley (south-west of Anina).

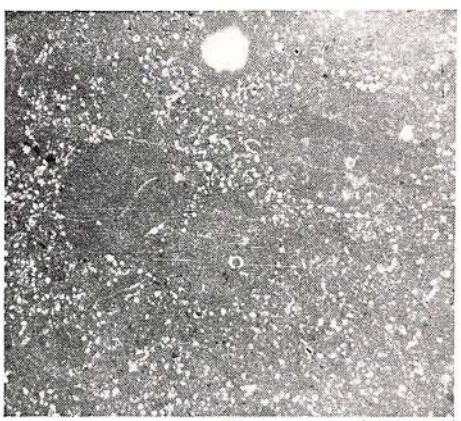
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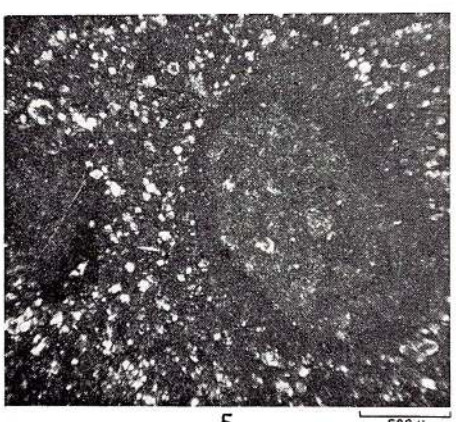
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LOWER GYPSUM FORMATION WEST OF CLUJ-NAPOCA — A SUPRATIDAL EVAPORITE; PETROGRAPHYCAL AND SEDIMENTOLOGICAL FEATURES [1]

BY

BOGDAN POPESCU [2]

Abstract

A supratidal sequence of evaporites (Lutetian in age) is described here. The petrographic evidences lead us to presume the following diagenetic event; replacement of carbonate sediments by 1—5 μ dolomicrite; recrystallization of dolomicrite to 5—10 μ dolomicrosparite and leaching of siltic and arenitic skeletal and nonskeletal grains; filling of pores and vugs by gypsum. Thus, the gypsum filling and the gypsum nodules growth take place in an already replaced and even recrystallized sediment. Then, in the early diagenetic stage, only euhedral dolorhombs, larger than 10 μ, fare formed around the pores. The complete sedimentological study shows that there have been many penetrations and retreats of the sea leaving behind thin sediments. By a supratidal exposure they underwent a penecontemporaneous sabkha diagenesis.

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Introduction

Evaporite formations from the Transylvanian Basin have since long focussed the scientific interest of many a geologists. Due to their

[1] Received on 2nd February 1973 and presented in the Meeting of March 9th 1973.

[2] Institutul de geologie şi geofizică, str. Caransebeş 1, 7000 Bucureşti.

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economic importance, the earlier studies have dealt with deposit conditions, physico-chemical features and ways of evaluation of these rocks.

There have been issued works covering a large range of problems concerning evaporite genesis (P a u c ă, 1967; D r a g o ş, 1969; and many others), or detailed works treating mineralogical and regional correlation aspects.

Reminding that well developed evaporites occur at three stratigraphical levels; Lutetian, Priabonian and Badenian in age, in this work we are going to describe only the major petrographical and sedimentological characters of the Lower Gypsum Formation (Lutetian) west of Cluj.

S t u d i e d A r e a

We have chosen as a study zone a rectangular area lying in the Fînului and Viilor Valley basins, bordered north by the Nadaş Valley and south by the Căpuşului Valley (Fig. 1).

Most complete outcrops in the Lower Gypsum Formation can be found in this area, this being also the only area where one can follow the sequence of evaporite facies from the crystalline basement of the Gilău Mountains (south) up to the zone where this strata go deep under younger formations. It may be mentioned that two quarries are to be found north and south of Leghia village. They offer excellent continuous outcrops where detailed field observations can be done in the area of maximum thickness of our evaporite formation.

W o r k T e c h n i q u e s

We have examined several profiles in the field, more than 150 samples being collected.

80 the sections and 14 polished samples have been studied in laboratory; 10 thin sections have been stained with Alizarine Red, following the F r i e d m a n (1959) techniques in order to establish the exact textural relations between calcite and dolomite; 8 Roentgen analyses have been made for calitative and cantitative evaluation of mineralogical components (Tab. 1)

Geological Setting

The stratigraphy of Eocene deposits of the Transylvanian Basin have been studied since the last century. K o c h (1894) was the first one to make a modern pattern of the Tertiary stratigraphycal sequence in Transylvania, including our study area as well. He separated two gypsum „horizons": lower gypsum horizon and upper gypsum horizon.

R ă i l e a n u and S a u l e a (1956) have first noticed that the Eocene suit can be symmetrically divided in two marine sequences of deposits separated by an intercalated continental sequence. According to these authors the evaporite formations are the first „horizons" in both

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marine series; the lower gypsum and *Anomia* bearing marly limestones horizon and the upper gypsum and oölitic limestone horizon respectivelly.

TABLE 1

Percentage of dolomite and calcite in Lower Gypsum Formation; Foidaș quarry.

No.	Sample no.	$CaCO_3$ %	$CaMg(CO_3)_2$ %	Other components %
1	4177/10	94–96	2–4	2–4
2	4177/12	90–92	5–6	2–3
3	4177/13	5–6	88–92	6–8
4	4177/17	4–6	90–96	1–3
5	4177/21	10–12	82–84	6–8
6	4177/24	2–4	90–96	2–5
7	4177/26	6–8	45–50	40–45
8	4177/28	2–4	94–96	2–6

In the present study only the Lower Gypsum Formation (Lutetian in age) is dealt with, which, as it has been mentioned above, attains

TABLE 2

Generalized stratigraphic succession of the Eocene deposits west of Cluj-Napoca

Age	Lithostratigraphic units
Priabonian	Hoia Formation Bryozoan Marl Formation Cluj Limestone Formation Upper Gypsum Formation Upper Red Clay Complex
Lutetian	Leghia Limestone Formation Mortănușa Marl Formation (emend) *Nummulites perforatus* Beds Lower Gypsum Formation
Ypresian-?Paleocene	Lower Red Clay Complex

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the maximum development in the studied region. Table 2 shows the frame of the Lower Gypsum Formation in the Eocene suit west of Cluj-Napoca [3].

We pointed out that the stratigraphical terminology is still in development, many other schemes being proposed.

The Lower Gypsum Formation is getting thicker and thicker (from 10 to 40 m) from south to north, where it disappears under younger deposits right near the Nadaşu Valley.

Petrographical Approach

The petrographic descriptions are mainly based on the two excellent outcrops of the Leghia and Foidaş quarries as of other additional profiles in the Finului and Viilor Valleys (Fig. 1 and 2). In our latter disscutions concerning the sedimentogenesis of evaporites some other incomplete successions scattered over the studied area will be taken into account.

Generally speaking (Fig. 2), three main lithological types can be noticed in the Lower Gypsum Formation; gypsum, gypsum-bearing rocks and dolomites alternating in different manners along the studied profiles.

In the lowermost part of evaporite sequence, above the red clays with intercalated microconglomerates, pebbles and scattered lens of quartzitic sandstones there lays a thin dolomitic bed, the first sign of evaporite regime. The dolomite is yellow-redish, porous, 75—85 cm thick (Fig. 2, 3621/14). It is made up of a 5—10 μ dolomicrospar [4] wherein there are scattered quartz grains and muscovite. The porous dolomicrosparitic mass is partly cemented by gypsum (Pl. I, Fig. 1,2).

The dolomicrospar is light brown with curved crystal boundaries, marked by a thin cryptocrystalline film, probably of an argillaceous

[3] B. Popescu. Stratigraphical Study of Paleogene Deposits between Căpuşul Mic and Aghireşu. 1970. Arch. of Institute of Geology and Geophysics.

B. Popescu. Stratigraphical Study of Paleogene Deposits between I Ticu-Mihăieşti Zone and II Valea Chioarului Zone. 1971. Arch. of Institute of Geology and Geophysics.

[4] This term would describe the same neomorphic features as those already pointed by Folk (1965) in the neomorphic transformation of the micritic calcite to microspar. In this work we try to demonstrate that after the replacement of the calcium carbonate (aragonite or Mg-calcite) sediments by dolomite; there results a dolomicritic mass (less than 5 μ in size) which passes via recrystallization to dolomicrospar (5—10 μ in size). All this processes are penecontemporaneous with the sedimentation of the lagoonal deposits as it results from the analogy with the recent evaporitic environments correlated with our petrographical data. The dolomicrospar is characterized by the great uniformity of the crystal size (5—10 μ) and by the ubiquity of the clay and organic? impurities. The crystals have curved to plane boundaries and form equant blocks. Tiny clay rims around the dolomicrosparitic crystals are frequent. The next step in this order of the dolomite recrystallization will be (probably in the early stages of diagenesis) the aggradation of the dolomicrosparite to dolopseudosparite. It is necessary to point out the genetic signification of these terms in opposition with the descriptive signification of Tood's microdolosparite and dolosparite (1966).

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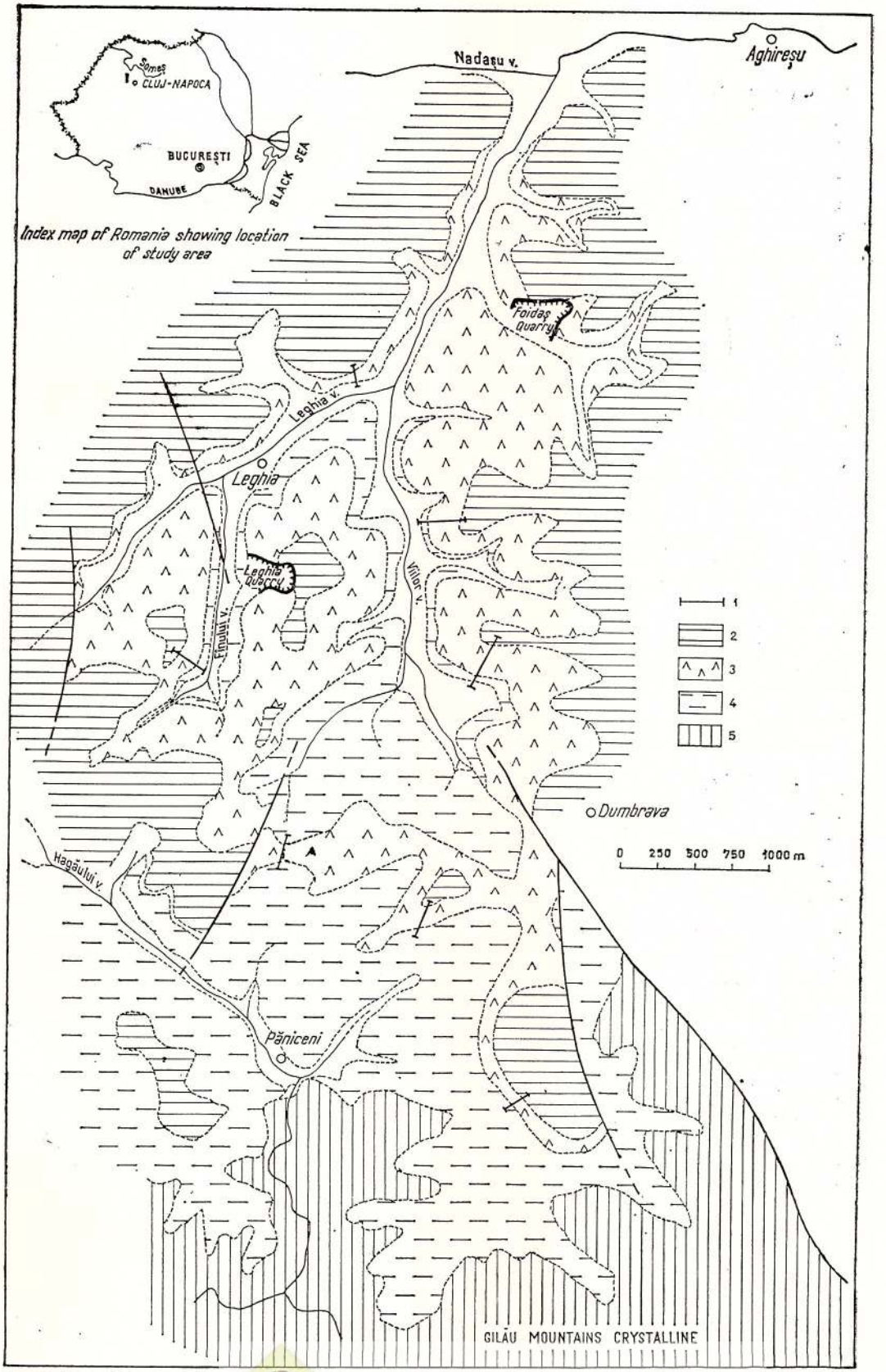


Fig. 1. Geologic map of the studied area. 1, studied outcrops; 2, postgypsum deposits; 3, Lower Gypsum Formation; 4, Lower Red Clay Complex; 5, crystalline basement.

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nature. This film was perhaps born within the recrystallization process of dolomicrite to dolomicrospar, the argillaceous matter being pushed aside to the crystal borders (Pl. XI, Fig. 2). The ferrous ion that is allways present in our dolomitizing solutions gives the brown colour of dolomicrosparite crystals. This ferrous ion is also present in solutions that follow the penecontemporaneous dolomitization as it is clearly noticed by cryptocrystalline patches of limonite which obscure the texture, covering both dolomicrospar and the gypsum cement.

Around the pores, 15—20 μ euhedral dolorhombs can be generally noticed, contrasting with the 5—10 μ dolomicrosparite mass. The larger pore dolorhombs have palimpsest structures. The outer clear dolomite envelope is built around a cryptocrystalline nucleus probably of an organic or clay nature. There are also 50—75 μ dolorhombs with a 15—20 μ brownish dolomite nucleus (the same as those formerly described), which grades into clear dolomite cover at this 50—75 μ crystal size. At high magnification on can notice that these crystals are not perfectly euhedral in shape, the growth of the clear dolomite cover being made at the expense of the dolomicrospar mass around it (Pl. XI, Fig. 3). Dolomicrosparitic inclusions are kept within the gypsum cement. Larger surfaces of these cement have the same optical arrangement.

Petrographical evidences as well as, we shall see, the sedimentological ones entitle us to suppose that this dolomite is the result of the penecontemporaneous dolomitization of upper intertidal and supratidal sediments. Dolomitization is the effect of capillary concentration of the interstitial solution. The fresh-marine waters moving landward become more and more concentrated being able to precipitate gypsum and to dolomitize the carbonate muds. This kind of process is common only for hot and dry regions where evaporation exceeds precipitation plus runoff. Nowadays, the recent sediments of the Persian Gulf (Illing et al., 1965; Butler, 1969; Kinsman, 1969; Kendall, Skipwith, 1969) or those of the Andros Island (Shinn et al., 1965) are the sites of the above mentioned process.

Gypsum precipitation precedes or is simultaneous with dolomitization (Butler, 1969). As the result of this first process a dolomicritic (less than 5 μ) mass with gypsum nodules and cement arises. Next to it, subaerial or submarine, that is directly influenced by the marine waters, the agrading recrystallization of dolomicrite to dolomicrospar takes place. This is a proper time when the argillaceous and organic material mixed with carbonate material in the initial sediment is pushed aside. Dolorhombs larger than 50 μ with brownish 15—20 μ dolomite nucleus are perhaps the last effect of the diagenetic event of this basal dolomite.

This diagenetic event and field data strongly suggest that we may speak about a transgressive level (=basal transgressive supratidal dolostone — Friedman, Sanders, 1967). Landward dolomitization is thus the effect of a slow covering process of the supratidal areas by the transgressive sea.

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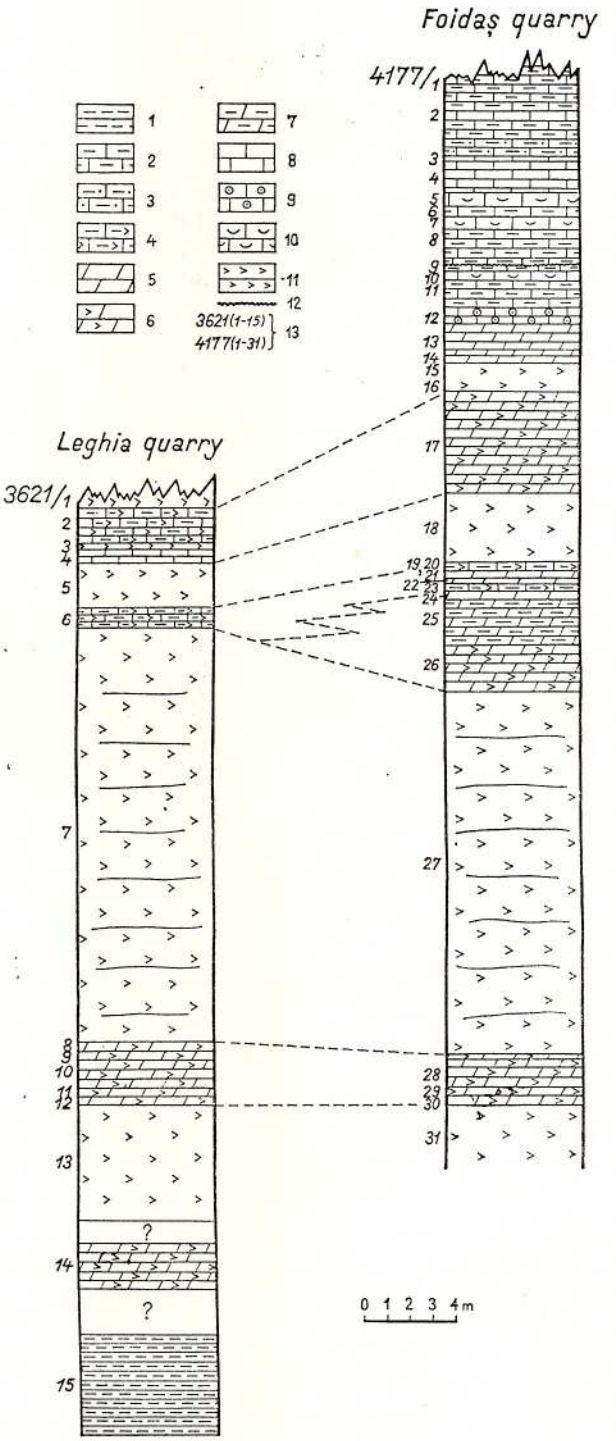


Fig. 2. — Lithologic columns through Lower Gypsum Formation of the Leghia and Foidaș quarries.

1, red clays; 2, marls; 3, silty marls; 4, gypsiferous marls; 5, dolomites; 6, gypsum bearing dolomites; 7, clayey dolomites; 8, limestones; 9, oölitic limestones; 10, bioclastic limestones; 11, nodular and bedded massive gypsum; 12, stromatolites; 13, number of the analysed sample.

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To the upper part of this dolomitic level, the detritic supplies increase; quartz grains become larger up to 5 mm in diameter showing a poor sorting and muscovite quantity is sensibly high (Pl. I, Fig. 4). There comes without any transition a thin pebble and sand layer with oblic stratification, and next to it red clays with lens shaped laminae of sands and silts.

Above this short terrigeneous episode there comes the first gypsum episode. This is a 5 m streaky-bedded mosaic[5] getting in the upper part nodular mosaic. The streaks are dark coloured of an argillaceous or bituminous origin. The texture of gypsum nodules is random or aligned felted, with patches of blocky mosaic textures.

There follows a 2.5—3 m thick allternance of finely laminated dolomites with nodular gypsum-bearing dolomites and nodular gypsum. The dolomites are very porous chalky, yellowish or light brownish in colour. If hammered they spread a bituminous smell.

In the lowermost part of this dolomitic bed there is to be found a very thin lamination (Pl. I, Fig. 1,4) due to the larger quantity of mixed iron oxides, clay and organic matter, which was in suspension in the fresh marine waters of depocenter. The laminated dolomites pass gradually to unlaminated white-greyish dolomites with some clay and ferruginous impurities but with gypsum diaclases and micronodules (Pl. I, Fig. 1 B). Above this already described microcycle one can find other microcycles of this kind or incomplete ones, always separated by a little erosional surface (Pl. I, Fig. 1D). Sometimes the lamination may be disturbed, the next microcycle moulding this relief. This disturbance is probably the effect of water movements in sedimentary medium, due to the high spring tides or abnormal storms.

The gypsum diaclases are normal to the bedding surface and are found either under the erosional surfaces or may transect (less common) more microcycles being larger than the former ones (Pl. II, Fig. 1). These structures suggest shrinkage cracks made up under subaerial exposure of dolomitized and gypsum bearing sediments. The resulted cracks have been subsequently filled by gypsum precipitated from concentrated waters of supratidal pans.

Overlaying the laminated dolomites alternance there are chalky, brown-yellowish gypsum nodules bearing dolomites or dolomites with 5—15 cm thick laminae. There can be also found gypsum diaclases often connecting the scattered gypsum nodules. The quantity of clay and ferruginous impurities is smaller here giving the rock a mottled aspect (Pl. IV, Fig. 1).

At microscope study we may notice the presence of gypsum only as cement within the fine varved brownish dolomites (Pl. XI, Fig. 2). The quantity of gypsum increases upwards within the microcycles where gypsum is to be found only as micronodules. They are mostly of an ovoidal or pris-

[5] In this work we follow the M a i k l e m et al. (1969) terminology proposed for anhydrite structures and textures. This is successfully applied by us in the field of gypsum structures and textures.

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matic shape of 75 to 500 μ in size and are „floating" in a 5—10 μ dolomicrosparitic mass (Pl. II, Fig. 2, 3). Around these micronodules as well as around the empty pores there are transparent 15—25 μ euhedral dolomites (Pl. XVI, Fig. 1). Some of these crystals have dark coloured, cryptocrystalline nucleus, probably of a clayey and/or organical nature (Pl. XVI, Fig. 1).

In nodular-mosaic gypsum beds and in nodular gypsum bearing dolomites we may find the same ovoidal and prismatic voids, subsequently filled by gypsum. The ovoidal micronodules have been considered as dissolved pellets or oölites and the prismatic ones as dissolved bioclasts. The dissolution of the sand and silt sized particles occurs in the last phases of the dolomitization process (I l l i n g et al., 1965; D e f f e y e s et al., 1965; M u r r a y, 1969). We may watch all the phases of pellet dolomitization. First larger anhedral dolomite crystals with palimpsest structures appear in dolomicritic pellets, leaving a porous mass behind. Then, this anhedral dolomite is pushed aside by gypsum infilling. Thus the photomicrograph 3 of Plate V is very instructive in this respect. One can see in it the following succession: under streaky nodular gypsum there is a thin pelletal bioclastic dolomicrospar; within this dolomicrospar the pellets dolomitized bearing larger 10—20 μ dolomite crystals with palimpsest structure; other pellets are completely filled by gypsum. In the underlying pelletal dolomicrite, there can be seen gypsum filled fissures (probably former mudcracks) and pores, as well as the thin rims of dolomicrospar around them (see also Pl. XIV, Fig. 4, Pl. XIII, Fig. 1,2).

The nodular-mosaic gypsum beds are intercalated at different levels in the dolomites. The gypsum nodules are of centimetric size and the clayey impurities lend them a grey colour and a dull luster. The dolomicrosparite was pushed aside by the growth of gypsum nodules. More nodules are often joined together, thin dolomicrosparitic streaks being left among them (Pl. V, Fig. 2,4). The gypsum nodules have a blocky mosaic and more seldom a felted texture. Sometimes in the dolomites there are to be found gypsum rosettes. The acicular gypsum of rosettes is mostly anhedral in shape but there are also well shaped crystal boundaries. Unaffected dolomicrospar is found within acicular gypsum (Pl. IV, Fig. 2,4).

Out of what we have mentioned above it results that the first step in dolomitization of carbonate sediments is the transformation of clay sized sediments into 1—5 μ dolomicrite. An advanced dolomitization causes the dissolution of dolomitization resisting particles (e.g. bioclasts and the great majority of pellets). Concomitently or just before there takes place the dolomicrite to dolomicrospar recrystallization. The resulting pores were subsequently filled by gypsum precipitated from supratidal highly concentrated waters. These transformations take place in the wet sediments in the penecontemporaneous stage of diagenesis. During the early diagenesis the precipitation of dolorhombs larger than 10 μ occured. This was probably possible owing to a long interaction of the dolomicrospar with interstitial solutions or by precipitation from the residual Mg-rich solution resulted after the gypsum precipitation. Thus the penecontemporaneous dolomitization and recrystallization occured at

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the same time or immediately before the gypsification process. In the early diagenesis only the larger dolorhombs will precipitate.

Unlike the first dolomitic level (3621/14, Fig. 2) at this level (3621/12-8, Fig. 2), due perhaps to a longer subaerial exposure the evaporitic diagenesis is more advanced. There occurred the dissolution of the resistent dolomization particles, the growth of gypsum nodules, and the subsequently filling of the resulted vugs by gypsum.

In both quarries as well as in other outcrops a thin varved marl with gypsum lenses of about lo cm thick beds overlies the lithological sequence already described.

There follows the main gypsum mass of the Lower Gypsum Formation. It is bedded massive (1—2,5 m thickness of beds) with thin varved green or blackish clay joints. The gypsum is chalky white, seldom greyish or reddish with black, probably bituminous laminae. The upper part of the main gypsum mass is blocky nodular-mosaic not bedded structure. Random felted and blocky mosaic textures occur. In the upper nodular gypsum the streaks are greyish and felted and blocky mosaic textures are quite common.

Over the main gypsum mass in Leghia quarry there is a gypsum bearing, finely laminated green clays. In the gypsum lens zones the lamination of the clays is disturbed due to the gypsum growth within muddy sediment.

In Foidaş quarry over the main gypsum mass there follows a 5 cm thick alternation of dolomites, gypsum bearing dolomites and gypsum nodules bearing marls as well (4177/26—19, Fig. 2). In the first 2 m there are white-yellowish chalky dolomites having variable amounts of clay and gypsum and nodules.

There follows a white porous dolomite of 1 m thickness (4177/25, Fig. 2). Here, within a dolomicrospar of 5—10 μ in size, around the pores, clear euhedral dolorhombs of 15 μ and larger are to be noticed as well. There can be also found mica, quartz and feldspar grains probably of an eolian origin. This texture is often obscured by argillaceous patches.

The white porous dolomite is overlain by a 10 cm thick brownish dolomite with smaller than one centimeter lenses of green-grayish pelletal dolomite (Pl. VII, Fig. 1). A large number of 100—500 μ ovoidal pellets, a smaller quantity of quartz grains (less than 5%) clay and limonite are to be found within the 5—10 μ dolomicrosparite mass. Most of the dissolved pellets are gypsum filled and the undissolved ones exhibit a different stage of dolomitization. The dissolved pellets leave ovoidal vugs which before gypsum filling are covered inside by a thin limonitic film (Pl. XIV, Fig. 1,2). After the gypsum filling this appears like a brownish cryptocrystalline envelope. Around the pores, as we have already mentioned before, there are the euhedral dolorhombs with or without palimpsest relicts.

In the dolomicritic tongues most of the pellets are undissolved (Pl. VII, Fig. 2,4) but for those on the dolomicrite/dolomicrospar boundary which are gypsum filled. The pellets of dolomicrite tongues are cryptocrystalline with scattered dolorhombs whose nuclei are dark probably of a mixed clay and organic nature.

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Gypsum microdiaclases start from dolomicrosparite to the dolomicritic tongues (Pl. VII, Fig. 2). They transect gypsum filled pellets or larger dolorhombs bearing pellets without no change in their texture.

Such being the case it results that this pelletal dolomitic bed is a review of all the diagenetic changes having occurred in the already described dolomites. Summing up the petrographical descriptions and ultimately the textural relations between different components of the mentioned dolomitic rocks, we can presume the following diagenetic events :

1. The dolomitization of carbonate mud which turn into a 1—5 μ dolomicritic mass. Penecontemporaneous recrystallization of dolomicrite to dolomicrosparite is supposed to occur in the unlithificated wet dolomicritic sediments. The capillary concentration and the seepage refluxion are suspected to be responsible with.

2. By the advancing of the dolomitization process, the resistant siltic or arenitic particles (all the bioclasts and some of the pellets) will be dissolved leaving an important vug porosity behind. Some of the undissolved dolomicritic pellets may bear larger dolorhombs with palimpsest relics of an organic nature.

3. In the newly created porosity it can be found clay, organic and ferruginous impurities which imbued the walls of the pores. These impurities will be first lost by decantation from the brines that are to be found in supratidal pans and lagoons. Then, by seepage of highly concentrated brines gypsum can precipitate filling thus most of the pores, shrinkage cracks and other cavities. If the sediments were exposed, in the dolomicritic but the most common in the dolomicrosparitic bulk of sediment the gypsum nodules will grow.

4. After a thin burial or due to the long contact of „residual" magnesium-rich solutions resulting from gypsum precipitation, the precipitation of clear euhedral dolorhombs larger than 10 μ takes place around the filled and unfilled cavities.

According to the latest dealing with supratidal „sedimentation" as well as with the changes occurring within the mass of such kind of sediments we presume that the 1, 2, 3 phases of our diagenetic event are penecontemporaneous. There are many evidences that these transformations take place in the soft wet sediment. The precipitation of clear euhedral dolomite larger than 10 μ is probably early diagenetic.

Overlying this much discussed pelletal dolomite there is a 30 cm of nodular gypsum bearing marls. The expansion of nodules in the soft sediment has disturbed the initial lamination. The muddy sediment between the nodules has been expelled, remaining only as thin streaks (Pl. VIII, Fig. 1, Pl. XIV, Fig. 3). The nodules are alligned or random felted with blocky mosaic patches in texture (Pl. VII, Fig. 3). In this felted texture there are larger than 250 μ euhedral anhydrite laths. Generally speaking, the anhydrite is precipitated from high temperature (more than 35C)° and chlorinities (more than 145%) interstitial solutions

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(Butler 1969). These very restrictive conditions are met in inner sabkha during the dry and hot season (Kinsman, 1969).

There follows a chalky yellowish free of gypsum dolomite bed. And here as we have always noticed up to now, within a 5—10 μ dolomicrosparite there are the common euhedral dolorhombs around the pores. As gypsum is missing here too, we may assert that in our supratidal sequence the first precipitate is dolomite. This fact is contradictory to the latest works on the recent sediments of the Persian Gulf (Illing et al., 1965; Butler, 1969; Kendall, Skipwith, 1969). This free of gypsum dolomite might represent the supratidal sediment of inner sabkha (Butler, 1969). The brines with highest Mg/Ca ratio come from the hydrostatic level which have already precipitated gypsum in the outer sabkha. Neomorphism e.g. the recrystallization of dolomicrite to dolomicrosparite does not depend on gypsum precipitation.

This dolomite is overlain by a 20 cm yellowish pelletal dolomite. The pellets are in different stages of dissolution (Pl. VIII, Fig. 4). The vugs thus created generally rimed by a thin limonitic envelope. Some of the pellets show a central relic surrounded by a gypsum envelope (Pl. XIII, Fig. 3, 4) suggesting a replacement of pelletal dolomicrite by gypsum.

Bedded nodular gypsum of a 15 cm thickness and yellowish chalky dolomite of a 5 cm thickness overlie the pelletal dolomite. They have the same textures and structures as those described for subjacent beds (see the description made for 4177/23 and 4177 /22).

Above them there lies a milky-white not bedded nodular-mosaic gypsum. The streaks between nodules are dark of an argillaceous-organic nature. And here the textures and structures are comparable with those of 4177/23, 4177/31, 3621/6, 3621/5 (Fig. 2). The 4177/18 may be correlated with 3621/5 and 3621/6 levels of the Leghia quarry.

In the Foidaş quarry above the nodular-mosaic gypsum there follows a 4,25 m yellowish chalky dolomite and greyish clayey dolomite alternance. The dolomites are highly porous being pelletal and bioclastic dolomites with scattered quartz grains (Pl. IX, Fig. 1). The textures are largely obscured by an important amount of clay. The rock mass is less than 5 μ dolomicrite. Around the micropores and scattered in the dolomicrite there are, following the rule, dolomicrospar patches.

Southwards, at the Leghia quarry, the equivalent of this dolomicrites is an alternance of chalky-clayey dolomites and greenish bearing gypsum nodule marls.

In both quarries there follows a not bedded, 1 m thick nodular-mosaic gypsum. It is much like those already described but here the streaks are only of white chalky clayey dolomite (Pl. XV, Fig. 4).

Further on the detailed references will regard only the Foidaş quarry as we have here the most obvious profile between the Lower Gypsum Formation and *Nummulites perforatus* Beds. In the Leghia quarry the uppermost part of the wall is blown off by the explosions during the exploatation.

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Thus in the Foidaş quarry the nodular-mosaic gypsum is overlain by a 1,80 m chalky, yellowish dolomite bed which gradually passes to clayey, brownish bioclastic dolomite. The bioclasts are dissolved and then filled with gypsum. Around the pores and the dissolved bioclasts there can be seen dolorhombs larger than 10 μ which markedly contrast with the dolomicrosparitic mass (Pl. IX, Fig. 4). Some of the pores are imbued with limonite and clay impurities.

Above the bioclastic dolomite is an oölitic bioclastic limestone of a 0,70 m thickness. The mud supported rock is a packstone according to D u n h a m ' s (1962) nomenclature (Pl. X, Fig. 1). This level is strongly transgressive. Southwards, the oölitic limestone becomes devoid of mud matrix, the bioclasts are subordinate; the cement is sparry calcite, the rock becoming a grainstone (Pl. X, Fig. 2). The scarce bioclasts play a sheltering role, the fabric under them being porous and incompletely cemented (Pl. XVII, Fig. 2). The oölits have a quartz, miliolid or pellet nucleus.

As in the Foidaş quarry the oölits are less than 15% and the associated matrix and particles (pellets, lumps) suggest a lagoon environment, we consider that the oölits have been splashed from the oölitic shoals into the lagoon, thus resulting a mud supported rock. Therefore, we may suppose that the lagoon was basinward as compared to the oölitic shoals (we mention that the crystalline basement and hence the land was southwards).

Overlaing the oölitic limestone without any transition there is a 1,5 m greyish marl bed. The marls contain the mollusc molds and miliolids. At the upper part the marls become more calcareous and highly bioclastic. The bioclasts are mainly forams, most of them being filled with sparry granular mosaic calcite. This kind of fabric is characteristic of modern-near shore calcareous sediments (A l e x a n d e r s s o n , 1972), being abundant in turbulent environments.

A thin stromatolitic level of 20 cm thickness ends the first cycle of *Nummulites perforatus* Beds. It is an ancient algal mat of „intertidal" origin[6]. Various kinds of grains, most common bioclasts are bound within the algal micrite (Pl. X, Fig. 4). Various neomorphic textures are to be found in these algal micrites (Pl. XVII, Fig. 3, 4).

Thus the *Nummulites perforatus* Beds started with a transgressive oölitic limestone followed by a lagoon bioclastic marls and by near shore bioclastic and stromatolitic levels. Over this short „intertidal" episod there comes the main marly mass of *Nummulites perforatus* Beds with a net marine fauna; nummulites, echinoids and molluscs.

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[6] In the Eocene time, the Transylvanian Basin was an arm of the Pannonian Sea. The tides probably were of small amplitude and consequently the intertidal space was quasiabsent as results form the lack of intertidal-structure sediments in our lithological sequence. We used he supra-, inter- and subtidal terms for making them easier to understand when referring to the carbonate sedimentology works.

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Discussions on the Environment of Deposition

The interpretation of the environment of deposition is based broadly on petrographical and stratonomic evidences given us by the two reference sections : Leghia and Foidaș. We will try to reconstruct the major events of sedimentary environment in the deposition time of Lower Gypsum Formation also adding the supplementary information furnished by many other sections scattered over the study area (Fig. 1)..

As we have already said, above the coarse deposits — microconglomerates, pebbles, sands and sandstones with red clays intercalations — of terrigenous origin, there comes transgressively the first occurrence of an evaporite regime in the region west of Cluj. The basal transgressive dolomite (common in many other world wide regions — Friedman, Sanders, 1967) is in fact a supratidal sediment which has become dolomitized under the action of capillary concentration process of marine solutions.

The basal dolomite (3621/14, Fig. 2) is an initial mixture of wind blown material (quartz and mica) and small amounts of carbonate sediments of supratidal origin. The interstitial water loss due to the high rate of evaporation is counterbalanced by the continuous flow of ground waters from the adiacent sea and by occasionally flood recharges (capillary concentration vs. flood recharge). This kind of sedimentation and their penecontemporaneous diagenesis is known in sedimentological literature as sabkha cycle.

It is known that in the supratidal spaces of hot and dry regions where evaporation exceeds precipitation and runoff, the concentration of interstitial solution is 10 times greater than that of fresh marine waters (Kinsman, 1969). Aragonite and gypsum can precipitate at this concentration and the replacement reaction between sediment and interstitial solutions can occur (Illing et al., 1965; Butler, 1969; Kinsman, 1969). Due to the aragonite and gypsum precipitation the Mg/Ca ratio will become sensibly high because of using Ca^{++} in $CaSO_4$ and $CaCO_3$ lattices. The growth of magnesium concentration in interstitial solutions will make possible the beginning of the carbonate mud dolomitization.

Illing et al. (1965) and Butler (1969) suggest that there is a direct quantitative relation between dolomitization and gypsum precipitation. Out of Butler's statements it results that this relation may be written as follows :

$$2CaCO_3 + Mg^{++} + SO_4^{--} = CaMg(CO_3)_2 + CaSO_4$$

sediment solution solution sediment sediment

A microcrystalline or cryptocrystalline sediment will result out of carbonate mud dolomitization. The siltic and arenitic particles are't resistent to dolomitization and will be leached by advancing of the dolomiti-

[7] *Op. cit.* point 6.

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zation process (I l l i n g et al., 1965; D e f f e y e s et al,. 1965). At the same time or just after, the penecontemporaneous recrystallization of dolomicrites takes place. The newly created pores can be subsequently filled by gypsum.

The basal dolomite records the first phases of penecontemporaneous diagenesis — the replacement of carbonate sediments by dolomite and its recrystallization to dolomicrospar. These are immediately followed by the cementation with gypsum (Pl. XII, Fig. 1). In the early diagenesis a precipitation of euhedral dolorhombs larger than 10—25 μ takes place. Then, these euhedral dolorhombs acquire a clear dolomitic envelope at the expense of the surrounding dolomicrosparite (Pl. XI, Fig. 3).

A new terrigenous supply is marked without any transition by cross laminated sands and pebbles. Above the sands there lies red clay, well laminated with sand lenses only. These deposits are thicker and thicker from north to the crystalline basement (south). The terrigenous deposits signify a regressive phase of the sea and the installation of continental sedimentary conditions. Probably many lakes receiving the torrential apports from the Gilău Mountains mainland were formed in the supratidal flat.

After this regressive episode the evaporite-marine regim is definitively established in the studied area. Thus, it is only north to the Dumbrava village (Fig. 1) that there is a nodular gypsum bed (Fig. 2, 3621/13 and 4177/31). It is a typical supratidal deposit. By a long time persistence of the evaporative conditions in the mid and inner sabkha, gypsum will largerly precipitate. It will grow in clay and silt sized sediments expelling and digesting the surrounding sediment. If sediments are carbonates, the dolomitization and recrystallization occur first. In Plates XIV, Figure 4, and XV, Figures 1,2, we can notice the tiny streaks and isolated crystals of dolomicrospar within gypsum nodules.

In its southward advance, the sea overlaps the nodular gypsum which will be overlain by subtidal pelletal and bioclastic muds. These deposits have about the same extent as the basal dolomite but not exceeding them.

The first deposits are finely laminated, pelletal bioclastic dolomites, followed by unlaminated, gypsum nodules and micronodules bearing dolomites. The structures and textures of these sediments strongly resemble with those of the coastal lagoon pans (K i n s m a n , 1969) and with well laminated brown clayey silts of the transition zone of the Colorado River Delta (T h o m p s o n , 1968). In the upper intertidal and supratidal area of the Colorado River Delta the gypsum is precipitated between the products of two floods. As in the Transylvanian Basin the tides where probably of little amplitude, the coastal and supratidal lagoons had a greater stability leading to a rapid concentration of waters and thus to gypsum precipitation. The petrographical sketches read that the dolomitization and then the recrystallization of the dolomicrite promotes from pores to the bulk of matrix (Pl. IX, Fig. 4, Pl. XII, Fig. 2). We presume that dolomitization started due to the seepage refluxion of thin films of suprasaturated solution from the shifting supratidal lagoons.

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After a subaerial exposure desiccation cracks may occur. The pores, the leaching grains and the desiccation cracks are filled by gypsum of a flood recharge origin (Pl. XIV, Fig. 4). If a new supply is absent the gypsum nodules growth is allowed and that is why the dolomicrite and the clay „matrix" is pushed aside. Before the concentration of the new marine supplies in the already mentioned cavities and above the substratum the impurities carried in solution by this flooded waters will be laid down.

The main gypsum mass evidences a new sea retreat, being deposited in a coastal lagoon with permanent water layer due to the quite regular periodic influx. The gypsum is thinner southwards and dissapears north of Păniceni village (Fig. 1). Near the Leghia village it attains 20 m in thickness being bedded massive with clayey joints and dark varved lamination. The varved and bedded structure suggests a stable evaporitic regime with periodic influxes. Each new marine supply was marked by the tiny clayey or clayey-bituminous laminae which represent the products of decantation of the suspended matter.

The mixture of sediments with the organic matter is reported in many areas of evaporite deposition. The dark colour of these laminae are due to strongly reducing conditions which are revealed by H_2S release (ex : Baja California, P h l e g e r, 1969).

On the upper part of bedded gypsum mass there appear the bedded nodular-mosaic structures which, as we have already said, are probably of an intertidal-supratidal origin. Thus, the lagoon filling leads its inclusion in the supratidal flat with its characteristic processes.

Nortwards, i.e. basinward, pelletal muds are deposited (Fig. 2, 4177/26—24). If exposed, they undergo diagenetic changes already mentioned. We may also say that in this pelletal dolomite the leached pellets coexist with the dolomicritic ones. I l l i n g et al. (1965) wrote about the occurrence of the same dolomitic mud fabric in the supratidal flats in the Persian Gulf. The leached pellets are subsequently filled by gypsum. The undissolved pellets are gradually digested and then filled by gypsum (according to the rule here presented, in early diagenetic stage arround the pores and the gypsum filled pellets, euhedral, clear dolorhombs larger than 10 μ were formed).

In the Leghia quarry and southwards, this pelletal dolomite is missing. This is in favour with the idea of a regressive phase which began at the same time with the main gypsum mass deposition.

A new marine ingression, overlaying the sediment outcropping in both quarries but not extending far to the south, is represented by an alternance of marly dolomites and finely laminated marls. Due to a subsequent retreat of the sea, these sediments have been subaerially exposed. Within their mass a whole range of sabkha processes takes place. Nodular gypsum displacive structures are to be found in the upper part of this regressive sequence (4177/18 and 3621/5, Fig. 2). It is only 1 km south of the Leghia quarry that these nodular gypsum are overlain by the sediments of a new marine ingression.

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These ingressive sediments are subtidal bioclastic muds, in the north and pelletal dolomites with nodular gypsum intercalations in the south (4177/17 and 3621/2—4, Fig. 2).

The last regressive phase in the time-span of the Lower Gypsum Formation emphasises the supratidal condition of deposition in the area. The scattered gypsum nodules within clay matrix are the only evidence left by this event.

Anomia bearing dolomites overlie most of the underlaying sediments, coming in direct touch with the main gypsum mass at the Dumbrava village and with the basal dolomite north of the Păniceni village.

The *Nummulites perforatus* Beds started with a transgressive oölitic limestone having about the same extent as the *Anomia*-bearing dolomites. The oölites suggest us a near shore deposition. They are mud supported, mixed with lumps, pellets and bioclasts northwards (i.e. basinward) and grain supported, cemented by fibrous sparite southwards (i.e. in the vicinity of the crystalline basement).

A new eustatic oscillation is recorded in the studied carbonate area, that is a thin stromatolitic level (4177/9, Fig. 2) representing an ancient algal mat of a probably upper subtidal-lower intertidal origin.

Conclusions

1. Petrographical and sedimentological evidences demonstrate the supratidal origin of the evaporites in the Lower Gypsum Formation.

2. Most of the carbonate sediments in this formation were deposited in normal marine waters. By supratidal exposure they underwent a sabkha penecontemporaneous diagenesis.

3. The diagenetic events seem to be the following:

a) The penecontemporaneous dolomitization of carbonate muds are probably due both to capillary concentration and seepage refluxion phenomenon. In this first stage the lime muds are replaced by a microcrystalline dolomite less than 5 μ. Unlike the penecontemporaneous supratidal diagenesis revealed by the recent sediments, the gypsum cement is seldom found in this stage.

b) After a longer subaerial and may be submarine exposure, the dolomitization process goes on. The pellets, probably of an initial aragonite mineralogy, will be dolomitized and the bioclasts will be dissolved. The recrystallization of dolomicrite (less than 5 μ) to dolomicrospar (larger than 5 μ but less than 10 μ) takes place in the same stage.

c) The gypsum filling of the newly created pores, the replacement of dolomicrite or dolomicrosparite by gypsum and the gypsum nodules growth take place at the same time or immediately after the above mentionated stage.

d) Euhedral dolorhombs are formed around the pores and gypsum filled pellets and bioclasts after a thin burial, probably due to the long contact of high concentrated magnesium solution during the early diagenetic stage. There is no evidence for late diagenetic changes.

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4. Within the Lower Gypsum Formation there are recorded several penetrations and retreats of the sea which left behind thin sediments. The first and the last penetration of the sea (i.e. the basal dolomite and the *Anomia*-bearing dolomite) had the greatest extent. The regressive phases brought about the subaerial exposure of marine sediments which bear the evaporite diagenesis.

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PLATE I

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PLATE I

Fig. 1. — Dolomicrospar partly cemented by gypsum. In the dolomicrosparitic mass there are scattered quartz-and mica grains as well larger dolorhombs (with dashed line in the center of the photo) with palimpsest structures. N +. Leghia quarry, 3621/14.

Fig. 2. — Same as Fig. 1. There are to be shown the gypsum cement (white milky patches) and the quartz-and mica grains (white like small pin points). N //. Leghia quarry 3621/14.

Fig. 3. — Dolomicrospar partly cemented by gypsum. Note the larger dolomicritic intraclast (lower right corner of photo). Basal transgressive dolomite. N//. Finului Valley.

Fig. 4. — Dolomicrite with large amount of quartz grains. Note the unsorting and the angularity of some grains. N //. Finului Valley.

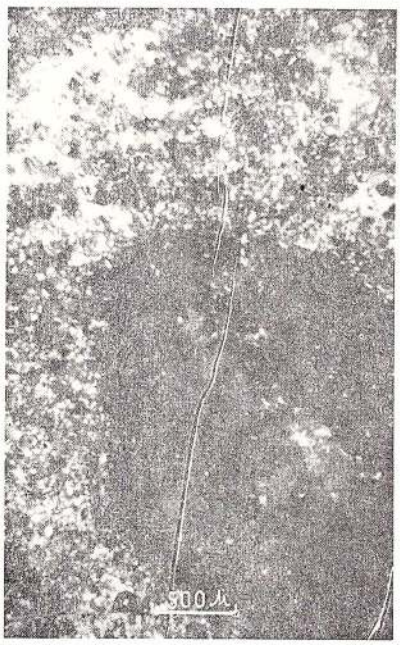
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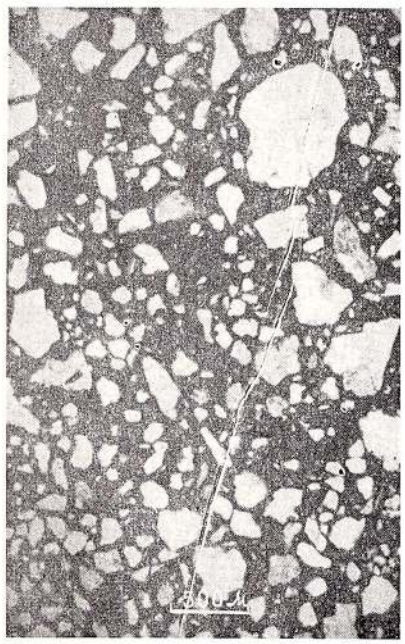
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PLATE II

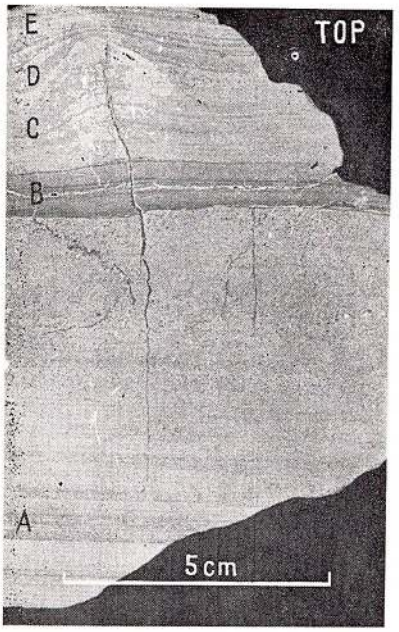
Fig. 1. — Polished slab in fine laminated dolomite. A, B, C, D, E are intervals controlled by thin sections. The lamination is due to the large amounts of iron oxide, clay and organic matter. Note the increase in gypsum micronodules (in fact leached pellets subsequently filled by gypsum) on the upper part of lower microcycle. Under a little erosional surface (C) there are tiny microdiaclases filled by gypsum, interpreted as former shrinkage craks. The upper part of the photo is an inverse microcycle with disturbed lamination. Leghia quarry 3621/12.

Fig. 2. — Dolomicrosparite with rare leached pellets. Note the large amount of cryptocrystalline clay and organic matter obscuring the dolomicrosparitic fabric. N//. Leghia quarry 3621/12 interval A in Fig. 1.

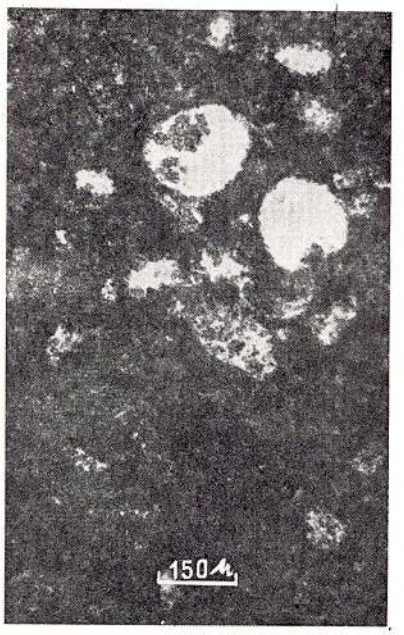
Fig. 3. — Pelletal dolomicrospar. Same features as in photo 2. N//. Leghia quarry 3621/12, interval B in Fig . 1.

Fig. 4. — Interval C in Fig. 1. The cryptocrystalline laminae are of an iron oxides, clay and organic matter origin. N//. Leghia quarry 3621/12.

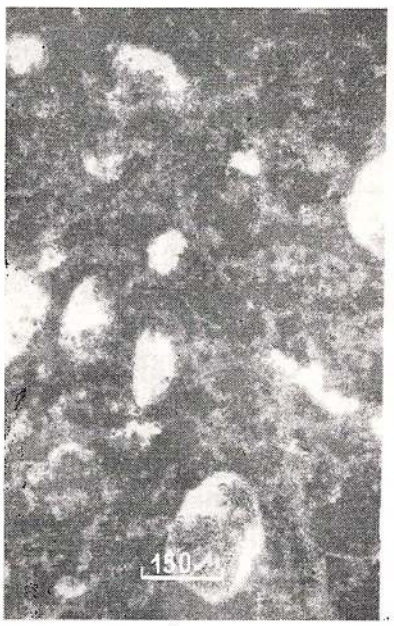
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PLATE III

Fig. 1. — Dolomicrosparite. The dark laminae are a mixture of clay iron oxides and bituminous material. N//. Leghia quarry 3621/12, interval D in Fig. 1, Pl. II.

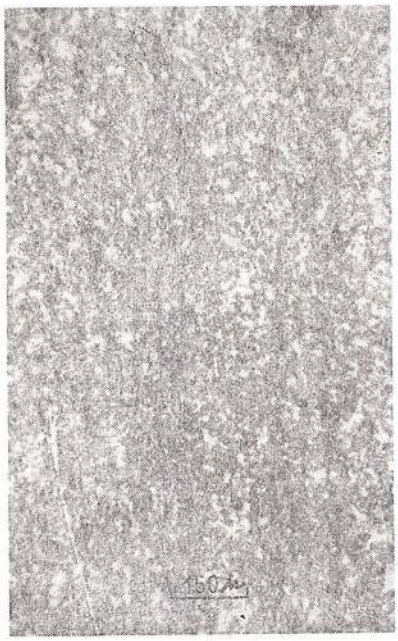
Fig. 2. — Leached ostracods shells subsequently filled with gypsum. The dolomicrosparitic fabric is largely obscured by cryptocrystalline clay iron oxides and bituminous material. N//. Leghia quarry 3621/12, interval E in Fig. 1, Pl II.

Fig. 3. — Same as Fig. 4. N +. Leghia quarry 3621/11.

Fig. 4. — Bioclastic dolomicrosparite. The leached shells are subsequently filled by gypsum. Note the definite boundary of the larger shell and the clear euhedral dolomite (arrows). N//. Leghia quarry 3621/11.

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B. Popescu. Lower Gypsum Formation West of Cluj-Napoca. Pl. III.



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PLATE IV

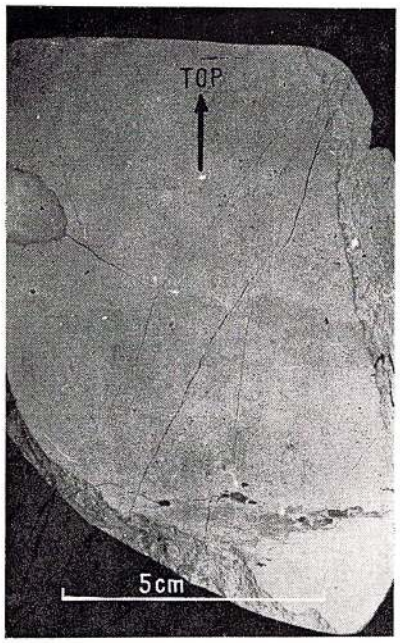
Fig. 1. — Polished slab, Dolomite with gypsum nodules and diaclases. The nodules are former cavities or macropores enlarged by subsequent circulation of gypsum saturated solutions. Note the fine channel bonds between nodules (lower part of photo). Leghia quarry 3621/10.

Fig. 2. — Gypsum rosettes in dolomicrosparite fabric. Most of gypsum needles are partly anhedral in shape. N//. Leghia quarry 3621/10.

Fig. 3. — Dolomicrosparite. The very porous fabric is partly cemented by gypsum. N//. Leghia quarry 3621/10.

Fig. 4. — Gypsum rosette. Within gypsum needle there are kept undigested crystals of dolomicrosparite. N//. Leghia quarry 3621/10.

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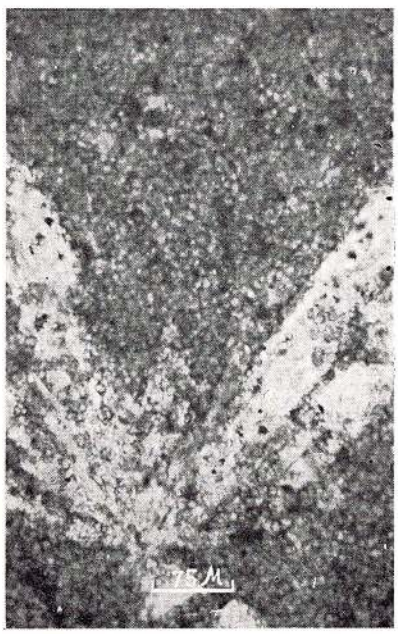
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PLATE V

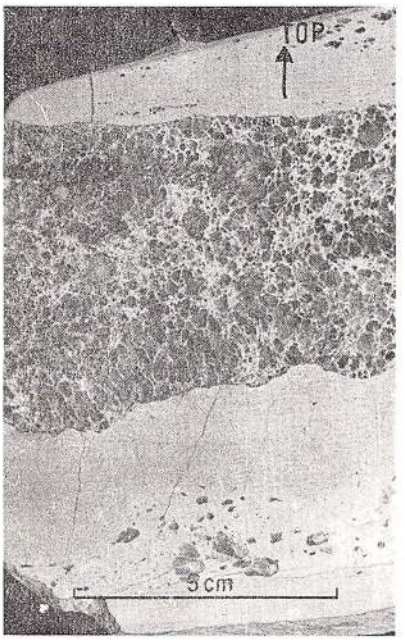
Fig. 1. — Polished slab. Bedded nodular-mosaic gypsum within brown-yellowish dolomite. Note the definite upper contact between the two laminae and the irregular one in the lower part. The lower contact is due to the subaerial exposure and to the gypsum growth, the upper contact is due to the sea landward ingression. Leghia 3621/9. A, B, C, D intervals are controlled by thin sections.

Fig. 2. — Gypsum nodules. The streaks between nodules are made of dolomicrosparite. There are isolated dolomicrosparite crystals kept within gypsum nodules. Thus, the recrystallization take place before the growth of nodules. N //. Leghia quarry 3621/9, interval A in Fig. 1.

Fig. 3. — Transition from nodular gypsum band (wite in the upper part of photo) to pelletal bioclastic dolomicrosparite and to the pelletal micrite (black in the lower part of photo). In the dolomicrosparitic layer most of the pellets are leached and filled with gypsum. There are also pellets (arrows) with scattered dolorhombs having cryptocrystalline nuclei. N//. Leghia quarry 3621/9, interval B in Fig. 1.

Fig. 4. — Gypsum nodules with tiny streaks of dolomicrosparite. The growth of gypsum nodules pushed aside the soft dolomicrospar around them. Compare with photo 2. N//. Leghia quarry 3621/9, interval C in Fig. 1.

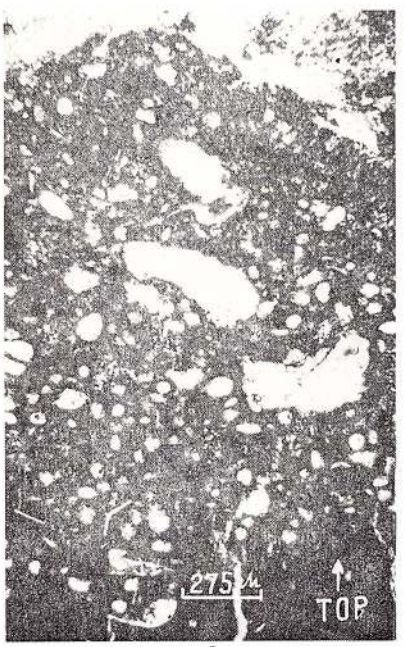
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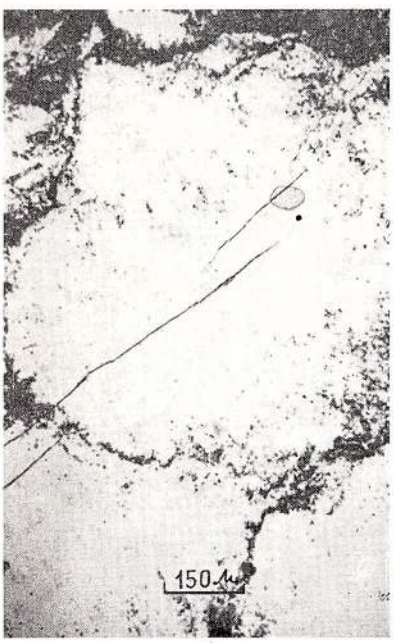
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PLATE VI

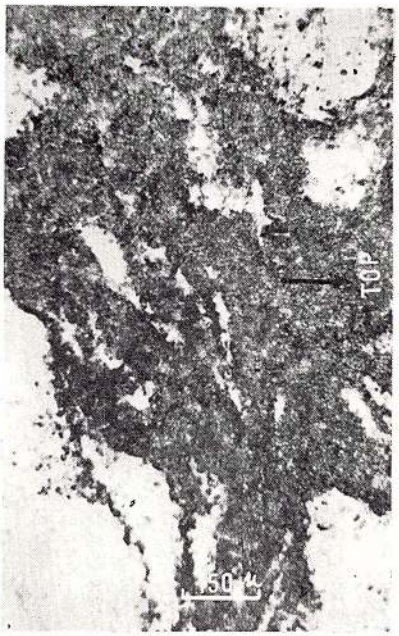
Fig. 1. — Nodular gypsum. In the dolomicrosparitic fabric there are ,,nuclei'' of future nodules. By expansion they can join expelling the dolomicrosparite around them. N//. Leghia quarry 3621/9, interval D in Fig. 1, Pl. V.

Fig. 2. — Dolomicrosparite with an important vug porosity. Some of pellets kepp larger dolorhombs with palimpsest structures and are partly filled by gypsum. N//. Leghia quarry 3621/8.

Fig. 3. — Pelletal dolomicrosparite. The pellets are leached and subsequently filled by gypsum. Some of them are ,,welded'' forming gypsum lenses. N //. Foidaș quarry 4177/28.

Fig. 4. — Same as Fig. 3. Here one can notice the line of penetration of the gypsum rich solutions. In the upper part of photo there are not filled pellets. N +. Foidaș quarry 4177/28.

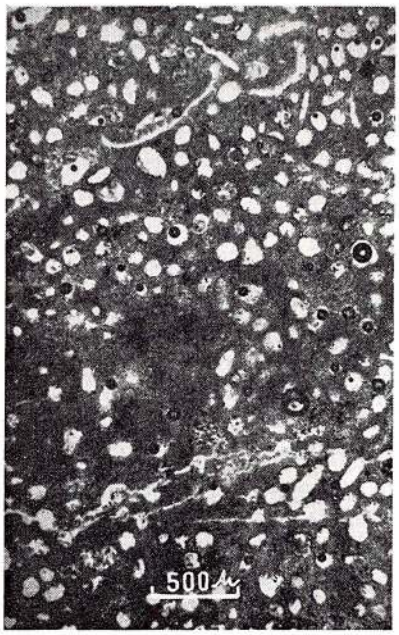
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PLATE VII

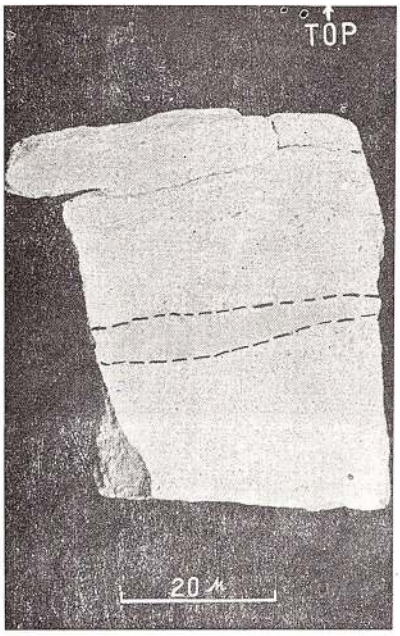
Fig. 1. — Polished slab. Brownish pelletal dolomicrosparite with lenses of slightly pelletal dolomicrite (with dased line). Foidaș quarry 4177/24.

Fig. 2. — Upper dolomicrosparite-dolomicrite boundary. Within the pelletal dolomicrite there are leached and not leached pellets. Some of leached pellets are subsequently filled by gypsum. From the pelletal dolomicrosparite start gypsum filled diaclases which transect the leached and then filled with gypsum pellets as well, not leached pellets (arrows) leaving no textural changes. Thus, the gypsum rich solutions are not related to the dolomitization process. N//. Foidaș quarry 4177/24.

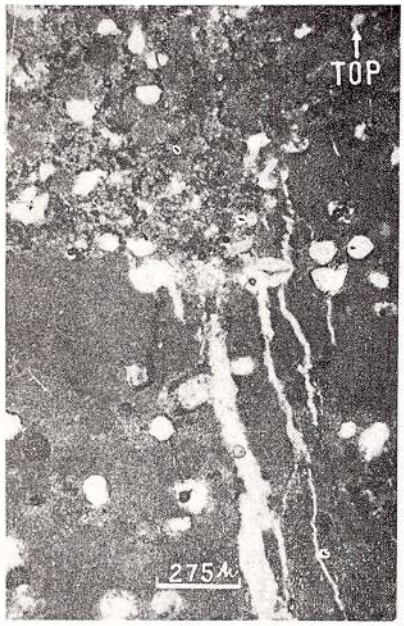
Fig. 3. — Lower dolomicrite-dolomicrosparite boundary. Note the well definite contact and the complete absence of pellets. N . Foidaș quarrry 4177/24.

Fig. 4. — Textural detail in the nodular gypsum. Blocky texture. N +. Foidaș quarry 4177/23.

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PLATE VIII

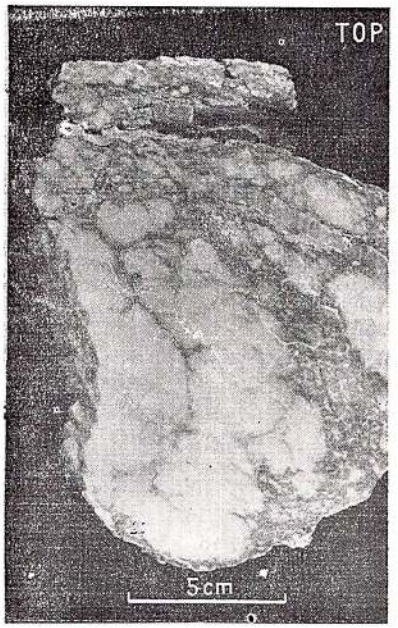
Fig. 1. — Polished slab. Nodular-mosaic gypsum. The growth of nodules have expelled the marly sediment surrounding them; some of the nodules joined together, leaving only tiny marly streaks. Foidaș quarry 4177/23.

Fig. 2. — Dolomicrite in the lower part of the photo, passing into dolomicrospar. Noted the highly porous fabric in the resulted dolomicrosparite area. N//. Foidaș quarry 4177/22.

Fig. 3. — Blocky gypsum mosaic. The euhedral prismatic crystals (arrows) are anhydrite. Foidaș quarry 4177/18. N +.

Fig. 4. — Pelletal dolomicrosparite. There coexists leached, not leached and partly leached pellets (arrows). Some of them are filled with gypsum. N //. Foidaș quarry 4177/21.

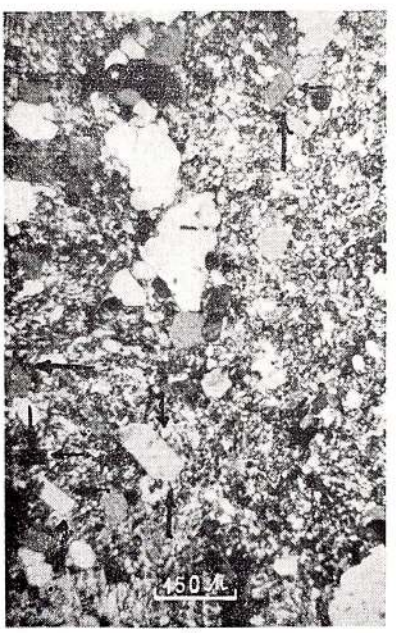
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PLATE IX

Fig. 1. — Clayey highly porous dolomite. Foidaş quarry 4177/17.
Fig. 2. — Random felted gypsum texture. N +. Foidaş quarry 4177/16.
Fig. 3. — Aligned felted gypsum texture. N +. Foidaş quarry 4177/15.
Fig. 4. — Compact dolomicrite (dark area) passes into porous bioclastic dolomicrosparite. The shells remaints have been leached, leaving vugs. N//. Foidaş quarry 4177/13.

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B. POPESCU. Lower Gypsum Formation West of Cluj-Napoca. Pl. IX.



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PLATE X

Fig. 1. — Oölitic packestone. The mud matrix supports various kinds of grains : oölits, bioclasts, lumps and pellets. N//. Foidaș quarry 4177/12.

Fig. 2. — Oölitic grainstone. There are only normal or superficial oölits cemented by sparry calcite. Viilor Valley, same level as in Fig. 1. N //.

Fig. 3. — Polished slab. Stromatolite. In the lower part there is a bulbous massive stromatolite layer overlain by a planar stromatolite. Foidaș quarry 4177/9.

Fig. 4. — Algal micrite with entrapped miliolid shell. Note the micritization of shell walls and the sparry calcite cement inside. N //. Foidaș quarry 4177/9.

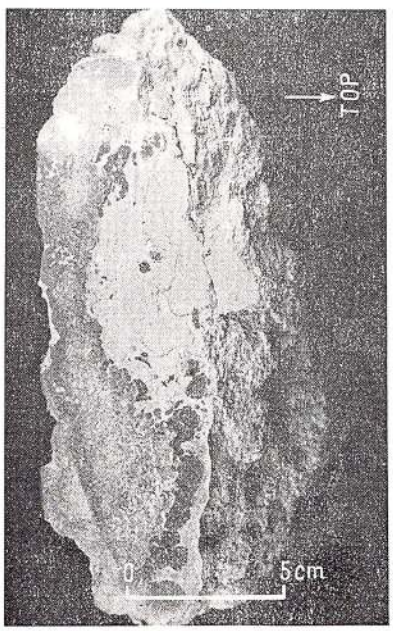
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PLATE XI

Fig. 1. — Typical dolomicrospar. Crystal average are about 6—9 μ in diameter. Note the uniformity of grain size and the equant shape of crystals. N//. Foidaș quarry 4177/14.

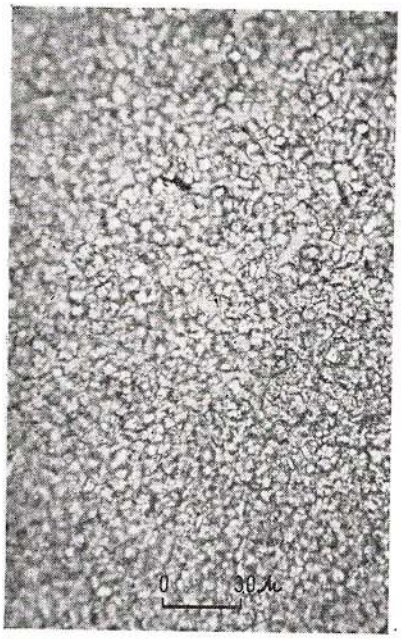
Fig. 2. — Dolomicrospar partly cemented by gypsum (arrows). Some of dolomicrosparitic crystals have palimpsest structures. N +. Leghia quarry 3621/8.

Fig. 3. — Larger dolomite crystal (with dashed line) within dolomicrosparite fabric. Around an euhedral dolomitic nucleus of 25 μ in size, there is a clear dolomitic cover. N +. Leghia quarry 3621/14.

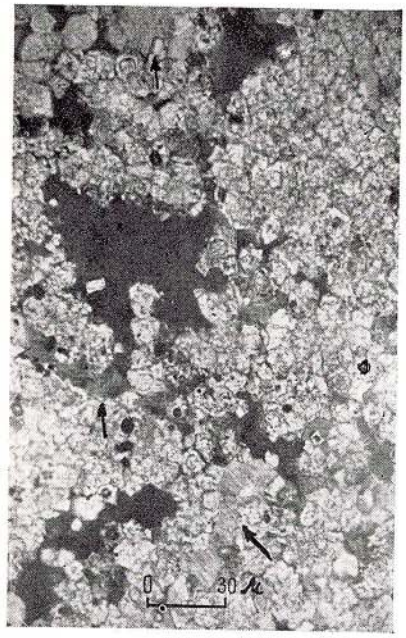
Fig. 4. — Dolomicrospar. In the lower right corner of the photo one can notice the cryptocrystalline material between the dolomicrosparitic crystals. It has been pushed aside by the recrystallization of dolomicrite to dolomicrospar. N //. Leghia quarry 3621/12.

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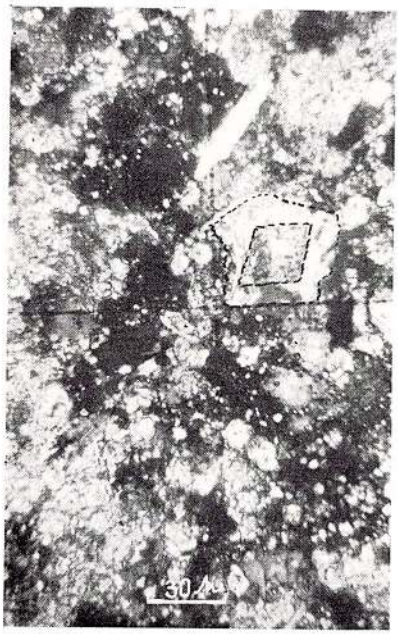
B. Popescu. Lower Gypsum Formation West of Cluj-Napoca. Pl. XI.



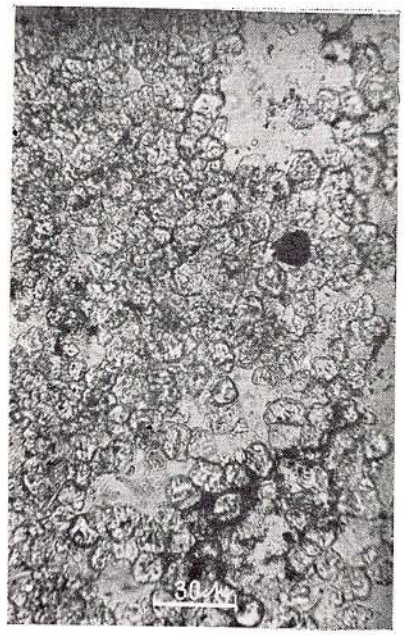
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PLATE XII

Fig. 1. — Dolomicrospar largerly cemented by gypsum. The dolomicrosparite crystals ,,float'' within gypsum cement. N //. Leghia quarry 3621/14.

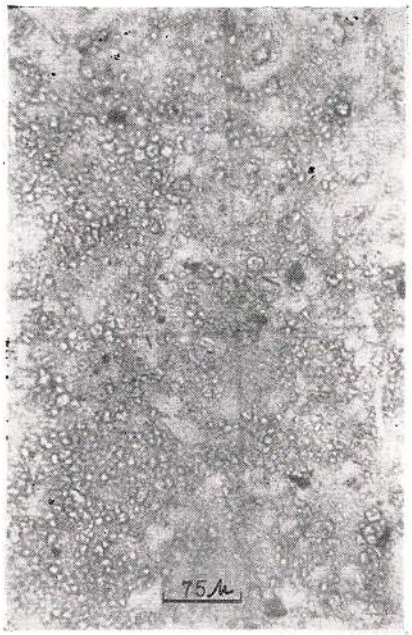
Fig. 2. — The transition from dolomicrite into porous dolomicrosparite. N +. Foidaş quarry 4177/13.

Fig. 3. — Undissolved dolomicritic pelletal remnant within a highly porous dolomicrosparite. N +. Foidaş quarry 4177/21.

Fig. 4. — The boundary between fine laminated dolomicrosparite and not laminated dolomicrosparite. The laminae are marked by the cryptocrystalline mixture of clay, ferrous oxides and organic matter. Within dolomicrosparite the cryptocrystalline material is only as spots and patches. N//. Leghia quarry 3621/12, interval B in Fig. 1 Pl. II.

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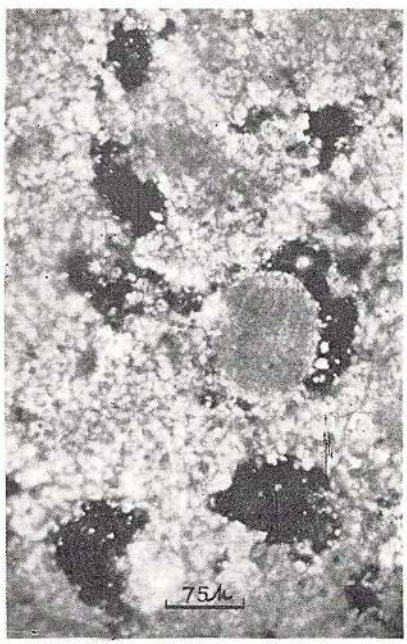
B. Popescu. Lower Gypsum Formation West of Cluj-Napoca. Pl. XII.



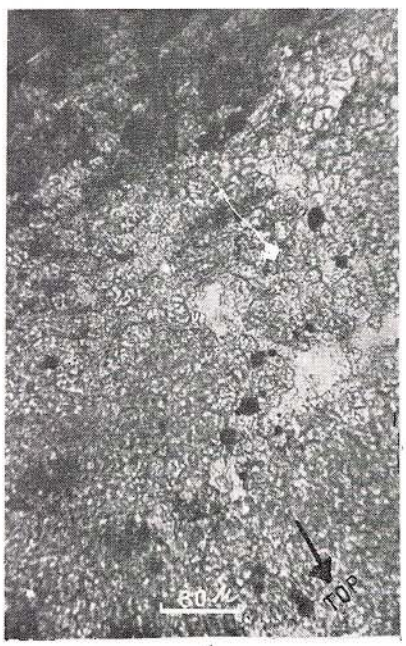
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PLATE XIII

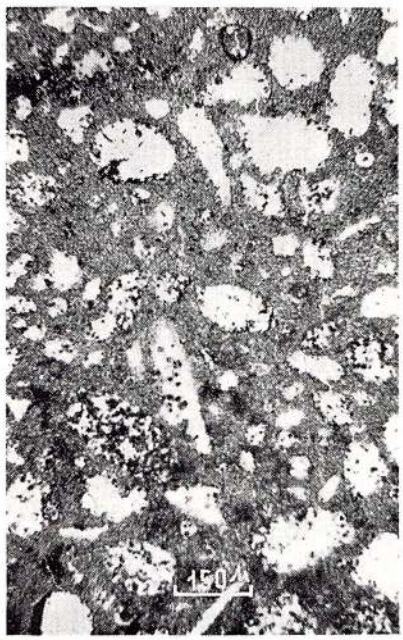
Fig. 1. — Pelletal bioclastic dolomicrospar. The leached pellets and bioclasts are filled with gypsum. Note the high number of dolomite crystals with cryptocrystalline nuclei scattered within gypsum filled pellets and bioclasts. N//. Leghia quarry 3621/9, interval B in Fig. 1, Pl. V.

Fig. 2. — Same as Fig. 1. N +.

Fig. 3. — Partly leached pellet. The vug has been filled with gypsum. An undissolved remnant is trapped within the gypsum filling. Note the cryptocrystalline envelope of this gypsum filled pellet. N//. Foidaș quarry 4177/21.

Fig. 4. — Same as Fig. 3. N +.

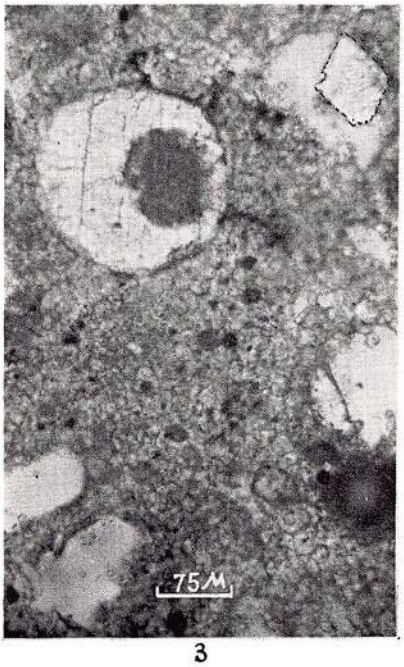
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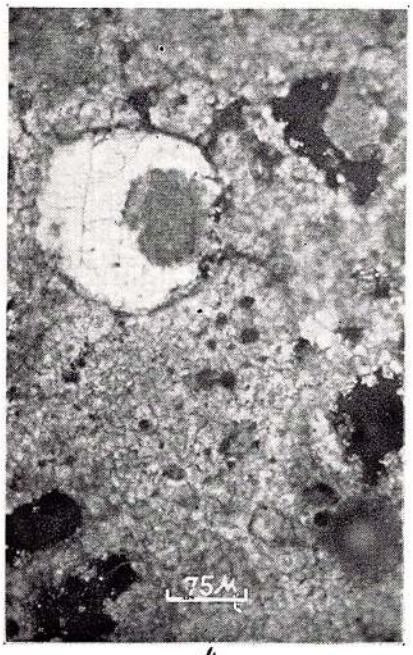
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PLATE XIV

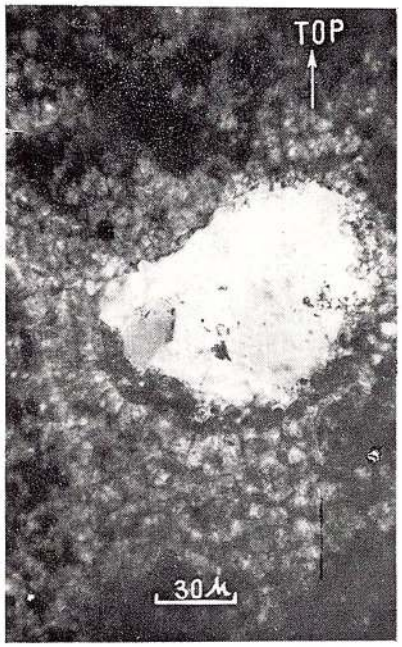
Fig. 1. — Dolomicrosparite with gypsum filled pore. Note the cryptocrystalline „floor" of the cavity and the same nature of this rim and of cryptocrystalline patches around them. N//. Foidaş quarry 4177/24.

Fig. 2. — Clear dolomicrosparite with gypsum filled pellet. The wall of the leached pellet is imbued with cryptocrystalline material. N//. Leghia quarry 3621/11.

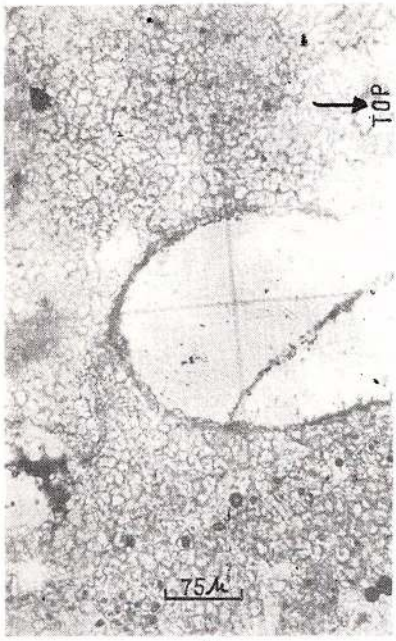
Fig. 3. — Tiny dolomicrosparitic streaks within nodular gypsum. N//. Foidaş quarry 4177/23.

Fig. 4. — Under streaky nodular gypsum there are a thin pelletal bioclastic dolomicrospar above pelletal dolomicrite. Within the pelletal dolomicrite the pellets are leached, some of them being filled with gypsum. There is any relation between the recrystallization of the dolomicrite and the gypsum filling. N //. Leghia quarry 3621/9 interval B in Fig. 1, Pl. V.

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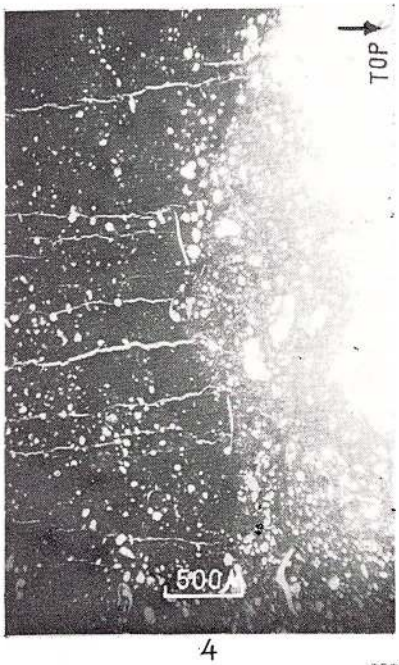
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PLATE XV

Fig. 1. — Gypsum nodule growth within dolomicrosparite. Note the loose boundary between gypsum and dolomicrospar and the streaks of dolomicrospar left within the gypsum nodule. N+. Leghia quarry 3621/9, interval D in Fig. 1, Pl. V.

Fig. 2. — Not ,,digested'' isolated crystals of dolomicrospar within gypsum nodules. N //. Leghia quarry 3621/9, interval A in Fig. 1, Pl. V.

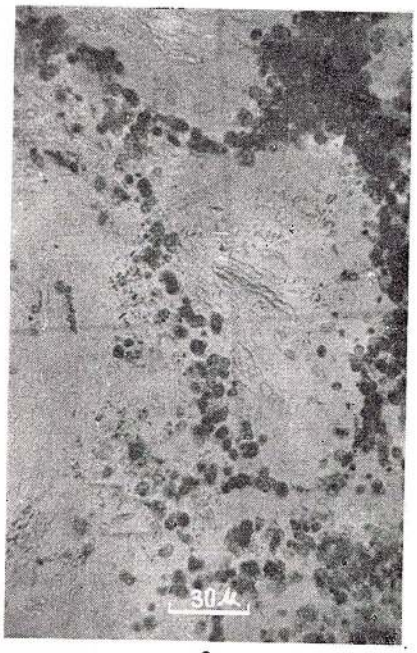
Fig. 3. — Blocky texture with larger partely euhedral anhydrite crystals (arrows). N+. Foidaș quarry 4177/18.

Fig. 4. — Felted texture. N+. Foidaș quarry 4177/15.

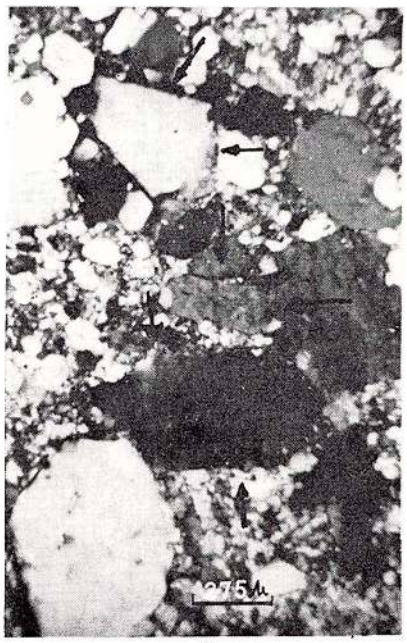
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PLATE XVI

Fig. 1. — Gypsum filled pore with larger euhedral dolorhombs. Some of them have cryptocrystalline relics. N//. Leghia quarry 3621/12, interval B in Fig. 1, Pl. II.

Fig. 2. — Around the pores there are clear euhedral dolorhombs. N//. Foidaș quarry 4177/13.

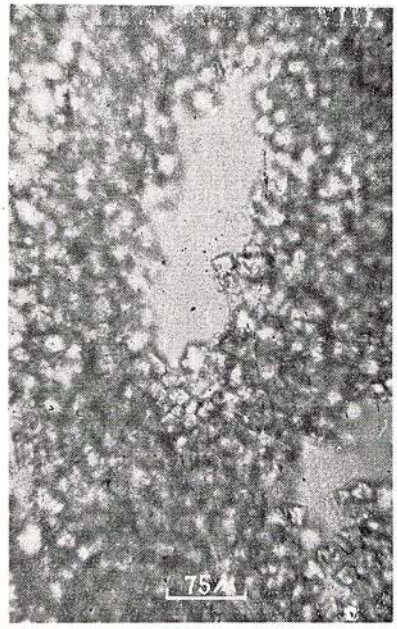
Fig. 3. — Pellets subsequently filled by gypsum and larger euhedral dolorhombs which „float" within the gypsum filling. N//. Foidaș quarry 4177/28.

Fig. 4. — Same as Fig. 2. N+.

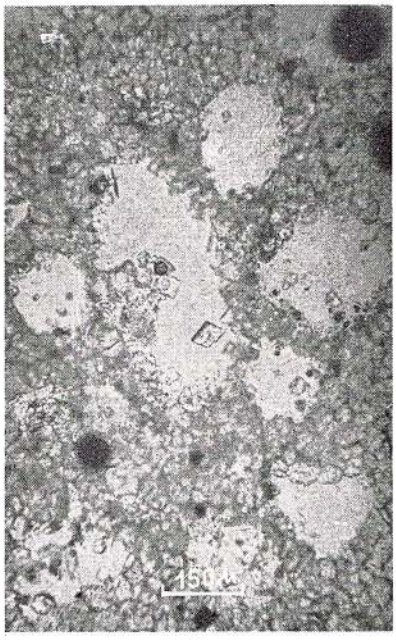
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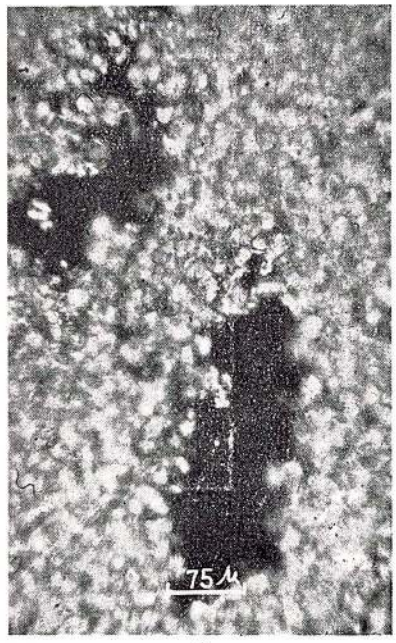
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PLATE XVII

Fig. 1, — Oölitic grainstone. Note the shelter effect of skeletal fragment and the sparry calcite cement beneath them N//. Viilor valley.

Fig. 2. — Oölitic grainstone. Note the loose packing beneath the shell sheltering. The pores are partly infilled with drusy calcite. N +. Viilor valley.

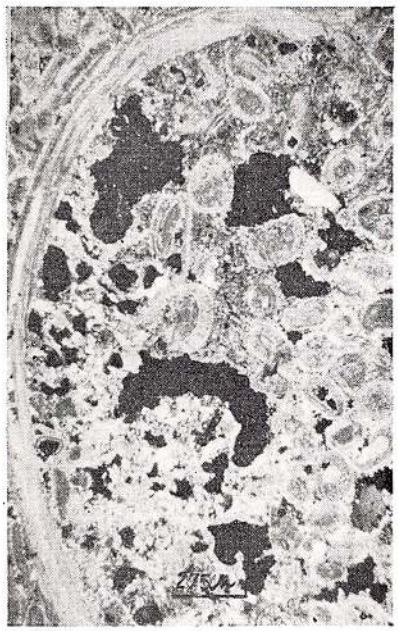
Fig. 3. — Pseudospar. Algal micrite with gradational contacts between micrite and pseudospar. N//. Foidaş quarry 4177/9.

Fig. 4. — Sparry calcite cement. Note the fibrous calcite perpendicular to the surface of the micritic wall (upper right corner) and the increase in grain size outward them. N//. Foidaş quarry 4177/9.

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SEDIMENTOLOGY OF PRIABONIAN CARBONATE ROCKS, JIBOU AREA, N—W TRANSYLVANIAN BASIN [1]

BY

BOGDAN POPESCU [2]

Abstract

The study of the field relations, paleontology and petrography of the Priabonian limestones which lie in Jibou area has lead to a reconstruction of the environment of deposition of the Turbuţa, Cluj, Brebi, Cozla, and Hoia Formations. Petrographic examination of polished slabs, thin sections as well as of X-ray and chemical data resulted in a grouping of carbonate rocks into environmentally controlled facies. The main components of the studied carbonates were grouped in : skeletal, nonskeletal, matrix and cement. The pelletal and oölitic grainstones or packstones, stromatolitic boundstones, gypsum-bearing deposits of the Turbuţa Formation suggest an inner sheltered lagoon and supratidal evaporite environments within the Ileanda-Turbuţa-Zalău embayment. North and south of this zone Turbuţa Formation deposits grade into continental-lacustrine deposits. With the Cluj and Cozla Formations it has been ascertained a clear marine sedimentation on the Transylvanian Paleogene Shelf. The well bedding, planar orientation of shells, high amount of terrigenous grains, of Cluj Formation packstones and grainstones, as well as the thin intercalations of silty and sandy marls show a rather turbulent depositional environment. Following this first transgressive event the studied area become a more stable carbonate platform. It support a great pure carbonate production with coral and coralgal boundstones on borders and with skeletal wackestones and packstones in center, which caracterize the Cozla Formation facies. South of this platform there was a large shelf lagoon with mixed terrigenous and carbonate sediments of the Brebi Marl Formation. Beginning with the Hoia Formation sedimentation in the Transilvanian Paleogene starts a regression which will last all over the Oligocene period. The carbonate facies of the Hoia Formation — oncolitic, skeletal grainstones and packstones — follow from north to south in general manner the actual shape of the Transylvania Basin.

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[2] Institutul de geologie şi geofizică, str. Caransebeş 1, 7000 Bucureşti.

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I. INTRODUCTION

It is rather strange that in about one hundred years of geological researches in the Transylvanian Basin no work on sedimentology of carbonate rocks has been made. Probably, the high attraction caused by the large bodies of Mesozoic limestones which surrounded the Transylvanian Basin could explain the scarcity of sedimentological works on Paleogene limestones. But the study of Tertiary carbonates and evaporites arises many and interesting questions, as we will show later.

The studied area is situated in the hydrographic basin of the Someş river and south of Ţicău — Preluca crystalline massifs.

In order to study the Priabonian carbonate rocks we have investigated and sampled many quarries and representative outcrops (Fig. 1 We have also examinated about 100 polished slabs and 500 thin sections. 25 samples were analysed X-ray diffraction and 15 by chemical methods. Additional data have been provided by the drillings bored after 1950 within the studied area.

In the next sections we will describe the stratigraphy and lithology the facies and depositional environments of carbonate rocks from the Jibou area.

II. STRATIGRAPHY AND LITHOLOGY

During the Eocene time the Transylvanian Basin was an arm of the Tethyan Geosyncline Realm. The Priabonian (Upper Eocene) lithologic succession begins with the Racoţi Sandstone Formation and continues with the Turbuţa Formation, Cluj Formation, Brebi Marl Formation and Hoia Formation[3].

[3] The Hoia Formation was considered as Lower Oligocene (Hofmann, 1879; Koch, 1894; Moisescu, 1967; Rusu, 1972) based on mollusc studies, or Upper Priabonian (Barbu, 1956; Bombiţă, 1963; Moisescu, 1966) based on nummulit, microforaminifer and echinoderm studies. The calcareous nannoplankton (Popescu, Gheţa,

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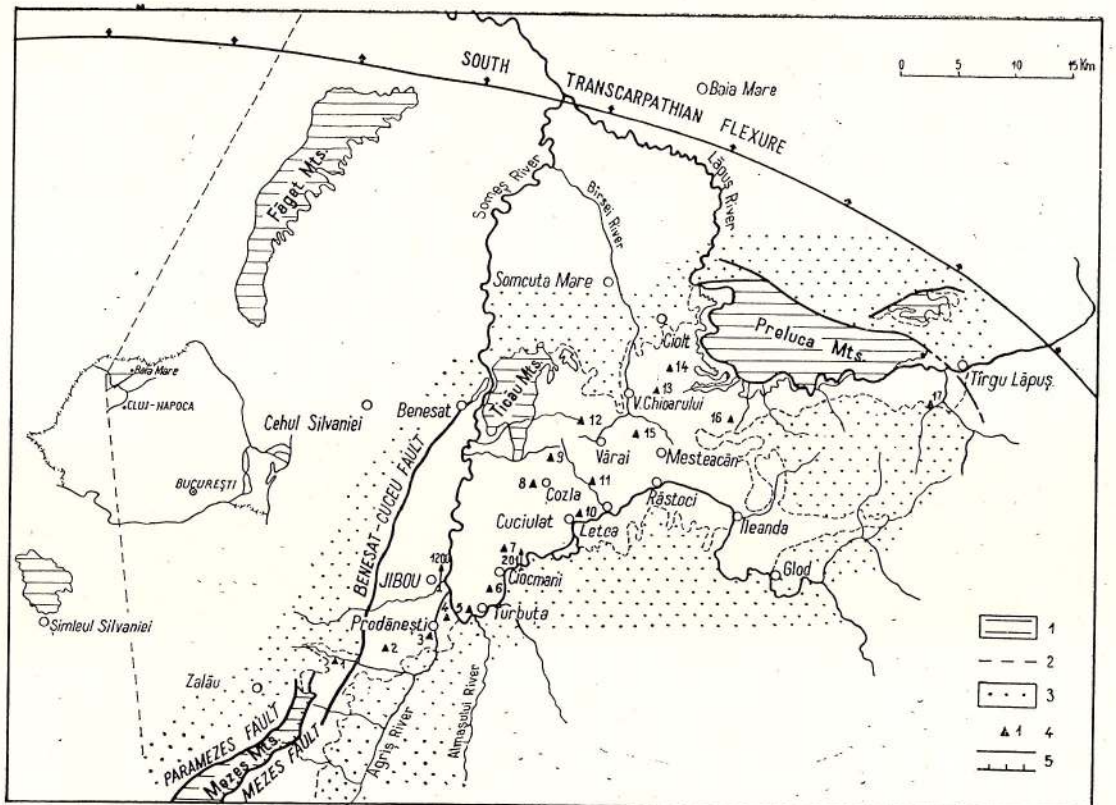


Fig. 1. — Generalized Geological Map of the Jibou Area, NW Transylvanian Basin. 1, crystalline massifs; 2, actual extension of the Paleogene deposits; 3, younger deposits; 4, major outcrops and quarries; 5, major faults. (Neogene in age).

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The Cozla Formation is the eastern facies transition of the Cluj Formation, Brebi Marl Formation and Hoia Formation (Table; Pl. XII).

TABLE

Schematic representation showing the relations between the Priabonian formations from the Mezeș-Jibou area and the adjacent areas

	Cluj Area*	Mezeș-Jibou Area	Țicău-Preluca Area
PRIABONIAN	Hoia Formation	Hoia Formation	Hoia's paleontologic association
	Bryozoan Marl Formation	Brebi Marl Formation	Cozla Formation
	N. fabianii level Cluj Formation *O. transilvanica* level	*N. fabianii* level Cluj Formation *O. transilvanica* level	N. *fabianii* level Marly Calcareous Member
	Upper Gypsum Formation	Upper Gypsum Formation	Upper Red Clay Complex
	Upper Red Clay Complex	Turbuța Formation	
	Leghia Formation	Racoți Formation	Racoți Formation

* The areas are the same as those separated for Oligocene deposits by A. Rusu (1970)

The dating of Upper Eocene formations was made by Bombiță (1963, 1968) who established that there are nummulits of a Priabonian age between the Racoți Sandstone Formation and the Hoia Formation.

A) **Turbuța Formation** (Hofmann, 1879). Overlying the Racoți Sandstone Formation, at the Hofmann's stratotype site (Fig. 2), without any lithological transition there follow few meters of green-blackish marls with molds of *Pitar* and *Tellina* and marine ostracods: *Bayrdia subdeltoidea* Münster, *Cytheretta elegans*, Ducasse, *Krithe bartonensis* Jones (the determinations of ostracods were made by Olteanu). There follow 20 m of dolomicrites, oölitic grainstones and packstones, pelletal packstones, stromatolites, gypsum-bearing marls with frequent intercalations of green marls and clays (Fig. 2, Pl. I and II).

West of Brebi (Fig. 1) above the Racoți Sandstone Formation there follow about 15 m of marly limestones, green marls and clays with thin intercalations of skeletal wackestones.

1972; Mészáros et al., 1973) shows the presence of terminal Priabonian (according with Tethyan Realm chronostratigrafic scale) = Lattorfian (according with the German Basin chronostratigrafic scale) within the upper part of the Bryozoan (=Brebi) Marl Formation and the Hoia Formation.

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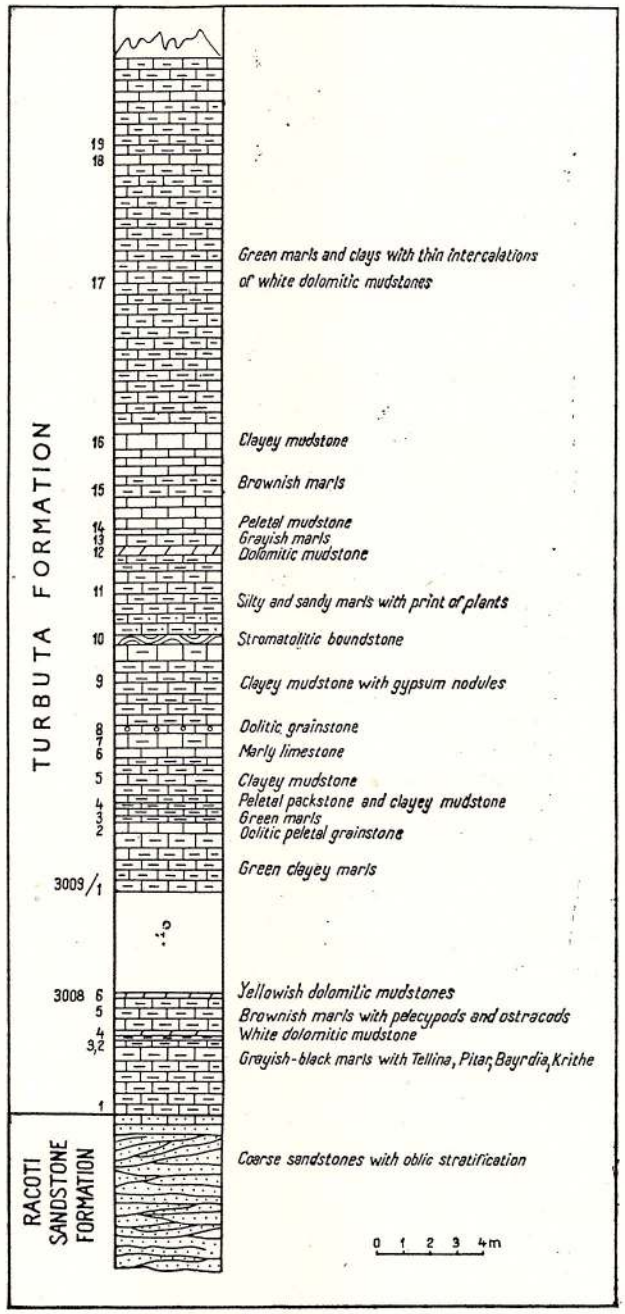


Fig. 2. — Lithostratigraphic column of the lower part of the Turbuţa Formation.

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Above this carbonatic succession there follows, on large areas, the typical lithology of the Turbuța Formation : green clays and marls with rare tiny intercalations of white dolomicrites and nodular gypsum lens.

On the upper part of the Turbuța Formation (north of Prodănești Fig. 1) there may be found thin beds of pelletal packstones and pelletal oölitic grainstones (Pl. I, Fig. 2).

North of Poeniţa parallel (Fig. 3,4), the Turbuța Formation becomes more and more terrigenous being constituted of red clays pebbles, silts and sands with a red clay matrix of a continental origin (= Upper Red Clay Complex).

South and west of the Someș river, the terminal part of the Turbuța Formation passes laterally to the Upper Gypsum Formation (as named within the Cluj zone). This evaporite formation exhibits the same sedimentological features as the Lower Gypsum Formation — supratidal deposit — from the Cluj zone (P o p e s c u , this volume).

The lithologic and paleontologic data clearly label the Turbuța Formation, at least on the stratotype site, as a marine deposit (from brackish marine to normal — and hypersaline marine deposits) and not of a continental origin as statuted by the foregoing authors.

B) **Cluj Formation, Berbi Marl Formation and Cozla Formation.** As within the studied area there is the interfingering of the Cluj Formation + Brebi Marl Formation facies with the Cozla Formation facies, we shall describe the lithology of these formations in the same section, starting from SW to NE, direction after the marly facies of the Brebi Marl Formation is replaced by the calcareous facies of the Cozla Formation.

1. *Cluj Formation* (H o f m a n n , 1879) or „Upper Coarse Limestone" (K o c h , 1894) is developed between the Parameseș Fault and the Poeniţa village (Fig. 1,5). Our studies show that the typical facies of the Cluj Formation near Cluj (the designated site of the stratotype) is to be found only in the lower part of the Cluj Formation within the Jibou area. Thus, H o f m a n n improperly named this lithologic unit, conferring it the features of the limestones within the Cluj zone. Their only unifying touch is the limits, both being placed between the Upper Gypsum Formation (the Turbuța Formation, respectively), and *Nummulites fabianii* paleontological level.

At the lowermost part of this formation the outcrops are scarce and incomplete. The vertical transition between the Turbuța Formation and the Cluj Formation is made by a thin alternance of skeletal grainstones and packstones (like those of „real" Cluj Formation), and sandy marls with *Anomia*, *Rimella*, *Vulsella*, *Terebellum*. There, follow the *Ostrea transsilvanica*-bearing marls, paleontologic and lithologic level very constant overall Transylvanian Basin. It disappears north and east of the Cozla village.

Within all the studied area we can see, above this level, few meters of skeletal packstones " type Cluj Formation" and then a succession of corallinacean miliolitic wackestones and packstones and corallinacean

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molluscan foraminiferal wackestones and packstones. Within this succession there are to be found, at different levels, coralgal — and/or coral boundstones of very variable thickness and length (Pl. III, Fig. 2).

At the top of the Cluj Formation there is the *Nummulites fabianii* level, marker level in Transylvanian Basin (Pl. IV, Fig. 4).

The paleontologic association from the Cluj Formation is dominated by coralline algae (*Lithophyllum, Corallina, Jania*), molluscs (*Anomia, Vulsella, Ostrea, Crasatella, Chlamys, Pleurotomaria, Campanile, Hippochrenes, Globularia, Rimella, Terebellum*), foraminifers (miliolids, nummulits, rotalids, orbitoids, *Halkyardia, Neoalveolina*), echinoids (*Leiopedina, Eupatagus, Echinolampas, Schizaster*), corals, worm tubes, etc.).

2. *Brebi Marl Formation* (H a u e r , S t a c h e , 1863) is developed within the same area as the Cluj Formation. It has a relative uniform lithology : yellowish white marls with rare intercalations of marly limestones and molluscan coquinas. The top of this formation is more lithified individualizing, west of the Someş river, thin clayey foraminiferal mudstones and wackestones. The thickness of the Brebi Marl Formation diminishes eastward and disappears at the Ciocmani-Poenița meridian.

The organic community is very rich both as genera and species. Foraminifers (miliolids, nummulits, globigerinids), ostracods, molluscs (*Chlamys, Spondylus, Pitar, Thracia, Panope, Corbula, Globularia, Cepatia, Ficus*), bryozoans, echinoids, are the main organic rests. Within the upper part of this formation there was determinated (by G h e ț a) a calcareous nannoplankton from *Istmolithus recurvus* and *Ericsonia? subdisticha* zones — belonging to the terminal Priabonian.

3. *Cozla Formation* is the main carbonatic formation from the studied area. Formely it was equalized with the Cluj Formation (H o f m a n n , 1879; K o c h , 1894) then with the Cluj Formation and the Brebi Marl Formation and named „Reefal Calcareous Series" (R ă i l e a n u and S a u l e a , 1956) or more simply "Calcareous Series" (D u m i t r e s c u , 1957). R ă i l e a n u and S a u l e a (1956) gave also the parallel name of "Culmea Cozlei Limestone". This latter nomenclature is adopted with a slight modification in this paper. It has the advantage to be in accord with the recommendations of the International Commission for Geological Nomenclature (1973) and it also eliminates the previous equivocal names.

This formation lies between the Țicău and Preluca crystalline massifs and north of the Someș river. The main facies are skeletal mud supported or bioconstructed. Depending on the amount of participation of skeletal grains, there are to be found : corallinacean miliolitic, corallinacean nummulitic, coral corallinacean, corallinacean molluscan wackestones or packstones. The boundstones are coraligenous or coralgal. The mounds are massive, thick and medium undulant bedded. The true grainstones and mudstones (1, 4, 5), are practically lacking within the Cozla Formation (Pl. III, Fig. 1, 6 ; Pl. IV, Fig. 1 ; Pl. V, Fig. 1, 4, 5).

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The lowermost part of this formation, above the Upper Red Clay Complex from the Ţicău-Preluca zone is an alternance of silty or sandy marls with thin corallinacean miliolitic packstones like those of typical Cluj Formation (Pl. IV, Fig. 1,3). It passes vertically to the skeletal packstones or wackestones with general appearance of the Cozla Formation. This is the Marly Calcareous Member of the Cozla Formation [4].

In the northern part (Chioarul Valley, Fig. 1) there may be also found scattered bioconstructed mounds, but the main mound area is situated in front of the Brebi Marl Formation (Fig. 5).

The top of the Cozla Formation includes the Hoia Formation too, which generally mimics the above described facies. It may be recognized only by their paleontological facies bearing some particular species of the: *Scutella*, *Cyrena*, *Ampulinopsis*, *Tympanotonos*, *Nummulites* as well as balanids, crab chelicerae.

The paleontological association of the Cozla Formation is not so diverse as those of the Brebi Marl Formation or of the Cluj Formation. The corallines (*Lithophillum*, *Mesophillum*, *Corallina*, *Amphipora*, *Jania*), foraminifers (miliolids, rotalids, nummulits, *Orbitolites*, *Chapmanina*, *Halkyardia*, *Eorupertia*) molluscs (*Pecten*, *Chlamys*, *Terebellum*, *Casidaria*) echinoids (*Echinolampas*, *Leiopedina*, *Laganum*, *Scutella*), corals (*Rhabdophillia*, *Pattalophillia*, *Hidnophillia*), crabs, are the main organic contributors of the skeletal fraction of these limestones.

C) **Hoia Formation** (K o c h , 1880). The typical development of the Hoia Formation is related to those of the Brebi Marl Formation. It is built up of an alternance of corallinacean nummulitic packstones or grainstones, molluscan grainstones, corallinacean coral packstones and marls. The total thickness varies from 4 to 8 m. (Pl. IX and X).

Within the Cozla Formation facies area, the Hoia Formation may be distinguished from the subjacent limestones by its well packed skeletal limestones and, as we said above, by its renewed paleontological association. This latter feature has decided many authors to introduce them within Oligocene parallelizing with North Paleogene basins (German, Anglo-Belgian-Parisian). This point of view is not in accordance with the geological situation, the Transylvanian Basin belonging in that time to the Upper Eocene of Tethyan geosyncline realm.

The nannoplankton already reported from the Brebi Marl Formation, correlated to that from the Hoia Formation near Cluj and with nummulits (*N. fabianii* P r e v e r , *N. budensis* H a n t k e n , *N. incrassatus* de L a H a r p e (B o m b i ţ ă , 1963) admit the Hoia Formation within the Priabonian stage.

The Hoia Formation is overlain by the Curtuiuş Formation coal bearing green and black clays and/or by the Ciocmani Formation — brackish and marine deposits of an Oligocene age.

[4] B. P o p e s c u. Stratigraphical Studies in NW Transylvanian Paleogene. 1971. Arch. of Inst. of Geol. and Geophys.

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III. SEDIMENTOLOGY

From the above description of the stratigraphic and lithologic succession it was revealed that whereas the Racoţi Sandstone Formation is detritic, the rest of the Priabonian deposits are carbonates, evaporites or mixed terrigenous carbonates.

The carbonate rocks of the Jibou area are predominantly biogenic, only the Turbuţa Formation being an admixture of terrigenous biogenic and nonbiogenic carbonates with clays, marls and evaporites.

A) **Main Petrographic Components.** In this section we will describe the two main categories of components differentiated during grain counting : the skeletal and nonskeletal grains to which one may add the third category of matrix and cement.

1. *Skeletal grains*

a) Coralline algae are the most common skeletal component of the carbonate rocks of the studied area. By far the Melobesioidae (crustose corallines)[5] are more abundant than the Corallinoidae (articulate coralline).

The crustose corallines show a great variety of the growth forms. Some are phylloid or well rounded individual grains but the majority form crusts : smoth, nodular, branched, undulate (Pl. VI, Fig. 2, 3, 4). They may be attached to molluscs, corals or may grow directly on the unlithified carbonate substrate.

The articulate corallines are rather scarce and were found mainly as fragments. We have not found yet remains of Dasycladaceae. Other green algae are suspected to have existed within depositional environment because sometimes there have been observed prints of carbonized seagrass debris in the Cozla Formation limestones. The great influence of seagrasses is indicated also by the high average content of matrix in all carbonate facies of the Cozla Formation. The carbonate mud can be trapped and stabilized by seagrass leaf baffle, the rizom and root mesh (Davies, 1970).

As it is well known the corallines usually flourish in shallow and very shallow marine sunlit waters where they were sediment suppliers and stabilizers. The coralline algae differ from other calcareous algae in that the $CaCO_3$ is secreted within and between the cell walls and around the plant tissue. The test mineralogy was initially Mg-calcite.

The corallines found in the Priabonian limestones of the studied area are usually partial or total micritized. Thus we may follow all the stages from a „fresh" to an „algal pellet" (Pl. XI, Fig. 4, 5). The conceptacles are as a rule filled with orthospar (Pl. VI, Fig. 2). The micri-

[5] In this paper we have used Johnson's (1961) classification of *Rhodophita*.

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tization of carbonate particles as well as the intragranular orthospar are fenomena related to the precipitation of calcium carbonate from shallow warm seawater (Bathurst, 1966; Winland, 1968; Alexandersson, 1972 a, b).

We cannot find a relation between the corallines habitat and the carbonate facies. They are spread overall carbonate area, in all facies. It is probable that most of corallines were transported from one facies to another. It is to be mentioned that corallines as binders of mound material or as building organisms were found generally within reef mound area.

b) Foraminifers are the second in importance after corallines as skeletal grain contributors. The main categories taken into discussion, in order of decreasing importance are: miliolids, macroforaminifers, planktonics.

Like corallines, the miliolids are present in all carbonate facies, sometimes being the single skeletal component (Pl. V, Fig. 3). In closely touch with miliolids there are other benthic foraminifers: *Cibicides*, *Rotalia*, *Pyrgo*, *Nonion*. *Cibicides* is a herbivorous foraminifer, the other benthonics being phytoplankton eaters.

These foraminifers have usually a porcellanous micritic wall structure. The test mineralogy was initially Mg-calcite or, less probable aragonite. The chambers are filled completelly or fringed with orthospar. When they are more or less transported the chambers are filled with a mixture of micrite and limonite from depocenter. They often exhibit a limonitic envelope, too. If the burial was rather rapid, the organic matter decay could create a reducing microenvironments and consequentelly the chambers would be filled with pyrite.

The macroforaminifers are common in all Priabonian sequences and very abundant at some levels (e.g.: *Nummulites fabianii* level and within the Hoia Formation — Pl. IV, Fig. 2, 4; Pl. X, Fig. 6). The main species is *N. fabianii* Prever and his transitional form to *N. intermedius*. Within the Cluj Formation there also appear *Borelis*, *Orbitolites* and within the Cozla Formation, *Chapmanina*, *Halkyardia*, *Orbitolites*, *Eorupertia*. These macroforaminifers are usually known as stenohaline forms. The test mineralogy was initially calcite (nummulits, *Eorupertia*, *Halkyardia*) or Mg-calcite (*Orbitolites*, *Chapmanina*, *Borelis*). The later have a micritic wall, others hialine. The calcitic forms were bored and then filled with micritic cement, finally forming a micrite envelope like those described by Bathurst (1966), Winland (1968), Friedman et al. (1971), Alexandersson (1972 b).

Among the planktonics, the globigerinids are the most important. They were found within the Brebi Marl Formation and within intraseptal micrites of some corals. The hialine wall, initially calcitic, is unaffected by recrystallization. The chambers are filled with micrite (within coraligenous facies) or with pyrite (within the Brebi Marl Formation).

c) Molluscs are very variated as genera as well as species. Generally speaking, we may notice a positive correlation between molluscs and bioaccumulated facies and negative with bioconstructed facies. Within coraligenous mounds and reef derived material we have found only

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species of a lithophag pelecypod (Pl. III, Fig. 3; Pl. VI, Fig. 1). In this respect they are more abundant in typical facies of the Cluj Formation and less frequent within the Cozla Formation.

Many of the molluscs show various degrees of fragmentation, abrasion, boring and wear (Pl. IX, Fig. 2, 4; Pl. X, Fig. 4, 5). They have a micritic envelope or a limonitic coat. Some of them (those with an initial aragonite or Mg-calcite shell) are completely recrystallized, others (those with an initial calcitic shell) remain unaffected.

d) Echinoderms are present mainly as spines and fragments of echinoids. The whole forms belonging to the *Echinolampas*, *Leiopedina*, *Laganum* are very scanty if they are reported to the great number of fragments.

The echinodermal grains have as a rule micritic material within their fenestral pore pattern. Some have also micritic envelopes and boring tubes filled with micrite. The grains without micritic envelopes are commonly enclosed by a syntaxial overgrowth rim (Pl. VII, Fig. 5).

e) Bryozoans are rather abundant in the Brebi Marl Formation and subordinated in all Priabonian limestones. Braga and Ghiurca (1969) have reported 133 species from the Transylvanian Priabonian. The zoarium has seldom a micritic envelope. The zoarium wall, probably initial calcitic, is unaffected by recrystallization. The zooecia are usually filled with orthospar.

f) Corals are found to form buildups or as isolated individuals within the Cluj and Cozla Formations. Many of them are encrusted by corallines or distructed by mechanical and/or biogenic action of the depositional environment (Pl. III, Fig. 3; Pl. VI, Fig. 5). The septae are invariably recrystallized like all initial aragonite skeletons. Within the intraseptal space it is to be found dark cryptocrystalline micrite or foraminiferal bearing micrite. These micrites have commonly scattered dolorhombs with a mean size of 20 μ.

2. *Nonskeletal grains*

The nonskeletal grains are not so widespread as the skeletal ones. They are related to the intertidal and supratidal facies of the Turbuţa Formation and less to the Hoia Formation. In the Cluj and Cozla Formations they are practically absent. Pellets, lumps oölites and oncolites as well as terrigenous grains will be described in this section.

a) Pellets are rather an uncommon component and have been found in the Turbuţa Formation and with less certainity in the Cozla Formation. They are fecal pellets and show two kinds of forms: some are cylindrical with long axis of 500—750 μ and the diameter of 100 μ, like Bahamian pellets (Illing, 1954; Cloud, 1962) others are ovoidal or subrounded with long axis of 250—300 μ and the diameter of 100 μ (Pl. II, Fig. 3; Pl. I, Fig. 2). Some have a tiny (superficial—Illing, 1954) or true oölitic coat. The pellets show an internal structure with boring tubes (2—5 μ in diameter) filled with sparry calcite (Pl. II, Fig. 3).

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Other „pellets" are of an algal origin and result from micritization of coralline algae, some from miliolids or other skeletal grains, as we have said in the previous sections. These may be tabulated as „pellets" (or more proper — peloids) or as cryptocrystalline grains (P u r d y, 1963) depending on the possibility of recognizing initial structure.

b) L u m p s were observed only in one thin section associated with fecal pellets in the Turbuţa Formation. They are irregularly shaped, with maximum diameter of 2—5 mm showing a great number of small patches of microspar.

c) O ö l i t e s were found only in the Turbuţa Formation when they formed thin mud-supported beds with oblique lamination. The nucleus is a pellet, quartz or skeletal grain. The cortex rarely exceeds 100—150 μ in thickness. The oölites have usually an elongate shape varying in size from 200 μ to 1,5 mm. The lamellae of clear radial calcite alternate with lamellae of brown cryptocrystalline matter. Asymmetrical oölits with the cortex unevenly distributed forming bosses like the quiet water oölits from Laguna Madre (F r e e m a n, 1962) were observed in the oölitic and peletal facies from the Turbuţa Formation (Pl. II, Fig. 2).

d) O n c o l i t e s like all nonskeletal grains from the Priabonian deposits of the Jibou area are scanty. They appear only in one of the facies of the Hoia Formation (Pl. XI, Fig. 1). The oncolites are sphaeroidal or ovoidal in shape with dimension range from 150 μ to 8 mm. Layering is of concentric brownish laminae (Type SS—C of L o g a n et al. 1964). The most common nuclei are nummulits, mollusc shells or fragments, lumps (Pl. XI, Fig. 3). Many of them have no identificable nucleus.

e) T e r r i g e n o u s g r a i n s are mainly quartz, feldspar and mica. Quartz grains are angular to subangular. Feldspars and micas are uncommon terrigenous grains being recorded only in few samples within the Cozla and Hoia Formations near Chioarul Valley village. The content of terrigenous grains, especially quartz, is greater in the Turbuţa Formation and in the Marly Calcareous Member of the Cozla Formation. Generally this content is under 5% in the Cluj Formation and under 3% in the Cozla Formation.

3. *Matrix and Cement*

The term matrix is used in this paper for clay size mass less than 4 μ in median size, which form the fabric of all mud-supported Priabonian limestones. The carbonate mud which has been the matrix of all above described grains is one of the most important petrographic constituent in studied limestones. Theoretically, it may be formed either by physico-chemical precipitation from seawater, by break-down of carbonate skeletons, or allohtonous by transport from other sites.

Excluding the matrix associated with oölitic, pelletal and stromatolitic levels from the Turbuţa Formation, which may content some precipitated carbonate, there is no proof for a precipitated matrix in the rest of calcareous formations from the studied area. As we will show in the next section, the long distance transported carbonate mud is hardly acceptable being the paleogeographic situation. In opposition there are

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some arguments for in situ production. Thus, the repartition of carbonate facies from the Cluj and Cozla Formations suggests us that between the Ţicău and Preluca massifs there has existed a barrier platform which was swept by sluggish currents. The presence of a great number of: skeletal grains with borings and micritic envelopes; tiny skeletons of foraminifers; echinoid and other vagrant benthic organisms which were important mud ingestors; the intact condition of majority of mollusc shells; the minor quantity of terrigenous grains supports the in situ production of the matrix by quiet water — biological reduction of grain sizes and less by the mechanical break-down.

The matrix suffers the majority of the diagenetic processes which extends from penecontemporaneous to the late diagenetic stages, e.g.: the aggrading neomorphism, the dolomitization, the dedolomitization, the stylolitization, the compaction, the grain fracturing and weald (Pl. VIII and IX).

The cement has been considered only the orthospar precipitated within hollow spaces of skeletons, intrareefal voids, interparticle pores. The microspar and pseudospar were considered as a result of the neomorphic process as Folk (1965) statuted.

B) **Facies and Depositional Environment.** In order to decipher the history of carbonate sedimentation in the Jibou area the next step will be the description of the facies categories and their rock type correspondents. For designating the rock types we will follow the terminology proposed by Dunham (1962) to which there were added the names for grain kind. Within the Priabonian sequence of limestones we have found all textural classes reported by Dunham. The boundaries used in classification are for all that arbitrary, because the relations among all compositional parameters are gradational. For example, within the mud-supported fabrics there are to be found all transitions from mudstones and wackestones to the grain-supported fabric of the packstones.

Within the Turbuţa Formation, the lowermost formation taken into discussion in this paper, there is an alternance of mud-supported and grain-supported fabrics (Fig. 2) overlain by marly-clayey deposits and finally by gypsum-bearing facies.

The pelletal packstones, pelletal oölitic grainstones, skeletal wackestones and stromatolites from the lower part of the Turbuţa Formation are interbedded with thin white dolomitic mudstones and clays and gypsum nodules. This sequence of deposits is common in hot, semiarid regions with a high rate of evaporation within protected and moderately exposed environments (Illing et al., 1965; Thompson, 1968; Kinsman, 1966; Butler, 1969; Kendall, Skipwith, 1968).

The excellent preservation of internal structure and shape of pellets, irregular form of many oölites and high amount of matrix, suggest us that west of the Turbuţa village (Fig. 3) there was an inner quiet and sheltered lagoon of the Priabonian sea. Sometimes, it could be moderately exposed

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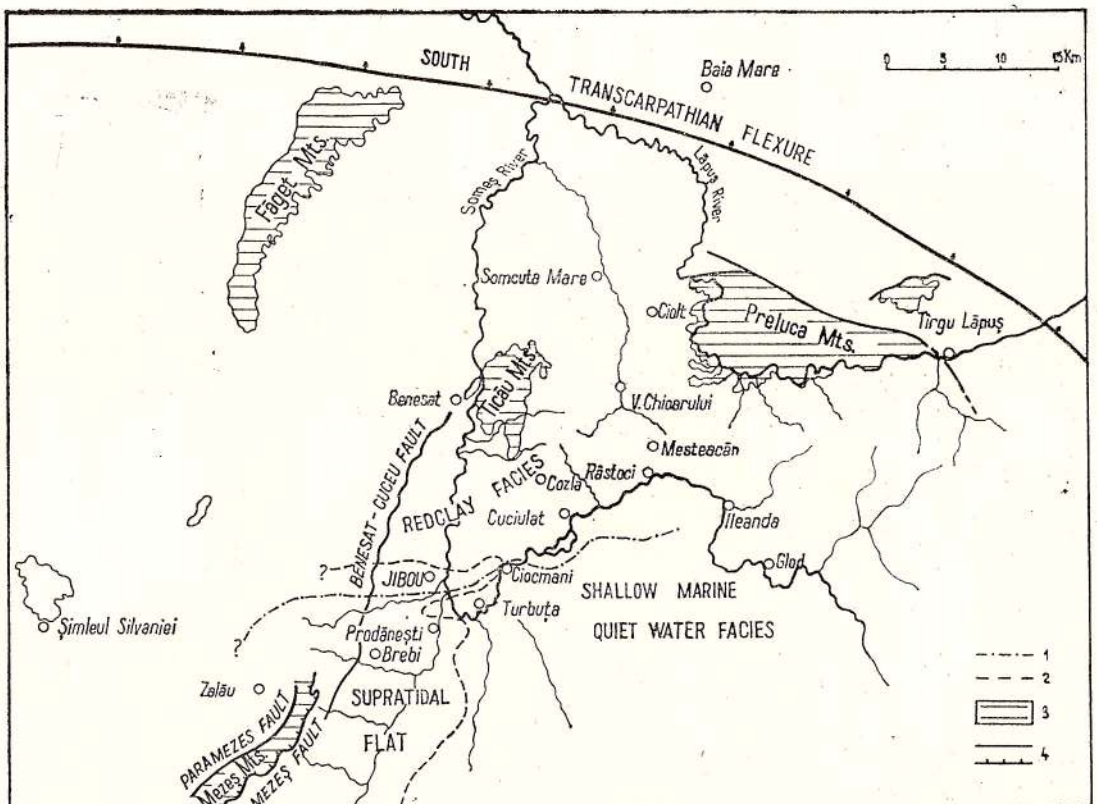


Fig. 3. — Facies distribution in the Turbuţa Formation.
1, northern boundary of the Zalău-Jibou-Ileanda embayment with quiet water facies; 2, the supratidal evaporitic area of the upper gypsum formation (upper Turbuţa Formation); 3, crystalline massifs; 4, major faults.

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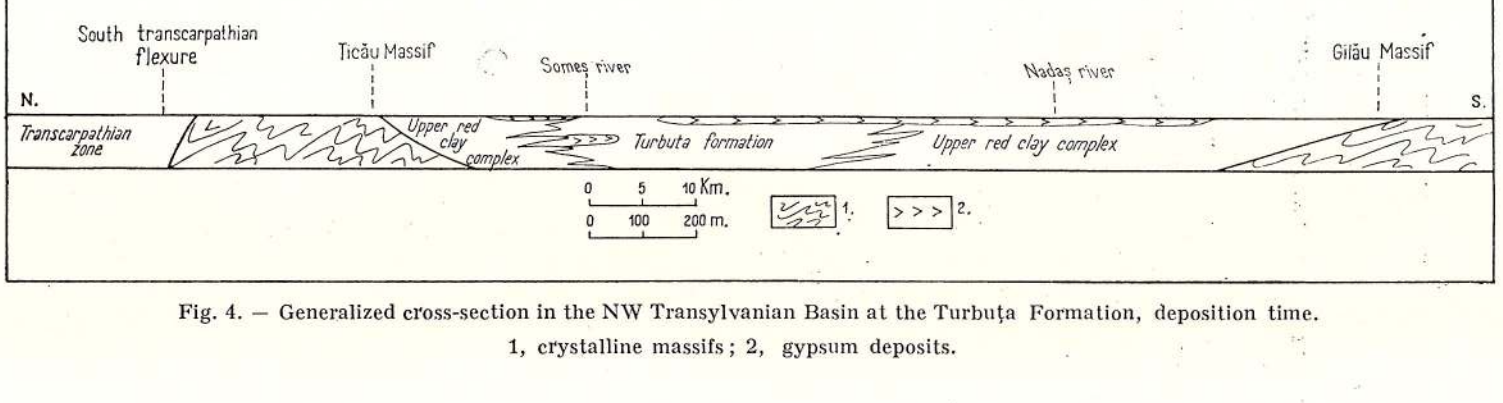


Fig. 4. — Generalized cross-section in the NW Transylvanian Basin at the Turbuța Formation, deposition time.
1, crystalline massifs; 2, gypsum deposits.

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to stronger action of environment, resulting the oölitic and oölitic pelletal grainstones.

After this intertidal and shallow subtidal lagoonal episod, the terrigenous influxes become richer and consequently the lagoonal carbonate facies shift westward.

The Turbuţa Formation ends with more or less thick supratidal gypsum deposits. They are lacking at the Turbuţa village parallel where there appear only the basal skeletal wackestones and dolomitic mudstones with *Anomia* (Pl. I, Fig. 6 ; Pl. II, Fig. 5). Northward the Turbuţa facies grades laterally to silty red and green clays with sands, pebbles and conglomerates of a lacustrine-continental origin. Southward, within the Cluj area we may find the same transition.

Thus, it may be supposed that during the Turbuţa Formation deposition, there was an approximativelly est-west embayment (on Ileanda-Turbuţa-Zalău direction (Fig. 3) with shifting shallow subtidal carbonate facies.

Therefore, the predominant fine low energy deposits of the Turbuţa Formation follow above the coarse high energy deposits of the Racoţi Sandstone Formation. Accordingly , we may consider that the Paleogene Shelf gradually rose after the Racoţi Sandstone Formation. The raising was differentiated being more rapid northward and southward from the Ileanda-Zalău embayment. This fact permits the development of northern and southern large zones with red continental deposits. While north of the Ileanda-Zalău embayment the raising was going on with the same speed and evaporite deposits were lacking, southward the speed of the raising was weacker allowing the appearance of the supratidal evaporitic conditions.

Within the time interval that succeeds to the Turbuţa Formation deposition it has been ascertained a clear marine sedimentation of the Cluj, Brebi, and Cozla Formations.

The areal development of the Brebi Marl Formation is related to those of the Cluj Formation. Both pass laterally east of Poeniţa meridian to the Cozla Formation (Fig. 5, 6). As we said in previous section, within the Cozla Formation area, the Hoia Formation is recognized only after their definite paleontologic features. Thanks to some constant paleontologic levels (*Ostrea transsilvanica* and *Nummulites fabianii* levels, Hoia's paleontologic association — Pl. XII) we may control the accuracy of petrological interpretations.

Firstly, the fresh marine sedimentation comes in the studied area with mixed terrigenous-carbonate sediments from the Marly Calcareous Member of the Cozla Formation and with skeletal packstones and grainstones bellow the *O. transsilvanica* level of the Cluj Formation. These sediments are well bedded, with a planar orientation of shells, high amounts of terrigenous grains and thin intercalations of silty and sandy marls. This kind of deposits have resulted in high energy environment of the transgressive Priabonian sea. It seems that the general direction of marine water penetration was from north (Transcarpathian zone) and eventually from

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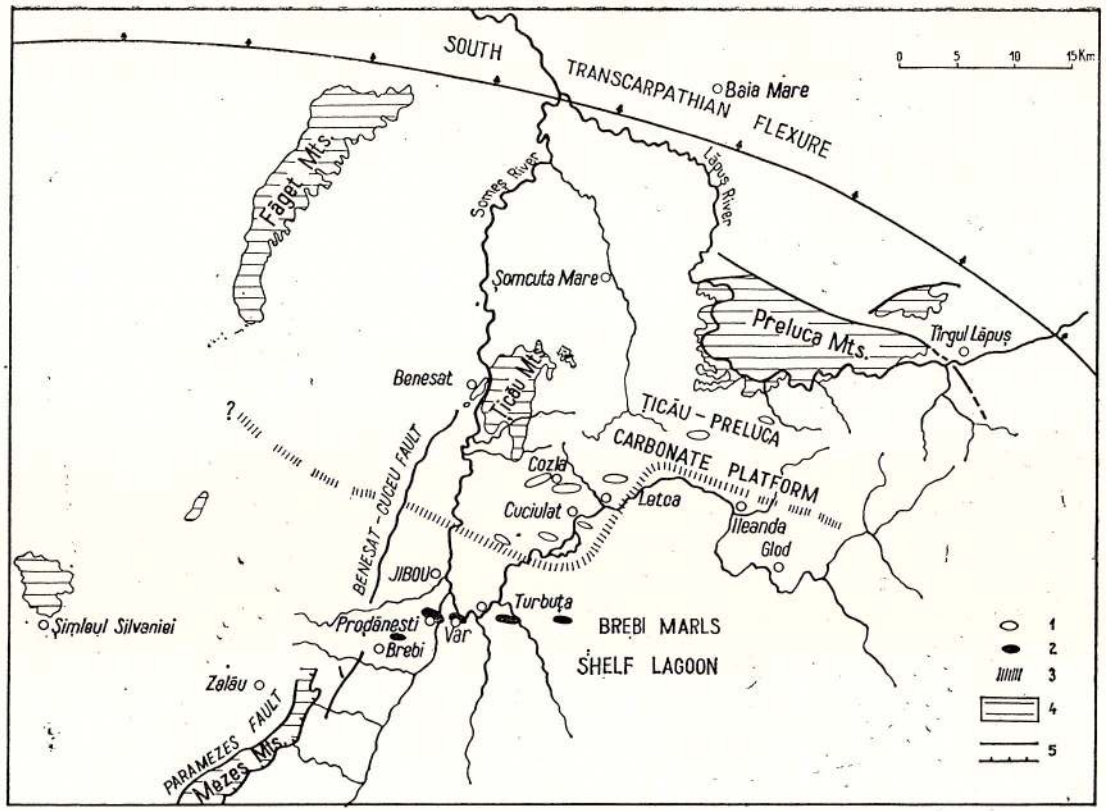


Fig. 5. — Facies distribution in the Cluj, Brebi and Cozla Formations.
1, coral and coralgal mounds within the Cozla Formation; 2, coral and coralgal mounds within the Cluj Formation; 3, the margin of the Țicău-Preluca carbonate platform; 4, crystalline massifs; 5, major faults.

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NE towards south (Cluj-Huedin zone), because the northward marine carbonate facies have southwar brackish and fresh water correspondants (Fig. 6).

Following this first transgressive event the Ţicău-Preluca area became a more stable carbonate platform like already described Barrier Platform from British Honduras (Matthews, 1966) or Barrier Reef Complex from W Australia (Maxwell, Maiklem, 1964). This platform will support a great pure carbonate production with biocostructed facies on borders and bioaccumulated facies in center (Fig. 5).

The bioconstructed facies from the southern border was an interrupted chain of coral and coralgal mounds and self-supported banks of skeletal sands. Mounds were ovoidal with long diameter of 25—100 m and the thickness of 5—10 m. They have the same dimension raports as the phylloid algal mounds described by Heckel and Cocke (1969) but smaller. In Figure 6 it may be shown that mounds have migrated in time from south to north and from the upper part of the Cluj Formation to the upper part of the Cozla Formation. A facies belt of the mounds is only a generalized concept because at different times they may shift laterally being recorded north of the mound area (Fig. 5, 6).

The corals and mutually encrusting corals and corallines are the organism which have formed a rigid framework. The open spaces were filled with mound associated skeletal mud (Pl. III, Fig. 2). The mounds are usually enveloped by thin layers of black shales which probably represent the interruption of mound growth (Pl. III, Fig. 4). The voids filled with sparry calcite, borings made by molluscs, dolomiztization of matrix, stylolitization, compaction and grain fracturing are the common features of these boundstones.

The intermound skeletal associated carbonates may form mound-like or bank-like structures which were stabilized by seagrasses (whose leaves prints are seldom carbonized).

The platform, north to the mound and bank area was, a wide zone of an intense skeletal-carbonate production. As we already say numerous mounds imbedded in corallinacean foraminiferic molluscan sandy muds and muddy sands were scattered overall platform. Locally the molluscs, bryozoans echinoderms, corals, nummulits combined with corallines and foraminifers offer various kinds of skeletal mud-supported textures.

The northern side of the platform is overlaid by Neogene deposits. Presumably, on the edge of the South Transcarpathian flexure there could be found a true reefal ridge like in modern areas with shelf edge reefs, or carbonate-terrigenous mixed facies.

South of the platform there was a large shelf lagoon (Matthews, 1966) or channel (Maxwell, Maiklem, 1964) with mixed terrigenous and carbonate sediments of the Brebi-, Bryozoan Marl-, and of the Cozla Formations.

For some time, in the lower part of Brebi Marl Formation this shelf lagoon was narrower. Back to the mound area (southward) there are skeletal packstones and grainstones with mound derived intercalations of

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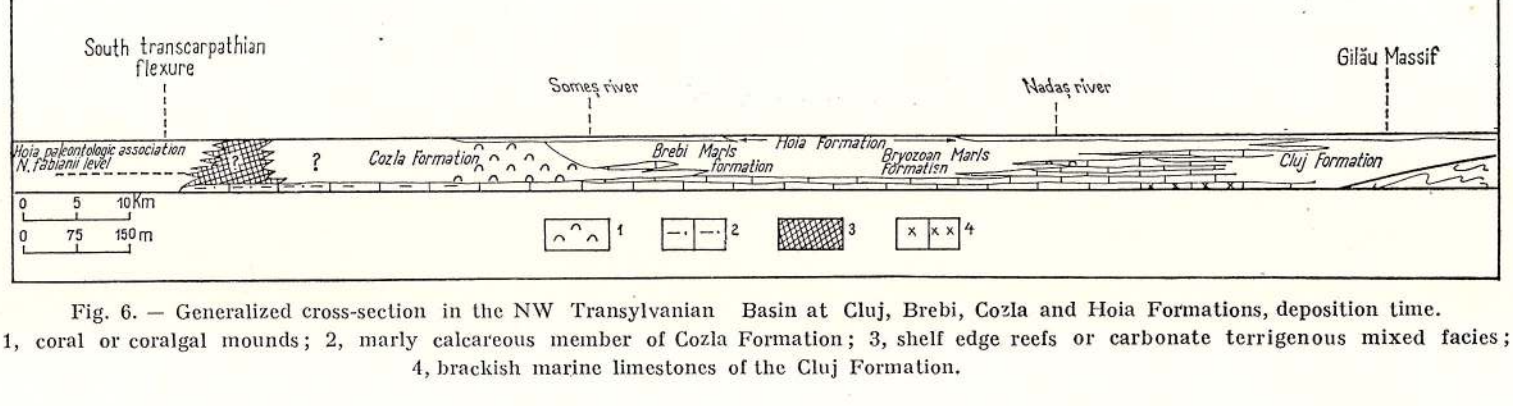


Fig. 6. — Generalized cross-section in the NW Transylvanian Basin at Cluj, Brebi, Cozla and Hoia Formations, deposition time. 1, coral or coralgal mounds; 2, marly calcareous member of Cozla Formation; 3, shelf edge reefs or carbonate terrigenous mixed facies; 4, brackish marine limestones of the Cluj Formation.

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the Cluj Formation. Above the *N. fabianii* level the transition between the Brebi and Cozla Formations is direct and rapid. Consequently, if we consider the lithologic, physiographic and climatic circumstances it results that the sedimentary supplies were mostly clays and they derived from a southern land towards a northern shelf.

The carbonate facies distribution allows us to suppose that in front of a wide shallow shelf (Cluj-Huedin-Zalău) with carbonate and mixed terrigenous-carbonate sediments of the typical Cluj Formation, there was a deeper shelf lagoon with longitudinal currents and clayey carbonate sediments of the Brebi and Bryozoan Marl Formations. Northward of this lagoon there was a wide platform with a powerful pure carbonate production (Fig. 6).

Unfortunately, we can not extent the suppositions either westward or eastward. From available data it results that the Paleogene flysch deposits common in the Transcarpathian realm are missing south of the Transcarpathian flexure (P a t r u l i u s et al, 1972)[6]. These authors consider the Rez-Măgura Șimleului area was raised during the Eocene time, non-excluding the possibility of the existence of eroded Eocene deposits in „Transylvan facies" in this area.

If we suppose that the northern part of the Mezeș Mts was submerse and the inner shelf with the Cluj Formation facies was an externe boundary it results that the shelf lagoon with Brebi and Bryozoan Marl Formation facies had a NW—SE direction (Fig. 5).

Beginning with the H o i a F o r m a t i o n sedimentation in the Transylvanian Paleogene zone there is a regression which will last — with little interruptions — all over the Oligocene period.

The carbonate facies from the Hoia Formation are not parallel with the NW—SE direction. It seems that it curves (Fig. 7) following in general manner the actual shape of the basin. Probably the Șimleu-Țicău zone was uplifted and the NW connection with the Transcarpathic realm interrupted.

Northward the studied area (Fig. 7) there is to be found an oncolithic grainstone or packstone facies. It shows excavations made by burrowing organisms filled with oncolithic mud from depocenter (Pl. XI, Fig. 2). The other nonskeletal grains are scarce, the skeletal ones being rather frequent. These last ones are the oncolithic nuclei.

Southward this facies is bordered by corallinacean mulluscan foraminiferal packstones and wackestones or corallinacean-molluscan coral packstones and wackestones of an moderate hydrodynamic regime. In the eastern part of this facies the molluscan wackestones and mudstones are predominant.

Next comes the corallinacean nummulitic facies. West of Prodănești there is to be found a corallinacean packstone or grainstone 6 m thick (Ciglean Limestone). The coralline are well rounded and sorted. The other

[6] D. P a t r u l i u s, A. D r ă g ă n e s c u, N. G h e ț a. Stratigraphic Correlations of the Sedimentary Deposits from NE Pannonian Depression. 1972. Arch. of Inst. of Geol. and Geophys.

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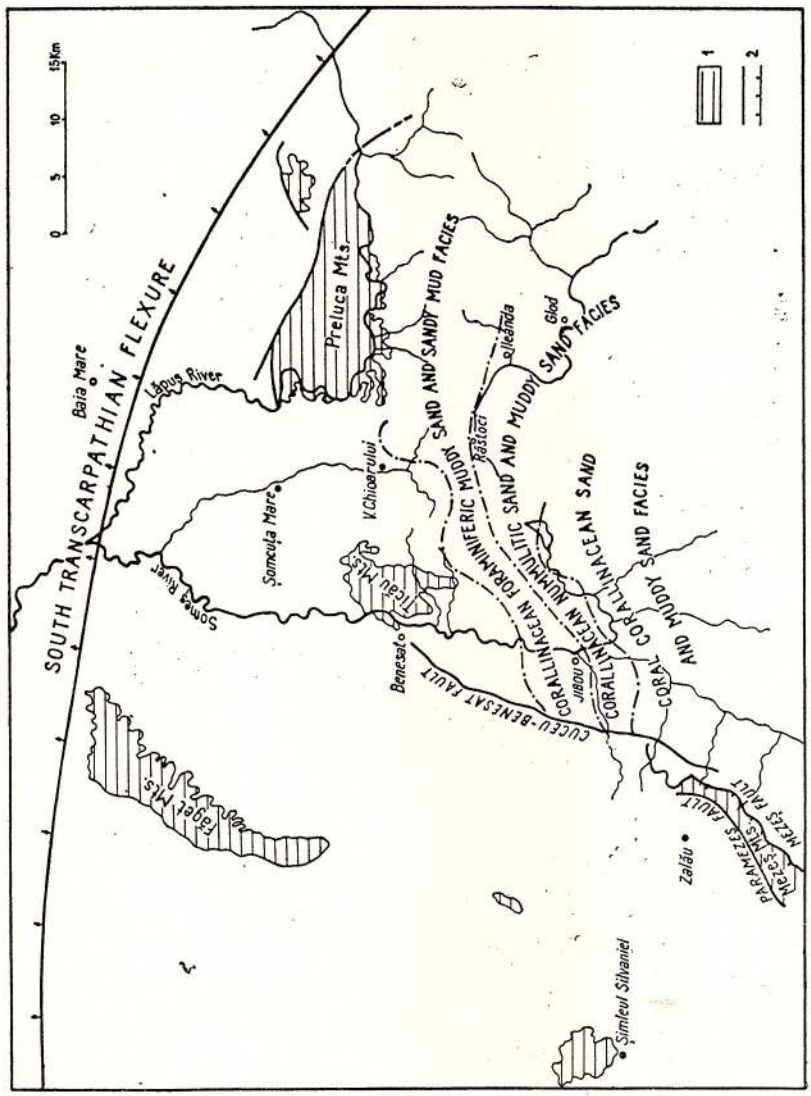


Fig. 7. — Facies distribution in the Hoia Formation.
1, crystalline massifs; 2, major faults.

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organisms are very subordinate. This facies is interpreted as an organic accumulation developed in shallow near shore turbulent water. Eastward this facies becomes richer in other than corallines skeletal grains (corallinacean nummulitic packstones or grainstones at Poenița, corallinacean nummulitic coral packstone at Letca) and in matrix (Pl. IX, Fig. 1, 3, 5; Pl. X, Fig. 2).

The southern most facies studied in this paper is an alternance of well packed molluscan packstones and grainstones, molluscan coral nummulitic grainstones and packstones with thin marly beds.

As we can see the turbulence of the depositional environment of the Hoia Formation increases from N to S and from E to W.

The strong packing, weak participation of matrix, excavations made by burrowing organisms, rapid variation of facies are the common textural features of the Hoia Formation deposits. An erosional relief between the Cozla and Hoia Formations has been rather frequently observed (Șoimușeni, Cozla, Răstoci).

The distribution of facies of the Hoia Formation as it is presented in Figure 7 is the result of summing of various small patchy facies not detailed in this paper. For an exhaustive description of the Hoia facies it would be divided in two parts (members) and their sedimentology separately presented. Because the paleontological content of the Hoia Formation from the Mezeș-Preluca zone is not exactly the same with those of the Hoia Formation from the Huedin-Cluj zone, the parallelizing must be made with great caution.

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PLATE I

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PLATE I

Fig. 1. — Grayish white planar stromatolite overlain by porous gray dolomitic mudstone. Lower part of the Turbuţa Formation. Piscul Ronei. Polished slab*

Fig. 2 — Pelletal grainstone. Welded pellets and oölits (bottom right). Upper part of the Turbuţa Formation, Runcu Valley. Thin section, N//.

Fig. 3. — Miliolitic wackestone-packstone. Note the good packing (top left) of the skeletal grains which grades into looser one (top right). The matrix is strongly recrystallized to microspar. Upper part of the Turbuţa Formation, west of Turbuţa village. Thin section, N //.

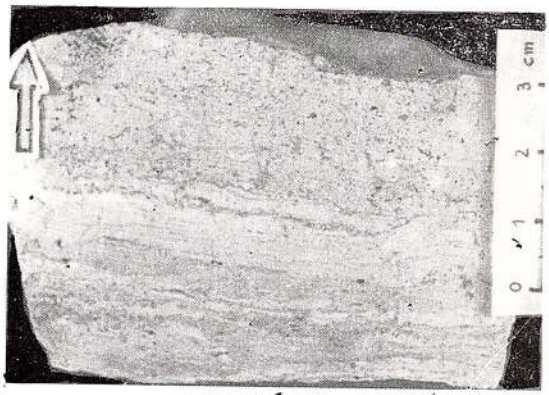
Fig. 4. — LLH stromatolitic boundstone. The thickness of the stromatolitic layer is of 3 cm. Lower part of the Turbuţa Formation, Piscul Ronei. Polished slab.

Fig. 5. — Porous mudstone with scattered foraminiferal remains. Upper part of the Turbuţa Formation, Runcu Valley. Thin section, N//.

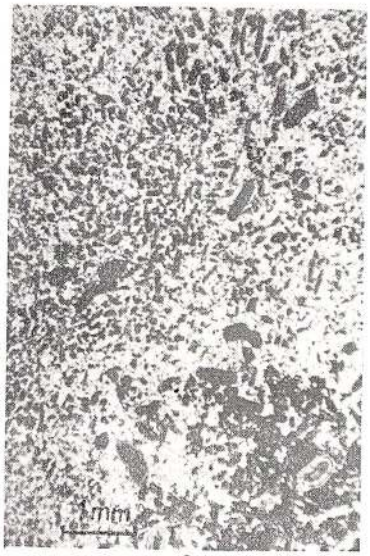
Fig. 6. — Dolomicrospar. The dark spots (top right) are of a clayey origin. Upper part of the Turbuţa Formation, Turbuţa village. Thin section, N//.

* The arrows point stratigraphically up.

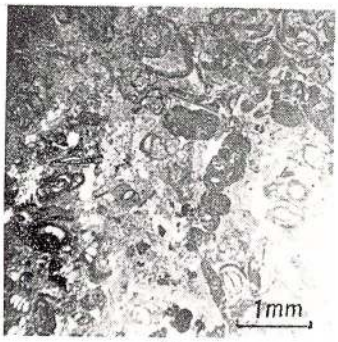
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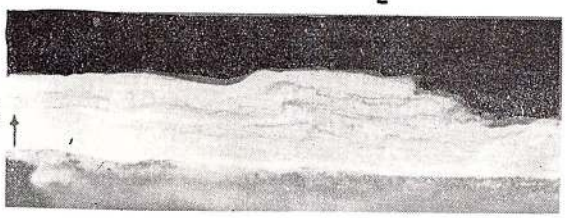
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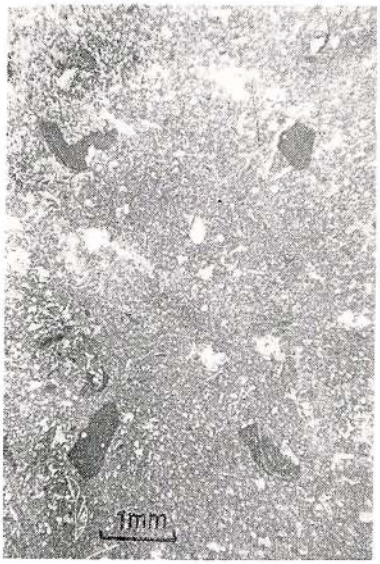
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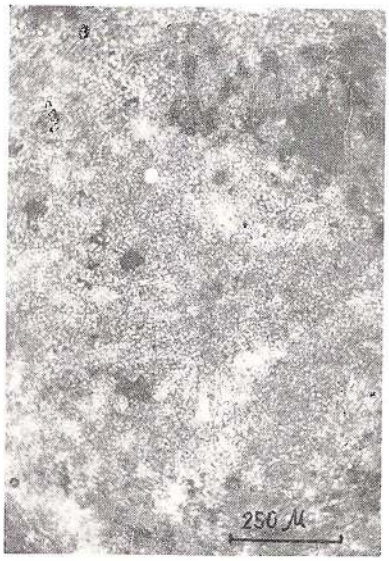
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PLATE II

Fig. 1. — Oölitic pelletal packstone showing various stages of the pellet oölitization. Lower part of the Turbuţa Formation, Piscul Ronei. Thin section, N//.

Fig. 2. — Asymmetrical oölits within pelletal packstone. The left oölit has a pellet within the oölitic coat. The oölitic coat shows bosses and gravitational elongation. Upper part of the Turbuţa Formation, Runcu Valley. Thin section, N//.

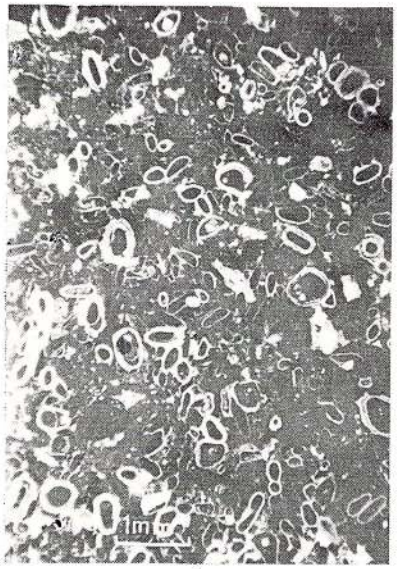
Fig. 3. — Fecal pellets showing tiny tubes and spots filled with sparry calcite. They have a superficial oölitic coat. Upper part of the Turbuţa Formation, Runcu Valley. Thin section, N//.

Fig. 4. — Corallines within dolomicritic matrix with scattered dolorhombs. Upper part of the Turbuţa Formation, west of Ciocmani village. Thin section, N//.

Fig. 5. — Foraminiferal wackestone. Upper part of the Turbuţa Formation, Turbuţa village. Thin section, N //.

Fig. 6. — *Nummulites* ex gr. *incrassatus* within dolomicritic matrix. Upper part of the Turbuţa Formation, west of Ciocmani village. Thin section, N//.

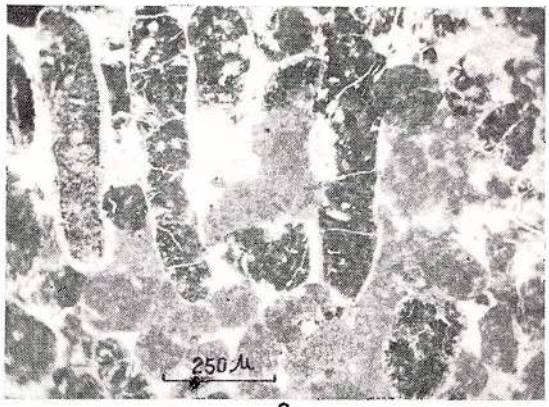
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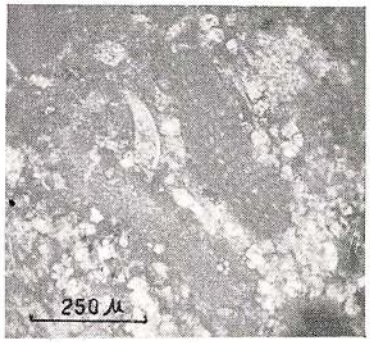
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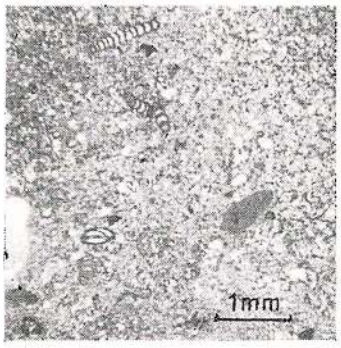
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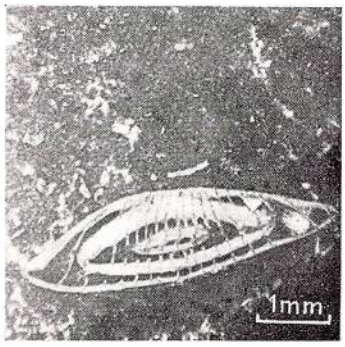
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PLATE III

Fig. 1. — Corallinacean wackestone of commonest aspect of the Cozla Formation, north of Poeniţa village. Polished slab.

Fig. 2. — Coral boundstone. Cluj Formation, Var village, Polished slab.

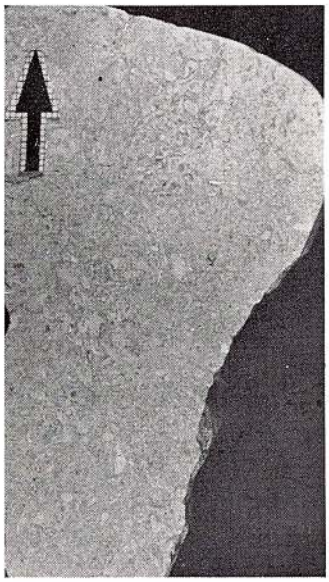
Fig. 3. — Coral head bored by a lithophag pelecypod. Cozla Formation, Prîsnel hill. Polished slab.

Fig. 4. — Marl bearing corallinacean wackestone from the top of coral mound. Cozla Formation, Cuciulat quarry. Polished slab.

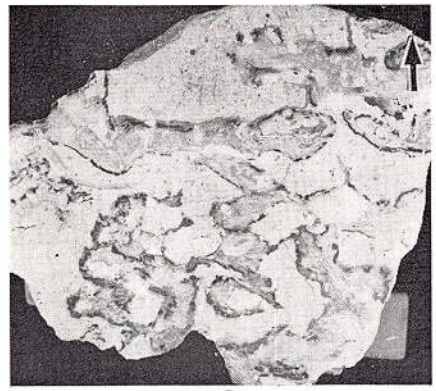
Fig. 5. — Mechanically made inlet within a coral head encrusted by coralline alga. The corallinacean crust was thereafter also destroyed. Cozla Formation, Prîsnel hill. Polished slab.

Fig. 6. — Bioaccumulation of corals which float within skeletal matrix. Cozla Fromation, north of Şoimuşeni village. Polished slab.

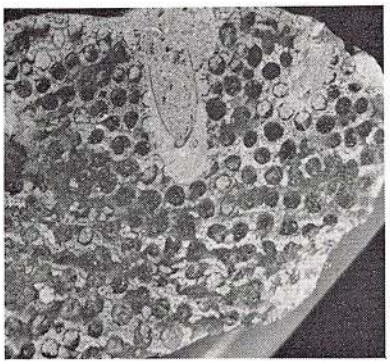
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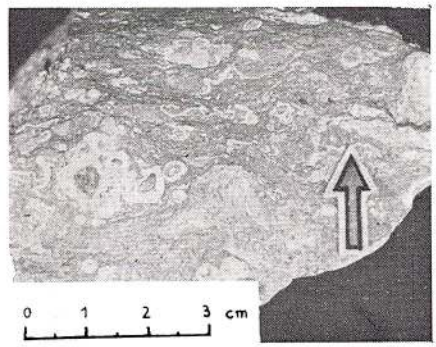
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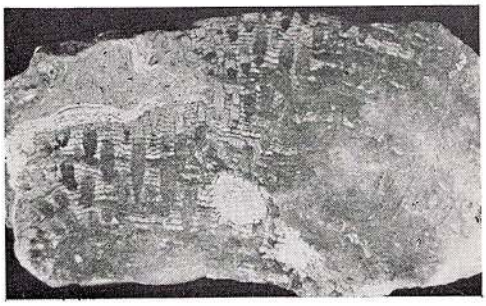
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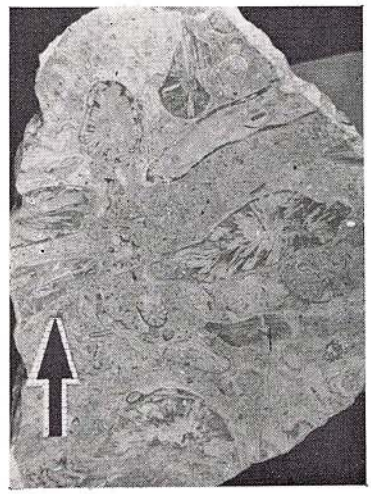
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PLATE IV

Fig. 1. — Foraminiferal wackestone from the lower part of the Cozla Formation. North of Şoimuşeni village. Thin section, N//.

Fig. 2. — Nummulitic wackestone from the *Nummulites fabianii* level of the Cozla Formation. Prisnel hill. This section, N//.

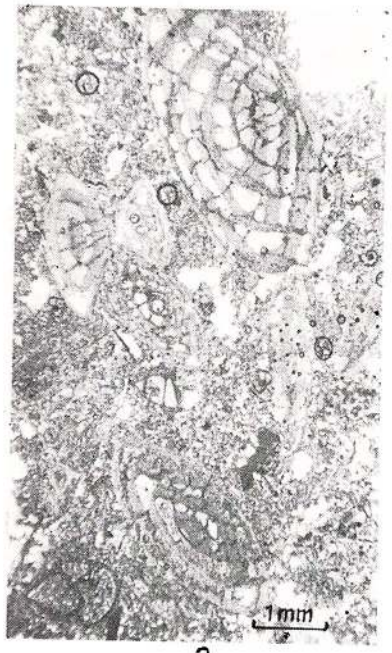
Fig. 3. — Foraminiferal wackestone from the Cluj Formation. Compare with Figure 1. Turbuţa village. Thin section, N//.

Fig. 4. — Nummulitic corallinacean packstone from the *Nummulites fabianii* level of the Cluj Formation. Giurgiului valley. Thin section. N//.

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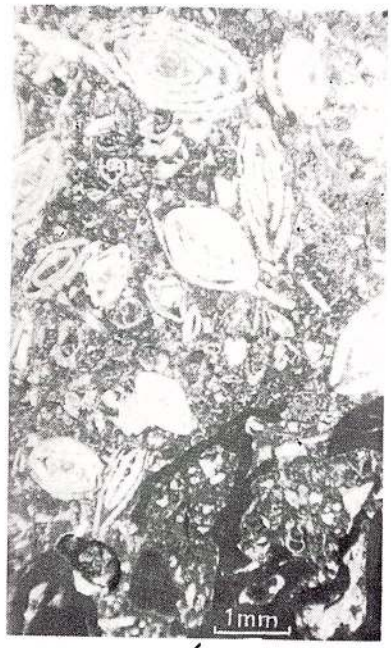
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PLATE V

Fig. 1. — Corallinacean foraminiferal packstone showing effects of compaction — grain fracturing and welding. Cozla Formation, Letca quarry. Thin section, N//.

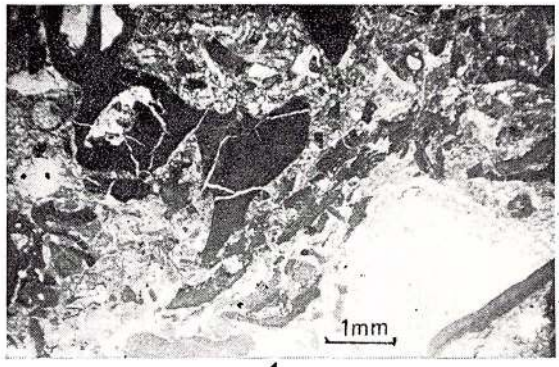
Fig. 2. — Corallinacean, foraminiferal echinodermal packstone. Matrix moderately recrystallized to microspar and pseudospar. Cluj Formation, north of the Ciocmani village. Thin section, N//.

Fig. 3. — Miliolitic packstone. Cozla Formation, Litard Valley. Thin section, N//.

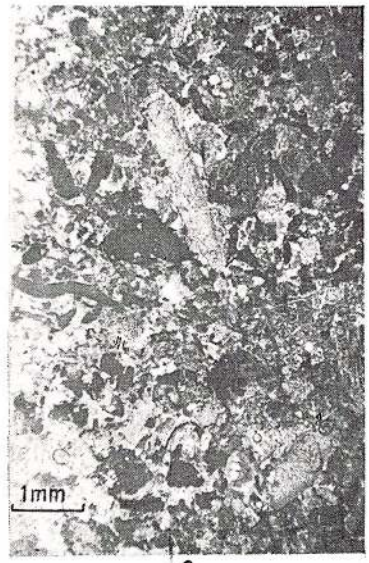
Fig. 4. — Corallinacean miliolitic wackestone. Note the poor sorting of the skeletal grains. Cozla Formation, Bîrsa Valley. Thin section, N//.

Fig. 5. — Corallinacean bryozoan foraminiferal packstone. Cozla Formation, north of Poeniţa village. This section, N//.

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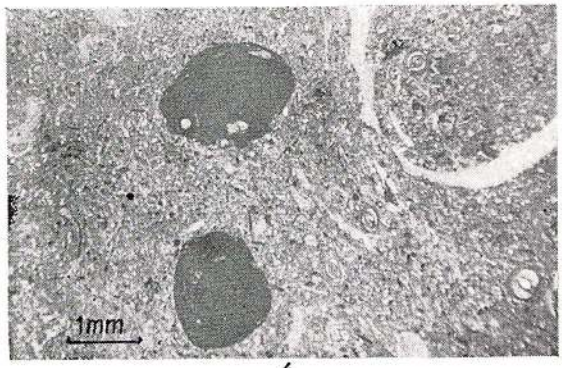
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PLATE VI

Fig. 1. — Lithophag pelecypod within a coral cavity. The coral septae are destroyed and the cavity subsequently filled with lime mud. Cozla Formation, Cozla village. Thin section. N//.

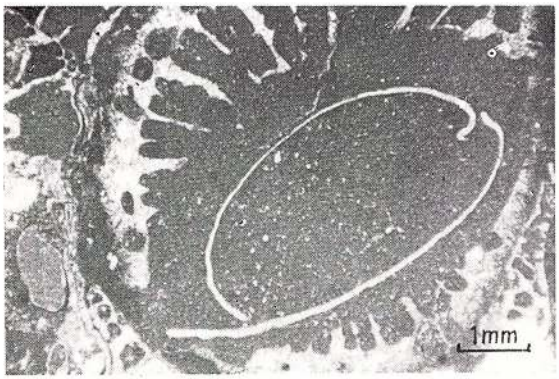
Fig. 2. — Corallinacean wackestone sediment trapped by a coralline alga. The conceptacles are filled with orthospar. Cluj Formation, Giurgiului Valley. Thin section, N//.

Fig. 3. — Intense fracturing of phylloid corallines in a mud supported texture. Cozla Fromation, Letca quarry. Thin section, N//.

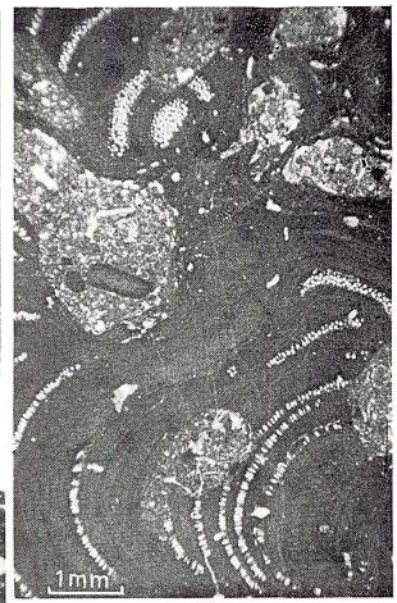
Fig. 4. — Coralline crust. Note the sheltering porosity done by coralline roof which favour the development of sparry calcite. Cozla Formation, Fundătura Valley. Thin section, N //.

Fig. 5. — Algal crust on coral fragment. Compare the strong recrystallization of the intraseptal micrite of the coral fragment with the poorer one of the micritic matrix. Cozla Formation, Cozla village. Thin section, N//.

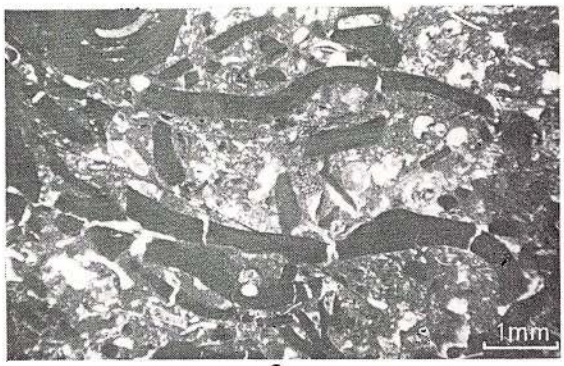
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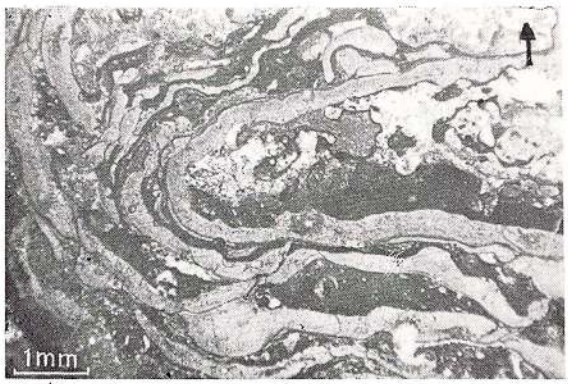
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PLATE VII

Fig. 1. — Internal cavities within coral boundstone. One of the cavities (right side) was first rimmed with orthospar, then with geopetal sediment and finally by neomorphic sparry calcite. Cozla Formation, Halăul hill. Thin section, N//.

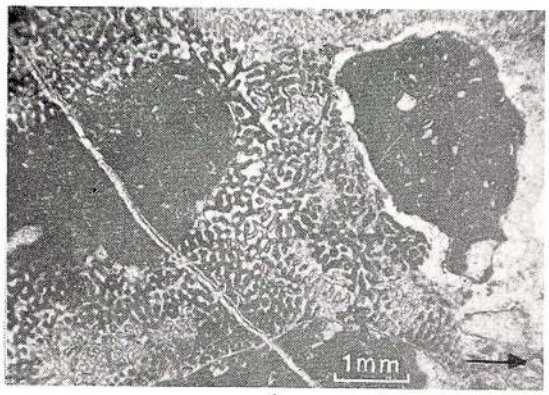
Fig. 2. — Stylolitic seam and related dolomitized micrite within a corallinacean wackestone. Cozla Formation, Cuciulat quarry. Thin section, N //.

Fig. 3. — Boundary between coral boundstone and foraminiferal corallinacean wackestone. Cluj Formation, Var village. Thin section, N//.

Fig. 4. — Dolomitized foraminiferic echinodermal wackestone. Cozla Formation, Cuciulat quarry. Thin section, N//.

Fig. 5. — Foraminiferal echinodermal molluscan wackestone. Note the gravitational tendency of the echinodermal syntaxial neomorphic calcite. Cozla Formation, Fundătura Valley. Thin section, N //.

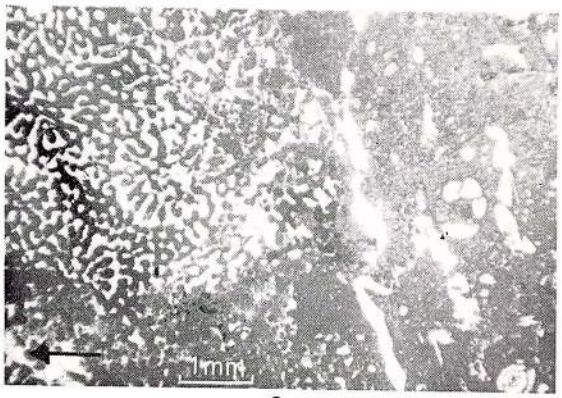
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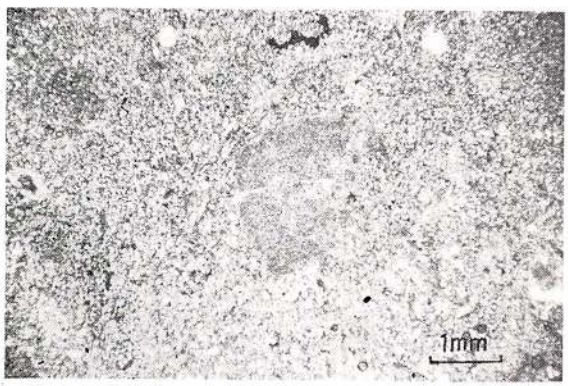
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PLATE VIII

Fig. 1. — Corallinacean coral wackestone with tiny argillaceous lamina and intense stylolitization. The stylolitic seam is marked by cryptocrystalline clay and organic matter. Cozla Formation, Cuciulat quarry. Polished slab.

Fig. 2. — Late diagenetic diaclases intersecting a coral with recrystallized septae. Cozla Formation, north of Poeniţa village. Thin section, N//.

Fig. 3. — Same as in Figure 1. The dolomitization of micritic matrix and skeletal grains is related with the stylolitic line. Cozla Formation, Cuciulat quarry. Thin section, N//.

Fig. 4. — Clayey porous dolomitized mudstone. Cozla Formation, Litard Valley. Thin section, N//.

Fig. 5. — Patch of recrystallized sparry calcite showing sharp angular boundaries possibly resulted from dolomitization-dedolomitization processes. Cozla Formation, north of Poeniţa village. Thin section, N //.

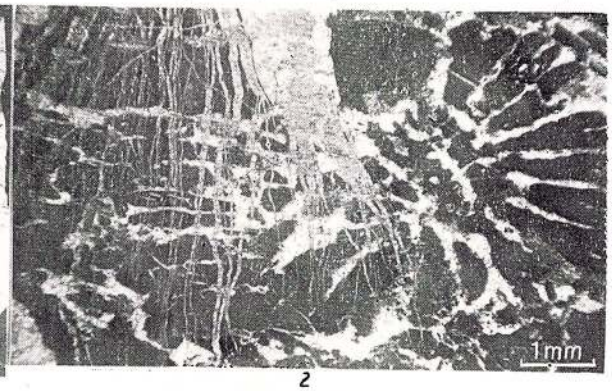
Fig. 6. — Dolomitized corallinacean wackestone. The larger dolorhombs show palimpsest relics and rims. Cozla Formation, Letca quarry. Thin section, N//.

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B. Popescu. Sedimentology of Priabonian Carbonate Rocks. Pl. VIII.



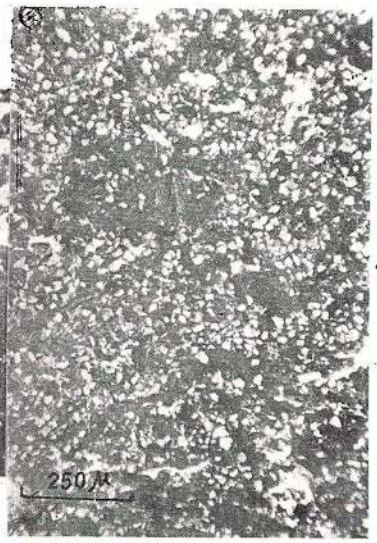
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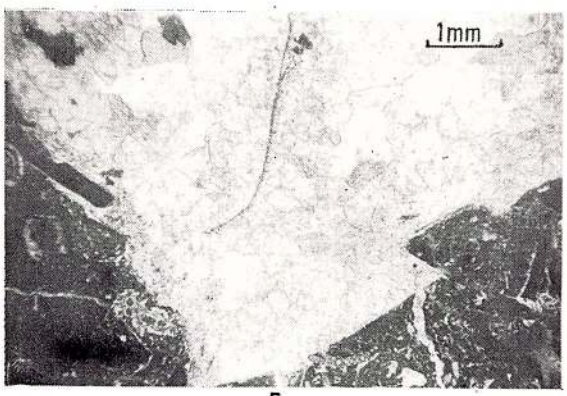
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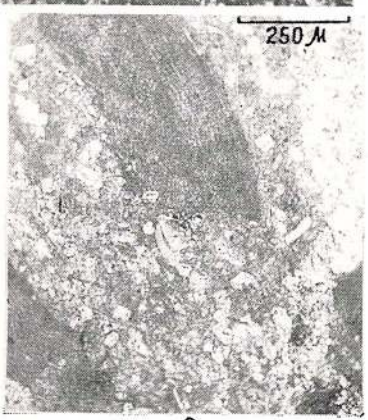
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PLATE IX

Fig. 1. — Corallinacean molluscan nummulitic grainstone showing poor lamination. Lower part of the Hoia Formation, west of Ciocmani village. Polished slab.

Fig. 2. — Same as Figure 1. Note the good packing and fracturing of skeletal grains (see the „corrugated" molluscan shell in the center of the photo).

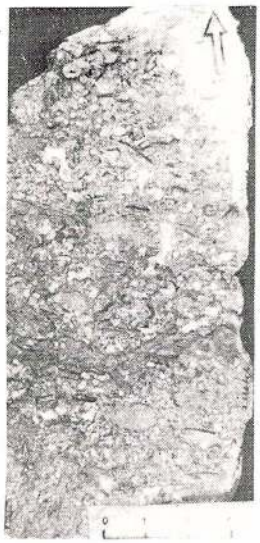
Fig. 3. — Corallinacean molluscan nummulitic packstone. Planar orientation of the molluscan shells. Hoia Formation, north of Poeniţa village at the top of the Cozla Formation Polished slab.

Fig. 4. — Same as Figure 3. Note the welding between the nummulit and coralline alga (center).

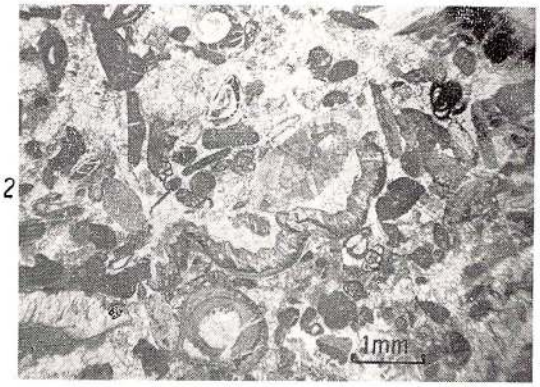
Fig. 5. — Corallinacean nummulitic packstone. Matrix becomes more abundant as in previous examples. „Hoia Formation", Letea quarry at the top of the Cozla Formation. Polished slab.

Fig. 6. — Same as in Figure 5. The good packing is preserved.

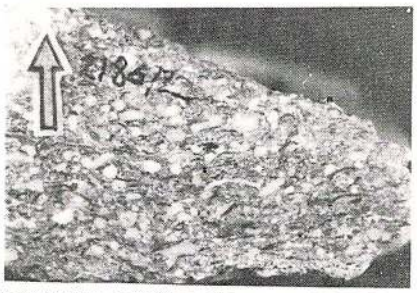
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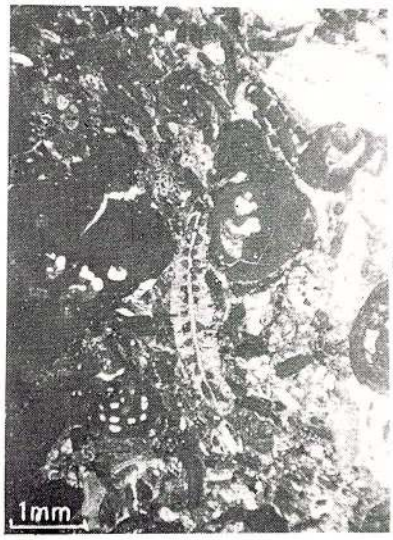
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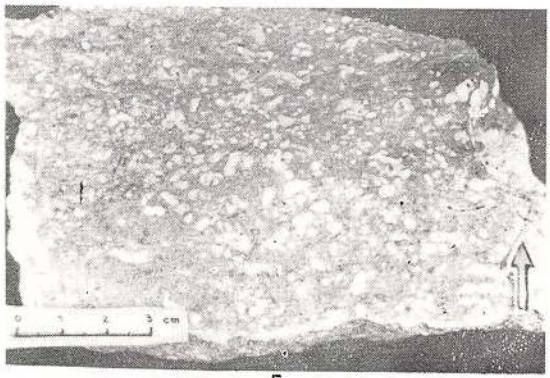
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PLATE X

Fig. 1. — Corallinacean nummulitic packstone, showing close packing and strong recrystallization of matrix. Hoia Formation (Ciglean Limestone), Ciglean village. Thin section, N //.

Fig. 2. — Corallinacean wackestone. Note the progressive increase in matrix content if compare with Figures 1, 3, 5 of Plate IX. ,,Hoia Formation'' Bîrsa Valley at the top of the Cozla Formation. Polished slab.

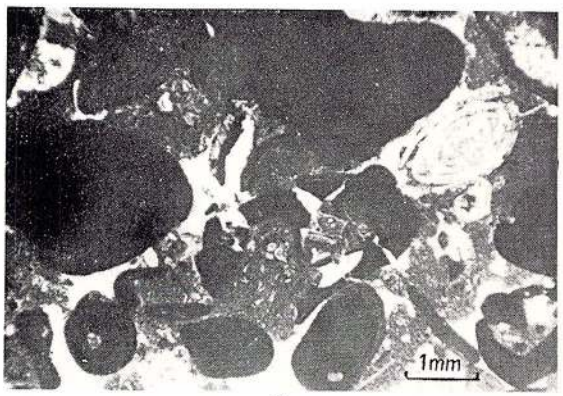
Fig. 3. — Corallinacean nummulitic packstone, Matrix strongly recrystallized to microspar and pseudospar. Upper part ot the Hoia Formation, Sacadicu hill. Thin section. N //.

Fig. 4. — Molluscan corallinacean grainstone. Overly close packing welded grains, good porosity. Lower part of the Hoia Formation, Brebi village. Thin section, N // .

Fig. 5. — Molluscan corallinacean grainstone. Close packing, fractured and welded grains. Note the shelterjng done by the molluscan shells. Lower part of the Hoia Formation, west of Ciocmani village. Thin section, N//.

Fig. 6. — Nummulitic wackestone. Lowermost part of the Hoia Formation, west of Ciocmani village. Thin section, N //.

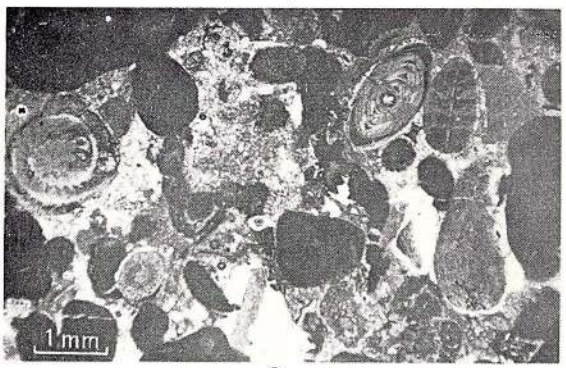
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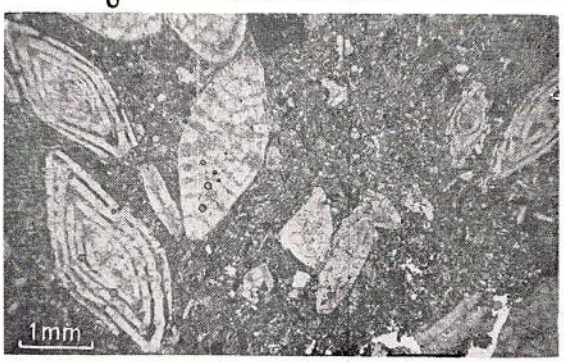


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PLATE XI

Fig. 1. — Oncolitic packstone. Matrix moderately recrystallized to microspar and pseudospar. Hoia Formation, west of Mestecăn village at the top of the Cozla Formation. Thin section, N//.

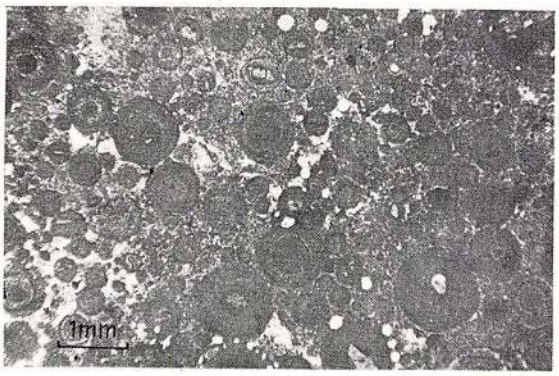
Fig. 2. — Pelecypod? burrow within oncolitic wackestone. The burrow filling is also an oncolitic packstone but moderately or not recrystallized. ,,Hoia Formation'', north of Vărai village at the top of the Cozla Formation. Thin section, N //.

Fig. 3. — Gastropod shell with an oncolitic coat within oncolitic packstone. Same locallity as in Figure 2.

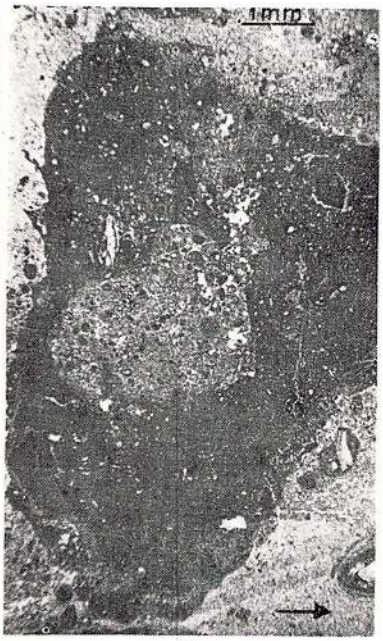
Fig. 4. — Corallinacean coral grainstone. Note the close resemblance of the micritized corallines with fecal pellets. Hoia Formation, upper part, west of Ciocmani village. Thin section. N//.

Fig. 5. — Same as Figure 4. At higher magnification one may sometimes recognize the corallinacean (C) or miliolitic (M) structure. Other corallinacean grains are wholly micritized (AP) becoming an algal pellet.

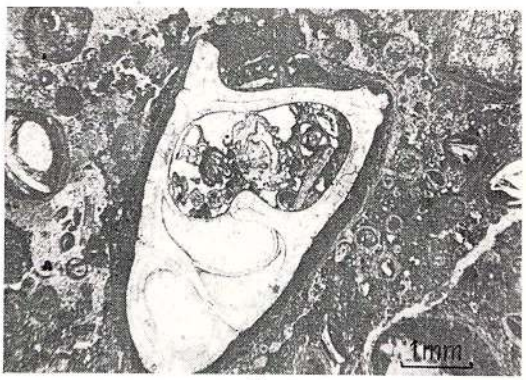
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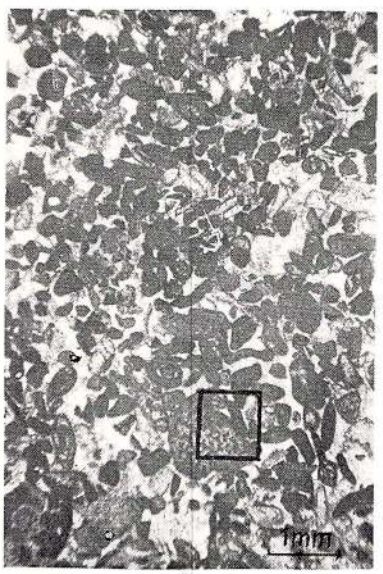
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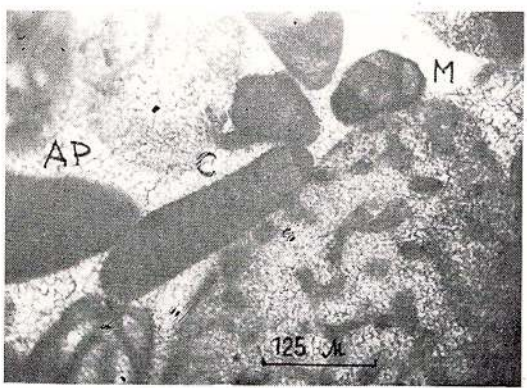
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5

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LITHOSTRATIGRAPHIC COLUMNS THROUGH THE PRIABONIAN CARBONATE FORMATIONS FROM JIBOU AREA

0 2 4 6 m

LEGEND

2- Giurgiului valley - Nr. of outcrop or quarry in fig. 1

3.572/1- Nr. of sample

- Skeletal wackestones and packstones
- Skeletal grainstones
- Coral and coralgal boundstones
- Marls bearing skeletal wackestones
- Limestone breccia
- Marls
- Sands and sandstones
- Fault

2. Giurgiului valley

3. Prodănești quarry

4. Var N.

6. Stejărei valley

6. Ciocmani W.

7. Poenița E.

10. Cuciulat quarry

9. Prisnel hill

CURTUIUȘ FORMATION

HOIA FORMATION

BREBI MARL FORMATION

HOIA FORM

BREBI MARL FORMATION

COZLA FORMATION

N fabianii

CLUJ FORMATION

COZLA FORMATION

Imprim. Atel. Inst. Geol. Geof.

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FACIES-ZONED CARBONATE SEDIMENTATION AT THE TIME OF THE HOIA LIMESTONE (UPPER TONGRIAN) IN NW TRANSYLVANIA (ROMANIA)[1]

BY

A. RUSU and A. DRĂGĂNESCU[2]

Abstract

La sédimentation carbonatique zonnée au niveau du Calcaire de Hoïa (Tongrien supérieur) dans le NW de la Transylvanie. L'analyse pétrologique des facies dépositionnels au niveau du Calcaire de Hoïa (Tongrien supérieur) des aires du Mezeş et de la Preluca (NW de la Tranylvanie) ont permis aux auteurs de mettre en évidence une intéressante zonalité faciale. Il faut souligner tout d'abord que le terme de „Calcaire de Hoïa" doit être retenu seulement pour le facies calcaire coralligène des Couches de Hoïa (s.l.). En même temps, on démontre que l'intervalle stratigraphique entre les Marnes de Brebi et les Couches de Curtuiuş de l'aire du Mezeş, de même que son équivalent (le facies calcaire algal = Calcaire de Ciglean) de la partie terminale de la Série calcaire de l'aire de la Preluca, correspond au Calcaire de Hoïa et à la partie basale des Couches de Mera (y compris le niveau inférieur à *Scutella*) de la région de Cluj-Napoca. Les coupes géologiques examinées relèvent l'existence de trois zones faciales, à savoir : la zone du facies sablonneux à corallinacées, occupant la plate-forme du shelf ; la zone du facies vaseux à miliolides et coraux, qui longe la bordure du shelf ; la zone du facies vaseux à globigérines et bryozoires, correspondant au domaine pélagique. Ce dernier facies est placé stratigraphiquement dans la partie terminale de Marnes à bryozoires (Marnes de Brebi) des zones plus internes du bassin. L'image stratigraphique régionale et la zonalité faciale mises en évidence nous portent à conclure que les dépôts du niveau du Calcaire de Hoïa correspondent à un moment de régression rapide. Le contenu paléontologique du Calcaire de Hoïa et des dépôts équivalents du bassin leur confère l'âge tongrien supérieur ; ce laps de temps doit être inclu soit à l'Oligocène moyen, si l'on considère le Lattorfien en tant qu'étage oligocène, soit à la base de l'Oligocène — si cet étage se range à l'Eocène supérieur. C'est cette dernière variante qui a été adoptée dans le présent ouvrage.

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[2] Institutul de geologie şi geofizică, str. Caransebeş 1, 7000 Bucureşti.

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The Hoia Beds stand for one of most interesting and controversial lithostratigraphic units of the Paleogene sequence of the Transylvanian Basin.

The achievement of a first facies zonation in the stratigraphic interval corresponding to the Hoia Limestone enables us to modify to a certain extent the previous manner of interpretation of this member.

The region discussed in the paper lies in the northwestern part of the Transylvanian Depression, in the proximity of the crystalline mountains of Mezeș, Țicău and Preluca. The concerned sector reaches more than 80 km in length and is centred by the locality of Jibou (Pl. XXI).

PREVIOUS WORK

The fauna of the Hoia time interval was first recorded in the pioneer's paper of H a u e r and S t a c h e (1863). Later, H o f m a n n (1879) separates at the base of the Oligocene sequence in the vicinity of Jibou, a distinct member under the name of „Untere marine molluskenreiche Schichten", mentioning that it corresponds to the Hoia Beds of the Cluj-Napoca region. Thus it would be for the first time now, that the name of Hoia Beds appears in the literature, because the paper of K o c h, the author which proposed it, was published an year later (1880, fide K o c h, 1894). From 1881 on, H o f m a n n used this term in his papers too, specifying that in the Preluca area the lower part of the Hoia Beds consists of limestones containing corals and „*Lithothamnium*" and their upper part — of limestones with molluscan shells.

In his famous monograph on the Tertiary deposits of the Transylvanian Basin, K o c h (1894) describes the section exposed in the Hoia Hill (Cluj-Napoca) pointing out that there the Hoia Beds encompass two layers of breccia-like marly limestones containing molluscan shells, corals, balanids and small nummulitids. In addition this author records several other equivalent outcrops in the Gilău area thus achieving the first correlation of the deposits in the stratigraphic interval of the Hoia Limestone.

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A fact interesting to be mentioned is the strange opinion of Mihăltz which considered the Hoia Beds to be transgressive and in places even uncomformably overlying the Brebi Marls (while actually, as will be proved further on, the under-discussion beds correspond to a phase of regression).

In 1956, Răileanu and Emilia Saulea correlated the coraligenous limestone of the Hoia Hill with the Mera Beds (Oligocene) — actually younger —, while Barbu (1956) asserted on micropaleontological grounds that the sequence separated as Hoia Beds in the Jibou region represents only a calcareous facies at the upper part of the Brebi Marls (Eocene). Thus we are confronted by two quite different interpretations, which led either to an erroneous correlation of these strata within the basin, or even to their elimination from the local stratigraphic sequence.

Rusu, initially by unpublished works (1962 —1963) and then by two published papers (1967, 1968) resumes the study of the Hoia Beds in the Jibou region, opinating for their quality of stratigraphic entity.

Subsequently, the situation in the stratotype area is clarified by the researches enterprised by Moisescu (1968) and Mészáros and Magdun (1972), which supply many convincing, detailed data. Quite recently, Mészáros et al. (1973) and soon afterwards Martini and Moisescu (1974) analyse the nannoplancton of the Hoia Beds, ascribing a Lattorfian and, respectively, a Lower Oligocene age to this member.

In spite of the lately achieved improvement of our knowledge on these strata, several aspects which led through the years to such different interpretations remained unclarified. Our belief is that the present paper will help to a better understanding of what represents and how must be interpreted the Hoia Limestone.

GENERAL DISCUSSION ON THE STRATIGRAPHY OF THE INTERVAL INCLUDING THE HOIA LIMESTONE

Inasmuch as the name of „Hoia Beds" was used in a broader meaning designating deposits rather different in facies and sometimes even different in age in comparison with those of the stratotype, we prefer the term of „Hoia Limestone" restricted exclusively to the coraligenous facies of this stratigraphic interval.

The type section of the Hoia Limestone was unfortunately ill-chosen because the outcrop on the Hoia Hill (Cluj-Napoca) does not expose the boundary with the underlying Bryozoa Marls, and, on the other hand, a red clay facies invades the overlying Mera Beds to a rather low level. The Hoia Limestone is, instead, well defined paleontologically (molluscs, corals, echinoids, foraminifera, nannoplancton) so that any references to the Hoia Beds must take into account the faunal assemblage of the stratotype.

Three main depositional areas of the Paleogene were distinguished (Rusu, 1970) in the northwestern part of the Transylvanian Depression :

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the Preluca area (developed around the crystalline massif of Preluca), the Mezeș area (flanking eastward the crystalline ridge of the Mezeș Mountains), and the Gilău area (bounding northward and eastward the crystalline massif of Gilău). The present paper discusses the Hoia Limestone and its stratigraphic equivalents only in the Preluca and Mezeș areas (Pl. XXI).

In the Jibou region (Mezeș area) two beds of limestones separated by a marly layer are individualized on a certain extension between the underlying Brebi Marls and the overlying Curtuiuș Beds ; the two limestone beds were together designated since H o f m a n n as Hoia Beds. The stratigraphical studies undertaken by R u s u pointed to the fact that only the lower, coral-bearing, limestone bed corresponds in age to the limestone of the Hoia Hill ; thus, only this one may be termed „Hoia Limestone" ; the upper, molluscan shell-bearing, limestone bed, for which the idea that it would be more typical for the Hoia Beds got accredited since H o f m a n n (1881, p. 327), actually corresponds to the lowermost part of the Mera Beds of the Cluj-Napoca region (Gilău area), namely to their *Scutella* Lower Level (Fig. 1).

The particular situation of the Jibou region (Mezeș area) was unjustly extended also to the Cluj-Napoca region (Gilău area), different authors reaching the erroneous idea of considering the *Scutella* Lower Level as the main element of the Hoia Beds in this latter region (M o i s e s c u , 1968 ; B o m b i ț ă and M o i s e s c u , 1968). Usually an equivalent of „Hoia" was looked for in the first sandy-calcareous deposits overlying the Bryozoa Marls ; but the nannoplancton analyses have established that the assemblage encountered at Cluj within the Hoia Limestone occurs, for instance at Mera, in the uppermost part of the Bryozoa Marls (M é s z á r o s et al., 1974). K o c h (1894) intuited partly this fact, considering that at Mera the deposits 2,5 m thick underlying the *Scutella* Lower Level belong to the Hoia Beds (p. 340), and showing that at Sard the place of the typical Hoia Beds is occupied by a layer of marly clays (2 m) containing fossils of the „Bryozoaentegels" (p. 320).

This interpretation was also put forward for the regions of the northernmost part of the basin. So, H o f m a n n (1881) contemplates the possibility for the limestones with corals and „ *Lithothamnium*" of the Preluca area to represent a littoral facies much reduced in thickness of the Brebi Marls, a possibility which, for lack of evidence, is rejected several lines further on (p. 327). B a r b u (1956) thinks that the strata separated as Hoia Beds in the Jibou region would represent only a calcareous facies of the upper part of the Brebi Marls. Later on, R u s u (1967) correctly remarks that only the lower calcareous bed (that is even the Hoia Limestone) could stand for a calcareous facies of the Brebi Marls, but he as well as B a r b u leaves to be understood that in this case the true Hoia interval would not be involved.

The recently undertaken studies led the present authors to the following picture concerning the stratigraphic interval including the Hoia Limestone (Fig. 1).

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In the Gilău area the Hoia Limestone Member is located stratigraphically between the Bryozoa Marls Member at the bottom and the Mera Beds (Member) at the top, the latter including in a subbasal position the *Scutella* Lower Level. In places the Hoia Limestone is lacking, being replaced by the Bryozoa Marls which rises up to the base of the Mera Beds; in this situation the calcareous *Scutella* Lower Level, by its geometrical position, gives the impression that it represents the Hoia Beds, a

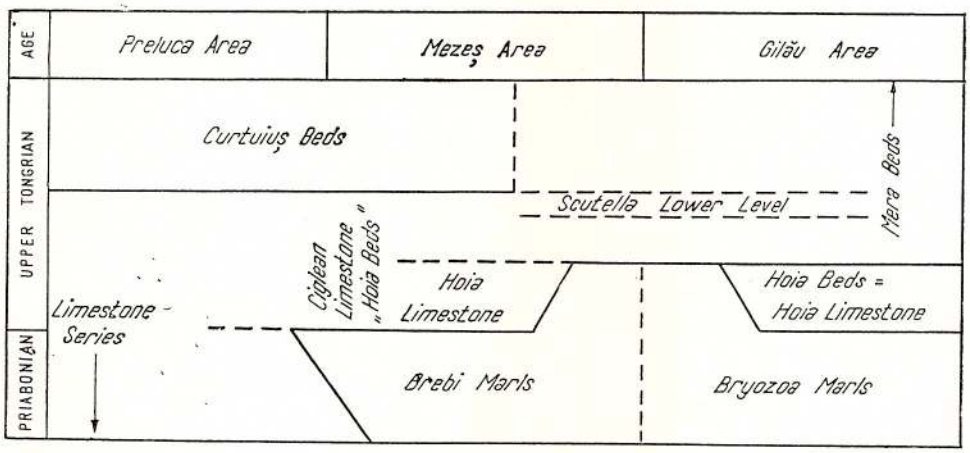


Fig. 1. — The Hoia Limestone position in the stratigraphic ensemble of the Transylvanian Basin in authors opinion.

fact which often misled the previous researchers in the stratigraphic separation of the Hoia Limestone. Therefore in the Gilău area the Hoia Limestone represents the true Hoia overlying the Bryozoa Marls and underlying the Mera Beds.

In the Mezeș area the „Hoia Beds" are located between the Brebi Marls Member (the northern equivalent of the Bryozoa Marls) at the bottom and the Curtuiș Beds (Member) at the top; but the lower boundary of the Curtuiș Beds lies stratigraphically higher than the lower boundary of the Mera Beds (the southern equivalent of the former), and the *Scutella* Lower Level is included in the uppermost part of the „Hoia Beds". In other words, in Mezeș area the „Hoia Beds" occupy a larger stratigraphic interval than the true Hoia Beds of the Gilău area. As regards its composition, the „Hoia Beds" in the Mezeș area appear heterogeneous, consisting of a lower calcareous coraligenous unit standing for the Hoia Limestone (the true Hoia Beds), and an upper unit composed of several marly beds with brackish-water faunas and crowned by the *Scutella* Lower Level, this upper unit as a whole representing the stratigraphic equivalent of the lowermost part of the Mera Beds. Therefore in the Mezeș area the Hoia Beds must be restricted to the Hoia Limestone Member, the overlying brackish member bounded between the Hoia Limestone and the Curtuiș Beds

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representing a distinct stratigraphic unit. In places the Hoia Limestone disappears, being replaced by the Brebi Marls which are directly overlain by the brackish member (the equivalent of the lowermost part of the Mera Beds). In other situations the Hoia Limestone and the overlying brackish member (underlying Curtuiuș Beds) are replaced by a calcareous corallinacean facies making up a single stratigraphic unit — the Ciglean Limestone — equivalent to the „Hoia Beds sensu lato" (alias sensu H o f m a n n).

In the Preluca area the Curtuiuș Beds (Member) are underlain by a comprehensive series — the Calcareous Series — including in its upper part corallinacean and coraligenous limestones standing for the stratigraphic equivalent of the Ciglean Limestone. Therefore the correspondent of the Hoia Limestone in this area must be stratigraphically placed in the subterminal part of the Calcareous Series, that is in the lower part of the uppermost interval of the Calcareous Series equivalent to the „Hoia Beds sensu lato".

In the present paper only the Hoia Limestone (the true Hoia Beds) interval in the Mezeș and Preluca areas will be discussed.

DESCRIPTION OF STUDIED SECTIONS

The main outcropping sections of the Hoia Limestone interval in the Mezeș and Preluca areas are described below. To have a systematic presentation, it is necessary to anticipate the next chapters and to state that in the Mezeș and Preluca areas three sinchronous, depositional, calcareous facies may be distinguished in the stratigraphic interval of the Hoia Limestone : a muddy, miliolid-coraligenous facies proper to the actual Hoia Limestone ; and two colateral facies (standing for the facies of the stratigraphic equivalents of the Hoia Limestone), namely a sandy, corallinacean facies in the uppermost part of the Calcareous Series, and a muddy, globigerinid-bryozoan facies developed in those sectors where the stratigraphic equivalent of the Hoia Limestone is located at the top of the Brebi Marls. The outcropping sections of each of these facies are described further on (Fig. 2).

A) T h e m u d d y , m i l i o l i d - c o r a l i g e n o u s f a c i e s appears in Mezeș area as a continuous strip following the line linking the localities Ciocmani-Brebi-Moigrad-Stîna-Ciumărna-Treznea, and in the Preluca area as scattered patches accommodated within the mass of the sandy, corallinacean facies.

In the Mezeș area there are three main sections within the mentioned strip of coraligenous facies : in the Jurteana valley (N of Stîna), în the Pîrîul Berculuí valley (Ciumărna), and in the Groapa Stiavului valley (Treznea).

In the section of the Jurteana valley (upper course, left side) it may be noticed that the Brebi Marls (*a*) are gradding upwards into

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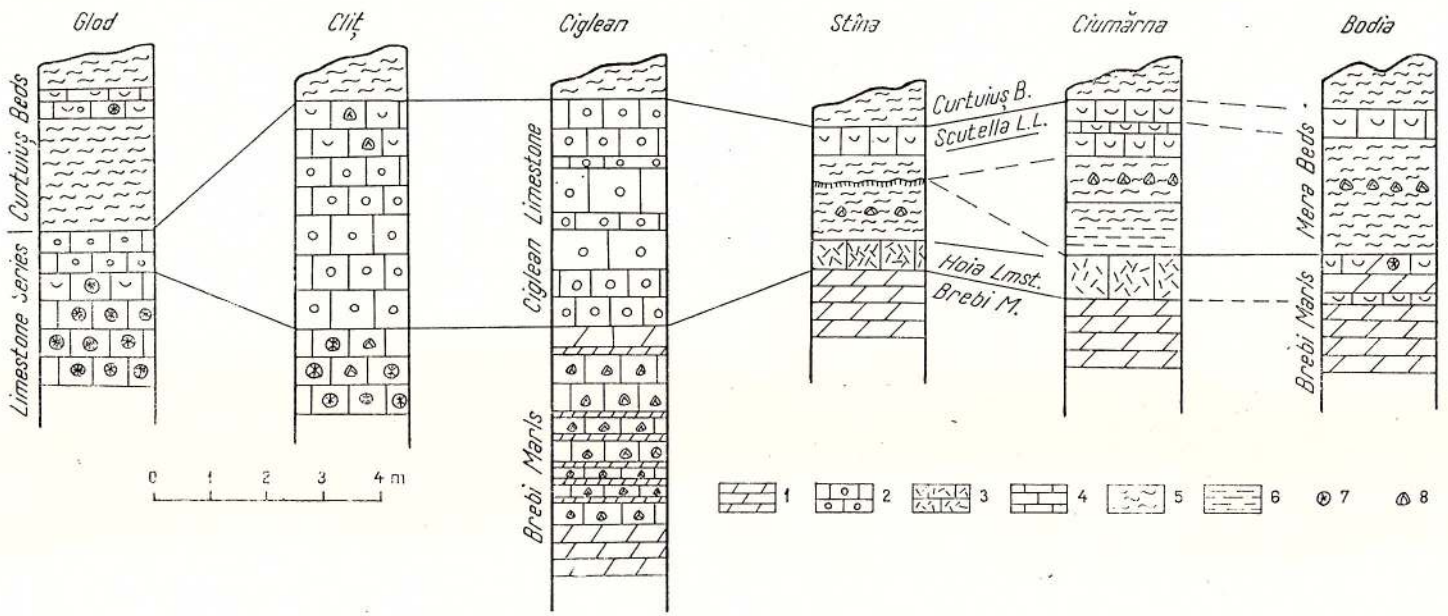


Fig. 2. — The correlation of several sections in the Preluca and Mezeș areas.
1, chalky marls ; 2, corallinacean limestones ; 3, coraligenous limestones ; 4,coquina limestones ;5, marls ; 6, clays ; 7, nummulitids ; 8, miliolids.

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b) an organogeneous, 50 cm thick, limestone containing corals of the *Rhabdophyllia* type, miliolids, nummulitids, *Chlamys*, small oysters, underlying,

c) 50—150 cm of white-greyish marls with miliolids and varied typical marine foraminifera,

d) 60 cm of yellow-greyish marls with brackish microfauna and, in addition, marine foraminifera reworked from the underlying bed,

e) 10 cm of greenish, fine clay,

f) 50 cm of coquina limestone with *Turritella*, *Ampullinopsis*, *Callista*, *Cardium*, followed by clayey marls belonging to the Curtuiuș Beds (*g*).

So far, the Hoia Beds were considered to encompass beds *b-f*. Actually only the limestone bed with corals (*b*) stands for the Hoia Limestone (true Hoia Beds), the upper, coquina limestones (bed *f*) corresponding to the *Scutella* Lower Level of the Mera Beds. The interval of the beds *d-e* of brackish facies represents the equivalent of the lowermost part of the Mera Beds. It is quite possible that a sedimentary discontinuity exists between the two marly beds (*c* and *d*), as suggested by a comparison with other sections (i.e. Ciumărna) where the stratigraphic sequence between the Hoia Limestone and the Curtuiuș Beds includes more lithologic terms, as well as by the reworking of marine microfaunas of the lower, marine bed (*c*) within the overlying, brackish bed (*d*). The possible sedimentary discontinuity between the two marly beds as well as the marine facies of the lower bed (*c*) versus the brackish facies of the upper one (*d*) give support to the affiliation of the lower, marine, marly bed to the Hoia Limestone interval. This marly bed displays a muddy, miliolid, calcareous facies strongly infested with clayey material; however, as will be seen later, the muddy, miliolid-coraligenous facies passes by gradual loss of the corals to a muddy, miliolidic variety ; thus, it is possible that the marly bed (*c*) stands for this coral-lacking variety and, consequently, from the facies point of view, its affiliation to the Hoia Limestone interval does not represent an impediment.

In the Pîrîul Berculuí valley (S of Ciumărna) the Brebi Marls (*a*), which end in calcareous marls containing *Chlamys*, *Panope*, small oysters and scarce nummulitids, are followed by

b) 75 cm of bioconstructed coraligenous limestone with *Rhabdophyllia* and *Hydnophyllia*,

c) 50 cm of green-greyish fine clays, which include at their bottom a centimetric crust of corals belonging to the *Hydrophyllia* type (that were stopped to keep developing by the unfavourable, brackish environment which was just installed), and in the remainder bed contain a brackish fauna consisting of individuals of *Polymesoda* and *Tympanotonos*,

d) 40 cm of slighty slaty, coarse marls,

e) 7 cm of sandy-marly limestone with *Cardium transsilvanicum* and scarce *Polymesoda convexa*,

f) 75 cm of sandy marls with rare cardiids,

g) 40—50 cm of sandy limestone with many individuals of *Cardium transsilvanicum*,

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h) 15 cm of limestone with miliolids bounded by marly joints,

i) 40 cm of coquina limestone with *Turritella*, *Babylonia*, *Callista*, and above, marls firstly with miliolids and then with brackish molluscs belonging to the Curtuiuș Beds (*j*).

From the interval of the beds *b-i* ascribed by the previous authors to the „Hoia Beds" we keep as representing the Hoia Limestone (true Hoia Beds), only the bioconstructed coraligeneous limestone (bed *b*) and the lowermost, coraligeneous part of the suprajacent, marly bed (*c*). A fact to retain is that the sequence between the Hoia Limestone and the Curtuiuș Beds is more complete here than in the Jurteana valley. The interval of the beds *c-i* would represent an equivalent of the lowermost part of the Mera Beds, the bed *i* standing for the *Scutella* Lower Level.

An interesting section is offered by the Groapa Stiavului valley (tributary to the Treznea valley, downstream the locality of Treznea). Here the Brebi Marls (*a*) ending in a hardened marly limestone with *Chlamys*, *Panope*, *Nummulites*, and a centimetric level of encrusting corals at the top, is overlain by

b) 75—80 cm of massive, coraligenous limestone mainly with *Rhabdophyllia*, *Hydnophyllia*, etc., covered by a sandy marl. A little way upstream on the valley, after an observation gap 1 m thick corresponding to a soft rock, the section continues as follows :

c) 20 cm of slaty, coarse marls with *Cardium transsilvanicum*, *Panope angusta*, which gradually passes upwards into

d) 30 cm of gresous-marly limestone with *Cardium transsilvanicum*, *Anodontia globulosa*, *Panope angusta*, *Sinodia incrassata*, *Ampullinopsis crassatina*,

another observational gap, but of a small ampleness (a few dm), and then

e) 10 cm of slaty, coarse marl with miliolids,

f) 15 cm of fine, marly limestone with small pelecypods,

g) 17 cm of slaty, coarse marl with miliolids and scarce pelecypods,

h) 35 cm of slightly coquina limestone with *Turitella biarritzensis*, *T. granulosa*, *Tympanotonos labyrinthum*, *Callista villanovae*, *Polymesoda convexa*, etc. and miliolids,

followed by the marls accommodating brackish fauna of the Curtuiuș type (*i*).

The Hoia Limestone may be easily recognized here (bed *b*) ; in this section it reaches the highest thickness for the Mezeș area. The sequence of the beds *c-h* stands for the equivalent of the lowermost part of the Mera Beds, here ending in the *Scutella* Lower Level (bed *h*).

In the area of development of the muddy, coraligenous facies of the Mezeș area is included as well the outcrop of sandy limestone with corals in the Peștera valley at Hodiș (southernmost part of the Mezeș area). Between Treznea and Hodiș the relief transects the deposits synchronous

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to the Hoia Limestone, situated inwards the basin and belonging to the muddy, globigerinid-bryozoan facies, the deposits of the coraligenous facies supposed to have developed westward being removed by the erosion.

Approximately 1 km WNW of the alignement of the coraligenous strip (above discussed), the stratigraphic interval of the Hoia Limestone accommodates the already mentioned facies variety of the muddy, miliolidic limestones. Outcrops cuting this facies variety occur NW of Ciumărna (on the left tributary stream of the Ciumărna valley), SW of Stîna (the uppermost course of Șipote valley), as well as in several points W of Moigrad.

In the muddy coraligeneous facies as a whole, the corals are represented mainly by branching and lamellous forms, subordinately by digitate forms. Rusu determined among these: *Rhabdophyllia tenuis* Reuss (branching corals) (Pl. III; Pl. IV, Fig. 1), and *Hydnophyllia* cf. *scalaria* (Catullo) (Pl. I, Fig. 1; Pl. II), *Cyathoseris apennina gibossa* (Prever) (Pl. IV, Fig. 2—3), *Stylocaenia* sp. (Pl. I, Fig. 3) and *Actinacis* (Pl. I, Fig. 2) (lamellous corals). The branching corals occur as bush-like colonies, those lamellous — as planar or nodular, crust-like colonies or as largely open, small, flat-fans, and those digitate — as short, massive, crust-like colonies. The coral colonies are usually in situ, in growth position, originating in biolithitic fabrics (bioconstructed limestones); in the case of the colonies of lamellous or digitate corals the biolithitic structure is continuous (welded biolithite), while in the case of the colonies of branching corals the biolithitic structure may be either continuous if the calicia are tangent (welded biolithite) or discontinuous if the calicia are spaced (spaced biolithite). Thus, at Ciocmani occur colonies of branching and lamellous corals, at Brebi and Moigrad — spaced colonies of branching corals, at Stîna — nodular, crust-like colonies of lamellous corals, at Ciumărna and Treznea — fan-like colonies of lamellous corals, crust-like colonies of lamellous or digitate corals, and welded or spaced colonies of branching corals.

In the Mezeș area the coraligeneous colonies of the Hoia Limestone do not make up a continuous biostrome, but within the continuous strip of muddy miliolid-coraligenous facies they seem to be irregularly distributed, giving rise to local concentrations as small bioherms, which considered at a regional scale may be assembled in a discontinuous biostromal level.

In the Preluca area, the muddy, miliolid-coraligenous facies is represented by scattered patches of bioconstructed coraligenous limestones consisting of colonies of lamellous (*Hydnophyllia*) and/or branching corals; such patches occur S of Soimușeni, in the quarry Letca (W of Toplița), N of Răstoci, N of Bizușa, and N of Dăbîceni; in places the coraligenous material appears triturated (i.e. E of Purcăreț).

B) In the area of development of the sandy, corallinacean facies, which corresponds to the outcropping area of the Calca-

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reous Series in Preluca area and may be watched southward up to Ciglean in Mezeș area, the sections are numerous but little various.

In the Pîrîul Pietrii valley (SW of Ciglean) the typical Brebi Marls (*a*) underlie :

b) 3,5 m of limestones with miliolids interbedded with marls,

c) 4 m of corallinacean limestones stratified in beds variable in thickness (0,2 — 1 m) ; these limestones contain, besides coralline algae, nummulitids, bryozoa debris, and scarce shells of *Pycnodonta gigantica* (S o l .).

In the authors' opinion the miliolidic limestone (level *b*) is attached to the Brebi Marls Member, the equivalent of the stratigraphic interval including Hoia Limestone + *Scutella* Lower Level (that is the „Hoia Beds sensu lato") being represented by the corallinacean limestones (level *c*), for this latter we thinking that it is useful to be kept the name of Ciglean Limestone proposed by I l i e s c u in an unpublished report (I l i e s c u et al., 1962)[3].

The Ciglean Limestone, essentially corallinacean, occurs also at N of Ciocmani, S of Piroșa, and in the quarry Băbeni. In the other points of outcropping of the uppermost part of the Calcareous Series, the Ciglean Limestone appears however atypical, strongly enriched in nummulitids. Thus, for instance :

In the Somes River cliff, N of Cliț, the coarse, nummulitic limestones typical of the Calcareous Series are overlain by :

a) 3 m of limestones with *Lithophyllum* and *Nummulites* (the Ciglean Limestone enriched in nummulitids),

b) 1 m of limestone with miliolids and shells of molluscs (corresponding to the *Scutella* Lower Level, here exhibiting lithological individuality).

At Topliță, on the hill slope near the village, at the top of the Calcareous Series 2—3 m of limestones occur containing abundant skeletons of *Nummulites* and *Lithophyllum* and subordinately echinoid remnants, some of them belonging to *Scutella subtrigona* (the Ciglean Limestone enriched in nummulitids, and clearly encompassing the Hoia Limestone unit and the *Scutella* Lower Level unit).

In the locality of Glod, near the spring of the village, the nummulitic limestones of the Calcareous Series which gradually discharge upwards of nummulitids underly :

a) 0,7 m of limestones with *Lithophyllum* and *Nummulites* (the Ciglean Limestone enriched in nummulitids),

b) 2 m of greenish grey marls with *Tympanotonos*, *Globularia*, *Polymesoda*, etc. belonging to the Curtuiuș Beds, followed by skeletal limestones with sorted elements represented by nodules of *Lithophyllum*, pieces of balanids, skeletons of nummulitids, associated with shells of *Tympanotonos*, *Globularia*, *Polymesoda*, whole this skeletal material being embedded in a

[3] O. Iliescu, Aura Naghel, Gh. Mărgărit, Maria Mărgărit, M. Gheorghian, Mihaela Ghe orghian, A. Naghel. Geological Report on the Prospections for Coal in Moigrad-Jibou-Stejerea-Ileanda-Răzoare Region (in Romanian) (1962). Arch. of Institute of Geology and Geophysics, Bucharest.

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marly-calcareous matrix ; these limestones, which may be easily confused with the Ciglean Limestone, stand for reworkings of skeletal material of the Ciglean Limestone in the Curtuiuș Beds. We take advantage on this occasion to draw the attention to the fact that resedimentations and reworkings in the lowermost part of the Curtuiuș Beds stand for a common phenomenon, especially in the Preluca area, being obvious at Var (Pîrîul Cheții valley), in the vicinity of locality of Poienița, and W of Glod quarry ; in these places, at the base or interbedded with the basal marls of the Curtuiuș Beds, a material of Ciglean type, fragmented and sorted, may be noticed together with brackish molluscs typical of the Curtuiuș Beds ; so far, this fact had not been pointed out, and often the limestone beds belonging to this type were erroneously taken for the Hoia Beds.

An other fact worth mentioning is that in places, in the area of development of the sandy corallinacean facies, the algal limestones are replaced by coquina limestones (pelecypods and/or gastropods) where the thalli of coralline algae occur sporadically. That is the situation found in the uppermost course of Brebi valley (NW of Brebi), in Pîrîul Haiturii valley (NW of Ciocmani), N of Piroșa, near Purcăreți, W of Răstoci, E of Valea Chioarului, etc. From all these points R u s u determined the following species : Gastropoda : *Turritella granulosa*, *T. biarritzensis*, *T. clumancensis*, *Tympanotonos labirinthum*, *T. trochlear diaboli*, *Ampullinopsis crassatina*, *Globularia* aff. *auriculata*, *G. patula*, *Polinices catena archatensis*, *P. semperi*, *Thericium intradentatum*, *T.* aff. *globulosum*, *Babylonia caronis*, *Diastoma costellatum elongatum*, *Pugilina bonnetensis*, *Bayania stygii*, *Architectonica subplicatula*, *Cancellaria excellens*, *Trochus trochlearis*, *Tectus lucasianus*, *Xenophora solida*, *Scaphander* aff. *jostisii*, *Collonia* sp., *Terebellum* sp., *Vexillum* sp., *Cassis* sp., etc. Pelecypoda : *Chlamys bellicostata*, *Callista* aff. *villanovae*, *Sinodia incrassata*, *Polymesoda convexa*, *Panope angusta*, *Cardium transsilvanicum*, *Crassostrea plicata*, *Pycnodonte gigantica*, *Anodontia globulosa*, *Bourdotia subornata*, *Glycymeris tenuisulcata*, *Caestocorbula henckeliusiana*. *Perna* sp., *Megaxinus* sp., *Pinna* sp., etc.

C) T h e m u d d y, g l o b i g e r i n i d-b r y o z o a n f a c i e s outcrops only in the Mezeș area as a strip bounding at SE the strip of the coraligenous facies on its Hodiș-Treznea segment. It may be ascertained in several sections among which more important are those of Bozna and Bodia.

In the village of Bozna, on a left tributary stream of Bozna valley, the Brebi Marls Member (*a*) ends in a coarser, indurated, calcareous marl, 0,5 m thick, macroscopically with miliolids and rare impressions of pelecypods ; this marl seems to represent the equivalent of the Hoia Limestone, inasmuch as 25 cm higher (covered interval) there are

b) 1,2 m of greenish grey marls with a faunal assemblage of Curtuiuș type (*Tympanotonos*, *Pirenella*, *Cyrena*),

c) 35 cm of marly limestone with miliolids and fine pelecypods,

d) 40 cm of marly limestone containing miliolids in the upper part,

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e) 35 cm of coquina limestone with *Turritella*, *Callista*, *Ampullinopsis* — the equivalent of the *Scutella* Lower Level,

f) 100 cm of marls belonging to the Curtuiuș Beds.

The interval of the beds *b-e* is considered the stratigraphic equivalent of the lowermost part of the Mera Beds.

In the Bodia valley, 1,3 km upstream of the Jibou-Ciucea road, the following sequence may be noticed :

a) the Brebi Marls are overlain by

b) 25 cm of grey calcareous marls containing abundant, sorted, skeletal material consisting of small nummulitids of *N. fabianii* group, fragments of pelecypods especially *Chlamys*, pieces of balanids, and unbrocken shells of *Chlamys bellicostata* and *Pycnodonte gigantica*; towards the uppermost of the bed the first individuals of *Tympanotonos* and *Cyrena* make their appearence,

c) 25 cm of bluish marls of Brebi type with small foraminifera,

d) 40 cm of well-bedded coquina limestone containing pelecypods (*Chlamys*, small oysters of *Cubitostrea* group) and nummulitids,

e) 5 cm of limestone with miliolids,

f) 200 cm of bluish grey marls with several miliolidic levels,

g) 40—50 cm of hard, skeletal limestone with miliolids, nummulitids, pelecypods, *Turritella*, plates of echinoids, etc., strongly suggesting to represent the equivalent of the *Scutella* Lower Level.

h) interbedded limestones and sandstones containing the faunal assemblage of the Mera Beds.

There is no doubt that a part of the deposits underlying the Scutella Lower Level stands for a correspondent of the Hoia Limestone Member. The intercalations bearing sorted skeletal material and occuring within a sequence of marls of Brebi type represent intraformational resedimentations of materials reworked from the near-shore areas and transported over the coraligenous zone eastward.

In the outcropping sector of the muddy, globigerinid facies in Mezeș area, the deposits, of the muddy, coraligeneous facies are missing, being removed by the erosion. The proof that these facies were initially juxtaposited is given northward, where the drillings performed east of the outcropping strip of the Hoia-Ciglean Limestone, namely at Lupoaia and Jac, entered from the Mera Beds directly the Brebi Marls (Mészáros et al., 1970). The Lupoaia boring was located approximately 4 km inward the basin from the outcrops of Ciglean Limestone at Ciglean, and the Jac drilling was bored approximately 4 km E of the alignement of Hoia Limestone near Moigrad. In these drillings, the first, coarser deposits, with *Turritella clumancensis*, *Cardium transsilvanicum*, *Tellina*, etc., overlying the Brebi Marls stand obviously for the correspondent of the *Scutella* Lower Level, and consequently the interval equilvalent to the Hoia Limestone must be included in the subjacent marls, that is in the upper part of the Brebi Marls Member.

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THE SEDIMENTARY FACIES OF THE STRATIGRAPHIC INTERVAL OF THE HOIA LIMESTONE

The petrological study of the sequences lying in the stratigraphic interval of the Hoia Limestone in the Mezeş and Preluca areas allowed to distinguish 3 synchronous, calcareous, depositional facies : the muddy, miliolid-coraligenous facies; the sandy, corallinacean facies; and the muddy, globigerinid-bryozoan facies. The sections analysed from each of these facies were discussed in the previous chapter. These facies are characterized by proper depositional petrologic features, each facies displaying peculiarities which individualize it sharply in comparison with the other facies. The successive petrological analysis of each facies is presented further on (Pl. VII—XX).

A) The muddy, miliolid-coraligenous facies is a marine, calcareous, muddy skeletal facies and represents the facies exhibited by the whole Hoia Limestone in the Mezeş area (alignement Hodiş-Treznea-Moigrad-Brebi-Ciocmani) (Pl. XIV, Fig. 3—6 ; Pl. XVI, Fig. 5—6 ; Pl. XVII, Fig. 1—3 ; Pl .XVIII—XIX), and by the limestones of several disparate patches (Soimuşeni, W of Topliţa, N of Răstoci, W of Bizuşa, Dăbîceni) in the Preluca area (Pl. XIII, Fig. 4—5).

Texturally, the limestones of this facies are lime wackestones and packstones accommodating coral biolithitic structures. The depositional petrographic components of these limestones are represented by the coral biolithitic structures and by the sedimentary accummulations within the available spaces inside the former (intrabiolithitic fillings) or between them (interbiolithitic sediments); the sedimentary accumulations consist of micrite matrix (lime mud) bearing abundant miliolids, fine pelagic foraminifera (globigerinids and globorotaliids), ostracods, fine pelecypods and sporadically bryozoa debris and coral fragments, to which locally and in minor amount nummulitids, coralline debris and echinoderm plates are added if this facies stays under the influence of the sandy, corallinacean facies. The skeletal grains show different degrees of micritization, the most micritized being sometimes the miliolid tests and allways the coralline fragments, both of them often appearing converted into cryptocrystalline grains. Though the main depositional components of these limestones are the coral biolithitic structures, the micrite matrix, the miliolids and the pelagic foraminifera, among these constituents only the micrite matrix and the miliolids stand for the all over spread and characteristic elements, because there are sectors where the corals and the pelagic foraminifera are missing (the zone lying west of the coraligenous alignement Treznea-Ciumărna-Stîna); therefore, within the muddy, miliolid-coraligenous facies a coral-lacking, miliolidic, muddy subfacies may be distinguished as variety.

To conclude, one may assert that this facies is characterized by more or less sandy lime muds containing scattered and undisturbed (in place) coral colonies.

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B) The sandy, corallinacean facies is a marine, skeletal calcareous facies and largely develops in the Preluca area occupying the whole mass of the uppermost part of the Calcareous Series equivalent to the Hoia Limestone (except those several scattered spots of a coraligenous facies) (Pl. VII—XIII), extending westward in the Mezeș area west of the coraligenous alignement up to Ciglean (the sector NW Brebi-Ciglean-W and N Ciocmani) (Pl. XIV, Fig. 1—2; Pl. XV; Pl. XVI, Fig. 1—4; Pl. XVII).

Unlike the previous facies, this facies consists exclusively of lime grainstones (originally lime sands), calcarenitic to microcalciruditic in grain-size, whose general and usually prevalent depositional petrographic component is represented by entire or brocken corallinacean nodules (almost exclusively species of *Lithophyllum* (Pl. V) among which Rusu identified *Lithophyllum simplex* Lemoine and *L. symetricum* Lemoine); as co-associated constituents are added in variable amounts: nummulitids and other large benthonic foraminifera (*Asterigerina*, *Spherogypsina*, *Rotalia*), miliolids, echinoderms, pelecypods, gastropods, plates of balanids, and bryozoas (Pl. VI, Fig. 1—2).

Several subfacies may be distinguished depending on the percentual participation of the secondary depositional constituents (varieties of the sandy facies): the corallinacean subfacies where the corallinacean tests make up at least 75% of the limestone bulk (Ciglean, N of Ciocmani, S of Piroșa, Băbeni); the nummulitid-corallinacean subfacies standing for the commonest subfacies, where both groups of organisms are abundantly represented; the coquina subfacies encompassing skeletal limestones consisting completely or mainly of shells of pelecypods (i.e. NW of Brebi), of gastropods (i.e. Valea Chioarului), or of both of them (i.e. Fundătura-Creaca, NW of Ciocmani, W of Răstoci); the miliolidic subfacies (W of Ciocmani, NW of Brebi on the uppermost course of the Brebi valley); and the oncolitic subfacies, a quite unusual subfacies, consisting of limestones whose skeletal grains appear coated with concentrical micritic layers and turned into pisoncoliths or macrooncoliths (to which microoncoliths are added in large amounts).

All the skeletal grains show micritizations, among them the miliolids and the corallinacean algae being submitted to strongest micritization; sometimes they display total alteration to a micritic mass evolving to pelets and micritic lumps. The skeletal grains appear completely merged into each other, resulting in a strongly compacted structure lending to the rock the feature of welded grainstone. If the corallinacean tests lie in a direct contact each other, their total micrization and welding may give the impression of micrite matrix in which different other skeletal grains (i.e. nummulitids, miliolids, etc.) float, the rock leaving the appearence of wackestone or packstone. The rocks show not only a total pore compaction resulting in a perfect merging of the grains into each other but also a strong grain compaction supported by the numerous contact-solution phenomena (ascertained between the grains) joining the deformation, interpenetration, and faulting of the grains (accompanied or not by the displacement of the resulting fragments). Therefore the micritization of the skeletal grains

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and mainly of the coralline tests was followed by a total pore compaction and then by a strong grain compaction, finally the rock displaying the grains, mostly micritized, merged and even intermingled. The diagenesis gave rise to a microspar (replacive, very finely crystalline, spar calcite) which developed as patches or networks covering irregular-shaped, large fields replacing equally the micritized corallinacean tests as well as the other micritized skeletal grains, however apparently consumming only the micritized areas of the different skeletal grains. This microspar may represent a neomorphic product but, alternatively, it may stand for the result of the diagenetic sequence including dolomitization on the micritized areas followed by dedolomitization, in this latter case being a dedolomitic product (D r ă g ă n e s c u , in preparation). This microspar, in places aggraded to larger pseudospar, is very finely crystalline (5—15 microns in crystal size) and contains the remaining unaffected skeletal grains as scattered or grouped welded grains floating within this matrix-mimicing microspar, leaving for the rock the appearence of a lime wackestone-packstone with the matrix slightly aggraded; as already stated, this microspar „matrix" actually stands for microspar irregular-shaped networks or patches developed at the expanse of micritized areas of the merged skeletal grains.

C) T h e m u d d y , g l o b i g e r i n i d - b r y o z o a n f a c i e s is a marine, marly, muddy skeletal facies, and appears developed in Mezeș area directly east of the coraligenous alignement Hodiș-Treznea-Moigrad-Brebi; it outcrops at Bodia and Bozna, northward being encountered in the drillings of Jac and Lupoaia (Pl. XX).

It represents the facies typical of the Brebi Marls Member and consists of marls with nummulitids, miliolids, globigerinids and globorotaliids embedded in a clayey-micritic matrix. Southward, in Gilău area, as will be seen later, this facies excels in the abundance of bryozoan tests accommodated in a marly matrix too (the Bryozoa Marls), reason through which generalizing for the whole NW sector of the Transylvanian Basin we have termed this facies: the muddy, globigerinid-bryozoan facies. Within its framework two subfacies may be distinguished: a northern, globigerinic subfacies included in the series of the Brebi Marls, and a southern, bryozoan subfacies accommodated within the series synchronous to the former and termed the Bryozoa Marls.

THE FACIES ZONATION WITHIN THE STRATIGRAPHIC INTERVAL OF THE HOIA LIMESTONE

Watching the distribution of the three facies distinguished over the surface of the Mezeș and Preluca areas, one ascertains that in the Mezeș area they are disposed as juxtaposited strips oriented NE-SW, while in the Preluca area the coraligenous and corallinacean facies make up a mosaic. We present further on the analysis of the areal distribution of the facies distinguished in the stratigraphic interval of the Hoia Limestone (Pl. XXI).

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In the Mezeș area the three facies form three, approximately parallel belts, oriented SW-NE, which from west to east succed as follows : the sandy, corallinacean facies zone ; the muddy, miliolid-coraligenous facies zone ; and the muddy, globigerinid-bryozoan facies zone.

The muddy, miliolid-coraligenous facies zone may be followed from the southern extremity of the Mezeș area, from Hodiș, to its northern extremity, to Ciocmani ; it appears as a continuous, muddy-miliolidic belt accommodating patches of coraligenous colonies along the alignement Hodiș-Treznea-Ciumărna-Stîna-Moigrad-Brebi-Ciocmani ; this zone is narrow (about 1,5 km wide) in the northern segment, in that southern, starting from the locality of Moigrad southward, widening gradually up to at least 2,5 km (westward eroded) on account of the attaching, westward of the strip of muddy miliolid-coraligenous facies, of a strip of coral-lacking, muddy-miliolidic facies (subfacies) making up a facies subzone within the muddy, miliolid-coraligenous facies zone. In the Brebi-Moigrad sector the coraligenous facies zone changes twice its direction displaying a lock. Inasmuch as northward of the locality of Ciocmani we did not find the outcropping muddy, miliolid-coraligenous facies making up a continuous strip, and on the other hand because northward and eastward of Ciocmani the sandy, corallinacean facies is present in outcrops, one may conclude that it is possible that starting from the locality of Ciocmani the coraligenous facies zone changes its direction and sinks under younger deposits, turning firstly towards SE (inscribing in a small gulf whose concavity is directed southward) and then directing eastward ; alternatively, it is possible that this facies zone disappears directly northward of Ciocmani.

The sandy, corallinacean facies zone lies directly west of the coraligenous zone, starting from the parallel of the locality of Ciglean towards north (southward being either eroded or, more probably, undeposited owing to the joining of the southern segment of the coraligenous facies zone to the mainland lying directly westward). This facies zone gets wider northward, largely linking to the facies of the Preluca area ; actually this facies zone stands for the southwestern termination of the area of development of the sandy, corallinacean facies which covers extensive portions within Preluca area.

The muddy, globigerinid-bryozoan facies zone develops directly east of the coraligenous facies zone, starting from the alignement Bodia-Bozna-Jac-Lupoaia and extending largely eastward, probably occupying the whole central part of the northwestern sector of the Transylvanian Basin.

In the Preluca area the outcrops show a single extensive facies zone — the sandy, corallinacean facies zone (represented mainly by the nummulitid-corallinacean subfacies) — containng patches of muddy, miliolid-coraligenous facies as small, scattered islands dispersed within the sand-sized corallinacean-nummulitic mass. Therefore the two facies — coraligenous and corallinacean — make up together a mosaic in which the coraligenous facies occurs as sporadic lenses floating in the mass of the corallinacean facies. The deposits of this facies zone occur starting from

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the top of the Preluca massif southward approximately up to the parallel of the locality of Glod ; south of this locality the deposits of the concerned stratigraphic interval sink under younger sediments. But certainly southward these facies are replaced by the muddy, globigerinid-bryozoan facies. It is possible that at the limit between the corallinacean facies zone with islands of coraligenous facies and the muddy, globigerinid-bryozoan facies, a more or less continuous, muddy, coraligenous facies zone be developed, similarly to the situation encountered in the Mezeș area.

As a conclusion, generally from NW towards SE one may consider that three facies zones are developed : an extensive, sandy, corallinacean facies zone able to rich more than 15 km in width in the Preluca area, geting thinner towards SW and disappearing south of Ciglean ; a narrow, muddy, miliolid-coraligenous facies zone ; and a broad, muddy, globigerinid-bryozoan facies zone covering all the remaining sectors of the central part of the NW Transylvanian Basin.

The published geological data reveal that the facies zonation distinguished in the Mezeș and Preluca areas verifies *in the Gilău area* too. Thus, the breccia-like limestones bearing molluscs, corals (skeletons of *Rhabdophyllia* and *Hydnophyllia*), balanids, and nummulitids of the Hoia Hill as well as those, less coraligenous, of Mînăștur Wood (Cluj-Napoca) are obviously included in the muddy, miliolid-coraligenous facies zone, very poorly represented in the Gilău area. The muddy, globigerinid-bryozoan facies zone, consisting here of bryozoa-bearing marls included in the uppermost part of the Bryozoa Marls Member and directly overlain by the Mera Beds (Member), is largely developed instead in this area, a fact which most often determined the previous researchers to separate, as Hoia Beds, deposits of the lowermost part of the Mera Beds.

The stratigraphic equivalent of the Hoia Limestone starting from Cluj towards the interior of the basin (towards NW) would be placed in different points at the following levels (Pl. XXI) :

At the mouth of the Popești valley, in an outcrop described by Koch (1894, p. 319—320) and Moisescu (1968, p. 493, 497), the stratigraphic equivalent of the Hoia Limestone lies under the *Scutella* Lower Level (lowermost part of the Mera Beds), that is in the interval 7—8 m thick consisting of an alternation of marls (of the Bryozoa Marls type) and skeletal marly limestones containing abundant fragments of *Chlamys*, balanids, crustaceans, solitary corals and nummulitids. Moisescu eroneously interpreted the *Scutella* Lower Level of the Mera Beds as representing the Hoia Limestone (true Hoia Beds).

In the already classical section on the Berecoaia valley, at Mera (Koch, 1894, p. 339—340; Bombiță and Moisescu, 1968, p. 710—712), a nannoplancton fauna occurring in the Hoia Limestone in the type section, is placed here within the Bryozoa Marls, whose upper part would therefore stand for an equivalent of the Hoia Limestone (Mészáros et al., in press). Worth mentioning is the fact that Koch (1894, p. 340), although he eroneously considered the first, indurated and coarser

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deposits in this section as equivalent to the Hoia Limestone, correctly placed the base of the Mera Beds at the first bed bearing *Scutella*, bed which later will be unjustly included by some authors in the Hoia Beds.

Following the same direction, at Şardu one ascertains, as K o c h (1894, p. 320) correctly remarked already, that the place of the typical Hoia Beds is occupied by a level of marly clays 2 m in thickness containing the shells of the Bryozoa Marls and many balanids.

Towards the center of the basin, in the drilling of Sîncraiu Almaşului, molluscs-bearing sandstones of a Mera Beds type directly overlie the Bryozoa Marls, lacking any megascopical element allowing to separate an equivalent of the „Hoia". In the ancient conception this equivalent would have been separated in the lowermost part of the Mera Beds; now it is proved that it must be placed in the uppermost part of the Bryozoa Marls.

The problem concerning the correlation of the deposits in the stratigraphic interval of the Hoia Limestone is not completely cleared up. However the additional details and specifications which will be got henceforth are expected not to change the general, here-proposed scheme of facies zonation.

THE GENETIC SIGNIFICANCE OF THE DISTINGUISHED FACIES ZONES

Taking into account the fact that the teritory of the Apuseni Mountains stood for a mainland area which included, in its northern part, the Gilău massif, the crystalline ridge of Mezeş and the Şimleul Silvaniei Basin (where the drillings proved the lack of the eocene-oligocene deposits), as well as the fact that the crystalline massives of Preluca and Ticău were shallow submerged areas together making up a large ridge representing the prolongation towards NE of the western mainland, one may conclude that the sandy, corallinacean facies zone was in a position juxtaposited to the shore, and the muddy, globigerinid-bryozoan facies zone was in an off-shore position.

The gradual loading in lime mud from the near-shore, corallinacean facies zone towards the off-shore, globigerinid-bryozoan facies zone, the high occurrence of the large benthonic foraminifera and of the coralline algae in the sandy, corallinacean facies zone and the high frequency of the globigerinids and globorotaliids starting from the outer part of the coraligenous facies zone and continuing in the globigerinid-bryozoan facies zone, the development of the hermatipic corals unde muddy facies being under the influence of both near-shore facies (the occurrence of the coralline algae and large benthonic foraminifera) and off-shore facies (the occurrence of globigerinids and globorotaliids), all these data suggest the following genetic interpretation on the three facies zones distinguished :

a) the sandy, corallinacean facies zone corresponds to the shelf-platform large area, standing for a corpusclar (sandy) skeletal marine facies (skeletal lime sands) of shallow, agitated water;

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b) the muddy, miliolid-coraligenous facies zone covered the shelf-edge and shelf-slope narrow areas, representing a muddy skeletal marine facies (coral bioherms in lime muds) of shallow, but quiet water;

c) the muddy, globigerinid-bryozoan facies zone stands for a basinal marine facies expressing quiet, rather deep water.

In other words, during the sedimentation of the Hoia Limestone, the shelf-platform areas were occupied by a sandy corallinacean facies, the shelf-edge (+slope?) areas were the sites of development of a muddy miliolid-coraligenous facies, and the basinal areas accommodated a mud, globigerinid-bryozoan facies. The reccurrences of mud coraligenous facies within the area of the sandy corallinacean facies zone may be explained by the occurrence of some small depressions with shallow, but quiet water which, reproducing the shelf-edge conditions, favoured the development of the mud coraligenous facies as tiny patches.

As a conclusion, the northwestern sector of the Transylvanian Basin represented, during the deposition of the Hoia Limestone, a gulf in whose center clayey-calcareous oozes of a basinal, globigerinid-bryozoan facies accummulated, and colaterally, on the margins, muddy, coraligenous deposits of a shelf-edge facies and skeletal (mainly corallinacean-nummulitic) sands were deposited, the latter covering the whole shelf platform between the shelf edge and the shore. Supposingly, the water depth of the sedimentary basin gradually increased from the sandy corallinacean facies zone (meters to a few tens of meters) towards the mud, globigerinid-bryozoan facies zone (50—60 meters or more).

CHRONOSTRATIGRAPHIC ASPECTS RELATED TO THE AGE OF THE HOIA LIMESTONE

Most of the researchers considered the Hoia Beds as representing the first Oligocene term of the Transylvanian Basin, terming them either Lower Oligocene (Hofmann, 1887), either Ligurian (Koch, 1894), or later on, Lattorfian. This age, established especially on the basis of the molluscan assemblage, seemed not to be in aggreement with that suggested by foraminifera. Bombiță, for instance, notes the occurrence in the Hoia Limestone of some evoluated forms of *Nummulites fabianii* standing for passage forms to *N. intermedius*, concluding that this limestone would represent a Priabonian member of passage to the Oligocene series (Bombiță, 1963; Bombiță and Moisescu, 1968).

Quite recently, the Oligocene age of the Hoia Limestone was reconsidered on the basis of the nannoplancton studies. Thus, Mészáros et al. (1973) describe from this limestone a nannofloral assemblage, ascribing it a lattorfian age (estimating the standard zone NP 21), and Martini and Moisescu (1974) point out the presence of the *Helicopontosphaera reticulata* zone (the standard zone NP 22), which is included by the authors together with the zone NP 21 in the Lower Oligocene.

Without having any doubt on the veracity of this last biozonation, which otherwise for the Bryozoa Marls corresponds to the zonation pre-

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viously accomplished by P o p e s c u and G h e ţ a (1972), we intend only to bring into the discussion the assigning of the standard zone NP 22 to the Lower Oligocene, a fact which in our opinion is wrong.

It is well known the fact that the Latdorf Sands as well as the Silberberg Formation proposed as neostratotype for the Lattorfian contain the *Ericsonia subdisticha* zone (standard zone NP 21) (M a r t i n i , 1969; M a r t i n i and R i t z k o w s k i , 1968). On the other hand, the Rupelian of D u m o n t encompasses as main component the Berg Sands and the Boom Clays, both of them included by M ü l l e r (1970) on the basis of the nannoplancton in the *Sphenolithus predistenthus* zone (standard zone NP 23). But the Rupelian of Belgium does not follow directly the Lattorfian, which here is separated under the name of Lower Tongrian (including the standard zone NP 21; M a r t i n i and M o o r k e n s , 1969), between them being interposed the Upper Tongrian (containing in the Vieux Joncs Sands the nannoplancton assemblage of the standard zone NP 22; M a r t i n i , 1969).

The question is where must the Upper Tongrian be attached, to the Lattorfian type or to the Rupelian type, or in other words where must this chronostratigraphic subdivision be included, in the Lower Oligocene or in the Middle Oligocene? To work out this question we remind the fact that B a t j e s (1958) proved, using the foraminifera, that the Upper Tongrian represents a lateral facies of the Early Rupelian, and that in Helmstedt region of Germany the deposits directly younger than the Silberberg Formation (Lattorfian) are for long time designated under the generic term of „Rupelton". This latter, in Brick yard Alversdorf, contains the nannoplancton of the *Cyclococcolithus margaritae* subzone (the lower part of the standard zone NP 22) (R o t h , 1970).

Taking into account these arguments it is evident that the Upper Tongrian and the equivalent deposits must be included in the Middle Oligocene, and together with them the *Helicopontosphaera reticulata* zone too (as already have proceeded some authors : R o t h , 1970; R o t h et al., 1971). In addition worth mentioning is the fact that this interpretation is in agreement with the original sense given by B e y r i c h to the Middle Oligocene.

As a consequence of the above-mentioned facts, the age ascribed by M a r t i n i and M o i s e s c u (1974) to the Hoia Beds on account of the presence of the standard zone NP 22 must be reinterpreted. They do not belong to the Lower Oligocene (Lattorfian), but to the Middle Oligocene (Rupelian s.l.) and correspond temporally to the Upper Tongrian of Belgium.

We have to mention that this age was already put forward by R u s u in an unpublished report (p. 38)[4] where the estimation was done as a result of the study of gastropods : the Hoia Beds and the Mera Beds belong

[4] A. R u s u. The Stratigraphy of the Oligocene Deposits of Treznea-Poiana Blenchii Area on the Basis of the Gastropod Fauna. (in Romanian) (1967) Arch. of Institute of Geology and Geophysics, Bucharest.

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to the Upper Tongrian (considered as the first term of the Oligocene series). In this sense, this author applied, then and later in a published paper (Rusu, 1972, Fig. 3), for the two members the term of „Lower Oligocene" (incorrectly used for the Upper Tongrian), considering the Lattorfian as belonging to the Eocene period.

An other problem to solve is the belonging of the Lattorfian to the Eocene or to the Oligocene; this matter keeps entailing much discussion and we think that the analysis of the Paleogene of the mesogean realm may bring some contribution to taking of a decision as adequate as possible.

If it is proved that the Lattorfian corresponds temporally to the whole interval of the Priabonian type (the situation encountered in the Transylvanian Basin pleads for this opinion), it will be adequate to consider it as Eocene. For the time being it may be indirectly deduced that the *Nummulites retiatus/problematicus* subzone which crowns the Priabonian (Cita, 1969), is equivalent to the Lattorfian. The above-mentioned subzone of nummulitids corresponds in terms of planctonic foraminifera to the *Globigerina gortanii gortanii* zone (Cita, 1969); but Roth et al. (1971), who studied concurrently the nannoplancton and the pelagic foraminifera of several paleogene sections of Italy, ascertain that the *Globigerina gortanii gortanii* zone contains the assemblage of the *Ericsonia subdisticha* zone (occurring also in the Lattorfian type). For this reason the cited authors opinate for the exclusion of the *Globigerina gortanii gortanii* zone from the Priabonian definition, proposal which, given the situation in the mesogean realm, has little chance to be adopted. More suitable it seems to be the placement of the Lattorfian within the Eocene column, as already recommended by Krutzsch and Lotsch (1957) and other authors, a fact which would be in a good agreement with the conception of Cavelier (1968) also stated in the report of this latter at the Colloquium on the Eocene in 1968 at Paris (Cavelier, 1969).

It is well known the fact that the West-German geologists hardly want to keep this stage included in the Oligocene period (maybe for this reason they broaden the Lower Oligocene interval unjustifiedly including here the standard zone NP 22 and thus giving to this subdivison a more oligoceneous appearence) and this situation is preserved as such sooner by virtue of the priority principle (the original definition of the Oligocene given by Beyrich), because the character of the faunas does not plead for this interpretation.

The adoption of the formula Lattorfian-Eocene would bring instead much advantage. On the one hand the Upper Eocene of the northern realm (Barthonian s.l.) would become equal to that of the mesogean realm (Priabonian), and on the other hand the delicate problem of the Eocene/Oligocene boundary would be solved. So, this boundary would correspond to the bionomic threshold given by both the macro- and microfauna of the Mediterranean bioprovince, by the molluscan fauna of the boreal bioprovince, and by the mammalian fauna („Grand Coupure" de Stehlin) of

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Europe. Placed over the Lattorfian, the Eocene/Oligocene boundary coincides with that traced between Ludian and Stampian (including at the base the Sannoisian facies) in the Paris basin, between Lower Tongrian and Upper Tongrian (Henisian) in Belgium, between Kievian and Harkovian in southern Ukraina, and at the top of the Priabonian in the mesogean realm (Fig. 3)

In the Transylvanian Basin this boundary would be placed, as suggested by the nummulitids and macrofauna — at the base of the Hoia Limestone and of its stratigraphic equivalents, and as required by the nanno-

	Mesogean Realm	Paris Basin	Belgium	N.W.Germany	S.Ukraina
OLIGOCEN	↑ Oligocen	↑ Stampian	U.Tongrian	↑ Rupelian s.l.	Harkovian
EOCEN	Priabonian ↓	Ludian ↓	L.Tongrian	Lattorfian	Kievian ↓

Fig. 3.—The chronostratigraphic correlation accepted in current paper around Eocene/Oligocene boundary. (A. Rusu).

plancton—a few meters below, at the level of disappearence of *Cyclococcolithus formosus* (if the limit between the standard zones of nannoplancton NP 21 and NP 22 is thought as representing the Lattorfian /Rupelian s.l. boundary).

GENERAL CONCLUSIONS

Emphasizing the fact that the content of the Hoia Beds must be restricted to the coraligenous calcareous facies (the Hoia Limestone) and that the *Scutella* Lower Level, which all the recent papers use to include in the Hoia Beds, actually belongs to the Mera Beds, the authors establish for the first time that the deposits overlying the Brebi Marls Member and underlying the Curtuiuș Beds (Member) and usually separated as „Hoia Beds" in the Mezeș area correspond both to the Hoia Limestone of the Hoia Hill (Gilău area) and to the lowermost part of the Mera Beds (including the *Scutella* Lower Level) of their stratotype section (Gilău area too). The facies of *Lithophyllum*-bearing Limestones, developed in the same stratigraphic interval starting from Ciglean northeastward, between Ciglean and Ciocmani being stratigraphically located between the Brebi Marls and the Curtuiuș Beds and from Ciocmani eastward representing the uppermost part of the Calcareous Series Formation directly underlying the Curtuiuș Beds, is proposed to be termed the Ciglean Limestone ; it displayes an other stratigraphical extent, namely corresponding to the interval encompassing the Hoia Limestone and the *Scutella* Lower Level. Therefore in the area of development of the Calcareous Series (Preluca area) the

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equivalent of the Hoia Limestone would be placed in a subterminal position, and the *Scutella* Lower Level — at the top of the Calcareous Series.

Presenting the outcropping sections in the concerned stratigraphic interval, the authors draw the attention on the reworkings of skeletal material of the Ciglean Limestone in the suprajacent Curtuiuș Beds, material which is able to originate in calcareous beds giving the false impression of standing for the Ciglean Limestone.

In the Mezeș area, the authors point out an obvious facies zonation including from the shore inward the depositional basin : the sandy, coralinacean, marine facies zone covering the shelf-platform areas (corresponding to the Ciglean Limestone) ; the muddy, miliolid-coraligenous, marine facies zone expressing shelf-edge conditions (corresponding to the Hoia Limestone) ; and the muddy, globigerinid-bryozoan, marine facies zone occupying the basinal environments (corresponding to the uppermost part of the Brebi Marls in those sectors where these latter substitute the Hoia Limestone). Thus it is proved the fact that *both the Hoia Limestone and the lower part of the Ciglean Limestone stand for correspondents of the uppermost part of the Brebi Marls*, replacing the latter ones in the marginal sectors of the basin in the Mezeș area.

The data of the literature suggest that the zonation ascertained in the Mezeș area keeps valid for the Gilău area too.

In the Preluca area, a single facies zone outcrops — the sandy, corallinacean facies zone — accommodating small islands of muddy, miliolid-coraligenous facies within its nummulitic-corallinacean sands ; it is to be supposed that the other facies zones distinguished in the Mezeș area are continued in this area too, under younger deposits.

Taking into account the genetic significance of the distinguished facies zones as well as the stratigraphic sequences in the investigated teritory, one may conclude that *the deposits of the stratigraphic interval of the Hoia Limestone correspond to a phase of rapid regression, in which phase the basinal marly Brebi facies withdraws inwards the basin, being replaced shoreward by calcareous, shelf facies* represented by lime muds accommodating small coral bioherms and by nummulitid-corallinacean lime sands.

The analysis of the available data concerning the age of the Hoia Limestone and especially of those data provided by the nannoplancton studies (the standard zone NP 22 established by M a r t i n i and M o i s e s c u , 1974), points to the fact that *the Hoia Limestone stratigraphic interval is temporally equivalent to the Upper Tongrian*, which must be included either in the Middle Oligocene if the Lattorfian is considered as belonging to the Oligocene period, or in the lower part of the Oligocene if the Lattorfian is placed in the Upper Eocene.

Noticing that the Lattorfian of the northern realm correlates with the uppermost part of the Eocene of the mesogean realm, R u s u opinates for the location of the Eocene/Oligocene boundary between the Lattorfian and the Rupelian s.l. (including the Upper Tongrian) ; in the Transylvanian Basin, this boundary would underlie directly the Hoia Limestone.

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QUESTIONS

B. Popescu: 1. Why do you exclude the NP22 zone from the Lattorfian (Lower Oligocene according to Martini's classification)?

2. If the limestones with *Scutella* and *Nummulites fabianii, N. intermedius, N. incrassatus, N. vascus* from the terminal part of the Calcareous Series (Cozla Beds) do not represent an equivalent of the Hoia Limestone, where might be found this equivalent within the Calcareous Series?

3. What is the significance of the large amounts of matrix in the limestones of the corallinacean sand facies which you consider to represent the materialization of an agitated environment, hence where this matrix could not have resisted the ,,washing"?

Answer: 1. The Lower Oligocene, as defined by Beyrich, corresponds exactly to the Lattorfian stage, and the latter includes in its stratotype only the association of the Nannoplankton Standard Zone NP 21. To include the NP22 Zone, characteristic of the Upper Tongrian, in the Lower Oligocene is an error, since it was proved that the Upper Tongrian represents the Early Rupelian, and consequently should be placed in the Middle Oligocene of Beyrich (or, in our acception, in the lowermost part of the Oligocene, provided the Lattorfian would be considered as Eocene in age). Hence the NP22 Zone should not be artificially enclosed in the Lattorfian.

2. The limestones with *Scutella* from the uppermost part of the ,,Calcareous Series" correspond to the *Scutella* Lower Level of the Mera Beds (Gilău area), and the equivalent of the Hoia ,,Limestones" is located in the ,,Calcareous Series" directly below it, i.e .in a subterminal position.

3. In the opinion of the authors the corallinacean facies are composed solely of skeletal corpuscles, being devoid of micrite matrix; the appearance of matrix is brought about either by micritization processes accompanied by the welding of the corpuscles, or by the development of very finely crystalline microspar as networks or irregular patches. This problem is approached in detail in the chapter reserved for the petrographic description of the sedimentary facies and illustrated in photo plates annexed to this paper.

M. Săndulescu: 1. The authors assert that the corallinacean facies closes towards south-west. How can be explained, according to this interpretation, the lack of the foregoing facies in the entire area south of Moigrad, where the coraligeneous muddy facies commes into a direct contact with the mainland (as interpreted by the authors)?

2. Recently B. Popescu presented a different paleogeographic image which presumes a westward prolongation of the facies zones oriented east-west. How do the authors consider this interpretation?

Answer: 1. The authors do not assert but just assume the ending of the coralline facies towards south-west, in accordance with the petrological evidences.

As shown also in the text of the paper, we think that the Preluca area, where the sandy corallinacean facies was largely developed, formed an extensive plateau covered by several meters of water and bordered westwards by a mainland with a low relief. These conditions which favoured the formation of real ,,algal strands", were no more accomplished south of Moigrad. We assume that here the depth of the basin increased faster, and the sea shore was higher supplying the terrigenous material which from north to south got more abundant and coarser. In the same direction the corallinacean thalli get gradually scarcer in the rock finally vanishing in the Stîna-Treznea sector. These data correlated with the outlines of facies zones suggest that here the corallinacean facies zone is missing, and the muddy coraligenous facies zone with miliolids directly joins the shore. Therefore, the rapid increase in the basin depth, and the development

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of some shelf-edge conditions in the close vicinity of the land would have represented the chief factor responsible for the facies configuration under discussion.

2. The paleogeographic image implying the westward prolongation of the facies zones presented by P o p e s c u refers to the stratigraphic interval located beneath the Hoia Limestone.

DISCUSSIONS

V. M o i s e s c u : We can refer the Hoia Limestone only to deposits cropping out in the Hoia Hill within the type section. A formation similar to it is no more encountered in the Transylvania Basin.

I do not agree with the facies interpretation put forward by the authors for the Hoia Beds. It remains only a working hypothesis.

Hoia Limestone is a regressive formation, still not marking a sharp regression.

Hoia Limestone belongs to the Lower Oligocene (post-Lattorfian) and to the Middle Oligocene although the molluscan fauna are partially resembling those from the superjacent deposits.

Taking into account its nannoplankton Hoia Limestone should pertain to the NP22 biozone, i.e. to the upper part of the Lower Oligocene. This biozone is confined to the Upper Tongrian, fact also admitted by the authors. The Upper Tongrian is, however, the chronostratigraphic equivalent of the upper part of the Lower Oligocene, and not of the Middle one as it is considered in this paper.

Answer : We do not share the opinion of V. M o i s e s c u, according to which the Hoia Limestone might be recognized only in the Hoia Hill. In our view any occurrence of coraligenous facies at this stratigraphic level should be tabulated as Hoia Limestone, i.e. at Stîna, Ciumărna or Treznea, in the Mezeş area.

Paleontological evidence bearing on the equivalence of Hoia Limestone with the terminal part of Bryozoa marls does exist even in the Gilău area. Thus M é s z á r o s, I a n o l i u and L e b e n s o n found the nannoplankton association of the Hoia Limestone from Cluj in the uppermost deposits of the Bryozoa marls (hence beneath the *Scutella* Lower Level) from the Mera type section (see also M o i s e s c u, M é s z á r o s, 1974, Plate VI). Therefore we are confronted with a facies variation from the border of the basin (coral-bearing miliolid lime muds) inwards (marls with globogerinids and bryozoas).

It seems that I was not sufficiently explicit, but in the paper it is clearly argued why the Upper Tongrian cannot be assigned to the Lattorfian, between them being located the Lower Oligocene/Middle Oligocene boundary in the northern realm, or the Eocene/Oligocene boundary in the Mesogean realm (as it is also the case of the Transylvania Basin).

B. P o p e s c u : 1. From the parallelism you presented as an answer to my question it results that the Hoia Beds of the Hoia Hill with nummulites of *fabianii-intermedius* and *incrassatus-vascus* type, corals still without *Scutella* have an equivalent in the sequence underlying the limestone of a similar paleontologic content but in addition with *Scutella* from the terminal part of the Calcareous Series (=Cozla Beds), which this time is parallelized with the lowermost part of the Mera Beds (Upper Tongrian-Rupelian s.s.) Thus you arrive at a conclusion entailing a situation unique of its kind, nowhere recognized, which implies the presence of forms of nummulites of the *fabianii* type in deposits assigned to the Middle Oligocene.

2. I likewise doubt as to existence of some reefs of the ,,patch reef" type in the Hoia Beds from the Mezeş area. In the course of our investigations, which the last years have compri-

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sed this zone too, I never encountered in the Hoia Beds corals in growth position. Numerous examples even of colonial corals do exist, but these are brocken and form accumulations yielding limestones with corals.

3. Finally I consider that the corals quoted in the corallinacean facies zone could not develop in depression zones formed in the skeletal sands and muds, as asserted by the authors. In zones with present-day sediments where fine-grained carbonate sediments develop, coral colonies do not grow in excavations, where they may be readily overlain by carbonate mud transported by the water movements from this facies zone (agitated hydrodinamic regime in this zone, according to the authors of this paper).

4. The paper presented by me last year led us to a different paleogeographic picture as regards the sedimentation of the Cozla Beds interval (=Calcareous Series), the Cluj Limestone, the Bryozoa Marls, the Brebi Marls. It is noteworthy that the facies of the regressive Hoia Beds mark a curve following, in general, the present-day shape of the Transylvania Basin, fact also asserted by the authors of this work. According to our interpretation the significance assigned to these facies was somewhat different, as we consider the shore as being located in the Gilău Massif Zone with a terrigenous carbonate shelf northwards. In front of the shelf a channel (=shelf lagoon) with pelitic sediments (Bryozoa Marls, Brebi Marls) and farther northwards a platform with a pure carbonate sedimentation are found.

Answer: We have to take into account that at the level of the Hoia Limestone, we are not anymore in the presence of some typical forms of *N. fabianii*; they actually stand for some mutations (according to B o m b i ţ ă other than *N. retiatus* or *N. problematicus*) which cannot be so far used as biostratigraphic arguments as even their age is to be established. The examination of other groups of fossil organisms enables us to state that the deposits including these nummulites correspond either to the Upper Tongrian of the nothern realm or to the lowermost Oligocene of the Mesogean realm.

The outcrops exposing the bioconstructed limestones are rather scarce and small, so that they may be easily overlooked.

G h . B o m b i ţ ă : The spirited character of discussions is explained by the bulk of data upon which each author is relying, data contained in their Ph. D. theses.

In Romania the Transylvania Basin is the most representative domain for the stratigraphy of the Paleogene, and any discussion referring to this domain are welcome. The disagreements which persist in defining the Eocene/Oligocene boundary are due to the frequent and exclusive utilization of one of the characteristic macro- and micro-fossil groups: mollusks, small or large foraminifera, nannoplankton, as well as the rigid use of data from some recent works which approach this problem, and whose conclusions are sometimes only apparently and theoretically definitive.

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PLATE I

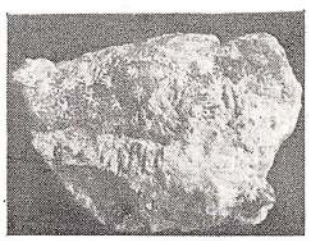
Fig. 1. — *Hydnophyllia* cf. *scalaria* (C a t u l l o) (x1,1) Hoia Limestone, Stina.
Fig. 2. — *Actinacis* sp. (×2), Hoia Limestone, Treznea.
Fig. 3. — *Stylocaenia* sp. (×2), Hoia Limestone, Treznea.

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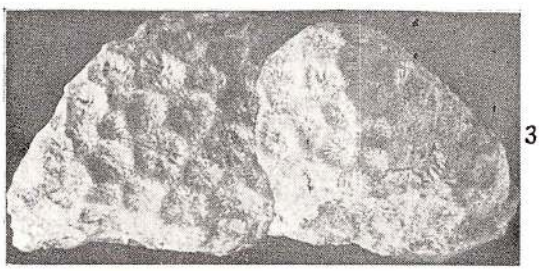
A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. I.



1



2



3

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PLATE II

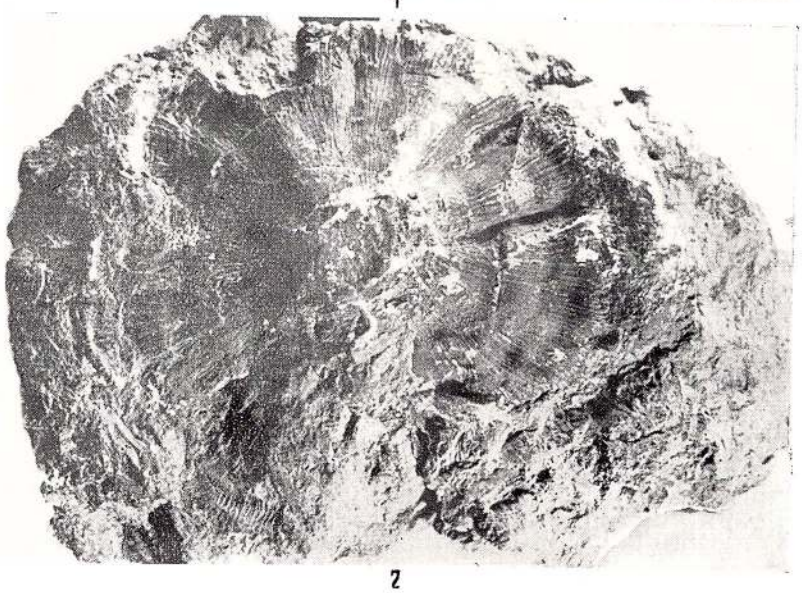
Fig. 1, 2. — *Hydnophyllia* cf. *scalaria* (C a t u l l o) (×1)
Hoia Limestone, Ciumărna. 1, the upper side. 2, the lower side.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. II.



1



2

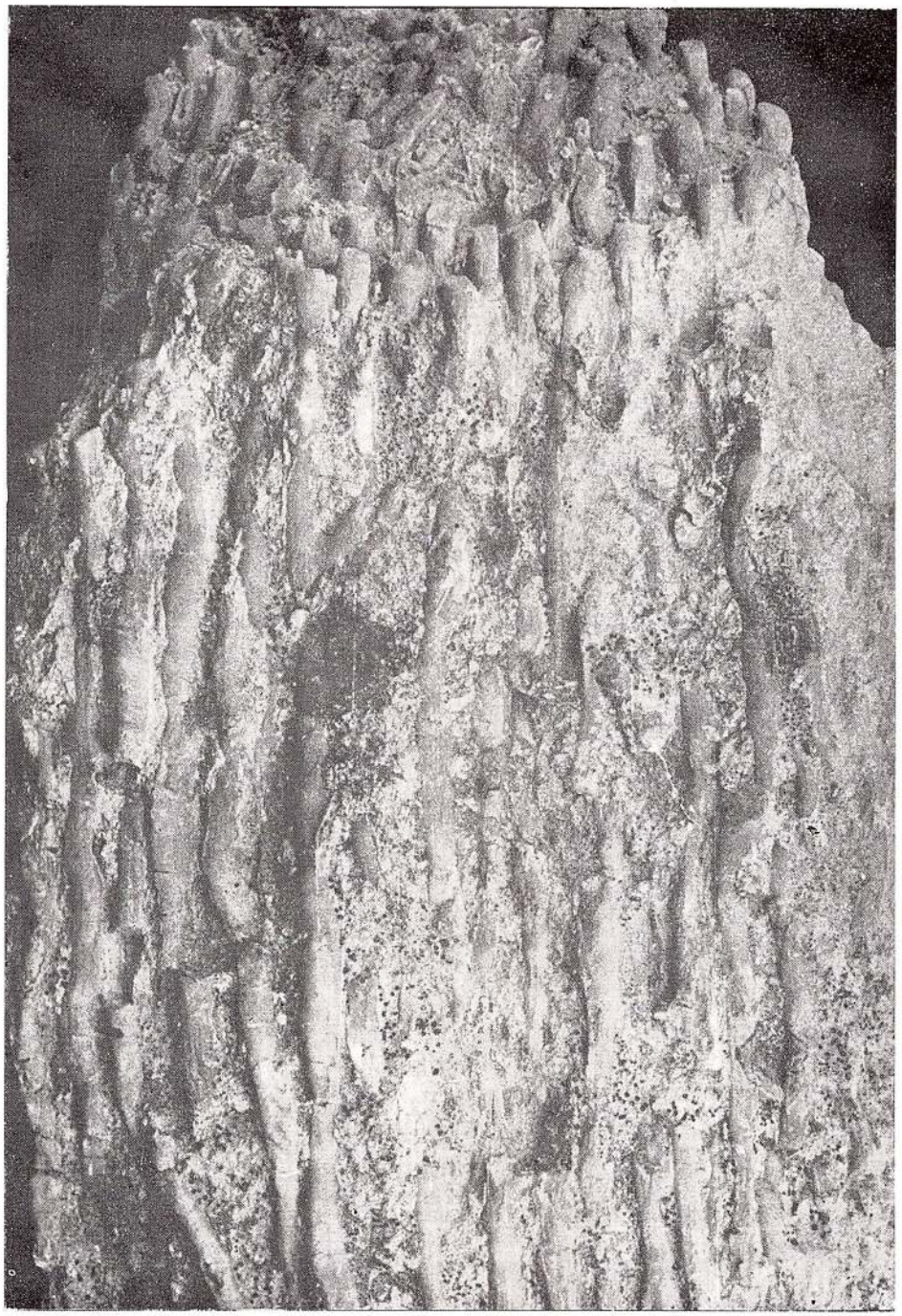
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PLATE III

Rhabdophyllia tenuis Reuss (x1), Hoia Limestone, Treznea.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. III.



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PLATE IV

Fig. 1. — Coraligenous limestone with *Rhabdophyllia tenuis* Reuss (×1). Hoia Limestone, Brebi.

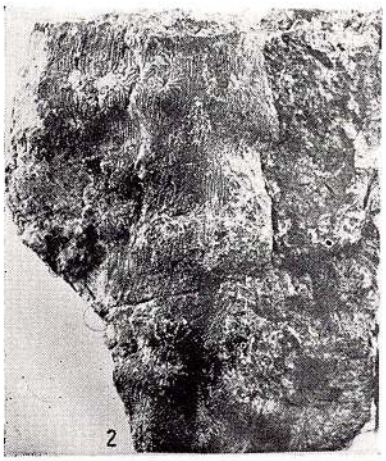
Fig. 2, 3. — *Cyathoseris apennina gibbosa* (Prever) (×1) Hoia Limestone, Treznea. 2, the upper side. 3, the lower side.

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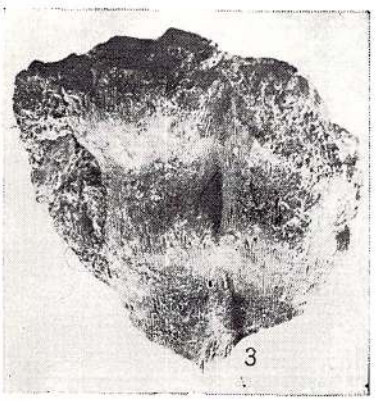
A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. IV.



1



2



3

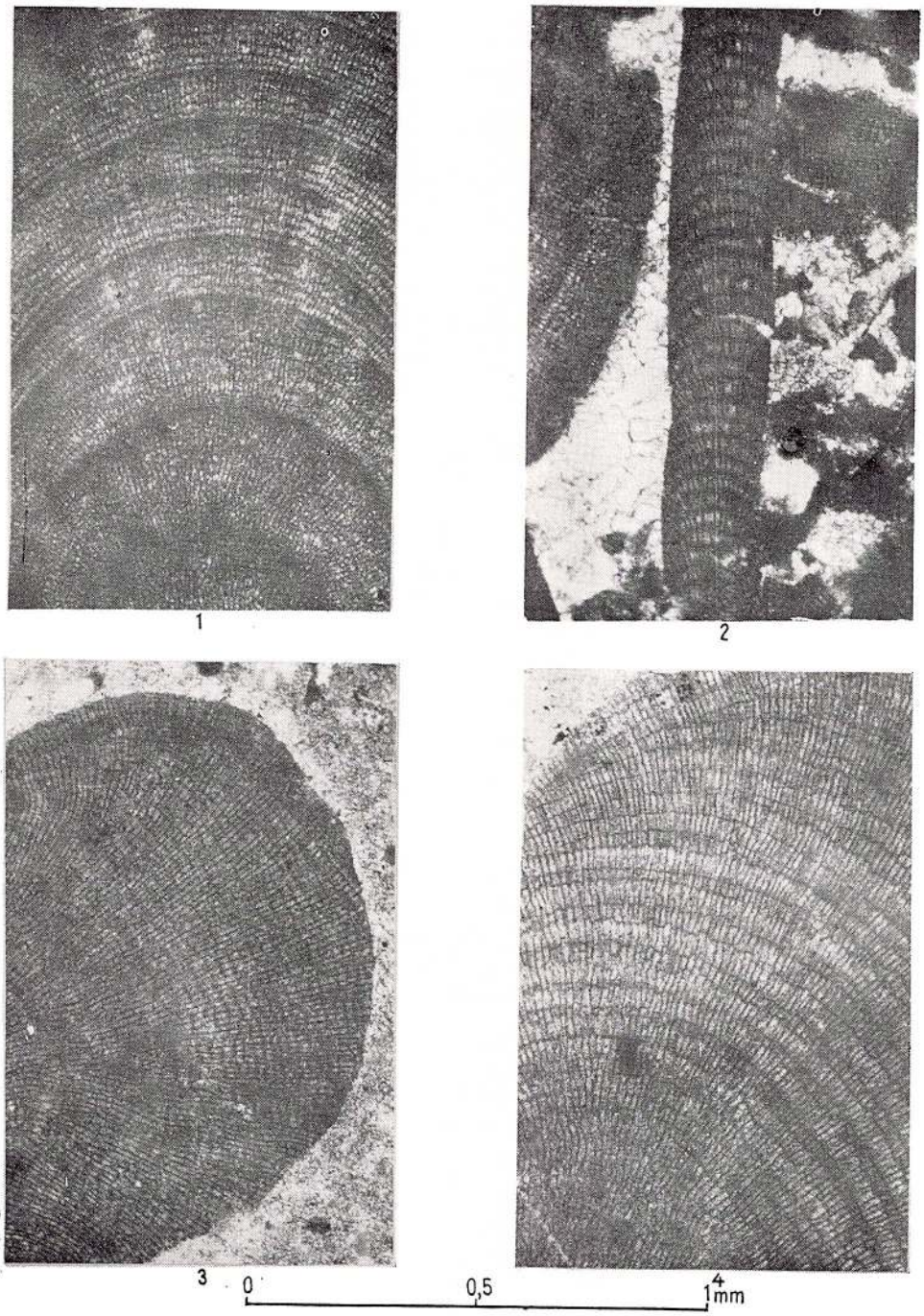
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PLATE V

Fig. 1. — *Lithophyllum symetricum* L e m o i n e (× 64). Ciglean Limestone. Ciglean.
Fig. 2. — *Lithophyllum simplex* L e m o i n e (× 64). Ciglean Limestone. Ciglean.
Fig. 3, 4. — *Lithophyllum* sp. (× 64). 3, — Hoia Limestone, Brebi. 4, — Limestone Series, Glod.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. V.



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PLATE VI

Fig. 1. — Individuals of *Globorotalia* sp. in Hoia Limestone (× 64). Fundătura (Creaca).

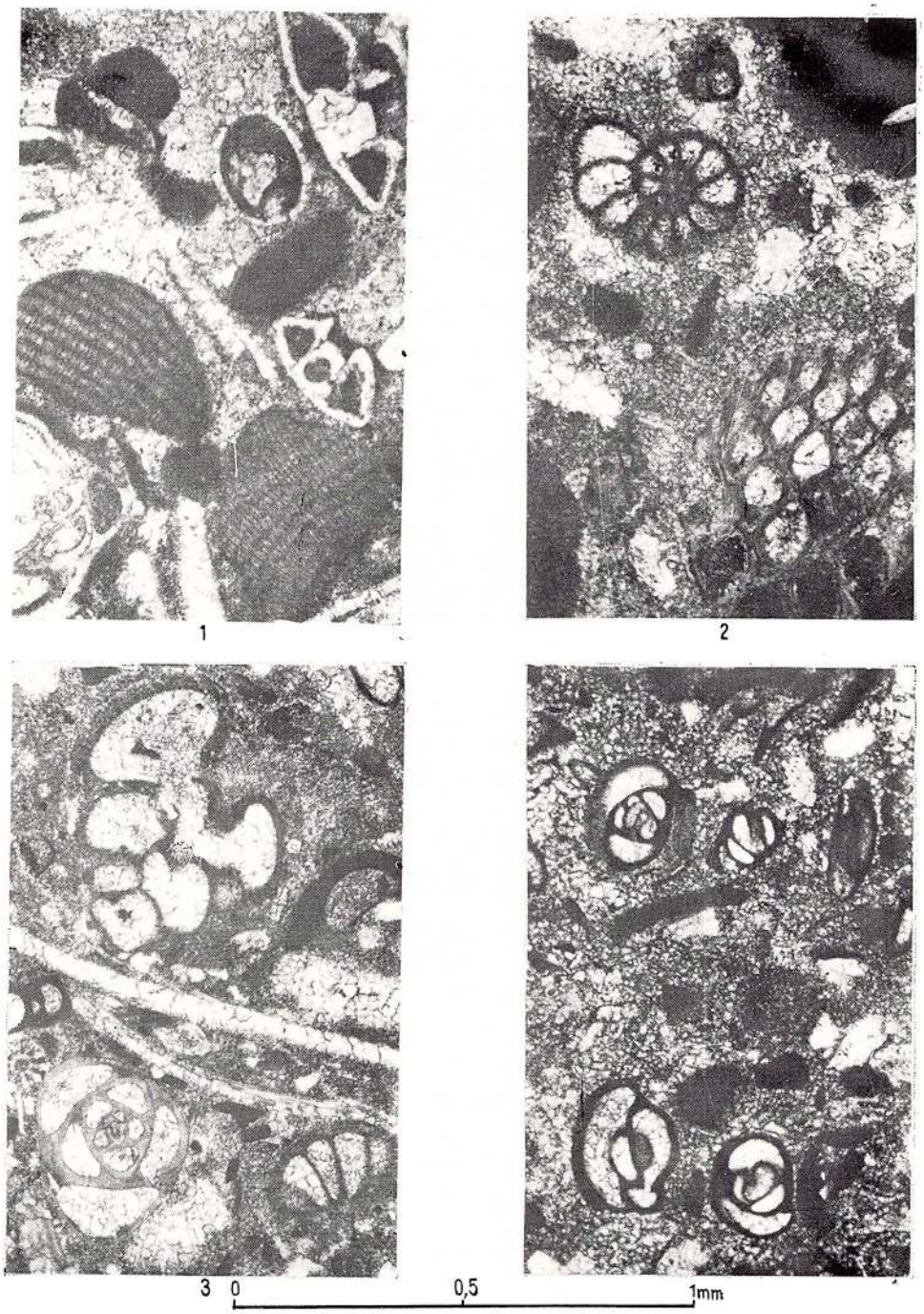
Fig. 2. — Bryozoan tests and *Rotaliids* in the uppermost part of the Limestone Series (× 64). Şoimuşeni.

Fig. 3. — Skeletal limestone with pelecypods and foraminifera (× 64). Hoia Beds sensu lato. The equivalent of the *Scutella* Lower Level of the Mera Beds. Stîna.

Fig. 4. — Sandy limestone with miliolids (× 64). Hoia Beds sensu lato. The equivalent of the lowermost part of the Mera Beds. Ciumărna.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. VI.



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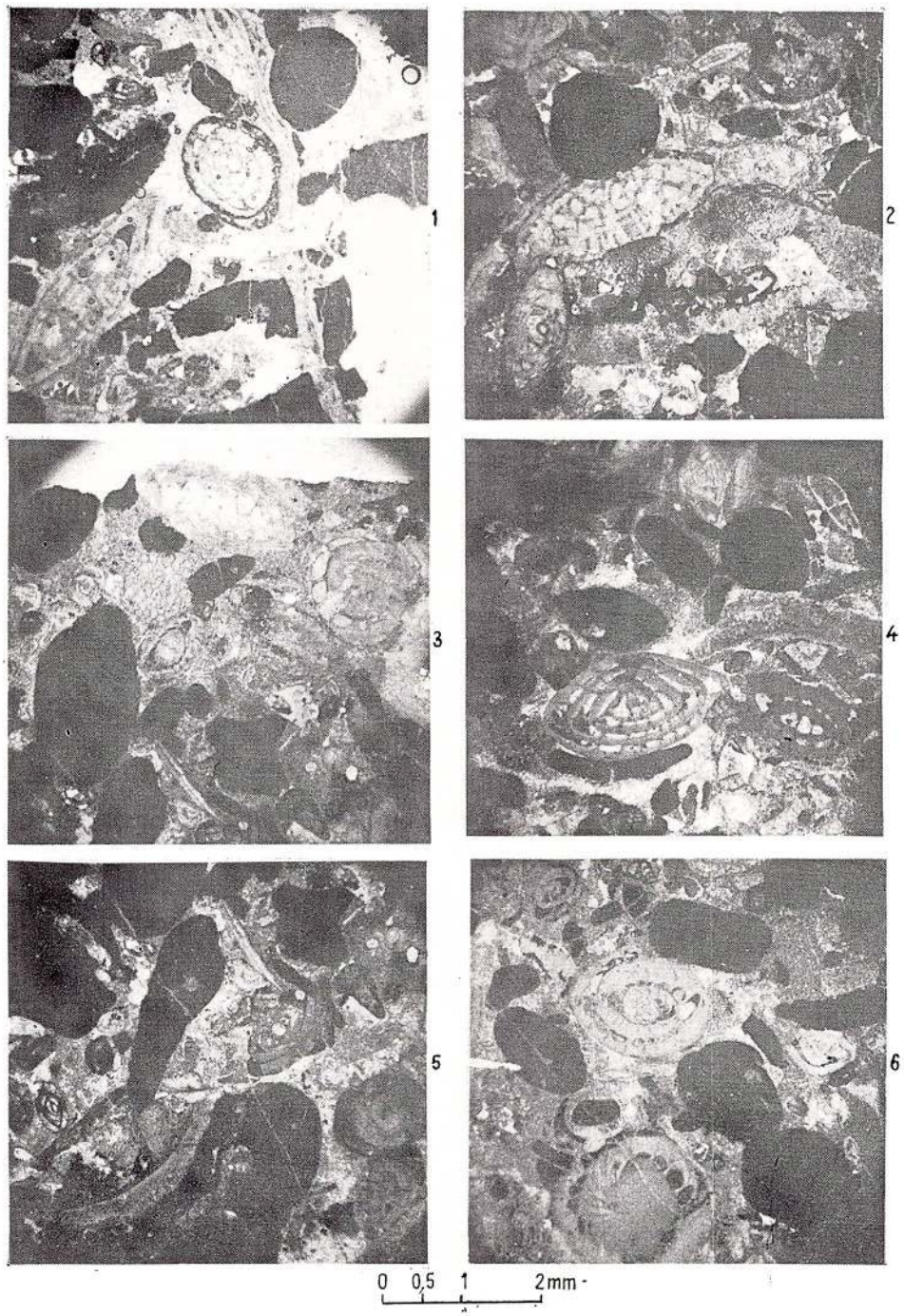
PLATE VII

The sandy, corallinacean facies zone.
Top of Calcareous Series. Preluca area : Glod.

Fig. 1—6. — Very coarse and coarse, nummulitid-corallinacean, welded grainstones consisting of coralline nodules (the black grains), nummulitids and other large benthonic foraminifera, echinoderm plates, bryozoas, miliolids. Corallinacean algae, miliolids and in places nummulitids are partly or wholly micritized. The grain welding most obvious in Fig. 2 ; hard grain compaction supported by grain interpenetration and contact-solution phenomena. In all other photomicrographs the originar welded texture is more or less obliterated by a diagenetic replacive spar calcite (microspar to pseudospar) developed at the expanse of the micritized skeletons, the spar calcite nibbling into, penetrating or transecting the different (especially coralline) grains.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. VII.



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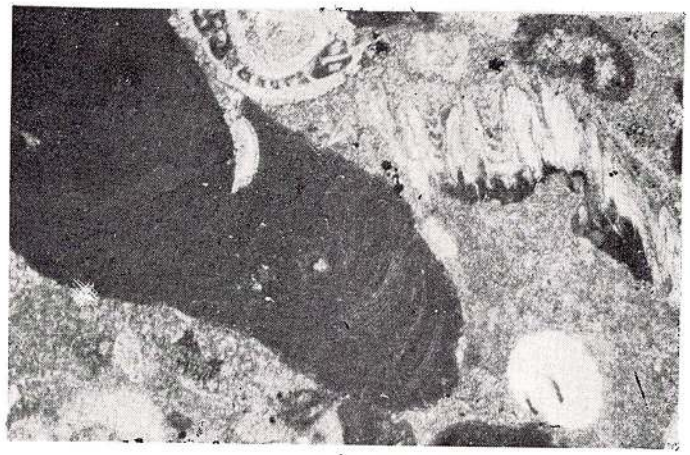
PLATE VIII

The sandy, corollinacean facies zone.
Top of Calcareous Series. Preluca area : Glod.

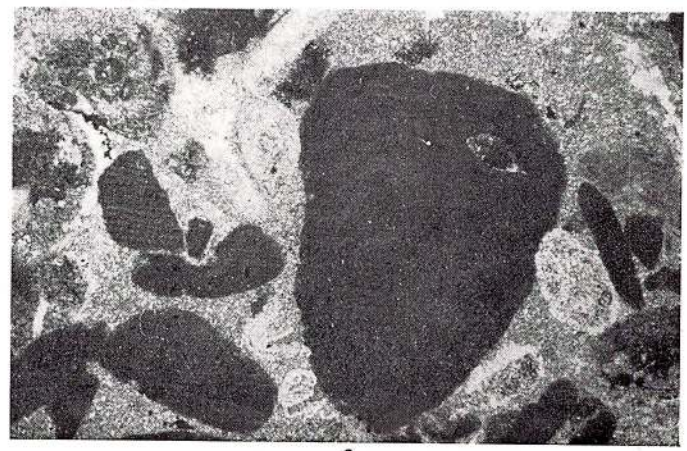
Fig. 1—2. — Details in very coarse and coarse, nummulitid-corallinacean,welded grainstones consisting of coralline fragments, nummulitids, balanids (Fig. 1),echinoderm plates (Fig. 2), etc. The photomicrographs show the matrix-mimmicing microspar clearly developed on the skeletal (mainly coralline) grains, which appear reoutlined and floating as isolated fragments into a microspar groundmass.

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A. RUSU, A. DRĂGĂNESCU. Sedimentation at the Time of the Hoia Limestone. Pl. VIII.



1



2

0 0,5 1 mm

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PLATE IX

The sandy, corallinacean facies zone.
Top of Calcareous Series. Preluca area : Boiu Mare.

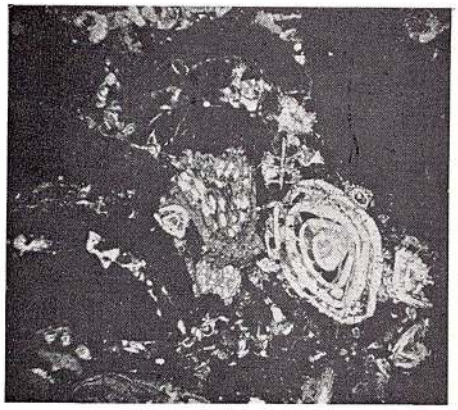
Fig. 1.—2.—Very coarse and coarse, nummulitid-corallinacean, welded grainstones consisting mainly of coralline nodules and nummulitids, and subordinately of bryozoas, miliolids, echinoderm plates. The welded texture is obvious ; strong grain compaction pointed out by grain interpenetration and contact-solution phenomena.

Fig. 3. — Detail in Fig. 2. Welding, solution and interpenetration of the skeletal grains is strongly manifest. All the black areas showing or not fine cellular fabrics stand for coralline algae.

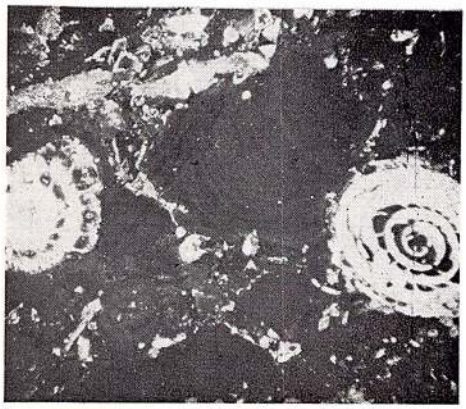
Fig. 4. — Detail in the same grainstones. The strong grain compaction revealed by the welding of the grains, is highly obvious.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. IX.

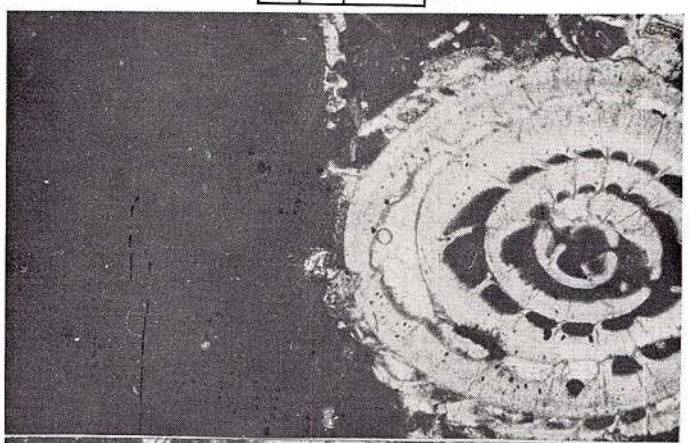


1

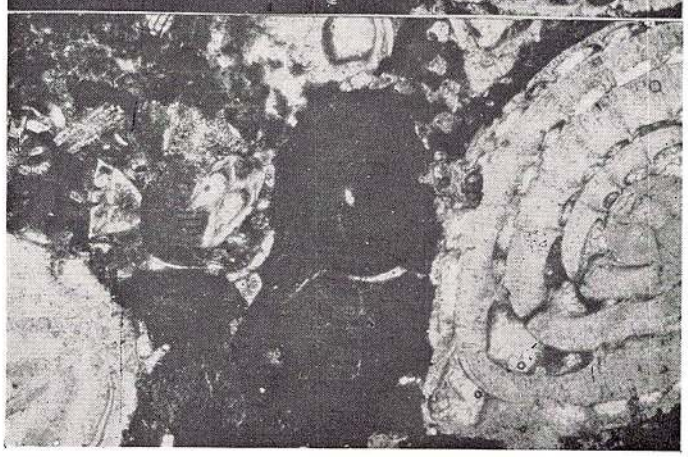


2

0 0,5 1 2 mm



3



4

0 0,5 1 mm

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PLATE X

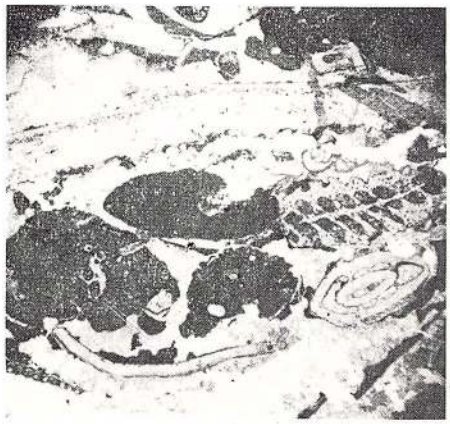
The sandy, corallinacean facies zone.
Top of Calcareous Series. Preluca area: Răstoci and Topliţa.

Fig. 1—3. — Răstoci. Fig. 3 is an enlargement in Fig. 2. Coarse and very coarse, nummulitid-corallinacean, welded grainstones. Welding of the skeletal grains obvious in Fig. 2 (detailed in Fig. 3), i.e. between a coralline grain and a nummulitid, or between bryozoan tests. Evident development of replacive spar calcite on account of skeletons.

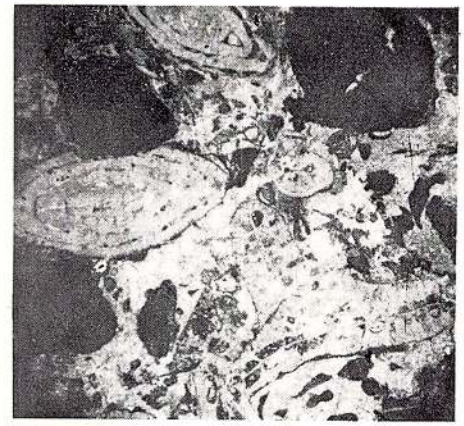
Fig. 4—5. — Topliţa. Fig. 5 is an enlargement in Fig. 4. Medium-grained, nummulitid-corallinacean, welded grainstones. Obvious welded texture. In photo 5 matrix-like microspar (in places aggraded to larger pseudospar) developed at the expense of the micritized corallinacean and foraminiferal welded tests.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. X.



1



2

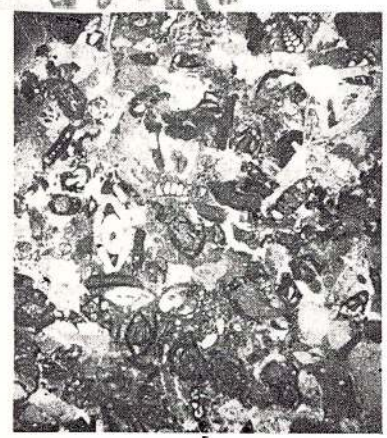


3



4

Fig. 1,2,4: 0 0,5 1 2 mm



5

Fig. 3,5: 0 0,5 1 mm

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PLATE XI

The sandy, corallinacean facies zone.
Top of Calcareous Series. Preluca area : Purcăreț, Băbeni, Cliț.

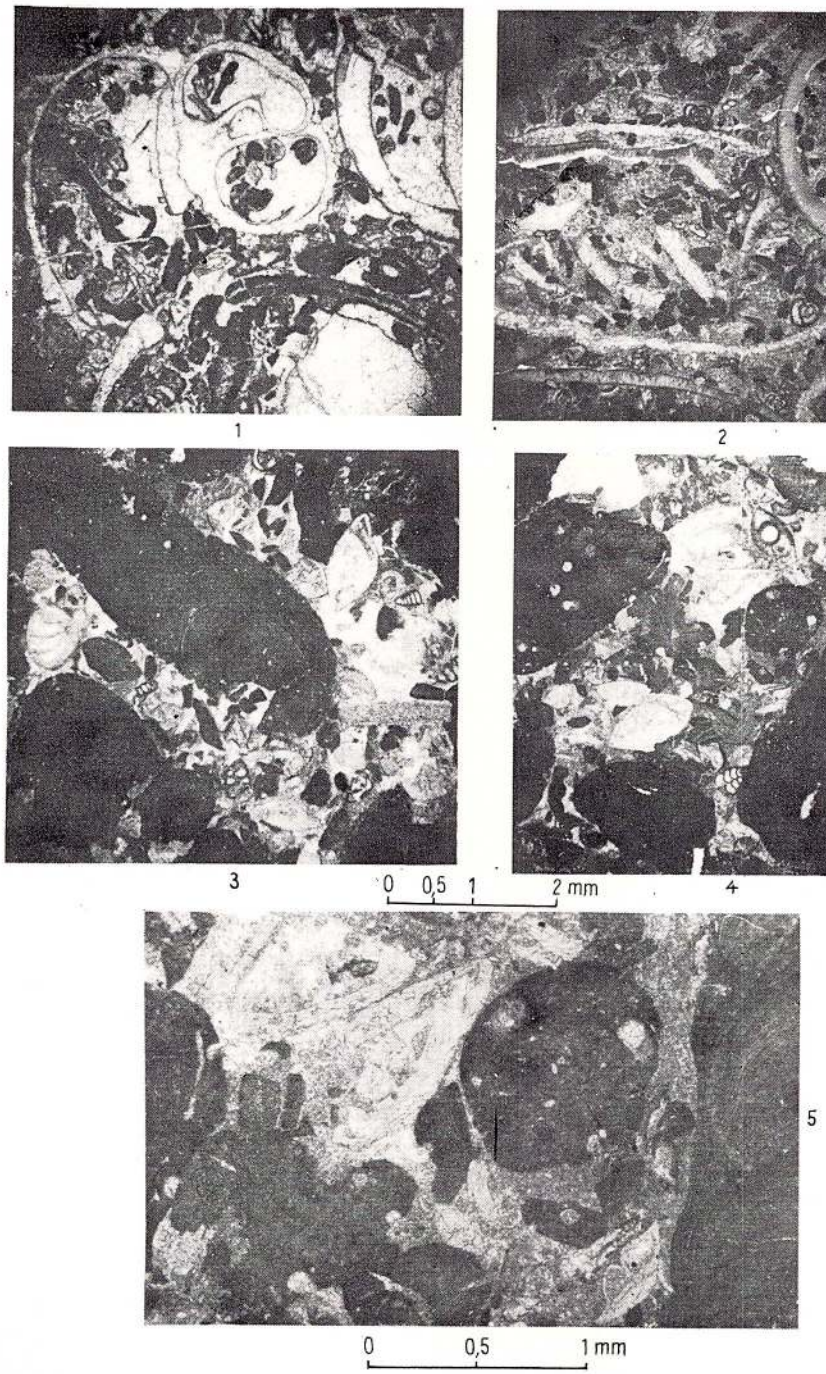
Fig. 1. — Purcăreț. Coquina welded grainstone consisting of gastropods, pelecypods, coralline tests, foraminifera. The gastropod cavities filled with skeletal grains too. Spar (pseudospar) calcite areas developed at the expense of the skeletal (mainly coralline) grains.

Fig. 2. — Băbeni. Coquina welded grainstone consisting chiefly of pelecypods and coralline fragments. The pelecypod shells often fractured, in place with the resultant fragments more or less displaced suggesting a strong grain compaction. Network-like microspar, partly aggraded to larger pseudospar, developed on the micritized skeletal (mainly coralline) grains.

Fig. 3—5. — N of Cliț. Fig. 5 is an enlargement in Fig. 4. Coarse to very coarse, nummulitid-corallinacean, welded grainstones. Obvious welding, melting (contact-solution) and intermingling of the skeletal grains. Development of microspar areas by consumption of the coralline nodules and nummulitids most manifest in Fig. 5; most of microspar in Fig. 2 aggraded to (mosaicate or monocrystalline syntaxial) larger pseudospar.

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A. Rusu, A. Drăgănescu. Sedimentation at he Time of the Hoia Limestone. Pl. XI.



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PLATE XII

The sandy, corallinacean facies zone.

Top of Calcareous Series (Ciglean Limestone). Preluca area : Băbeni and Piroșa.

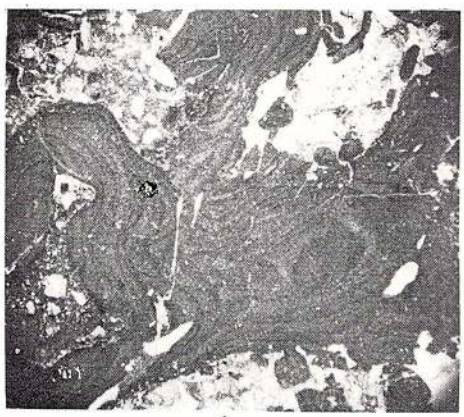
Fig. 1—2.— Băbeni. Very coarse, corallinacean, welded grainstones consisting chiefly of large nodules of coralline algae. The skeletal grains are welded, showing contact-solution phenomena and interpenetrations. The spar areas are replacive calcite mainly on coralline tests.

Fig. 3. — Băbeni. Oncolitic welded grainstones consisting of microoncoliths and pisoncoliths. The spar areas are spar calcite replacements on fields of welded microoncoliths.

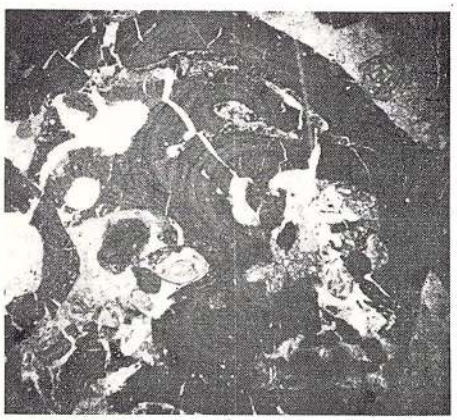
Fig. 4—5. — Piroșa. Fig. 5 is an enlargement in the central part of Fig. 4. Coarse to medium-grained, corallinacean, welded grainstone displaying the component skeletal (mainly coralline) grains completely merged into each other. The abundance of the coralline tests, their stronger or slighter micritization and their perfect welding originated in an appearence of micrite matrix, at low magnifications; at higher magnifications, as photo 5 proves, the ,,matrix'' dissociates optically in welded coralline fragments.

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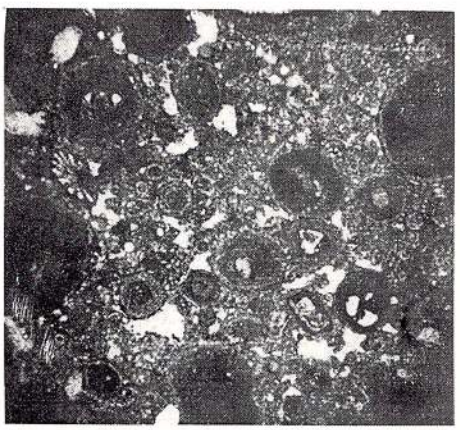
A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XII.



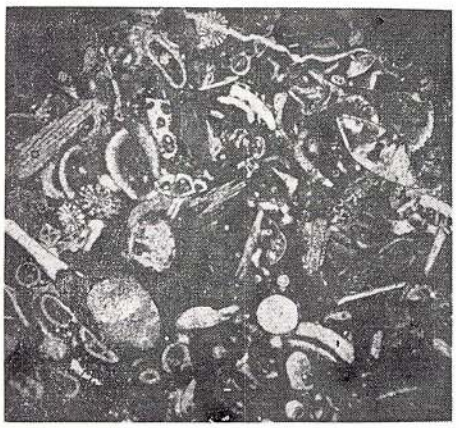
1



2

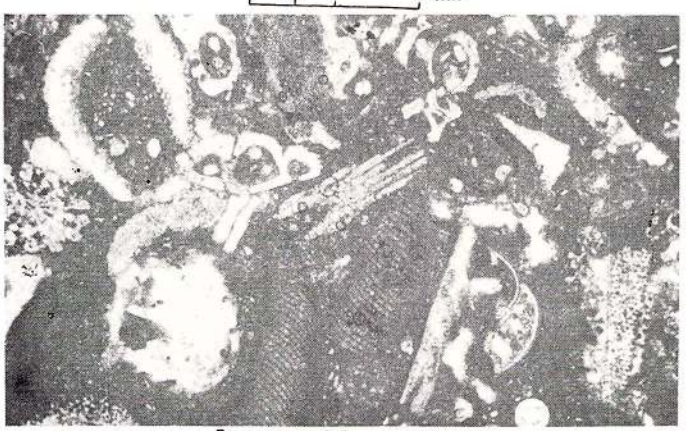


3



4

0 0,5 1 2 mm



5

0 0,5 1 mm

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PLATE XIII

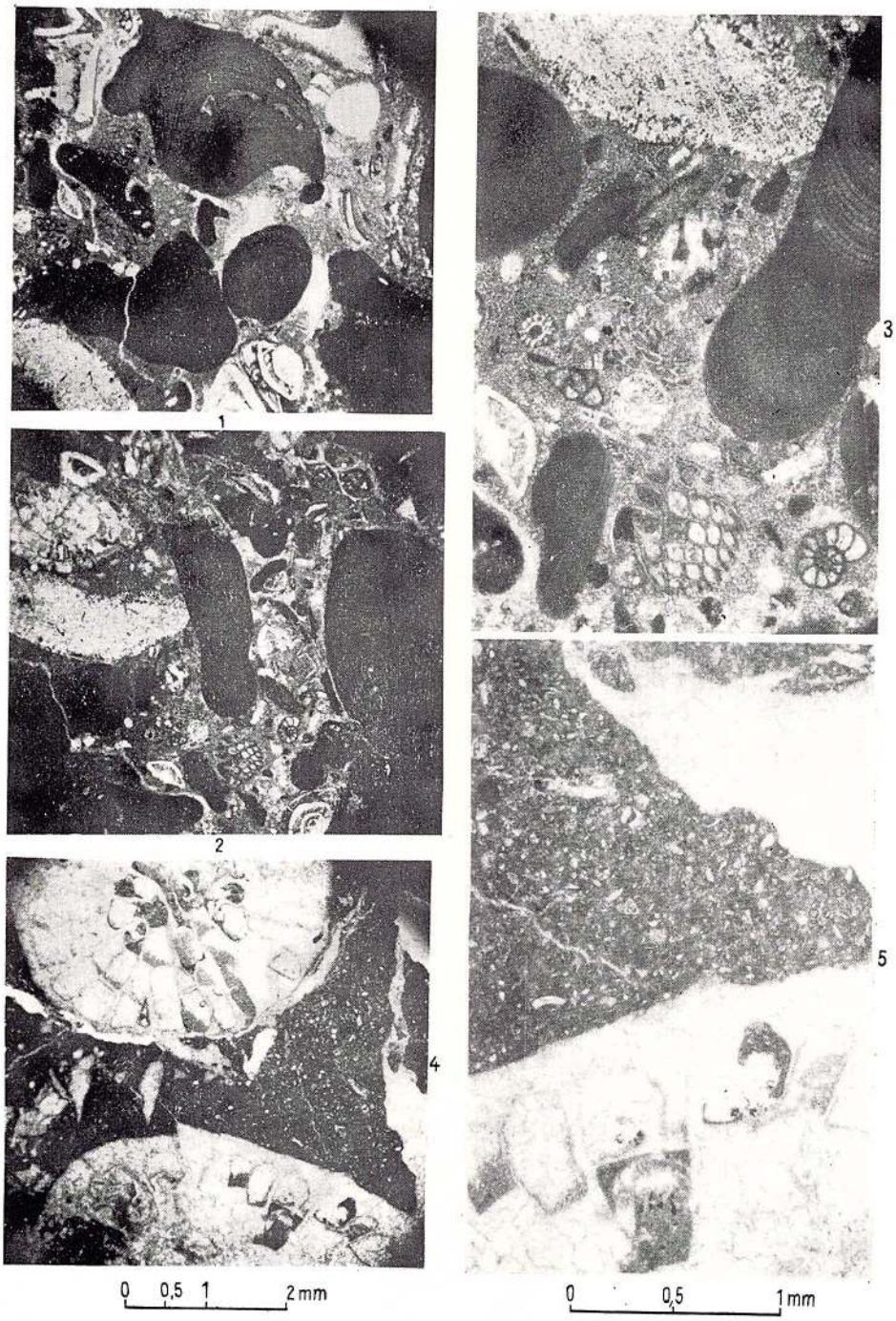
The sandy, corallinacean facies zone.
Top of Calcareous Series. Preluca area : Soimușeni.

Fig. 1—3. — Fig. 3 is an enlargement in the bottom of Fig. 2. Coarse to very coarse, nummulitid-corallinacean, welded grainstone. The skeletal grains show contact-solution phenomena. The network-like microspar in which the skeletal grains seem embedded is not a slightly aggraded micrite matrix but, as clearly pointed out by Fig. 3, it stands for a replacive, very finely crystalline spar calcite superimposed on the original welded skeletal grains (now more or less ,,melted'' within the microspar groundmass).

Fig. 4—5. — Island of mud ly, coraligenous facies within the sandy, corallinacean facies zone. Fig. 5 is an enlargement in the bottom of photo 4. Globigerinids- and globorotalliids-bearing wackestones, accommodating coral biolithitic structures.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XIII.



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PLATE XIV

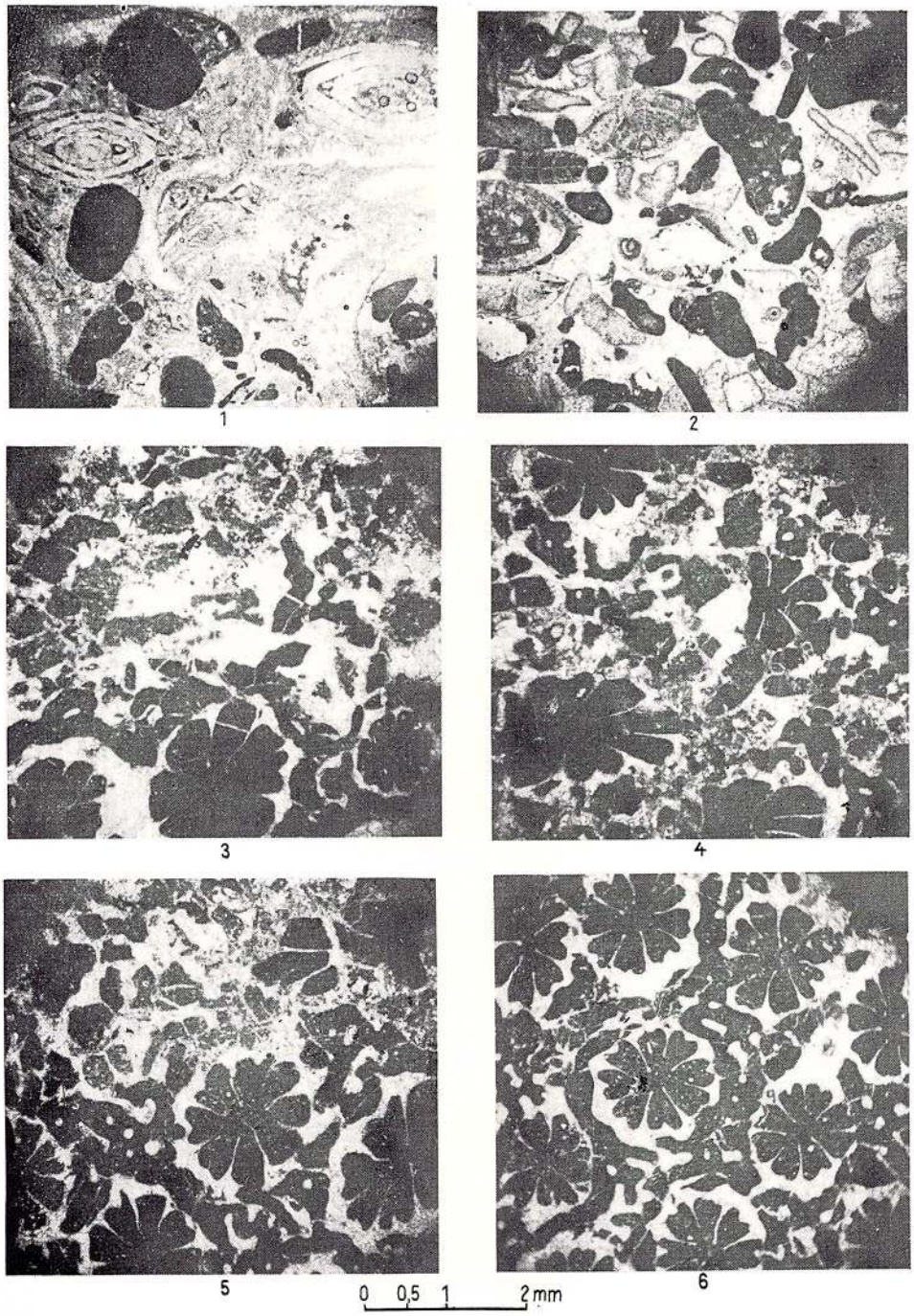
Preluca area : Ciocmani.

Fig. 1—2. — N and, respectively, E of Ciocmani. The sandy, corallinacean facies zone. Ciglean Limestone enriched in nummulitids. Medium-grained to coarse, nummulitid-corallinacean, welded grainstone showing the skeletal grains apparently loosely packed owing to the development of a network-like replacive spar calcite (initially microspar, subsequently aggraded to larger pseudospar) which ingested many of the original, welded skeletal grains ; the unaffected grains show contact-solution phenomena and slight interpenetrations, supporting the welded grainstone texture of the original rock.

Fig. 3—6. — Locality of Ciocmani. The muddy, coralligeneous facies zone. Hoia Limestone. Coral biolithites filled with allochem-free (algal?) micrite. Thin-sections in massif, crustous colonies of digitate corals.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XIV.



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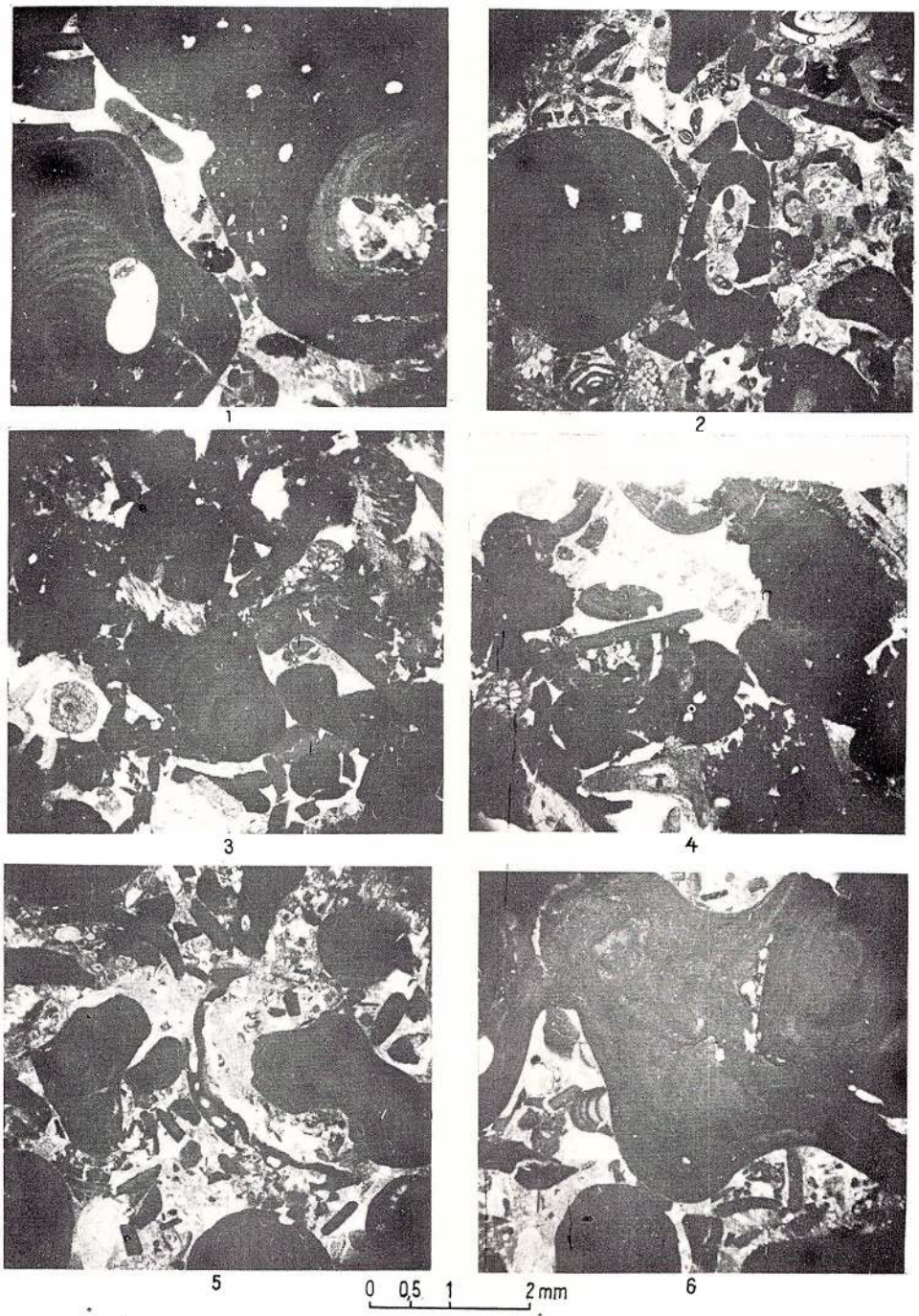
PLATE XV

The sandy, corallinacean facies zone.
Ciglean Limestone. Mezeş area : Ciglean.

Fig. 1—6. — Coarse to very coarse, corallinacean, welded grainstones consisting of merged corallinacean nodules and fragments, and in minor amounts, foraminifera, echinoderm plates (with replacive monocrystalline syntaxial overgrowths). The spar calcite areas, varying from very finely to medium crystalline, stand for replacive diagenetic calcite (microspar and pseudospar) formed on account of the coralline tests.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XV.



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PLATE XVI

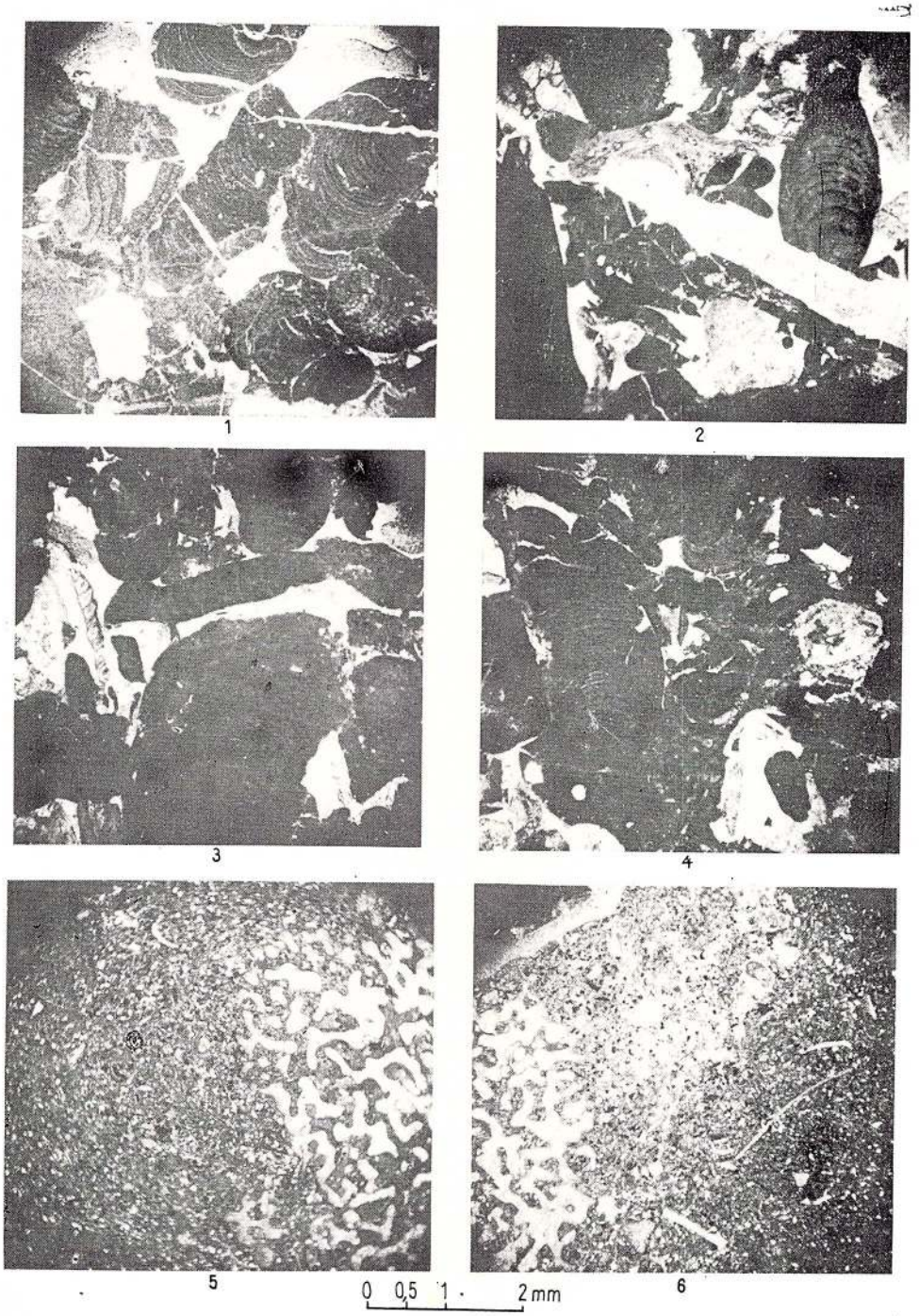
The Mezeş area.

Fig. 1—4. — The sandy, corallinacean facies zone. Ciglean. Ciglean Limestone. Corallinacean welded grainstones showing interpenetration of the grains and obvious contact-solution phenomena (Fig. 2) accounting for the infered strong grain compaction.

Fig. 5—6. — The muddy, miliolid-coraligenous facies zone. Brebi (Pîrîul Ortelec valley). Hoia Limestone. Lamellous coral biolithites accommodated in a micrite matrix bearing miliolids and pelagic foraminifera (see photographic details on the matrix in next plate).

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XVI.



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PLATE XVII

The Mezeș area : Brebi.

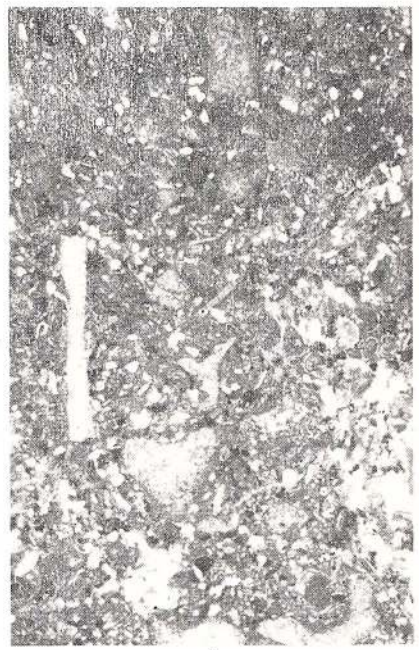
Fig. 1—3. — Brebi (Pîriul Ortelec valley). The muddy, miliolid-coraligenous facies zone. Hoia Limestone. Details in the matrix of the limestones presented in Pl. XVI ; Figs. 5—6. Wackestones consisting of miliolids, globigerinids, globorotaliids, ostracods and pelecypods embedded in a micrite matrix.

Fig. 4. — NW of Brebi (uppermost course of Brebi valley). Top of Brebi Marls Member : packstone containing abundant miliolids, coralline fragments, pellets, fine-grained quartz, diverse foraminifera.

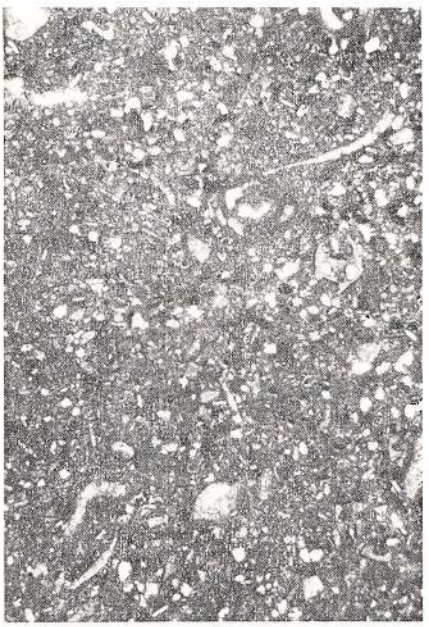
Fig. 5. — NW of Brebi (uppermost course of Brebi valley). The sandy, corallinacean facies zone. (Hoia Limestone interval). Coquina welded grainstone consisting of merged pelecypod shells.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XVII.



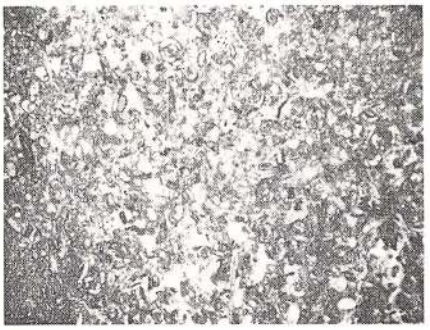
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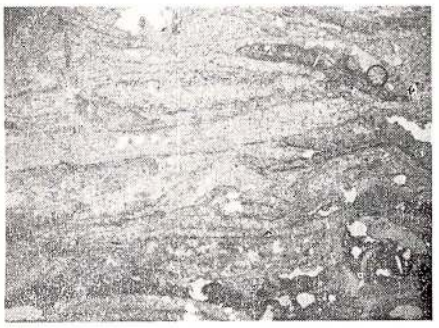
2



3



4



5

0 0,5 1 mm

Fig. 4,5: 0 0,5 1 2 mm

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PLATE XVIII

The muddy, miliolid-coraligenous facies zone.
Hoia Limestone. Mezeș area : Moigrad, Stîna and Ciumărna.

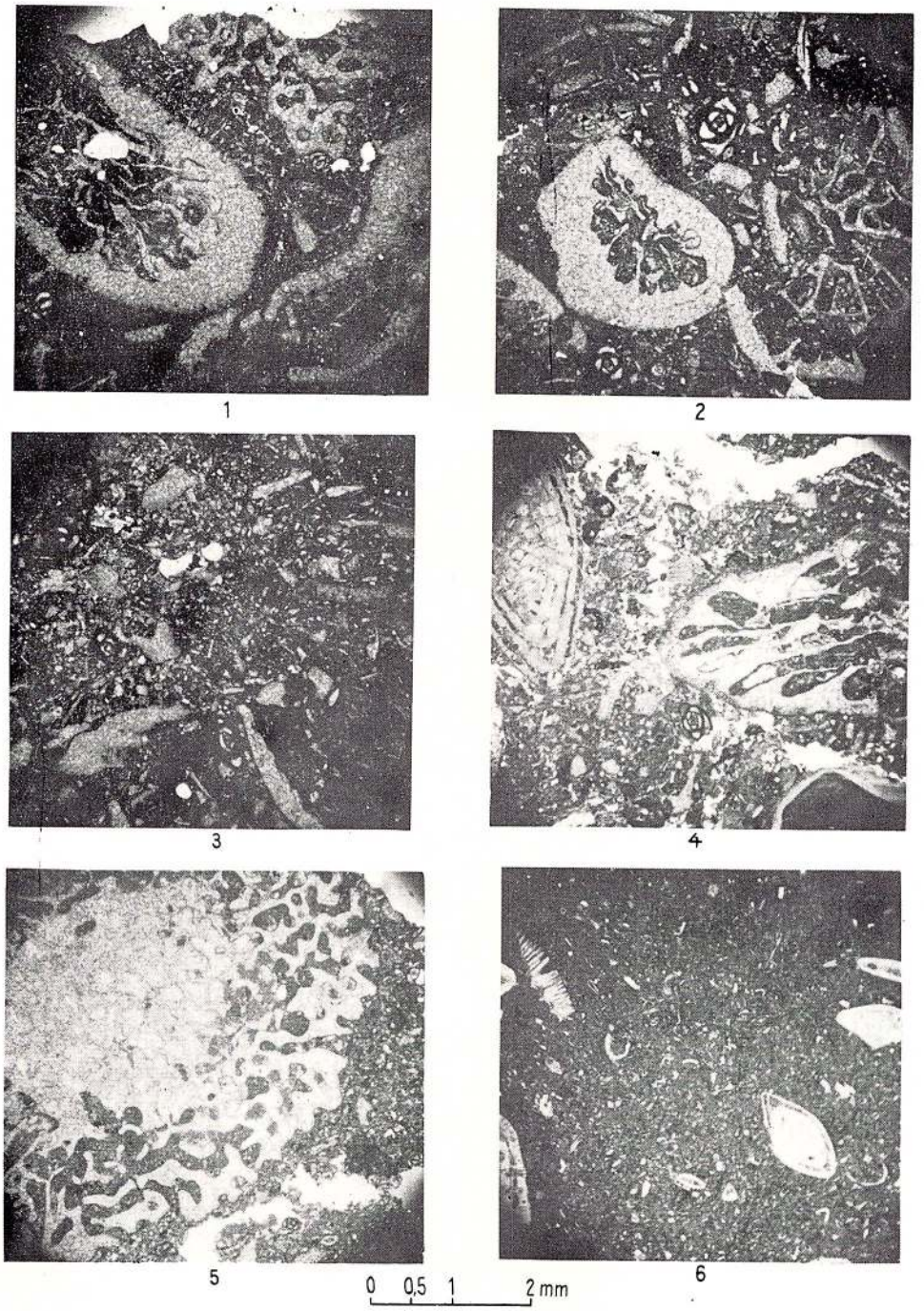
Fig. 1—3. — Moigrad. Lime wackestones and packstones accommodating coral biolithitic structures. The host sediment consists of coral debris, miliolids, globigerinids, globorotaliids, etc. embedded in a micrite matrix.

Fig. 4—5. — N of Stîna (Jurteana valley). Coral biolithitic structures in a lime wackstone or packstone composed of micrite matrix, miliolids, coral debris, nummulitids, etc.

Fig. 6.— Ciumărna. Top of Brebi Marls, underlying directly the Hoia Limestone. Marl containing nummulitids, pelagic foraminifera, etc.

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XVIII.



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PLATE XIX

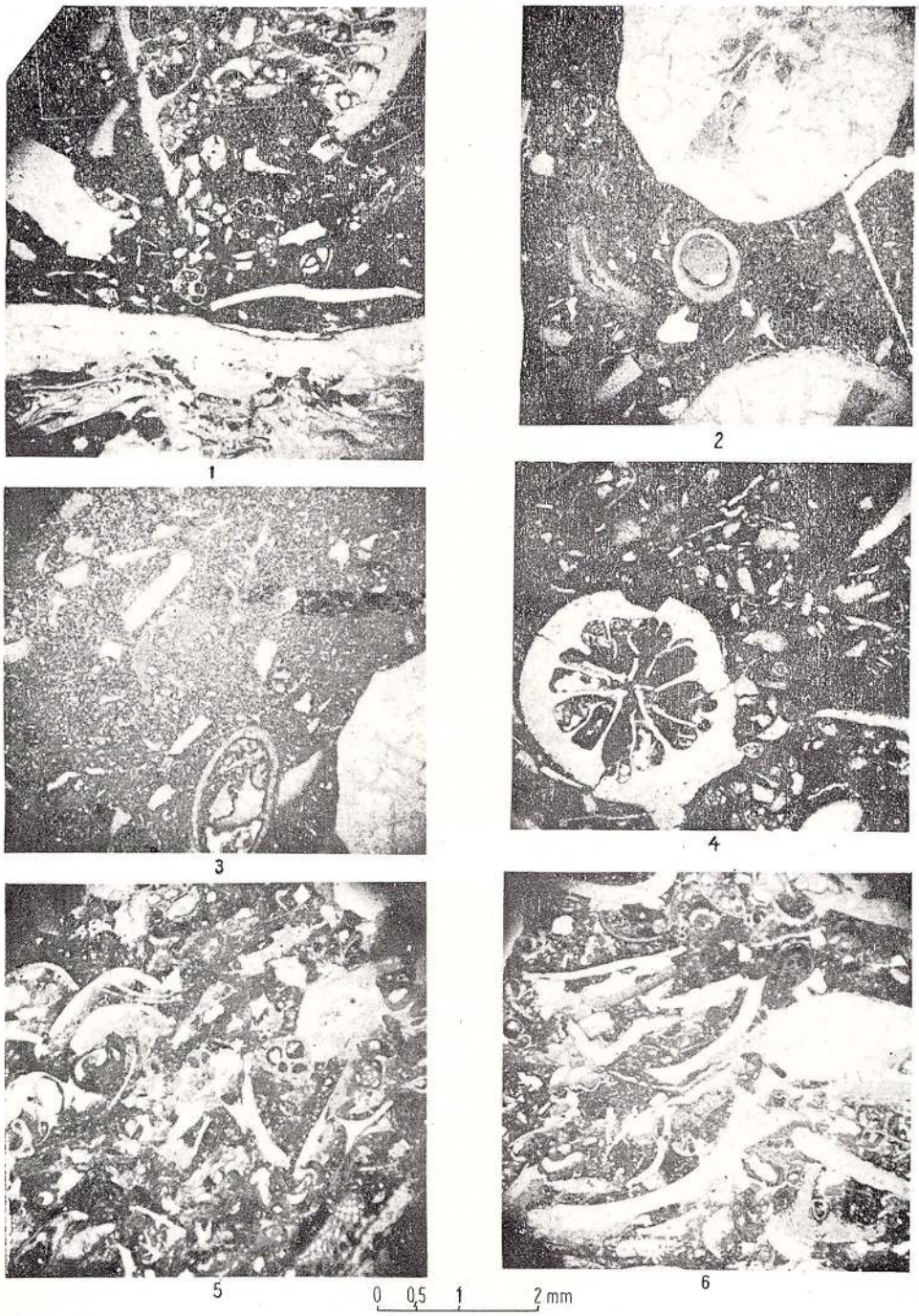
The muddy, miliolid-coraligenous facies zone.
Hoia Limestone. Mezeș area : Treznea and Stina.

Fig. 1—4. — Treznea. Coral biolithitic structures accommodated in lime wackestone or packstone composed of micrite matrix, coral fragments, miliolids, pelagic foraminifera, etc.

Fig. 5—6. — Stina (Șipote valley). The muddy, pelecypod-miliolidic subfacies consisting of lime muds containing chiefly pelecypods and miliolids (local variety of the muddy, miliolid subfacies).

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XIX.



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PLATE XX

The muddy, globigerinid-bryozoan facies zone.
Top of Brebi Marls (equivalent to the Hoia Limestone). Mezeș area : Bozna.

Fig. 1—3. — Marls consisting of clayey micrite matrix (partly replaced by a microspar or pseudospar calcite) containing miliolids, pelagic foraminifera, ostracods, and cryptocrystalline grains (lumps and pelets, mainly proceeding from micritized miliolids).

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A. Rusu, A. Drăgănescu. Sedimentation at the Time of the Hoia Limestone. Pl. XX.



1



2



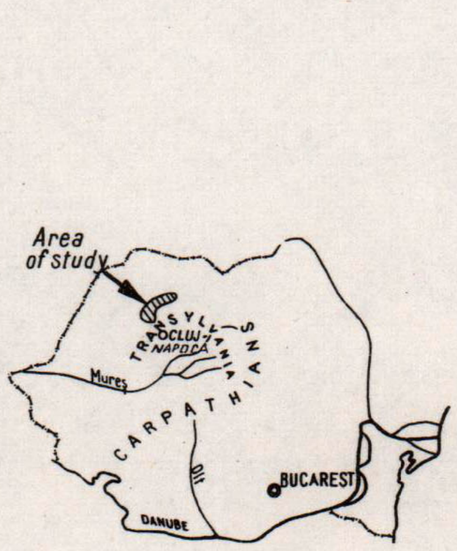
3

0 0,5 1 mm

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FACIES DISTRIBUTION DURING THE HOIA LIMESTONE MEMBER DIPOSITION IN THE NORTHWESTERN TRANSYLVANIA (ROMANIA)

0 2 4 6 8 10 Km



LEGENDUM

- Outcropping metamorphic and magmatic basement
- Corallinacean-nummulitic, coquina and oncolitic grainstones
- Corallinacean grainstones
- Miliolidic wackestones-packstones accommodating lens-shaped coral boundstones
- Coral-lacking miliolidic wackestones-packestones
- Marls containing globigerinids and globorotaliids

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Redactor: ILONA SANDU
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TOME XLVIII

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