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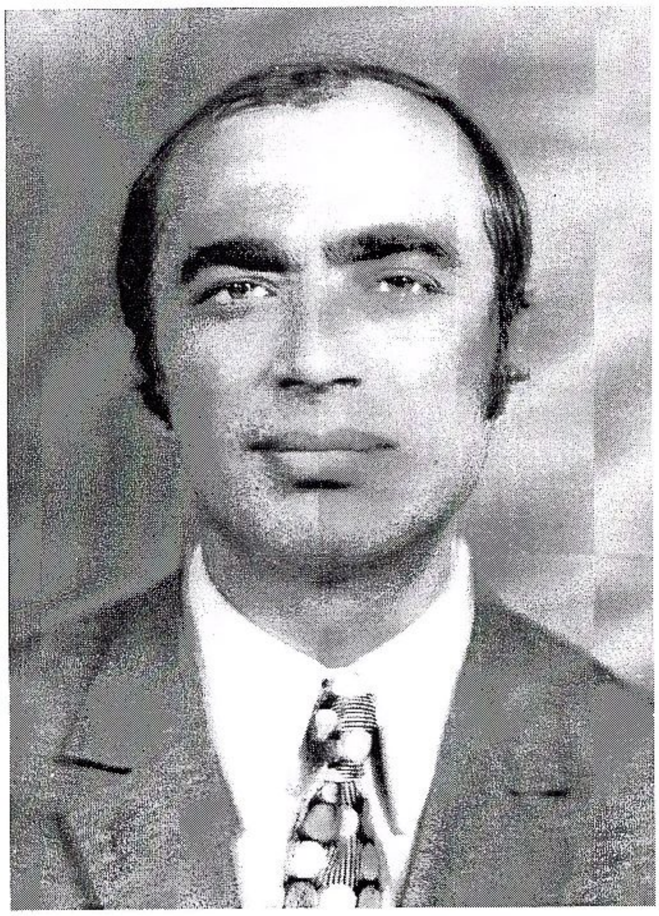

GRIGORE POP
(1935-2000)

Rom. J. Stratigraphy, 79, p. 3-10, Bucureşti, 2001

In memoriam
GRIGORE POP
(1935-2000)

La 30 ianuarie 2000 (în ziua Sfinţilor Trei Ierarhi: Vasile, Grigore şi Ioan) comunitatea geologică românească pierdea pe unul dintre cei mai straluciţi reprezentanţi ai generaţiei sale: geologul Dr. Grigore Pop, cercetător ştiinţific principal la Institutul Geologic al României.

Născut la 23 noiembrie 1935 în Albeştii Bistriţei (judetul Bistriţa-Năsăud), Grigore, al treilea din cei patru copii ai familiei Pop, va înscrie în analele ştiinţelor geologice numele moştenit de la tatăl său, gospodar de frunte al satului şi diacon la biserică.

Pop Grigore al lui Grigore (cum îl numeste primul său învăţător din localitatea natală) a absolvit gimnaziul la Teaca şi Şcoala Pedagogică la Năsăud (o şcoală vestită în această parte a Transilvaniei), obţinând diploma de învăţător în 1953. Rezultatele foarte bune la învăţătură şi înclinaţia pentru ştiinţele naturii îl determină să urmeze cursurile universitare la Facultatea de Geologie-Geografie (secţia Geologie) a Universităţii Bucureşti, pe care le absolvă cu succes în 1958.

Remarcat încă de pe băncile facultăţii de către profesori pentru aptitudinile sale pentru cercetare, este repartizat în acelaşi an la nou înfiinţatul Institut de Geologie şi Geografie al Academiei R.P.R. În fişa sa de atestare după primul an de activitate, mentorul sau Prof. Grigore Răileanu consemnează: "aptitudinile sale lasă să se întrevadă o frumoasă posibilitate de dezvoltare". În caracterizarea făcută mai tárziu (1967) de Prof. Virgil Ianovici (directorul institutului), în vederea promovării în funcţia de cercetător ştiinţific principal, Grigore Pop este apreciat ca "... un cercetător minuţios care işi organizează în mod eficient munca de cercetare şi are un orizont ştiintific bine conturat".

Seriozitatea pregătirii sale este remarcată foarte repede, astfel că încă din primii ani de cercetare semnează lucrări ştiintifice împreună cu personalităţi geologice de prestigiu precum: Acad. Mircea Savul, profesor la Universitatea din Iaşi (cu care publică prima sa lucrare în 1962), Prof. Acad. Alexandru Codarcea, preşedintele Comitetului Geologic, Prof. Dr. Grigore Răileanu, directorul Institutului Geologic etc.

Începând din 1970 (anul fuzionării I.G.G. al Academiei cu Institutul Geologic al Comitetului de Stat al Geologiei) şi până la sfârşitul vieţii geologul Grigore Pop şi-a desfăşurat activitatea în cadrul Institutului Geologic al României, ajungând datorită calităţilor sale, a pasiunii şi a sacrificiilor, un specialist de marcă ce a făcut cinste institutului, geologiei româneşti şi ţării.

În 14 martie 1973 obţine titlul de doctor în geologie cu teza de doctorat intitulată "Depozitele mezozoice din muntii Vîlcan", avându-l conducător ştiinţific pe eruditul geolog şi exigentul profesor care era Acad. George Murgeanu. Tiparită în acelaşi an în editura Academiei R.S.R. teza sa devine o lucrare de referinţă pentru regiunea studiată.

Activitatea ştiinţifică efectuată timp de 42 ani fără întrerupere a fost consacrată cu precădere studiului formaţiunilor paleozoice şi mezozoice, precum şi resurselor minerale asociate (cărbuni, bauxită, minereuri de fier, calcare, dolomite etc) de pe teritoriul ţării, din Cuba şi din Madagascar.

La curent cu progresele ştiinţifice mondiale în domeniul terenurilor sedimentare, pregătirea sa polivalentă îi permite să abordeze cu uşurinţă probleme de petrologie sedimentară, litostratigrafie, biostratigrafie, paleontologie, tectonică şi gitologie pe care le rezolvă exemplar.

În anul 1970 (11 aprilie-7 iulie) a studiat petrografia şi microfaciesurile unor formaţiuni carbonatice jurasice şi cretacice din provinciile Pinar del Rio şi Las Villas (Cuba) conform protocolului de colaborare dintre Academia R.S.R. şi Academia de Ştiinţe din Cuba, publicandu-şi o parte din rezultatele ştiinţifice.

În perioada 1 noiembrie 1975-31 decembrie 1978 Dr. Grigore Pop a lucrat în R.D. Madagascar în calitate de expert tehnic al Programului Naţiunilor Unite pentru Dezvoltare (PNUD), în proiectul "Studiu asupra Industriei Petroliere şi asupra Resurselor Miniere", efectuând studii geologice, prospecţiuni şi explorări în ariile a două zăcăminte de fier (Bekisopa şi Betioky). Şi de această dată rezultatele ştiinţifice cele mai importante apar publicate în reviste de specialitate.

În domeniul petrologiei sedimentare a efectuat studii asupra unor formaţiuni siliciclastice şi carbonatice de diverse vârste. Au fost cercetate formaţiuni siliciclastice marine (de tip prefliş, fliş, olistostroma, molasă) şi continentale (de tip "red beds", Gresten) din Carpaţii Meridionali (Pânza Getică, Domeniul Danubian, Unitatea de Severin) şi din Madagascar. Studiile asupra formaţiunilor carbonatice, în special mezozoice (care au constituit o preocupare permanentă), au avut în vedere atât faciesurile platformelor carbonatice, cât şi depozitele bazinale (fie de origine pelagică, fie turbiditică). Astfel, depozitele carbonatice de tip platformă au fost cercetate în Domeniul Danubian extern, Pânza Getică, unităţile de Bihor şi Vălani ale Apusenilor de Nord şi în Dobrogea. Depozitele carbonatice bazinale (s.s.) au fost studiate în domeniul Getic (zonele

Reşita, Haţeg), Domeniul Danubian intern (Unităţile de Sirinia, Dubova şi Herculane), în unele unităţi central est-carpatice, Klippele Pienine (de la Poiana Botizei), Unitatea de Severin şi în câteva unitâti din Cuba. Au fost abordate, de asemenea, aspectele sedimentologice şi litogenetice ale depozitelor pe baza cărora s-au interpretat paleomediile de sedimentare şi evoluţia acestora sub influenţa factorilor tectonici şi paleogeografici.

În domeniul biostratigrafiei efectuează studii complexe de datare biocronologică şi corelare a formaţiunilor sedimentare jurasic superioare-cretacic inferioare pe bază de calpionele, calcisfere, foraminifere şi alte organisme fosile.

Un loc important în preocupările lui l-au avut cercetările asupra calpionelelor, pentru care face o adevărată pasiune. Incepe prin a studia cele mai reprezentative secţiuni geologice în depozitele carbonatice (bazinale) tithonic-neocomiene cu calpionele din România precum şi câteva secţiuni din Cuba şi Spania (Cordiliera Betică). Cunoaşte practic tot ce s-a scris despre acest misterios grup fosil de ciliate şi împătimit îi dedică multe mii de ore de examen microscopic, ajungând astfel unul dintre cei mai apreciaţi specialisti în calpionele din lume.

Rezultatele cercetărilor sale în acest domeniu constau în elaborarea unei scheme de biozonare de înaltă rezoluţie (fig. 1) - bazată pe evenimente evolutive (filogenetice) - şi în importante contribuţii de ordin taxonomic.

În cadrul acestui grup de organisme planctonice, considerat bine cunoscut, creează 8 genuri noi şi 13 specii noi. Acestea sunt: *Praecalpionellites* Pop, 1986; *P. murgeanui* (Pop 1974); *P. sirianensis* Pop, 1986; *Remaniella filipescui* Pop, 1994; *R. borzai* Pop, 1994; *R. duranddelgai* Pop, 1996; *R. colomi* Pop, 1996; *R. catalanoi* Pop, 1996 (din familia Calpionellidae) şi *Dobeniella* Pop, 1997; *Longicollaria* Pop, 1997; *Borziella* Pop, 1997; *Almajella* Pop, 1998 (nom. nov. pentru *Cubanella* Pop, 1997); *Aninella* Pop, 1998 (nom. nov. pentru *Cylindrella* Pop, 1997); *Carpathella* Pop, 1998; *Daciella* Pop, 1998; *Chitinoidella elongata* Pop, 1997; *Carpathella rumanica* Pop, 1998; *Daciella banatica* Pop, 1998; *D. almajica* Pop, 1998; *D. danubica* Pop, 1998; *D. svinitensis* Pop, 1998 (din familia Chitinoidellidae). Prin aceşti taxoni de calpionele numele lui Grigore Pop devine nemuritor pentru ştiinţă.

Rezultatele de ordin petrologic, paleontologic şi stratigrafic obţinute i-au permis cunoaşterea mai avansată a unor evenimente geologice cu caracter local, regional sau global privind condiţiile de sedimentare, schimbările relative ale nivelului marin şi dinamica crustală. Analiza megasecvenţelor sedimentare pe baza acestor date, alături de observaţiile asupra aspectelor deformaţionale, relaţiilor stratigrafice şi tectonice, a fost favorabilă interpretării mai aprofundate a evoluţiei bazinelor de sedimentare. Astfel de studii au avut drept rezultat evidenţierea unor noi unităţi tectonice cum sunt Unitatea de Iloviţa în cadrul Pânzei Getice şi unităţile laramice de Dubova şi de Herculane în Domeniul Danubian.

Ridicările geologice, însoţind majoritatea studiilor efectuate, au fost legate în principal de programul de elaborare a Hărţii Geologice a României scara 1:50000, la care dr. Grigore Pop a contribuit prin redactarea a 14 foi de hartă din Carpaţii Meridionali şi Dobrogea de Sud.

A coordonat şi a colaborat la studii cu caracter economic asupra unor resurse minerale (cărbuni, bauxită, calcare, dolomite) de pe teritoriul ţării, în vederea evaluării potenţialului lor economic şi a măsurilor necesare pentru protecţia patrimoniului mineral naţional.

În decursul timpului desfăşoară şi o bogată activitate ştiinţifică internaţională legată de programele unor grupuri de lucru (pentru limita Jurasic-Cretacic, pentru limita Kimmeridgian-Tithonic şi pentru limitele Valanginianului) şi de proiectele IGCP 262 - Corelarea Cretacicului Tethysian şi IGCP 362 - Corelarea Cretacicului Tethysian şi Boreal.

Participă la numeroase reuniuni ştiinţifice internaţionale organizate în ţară sau în străinătate, unde lucrările prezentate de el se bucură de aprecieri elogioase.

Din 1981 până în 1989 s-a ocupat de redactarea Buletinului de informaţii ştiinţifice şi tehnice al Comisiei inginerilor şi tehnicienilor din I.G.G. Tot din anul 1981, dr. Grigore Pop face parte din Comitetul de redacţie al institutului, devenind editorul ştiinţific al revistei "Romanian Journal of Stratigraphy" editată de I.G.R.

Dr. Grigore Pop a fost mulţi ani referent de specialitate la revista "Geologica Carpathica" (Bratislava), ultima solicitare de avizare a unei lucrări privind calpionellidele din Albania sosindu-i după deces. Iată reacţia D-nei Eva Chorvatova - managing editor al revistei - la aflarea tristei veşti; "Dr. Grigore Pop was a favour referee of our journal for his professional and prompt reviews. We shall be missing him" (24.VII.2000).

Rezultatele sale ştiinţifice sunt consemnate în 57 de lucrări publicate în ţară şi în străinătate, în cele 14 foi (6 tipărite şi 8 machete) ale Hărţii Geologice a României scara 1:50 000 şi în peste 100 de rapoarte geologice existente în arhivele Institutului Geologic al României (Fondul Geologic), Institutului de Geologie al Academiei de Ştiinţe din Cuba, Serviciului Geologic din Madagascar şi ale PNUD (New York).

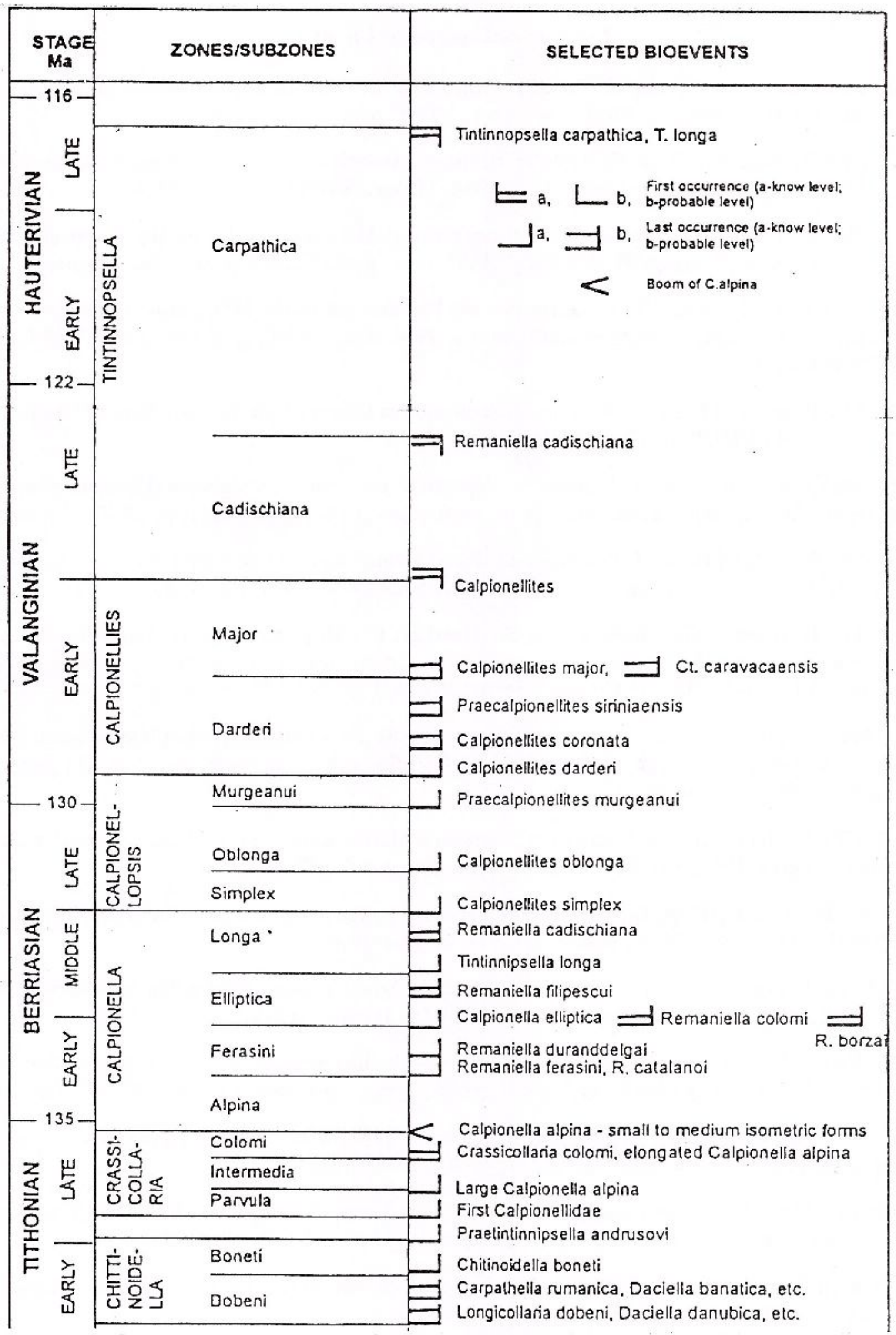

Fig. Biozonarea Tithonic-Neocomianului pe bază de Praecalpionellidae şi Calpionelidae (Pop, 1997)

Grigore Pop lasă pentru posteritate o operă durabilă, edificată cu perseverenţă şi intuiţie. Prin dispariţia sa prematură, geologia românească devine mai săracă, iar cei ce l-au cunoscut pierd un prieten sincer şi un coleg respectat pentru calităţile sale de om şi de cercetător talentat.

Dr. Anatol Rusu

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Rom. J. Stratigraphy, 79, p. 11-12, Bucureşti, 2001

DATE INEDITE ASUPRA PÂNZEI-SOLZ DE NICULIŢEL

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Key words: Tectonics. Triassic. North Dobrogea

Abstract: *New data concerning the Niculiţel thrust-fold (North-Dobrogea).* An attempt to interpret the unpublished geological map (O. Mirăuţă), elaborated during 1965-1966, shows that the North Dobrogean Niculiţel thrust-fold can be subdivided in three subunits: the inner subunit contains limestones (Illyrian - Upper Fassanian), basalts (Bithynian - Pelsonian) and sandstones (Norian); the outer subunit, with Lower to Upper Triassic sediments, includes basalts (Lower Bithynian) and sandstones (Norian); the middle, shally subunit (Upper Ladinian to Upper Norian), here named the Coasta Hill Formation.

În cuprinsul pânzei-solz de Niculiţel se pot separa trei subzone (subunităţi) între care există raporturi tectonice.

A. Subzona internă se caracterizează prin:

a) prezenţa, în baza succesiunii litologice, a Stratelor de Izvoarele (An_3 - $Ld^2{}_1$);

b) prezenţa, la limita Ladinian inferior - Ladinian superior, a unor corpuri bazaltice, denumite "bazalte superioare de Niculiţel". Ele conţin enclave sincrone de calcare în faciesul pelagic roşu, cu amoniţi la nivelul Bithynian - Pelsonian;

c) prezenţa Gresiei de Alba, transgresivă şi discordantă peste diverşi termeni litologici, inferiori ca vârstă; local, ea stă peste şisturi silicifiate, cu calcare pelagice norian superioare. În general, gresia descrie anticlinale şi sinclinale normale.

B. Subzona externă se caracterizează prin:

a) continuitate de sedimentare din Triasicul inferior până în Norian (probabil);

b) prezenţa "bazaltelor inferioare de Niculiţel" care includ enclave sincrone de calcare în faciesul calcarelor cu cherturi la nivelul Bithynian-Pelsonian;

c) prezenţa Gresiei de Alba în continuitate de sedimentare, având în bază calcare carnian superioare şi urcând, probabil, în Norian. În general, gresia apare în sinclinale deversate spre NE sau N.

C. Subzona mediană se caracterizează prin prezenţa predominantă a faciesului şistos, intens cutat, cu nivele de calcare cu vârste cuprinse între Ladinianul superior şi Norianul superior; faciesul grezos de la partea terminală ar putea reveni Rhaetianului. Am denumit această secvenţă predominant şistoasă "Formaţiunea de Dealul Coasta".

Poziţia acestor trei subzone / subunităţi şi raporturile dintre ele sunt următoarele:

Subzona externă apare în fruntea Pânzei-solz de Niculiţel, în partea ei nordică şi nord-estică, unde este mult avansată peste Pânza de Tulcea (Promontoriul Isaccea, Dealul Meseaua Roşie, Dealul Sarica, Dealul Chilizbaş).

Subzona internă se află la vest de cea externă şi este paralelă cu ea. Subzona internă încalecă peste cea externă, având în baza secvenţei Strate de Izvoarele care apar fie în anticlinale deversate spre NE, fie ca flanc normal al acestor anticlinale. Contactul tectonic dintre cele două subzone are loc fie (1) între bazaltele superioare de Niculiţel (situate deasupra Stratelor de Izvoarele) din subzona internă şi Gresia de Alba (în sinclinale deversate) din subzona externă; fie (2) între stratele de Izvoarele (din subzona internă) şi Gresia de Alba (din subzona externă).

Subzona externă dispare la S de Dealul Chilizbaş.

Subzona mediană apare la zi în axul unui anticlinal deversat spre NE în valea dintre Dealul Piatra Edirlen şi Dealul Chilizbaş. Aici apar cele trei subzone în poziţia pe care o definesc: la W, subzona internă, la E cea externă, la mijloc, între ele, subzona mediană. Spre S, în urma dispariţiei subzonei externe, în fruntea Pânzei-solz de Niculiţel şi în contact tectonic cu formaţiunile Pânzei de Tulcea de la E, rămâne Formaţiunea de Dealul Coasta din subzona mediană. Din anticlinalul deversat spre NE se mai păstrează, spre S (Dealul Cartalu - Muchia Verde şi Dealul Cricman, apoi Dealul Caracuş şi Bujor) numai flancul normal, vestic, care intră sub Stratele de Izvoarele din baza subzonei interne. Este posibil ca şi în Dealul Coasta Formaţiunea de Dealul Coasta să descrie tot un ax de anticlinal deversat spre NE. Această formaţiune, cu direcţii ale stratelor N-S, NW-SE sau NE-SW, deci conforme cu cele care caracterizează Pânza-solz de Niculiţel, este afectată de cutările E-W ale depozitelor jurasice din Pânza de Tulcea. Ea apare şi în axul unui anticlinal de vale situat între Dealul Coasta şi Dealul Caracuş, în timp ce Gresia de Alba din cele două dealuri apare în sinclinale suspendate.

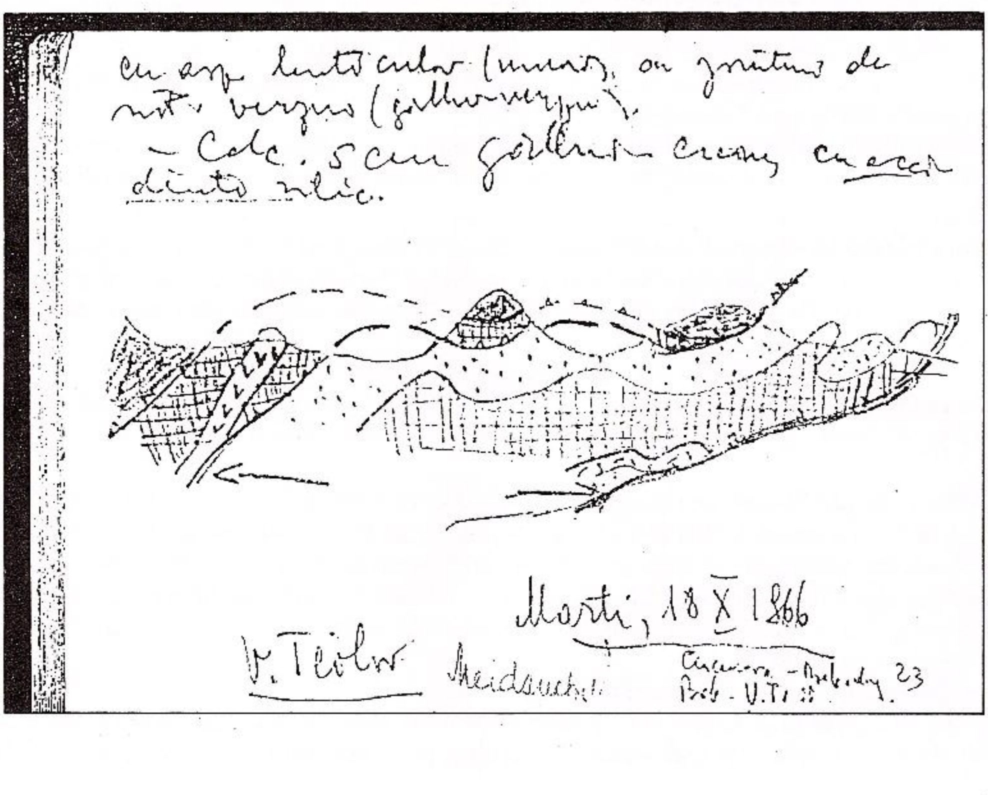
Este posibil ca Formaţiunea de Dealul Coasta să fi servit ca lubrifiant pentru avansarea spre E a Pânzei-solz de Niculiţel (?) sau numai a subzonei interne, ceea ce ar explica cutarea ei intensă. Este posibil, de asemenea, ca şi ea să fi fost antrenată în avansarea spre E, încălecând deci peste formaţiunile Pânzei de Tulcea.

Apartenenţa solzului Ghel Tepe este mai dificil de stabilit. Faptul că el include o succesiune litologică relativ continuă de la Triasicul inferior la Carnianul inferior (cu filamente şi influxuri de Gresie de Alba) ne-ar determina să-l atribuim subzonei externe. Absenţa bazaltelor inferioare de Niculiţel - reduse la filoane bazice în calcarele spathiene - precum şi aspectul calcarelor pelsoniene leagă acest solz de Pânza de Tulcea.

Indiferent de apartenenţa lui, depozitele care îl constituie definesc, în E, un anticlinal cu axul N-S care e tăiat, la N, de linia de încălecare care îl mărgineşte; la W depozitele de pe flancul vestic al anticlinalului intră sub Formaţiunea de Dealul Coasta din subzona mediană care se interpune astfel între subzona internă de la W şi solzul Ghel Tepe, de la E.

Lucrarea de faţă reprezintă o încercare de interpretare a hărţii geologice elaborată de O. Mirăuţă în 1965 şi 1966. Această interpretare se bazează pe datele conţinute în caietele de teren ale lui O. Mirăuţă şi pe datele de vârstă furnizate mai ales de asociaţiile de conodonte (Elena Mirăuţă) dar şi foraminifere (Doina Gheorghian) şi macrofaună (Magdalena Iordan), studiate în deceniile 70 şi 80. Concepţia lui O. Mirăuţă privind structura în pânze a Dobrogei de Nord este ilustrată de facsimilul alăturat, care redă o secţiune geologică ipotetică, desenată pe un caiet de teren, în toamna anului 1966.

Mulţumesc colegelor mele, precum şi Dr. K. Budurov pentru sprijinul pe care mi l-au dat în obţinerea acestor date de vârstă.



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AMMONITE ASSEMBLAGES OF THE ECLEJA FORMATION (NORTHERN APUSENI MTS, ROMANIA)

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Key words: Biostatigraphy. Ammonites. Lower Cretaceous. Ecleja Formation. Pădurea Craiului, Bihor Mts.

Abstract: The Ecleja Formation is here accepted as it has been emmended by Bordea et al. (1986), i. e. as containing four members, as follows (from bottom to top): (1) gray marlstones and dark spathic, organoclastic limestones; (2) marlstones and siltstones, with gray and black limestone interbeds; (3) "the median limestone with pachyodonts"; (4) once more marls and marly siltstones. All the ammonites recorded so far from this formation are revised and commented. These ammonites come from 15 fossiliferous sites, located in the Pădurea Craiului Mts (14) and in the Bihor Massif (1 of them). They point to the uppermost Bedoulian (sensu Erba, 1996, i. e. including the Furcata Zone) - Upper Gargasian, in places even lowermost Clansayesian age of the members 1+2 (supra). Moreover, they argue for a condensed sedimentation during the uppermost Bedoulian and lowermost Gargasian (the Martinoides Zone) in the whole region. Besides, some new data concerning the ranges of some ammonite genera have been documented, and three new species have been proposed: *Toxoceratoides stefanescui* n. sp., *Mathoceras istocescui* n. sp. and *M. leymerielliforme* n. sp.

1. Introduction

The age of the marly-silty and bioconstructed calcareous succession, lying on the "lower limestone with pachyodonts", known as "Ecleja Formation" and largely exposed in the Pădurea Craiului Mts and in the Bihor Massif (Northern Apuseni Mts, Fig. 1), was successively considered Neocomian (Preda, 1963) or Aptian (Patrulius et al., 1968; Istocescu, 1970; Istocescu et al., 1970; Istocescu, Bordea, 1970; Bordea et al., 1986), and restricted later to the Middle-Upper Aptian (Patrulius et al., 1982).

This paper takes into consideration all the ammonites gathered from these strata in the whole Bihor structural unit, in the Pădurea Craiului Mts (by I. Preda, D. Istocescu, S. Bordea, E. Avram, M. Ştefănescu) and in the Bihor Mts (by R. Huza). Only the Aptian ammonite published by Manea, Chivu (1973), and the Upper Albian fossils announced by Fisch (1924) are lacking: the former of themwas destroyed in the earthquake from 1977 and the latter are not preserved in Romanian repositories and, probably, were found in the next in succession lithostratigraphic unit, namely "the glauconitic sandstone complex". Up to now, the Ecleja Formation deposits offered no ammonite remains in the Vălani Nappe of the Pădurea Craiului Mts, so they are out of this paper subject.

The largest exposures of the formation are in the central and western part of the Pădurea Craiului Mts (Fig. 1 A), in the Fâsca-Vârciorog area and southwards (the Râu Valley basin, including its tributaries - the Copilu, Urzicaru and Rogojele valleys; the Vida/Blajul Valley springs; the Văsii and Măgura valleys basins). In the Bihor Massif it crops out only in the left slope of the Galbena Valley (Fig. 1 B), but is also developed westwards, in subsurface (under the Ferice and Arieşeni nappes successions), where it was met in the Juleşti gallery.

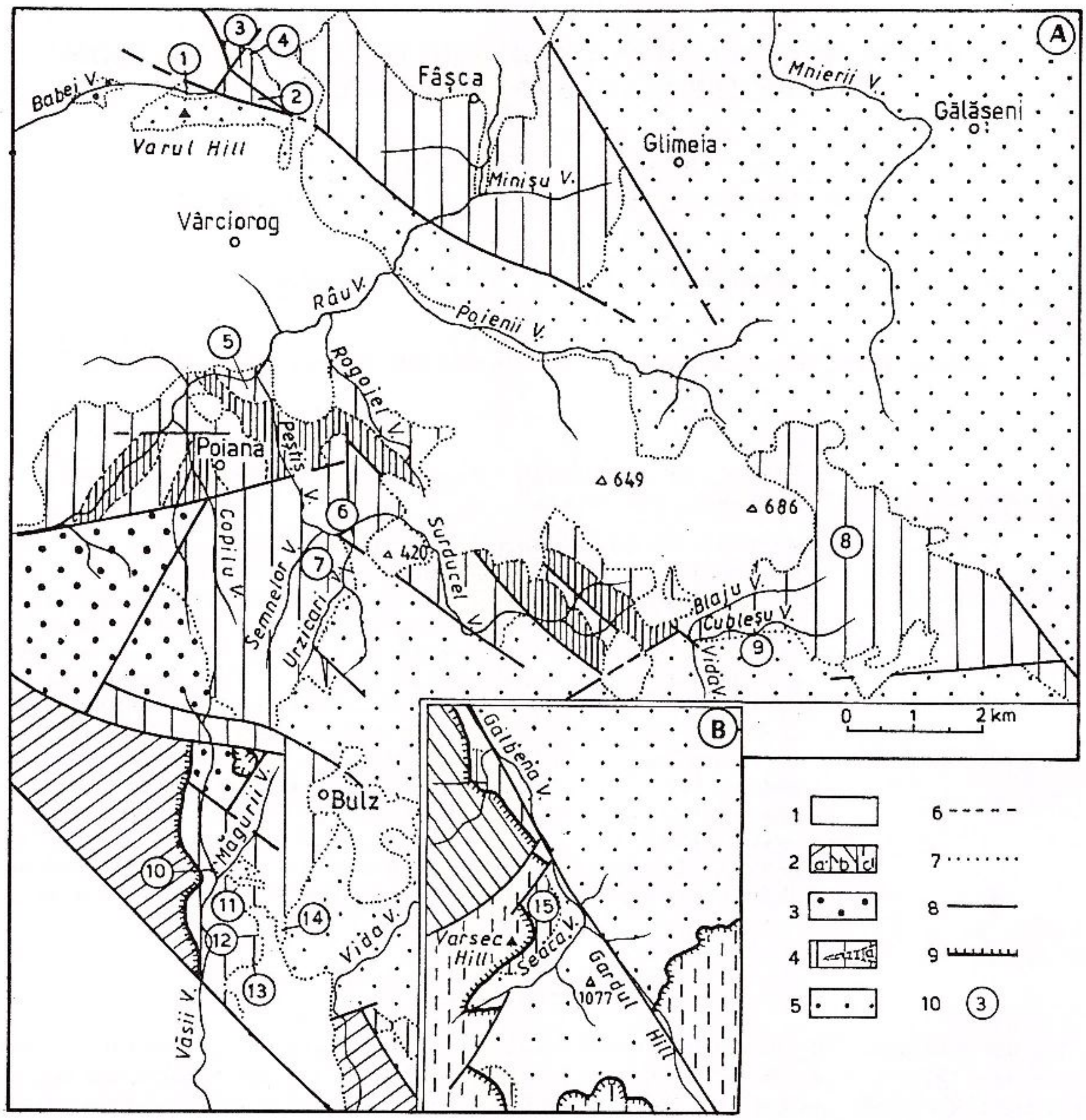

Fig. 1 – Sketch map of the Ecleja Formation exposures in the Pădurea Craiului Mts (A) and in the Bihor Massif (B) (after S. Bordea, J. Bordea and G. Mantea, unpublished). 1, Quaternary and Neogene deposits; 2, Codru Nappe system: a= Finiş Nappe, b=Ferice nappe, c=Arieşeni nappe; 3, post-Aptian deposits; 4, Ecleja Formation (a=the middle limestone with pachyodonts); 5, pre-Upper Bedoulian deposits; 6, lithostratigraphic boundary; 7, unconformity; 8, fault; 9, overthrust; 10, fossiliferous site.

The authors' contributions to the present paper are, as follows: S. Bordea realized the regional stratigraphic framework, in places with I. Cociuba's assistance; all the authors have put in common their fossils gathered from the Ecleja Formation; and E. Avram identified and revised the ammonite species of these fossils.

A special contribution to this paper was carried out, by their rich collections of fossils, by two other colleagues: Dr. Mihai Stefănescu, and the late colleague and friend Dr. Dumitru Istocescu; two new species described here below, devoted to these colleagues, underline their help in the accomplishment of the present study.

Almost all the fossils presented below are housed in visitable collections: a large part of them - in the Geologic Museum of the Geological Institute of Romania, Bucharest (IG P); some other - in the Cluj University repository (CU); and some - in the "Crişurilor" Museum (CM), in Oradea.

2. The Ecleja Formation as a lithostratigraphic unit

The first mention in literature of the rock-sequence subsequently defined as "Ecleja Formation" belongs to Fisch (1924) who recognized in the Lower Cretaceous deposits succession exposed north of the Dobreşti locality a "lower group of gray marls and marlstones, followed by a calcareous group", which correspond, more or less, to the formation here discussed.

The name of the Ecleja Formation was introduced by Patrulius (Patrulius et al., 1968), who considered it as being built of "a 700-800 m thick succession of marls and marly-siltstones, its base being made up of organoclastic marly limestones; it is followed by a second, thick bed of gray limestones, rich in pachyodonts".

Later, a breccia ("the Gugu breccia") was recognized immediately below the marly/siltstone succession, proving the transgressive position of the succession above the underlying limestone (Patrulius et al., 1982). Finally, Bordea et al. (1986) proved the real position of the second level of bioconstructed limestones (so called "the median limestone with pachyodonts") as being included in the upper part of the formation, but not constituting its top. This position is obvious in the Râu, Copilu, Urzicari and Rogojel valleys (in the western part of the Pădurea Craiului Mts), where the uppermost part of the Ecleja Formation, developed above the middle limestone with pachyodonts, is still preserved.

Thus, as accepted by the last authors (supra), the Ecleja Formation is constituted of some 800 m thick succession, including four members:

(1) A 5 to 24 m thick succession of gray marlstones and dark spathic, organoclastic limestones, in places fossiliferous (with uppermost Lower Aptian ammonites).

(2) A thick (500-800 m) alternance of marls, marlstones and siltstones, with interbeds up to 10 m thick of light gray or black, bedded limestones. The former of these limestones are rich in miliolids and also contain rare pachyodonts; locally, they are involved in the whole thickness of the formation.

(3) The median limestone with pachyodonts, developed in the upper third of the formation in almost the whole region, and reaching up to 200 m thickness in the Urzicari and Copilu valleys area. It was considered by the first researchers of the region (Patrulius et al., 1968; Istocescu, Bordea, 1970; Patrulius, in Ianovici et al., 1976) as an independent lithostratigraphic unit, next in succession to the Ecleja Formation.

(4) A marly and marly-siltstone succession, with thin sandstone interbeds. It is developed in only two restricted areas: round the Poiana Hill (on the watershed between the Copilu and Urzicari valleys), and in the eastern end of the Varu Hill.

In the Bihor Massif, the Ecleja Formation is exposed discontinuously, along the left slope of the Galbena Valley (Fig.1 B), where it is built up of gray/rusty marls, marlstones, siltstones and sandstones, completely deprived of calcareous interbeds.

3. The fossiliferous sites

The fossiliferous sites displaying ammonites are far more frequent in the Pădurea Craiului Mts, by comparison to the Bihor Massif, in relation with the larger exposures of the Ecleja Formation in the former of these regions. They are situated (Fig. 1 A), from north to south, as follows:

Site 1: on the right slope of the Babei Valley, some 700 m downstream its source in the Varu Hill (large outcrop shown by Istocescu et al., 1970, in their map and exposing a continuous succession from the Upper Bedoulian up to the ?Middle Gargasian). The ammonites were gathered by the cited authors at several levels, unfortunately without any possibility to restore now their bed-by-bed biostratigraphy. Nevertheless, at least two assemblages seem to be superposed:

- in the lower part of the succession - *Dufrenoyia* cf. *discoidalis* CASEY (a fragment from Istocescu's coll., no. IG P 19182 = Pl. II, Fig. 4), *D. notha* CASEY (same collection, IG P 19180 = Pl. II, Fig. 10), and *Dufrenoyia* sp. (cf. *D. subfurcata* KASANSKY; same collection, IG P 19181 = Pl. II, Fig. 11);

- in the upper part of the succession - *Pseudocrioceratites* cf. *pseudoelegans* EGOIAN (Bordea's coll., IG P 19166 = Pl. II, Fig. 1), *Ammonitoceras* aff. *lahuseni* SINZOW (same collection, IG P 19167 = Pl. II, Fig. 3), *Tonohamites* sp. ex gr. *T. limbatus* CASEY (Istocescu's coll., IG P 19173 = Pl. I, Fig. 9), *Ancyloceras ?* sp. (a young fragment displaying lateral ribbing similar to that of *Epancyloceras coquandi* (MATHERON), same collection, IG P 19171 = Pl. II, Fig. 2), *Simionescites ?* sp. (a tuberculated helix only; same collection, IG P 19172 = Pl. I, Fig. 3), *Cheloniceras* cf. *mackesoni* CASEY (a single, mature half-whorl, with dense ribbing and convex unsulcate venter; same collection, IG P 19178), *Colombiceras (C.) tobleri discoidalis* SINZOW (same collection, IG P 19176 = Pl. III, Fig. 4), *Colombiceras (C.)* sp. (a crushed specimen; same collection IG P 19177), *Mathoceras* cf. *sucre* RENZ (same collection; IG P 19183 = Pl. III, fig. 13), *M. istocescui* n. sp. AVRAM (same collection, IG P 19184-19185 = Pl. III, Figs. 9-11), *M. leymerielliforme* n. sp. AVRAM (same collection; IG P 19186 = Pl. III, fig. 12), *Mathoceras ?* sp. (deprived of any lateral tubercle; same collection, IG P 19187 = Pl. III, fig. 14), *Epicheloniceras* sp. aff. *E. martini* (D'ORBIGNY) (same collection, IG P 19179 = Pl. II, Fig. 13), *Uhligella ?* cf. *clansayense* JACOB (same collection, IG P 19189 = Pl. III, Fig. 24; specimen labeled as coming from the "north-west of the Varu Hill", but much younger in age that the other species of the assemblage).

Site 2: on the Hotar Valley, some 600 m upstream the mouth of its tributary on the right, or 450 m below the Varu Hill summit (of Uppermost Gargasian ? - Clansayesian age; also noted by Istocescu et al., 1970, on their map). The ammonite content is: *Holcophylloceras* cf. *kiliani* (SAYN) (a small, smooth example, with typical constrictions; Avram's coll., IG P 19197), *Tetragonites ?* sp. cf. *T. heterosulcatus* ANTHULA (same collection, IG P 19198 = Pl. I, Fig. 1), *Tonohamites ?* cf. *aequicingulatus* (v. KOENEN) (strange late occurrence of an uncoiled, equally ribbed, ventrally tuberculated fragment; same collection, IG P 19199 = Pl. I, Fig. 10), *Acrioceras* sp. ind. (same collection, IG P 19200 = Pl. I, Fig. 4), *Hypacanthoplites* cf. *uhligi* (ANTHULA) (a crushed specimen preserving its right side, from the same collection, IG P 19201 = Pl. III, Fig. 15; besides, a fragment from Istocescu's collection, IG P 19196), *Uhligella ?* sp. (a very fragmentary example, Avram's collection, IG P 19202).

Site 3: on the right tributary of the Hotar Valley, 215 m upstream its mouth. Only a fragmentary and wrongly preserved specimen of *Acanthohoplites* sp. (Avram's coll., IG P 19203) was identified; it prove the Upper Gargasian age.

Site 4: on the same valley, 160 m upstream the site 3, but dislocated (Upper Gargasian or Lower Clansayesian in age). There were identified: *Acanthohoplites* cf. *bigoureti* SINZOW non SEUNES (Avram's collection, IG P 19204 = Pl. III, fig. 18), *A. bigoti levicostatus* EGOIAN (a crushed young specimen, comparable to that figured by Egoian, 1969, on his Pl. XII, Figs. 5a-c; from the same collection, IG P 19205 = Pl. III, fig. 19), *Acanthohoplites* sp. (a fragmentary individual and three very small examples, early loosing their lateral tubercles; same collection, IG P 19206).

Site 5: on the Râu Valley, 400 m downstream the Peştiş Valley mouth (Upper Gargasian). It yielded the species: *Tonohamites ?* sp. (a very incomplete and crushed specimen, from Istocescu's collection, IG P 19192), *Acanthohoplites* sp. (small, stressed specimens, from the same collection, IG P 19191 = Pl. III, Fig. 20), *A.* cf. *aschiltaensis* (ANTHULA) (same collection, IG P 19190 = Pl. III, Fig. 16), *Sanmartinoceras ?* (*Sinzowia ?*) sp. (a very small, widely umbilicate specimen, with only a beginning of typical ornamentation, comparable to that of *S. (S.) stolleyi* CASEY; same collection, IG P 19193= Pl. III, fig. 21).

Site 6: in the left bank of the Surducel Valley, 250 m upstream its mouth. Only *Colombiceras (C.)* sp. (from Cociuba's collection, CU 23235) supported a Gargasian age.

Site 7: in the left bank of the Peştişel Valley (small tributary of the Peştiş Valley downstream its confluence with the Urzicaru Valley). Its Gargasian age is proved only by a specimen of *Colombiceras (C.)* sp. (Cociuba's collection, CU 23236).

Site 8: on the ridge on the left of the Mnierii Valley, 600 m south of the upper boundary of the spathic limestone (saddle on the crest, with two roads). There, *Helicancylus ?* sp. (very fragmentary; Avram's coll., IG P 19209), *Simionescites* sp. (only the end of a phragmocone, with subtriangle-shaped whorl section and fine, dense ribbing starting from periumbilical tubercles; same collection, IG P 19210), *Uhligella ?* sp. (a complete, but very crushed example; same collection, IG P 19213 = Pl. III, Fig. 25), *Colombiceras (C.)* sp. (young individual, displaying rare main ribs, trifurcate from the lateral tubercles, and numerous intercalatories, then ribbing becoming uniform, tubercleless from a relatively small diameter; same collection, IG P 19211 = Pl. III, fig. 8), *Acanthohoplites* sp. aff. *A. bigoti levicostata* EGOIAN (a very crushed and deformed specimen, bearing fine, dense ribs; same collection, IG P 19212 = Pl. III, Fig. 17) argue for the Gargasian age, too.

Site 9: on the Cubleşu Valley, 1 km upstream the Blajul Valley mouth. This site yielded only a *Parahoplites* sp. (Bordea's coll., IG P 19169), proving the Gargasian age.

Site 10: several outcrops on the Vasii Valley, some 250 m upstream the Măgurii Valley and on its watershed to the latter. They expose the lowermost fossiliferous level of the formation, immediately above the lower limestone with pachyodonts[1]. There, the ammonite assemblage is indicative for the uppermost Bedoulian-lowermost Gargasian interval: *Holcophylloceras* sp. (a single, very deformed specimen, displaying 8 constrictions on the outer whorl and relatively rare ribs, projected on venter like in *H. guettardi* (D'ORBIGNY) figured by Anthula, 1899, in Pl. IV, Figs. 5a-b; Preda's coll., IG P 19218), *Phylloceras* s. l. sp. (crushed large examples; Bordea's coll., IG P 19158; Cociuba and Bucur's coll., CU 23237), *Eotetragonites* sp. (a very deformed specimen; Preda's coll., IG P 19219 = Pl. I, Fig. 2), *Toxoceratoides stefanescui* n. sp. AVRAM (three specimens from Ştefănescu's and Avram's coll., IG P 19153 = Pl. I, Fig. 7, and 19214 = Pl. I, Fig. 6, respectively), *Helicancylus aequicostatus* (GABB) (several fragmentary individuals, from Ştefănescu's, Preda's and Cociuba's collections, nos IG P 19152, 19221 = Pl. I, Figs. 12, 13, and CU 23238, respectively). *Acrioceras* sp. ex gr. *A. sarasini* SARKAR (a fragmentary specimen; Ştefănescu's coll., IG P 19151 = Pl. I, Fig. 5), *Dufrenoyia* sp. (large, indeterminable examples from Avram's, and Cociuba and Bucur's colls., nos IG P 19216 and CU 23239, respectively), *D.* cf. *lurensis* (KILIAN) (fragmentary; Cociuba and Bucur's coll., CU 23240 = Pl. II, Fig. 5), *Colombiceras (C.)* cf. *tobleri* (JACOB & TOBLER) (Cociuba and Bucur's coll., CU 23241), *C. (C.)* cf. *tobleri discoidalis* (SINZOW) (Bordea's coll., IG P 19161 = Pl. III, Fig. 3). *C. (C.)* cf. *subpeltoceratoides* (SINZOW) (six incomplete specimens, from Preda's, Bordea's and Ştefănescu's colls., IG P 19222 = Pl. III, Fig. 5, 19160 = Pl. III, Fig. 6 and 19154, respectively), *Colombiceras (C.)* sp. (Bordea's coll., IG P.19162), *C. (Egoianiceras)* cf. *multicostatum* AVRAM (five specimens, from Bordea's and Preda's colls., IG P 19163 and 19223, respectively), *Pseudohaploceras* cf. *matheroni* (D'ORBIGNY) (a single specimen, Avram's coll., IG P 19217 = Pl. III, fig. 23), *Pseudohaploceras* sp. (fragmentary examples from Bordea's, Cociuba and Bucur's, and also Preda's colls., nos. IG P 19165, CU 23242 and IG P 19225, respectively), *Cheloniceras* sp. (very crushed example from Bordea's and Preda's colls., IG P 19164 and 19224).

Site 11: on a small tributary of the Măgurii Valley, at the same level as the site 10. It yielded the species: *Helicancylus* cf. *aequicostatus* (GABB) and *Hamiticeras* cf. *pilsbryi* ANDERSON, both from Ştefănescu's coll. (IG P 19155 and 19156 = Pl. I, Fig. 11, respectively), of the lowermost Gargasian age.

Site 12: on the Runc valley. Two outcrops, very near each other, in the middle course of the valley offered: *Eotetragonites* sp. (very crushed), *Toxoceratoides* cf. *royerianus* (D'ORBIGNY) (fragmentary), *Colombiceras (C.) sinzowi* (KASANSKY), *Colombiceras (C.)* sp., *Cheloniceras* cf. *proteus* CASEY (all in Cociuba's coll., CU 23243, CU 23244 = Pl. I, Fig. 8, CU 23245 = Pl. III, Fig. 2, CU 23246 and CU 23247 = Pl. III, Fig. 1, respectively), proving the uppermost Bedoulian and lowermost Gargasian age.

Site 13: west of Poniţa Peak, between the Runc and Mariei valleys springs. There, Cociuba recorded the species: *Dufrenoyia* sp. (a small, fragmentary mould, no. CU 23252) and *Colombiceras (Egoianiceras)* cf. *multicostatum* AVRAM (no. CU 23248), at the Bedoulian-Gargasian boundary.

Site 14: in the Osiel Hill, near the Coposeni village. The Gargasian age of that exposure is documented by the species: *Cheloniceras* sp. (small and fragmentary, no CU 23249), *Aconeceras* cf. *nissus* (D'ORBIGNY) (small, with relatively wide umbilicus; no. CU 23250 = Pl. III, fig. 22), both in Cociuba's collection.

In the north-western areal of the Bihor Massif (Fig. 1 B) a single fossiliferous **site (15)** was identified, on the left slope of the Galbena valley, some 450 m NW of its confluence with the Seaca Valley. There, the ammonite species (all collected by R. Huza): *Macroscaphites* cf. *yvani striatisulcatus* (D'ORBIGNY) (CM 20759), *Dufrenoyia* cf. *lurensis* (KILIAN) (CM 20761), *D.* cf. *notha* CASEY (CM 20764-20766 = Pl. II, Fig. 6), *D.* cf. *scalata* CASEY (CM 20763 = Pl. II, Fig. 7), *D. praedufrenoyi* CASEY (CM 20767 and 20769 = Pl. II, Figs. 8, 9), *Cheloniceras* cf. *quadrarium* CASEY (CM 20758 = Pl. II, Fig. 12), *Colombiceras (Egoianiceras)* cf. *multicostatum* AVRAM (CM 20762 = Pl. III, Fig. 7) prove the uppermost Bedoulian-lowermost Gargasian age. As mentioned above, the only ammonite previously recorded from this site by Manea, Chivu (1973), *Cheloniceras cornuelianum* (D'ORBIGNY), was destroyed.

4. Biostratigraphy

Almost all the ammonites recorded and published previously from the Ecleja Formation, by Preda (1963), Patrulius et al. (1968), Istocescu et al. (1970) and Patrulius et al. (1982) constitute the object of the present revision; consequently, the biostratigraphic conclusions now grow better these papers inferences.

On the other hand, it is to note that a large part of the ammonites are young or dwarfish and crushed, so that they could be identified only with caution; moreover, in at least one case (Site 1) the common list of fossils, filled in after the main collector (D. I.) death, made impossible a correct restoration of the faunal

[1] After one of the authors (I. C.) this limestone is in fact the median limestone with pachiodonts.

succession. For these reasons, the comments here below are based on the stratigraphic range of only the ammonite genera.

Lowermost in succession is the fossiliferous site 10, where the presence of the *Dufrenoyia* representatives is indicative for the zone with *D. furcata* (SOWERBY) (= Bowerbanki Zone by Casey, 1961), at the top of the Bedoulian substage, sensu Erba (1996). The most important fact of this assemblage is the association, in the basal sequence, no thicker than 2-3 m, of the last *Toxoceratoides* with species of *Dufrenoyia, Colombiceras (C.) (Egoianiceras)* and *Helicancylus*, last genera being considered in literature as newer (Drushchits, Kudrjavcew, 1960; Casey, 1961-1965; Avram, 1974; Demay, Thomel, 1986). This fact proves that the bottom of the Ecleja Formation is condensed, including both the Furcata Zone and a part of the Lower Gargasian Martinoides Zone.

The same sequence was recognized, especially by the presence of *Dufrenoyia* species, in the fossiliferous sites 1 (lower part), 13 and ? 12, while in site 15 the assemblage *Dufrenoyia-Colombiceras (C.)* and *C. (Egoianiceras)* of the site 10 is found again.

Site 1 introduced in the Romanian literature the genus *Mathoceras*, the range of which is Gargasian (the Martinoides Zone) in Venezuela (fide Renz, 1978), but also Clansayesian (the Nolani and Jacobi zones) in Bulgaria (fide Stoykova, 1990). In Romania, its association with *Colombiceras (C.), Ammonitoceras* and especially with *Epicheloniceras* argues the first age (supra). In the same assemblage one fact is also to be signaled: the presence in association of some species of the genera *Pseudocrioceratites* (described from the Clansayesian successions of Caucasus by Egoian, 1969, but also present in the Albian assemblages of the Sinai Peninsula published by Douville, 1916, and also in the Barremian and Lower Bedoulian of Romania - unpublished data), *Simionescites* (described form the Barremian-Bedoulian boundary beds in Caucasus and in Romania) and *Tonohamites ?* (in fact straight fragments of shaft, similar but not identical to it; the known range of the genus being Upper Barremian-lowermost Aptian).

In the site 14, the assemblage *Aconeceras* and *Cheloniceras* leads to a comparable age, i. e. approximately the Furcata Zone of the uppermost Bedoulian.

Sites 3, 4, 5, 8, 9, by the common presence of the *Acanthohoplites/Parahoplites* representatives are located above, in the Gargasian or even in the lowermost Clansayesian; by their position below the middle limestone with pachyodonts, they establish the new stage of the lower part of the Ecleja Formation (member 2 on page 2).

Finally, site 2 is Clansayesian in age as proved by *Hypacanthoplites*; there, the presence of some *Tonohamites ?* representatives lets us suppose the existence all along the Aptian stage of some uncoiled ammonites displaying blunt and simple ribbing, referable to or, at least, comparable to this genus.

5. Conclusions

The Ecleja Formation is accepted here as it was emended by Bordea et al. (1986), i. e. to include both the median limestone with pachyodonts and the upper marly and siltstone succession, beside the basal marlstone-limestone sequence and the main member, of marls, marlstones and siltstones.

The ammonite assemblages point to its Upper Bedoulian (sensu Erba, 1996)-Clansayesian age; all these ammonites were yielded by the lower part of the formation, below the middle limestone with pachyodonts.

A new species of the genus *Toxoceratoides*: *T. stefanescui* n. sp., and two species of the genus *Mathoceras*: *M. istocescui* n. sp. and *M. leymerielliforme* n. sp. were identified and described. In addition, new data concerning the range of some ammonite genera were documented: the presence of *Tonohamites* (or a similar genus) across almost the whole Middle and Upper Aptian; the presence of *Pseudocrioceratites* in the Middle Aptian. Finally, the presence of *Colombiceras (C.) - C. (Egoianiceras)* and *Helicancylus* in the beds with *Dufrenoyia* is a proof that the base of formation is condensed.

6. Systematics

Family Ancyloceratidae MEEK, 1876
Genus *Toxoceratoides* SPATH, 1924
Toxoceratoides stefanescui n. sp. AVRAM
Pl. I, Figs. 6a-b, 7

Holotypus: the specimen figured in Pl. I, Fig. 7 (Ştefănescu's coll., IG P 19153).

Derivatio nominis: in honour of the colleague Dr. Mihai Ştefănescu, renowned Romanian geologist and structuralist.

Locus typicus: left slope of the Vasii Valley, immediately above the Măgurii Valley mouth (site 10, supra).

Stratum typicum: bottom of the Ecleja Formation, Dufrenoyia furcata Zone of the Lower Aptian.

Material: the holotype and two other specimens (Avram's coll., IG P 19214 = Pl. I, Figs. 6 a-b), all recorded in the same stratigraphical interval.

Description. Although very crushed, the specimens here described are apart from all the other species of the genus by evolution of their ornamentation: on the shaft with strong trituberculate, slightly prorsiradiate main ribs (with septate tubercles) and 2-5 thin intercalatories in between (thus resembling the species *Ancyloceras matheronianum* (D'ORBIGNY)), but on the bend and the hook with single, equal, slightly rursiradiate ribs only, deprived of tubercles except rare and discontinuous rib-thickenings at the base of the flanks. The aperture is equipped with a constriction, bordered by two ribs stronger than usual. The holotype illustrates the rapid grow of the shaft, like in the typical species of the genus.

Family Deshayesitidae STOYANOW, 1949
Subfamily Mathoceratinae CASEY, 1964
(incl. Venezuellinae KVANTALIANI, 1980)
Genus *Mathoceras* CASEY, 1964
Mathoceras istocescui n. sp. AVRAM
Pl. III, Figs. 9, 10, 11

Holotypus: the specimen figured in Pl. III, Fig. 9 (Istocescu's coll., IG P 19184).

Derivatio nominis: in the memory of the late colleague Dr. Dumitru Istocescu, one of the first explorers of the Ecleja Formation.

Locus typicus: the Pădurea Craiului Mts in site 1 (supra).

Stratum typicum: Middle Aptian, in the Ecleja Formation.

Material: three fragmentary and oblique compressed examples, all recorded in the same fossiliferous site and level (IG P 19184, 19185).

Description. The holotype preserves almost half a whorl of the body chamber, at a diameter of about 40 mm; this outer whorl displays flat sides, narrow and concave ventral area, rounded umbilical margin and very steep, short umbilical wall, rounding a relatively large (some 1/3 of the diameter) shallow umbilicus. Its ornamentation consists of 9 very sigmoid (almost falcoid) main ribs (probably some 20 in all on a complete whorl), with two short and debile intercalatory ribs in between on all interespaces, stronger only near the external border. The main ribs bear slight umbilical and lateral bullae (the latter located at 2/3 of the sides height) and also oblique perisiphonal clavi at the border of the smooth siphonal band.

Variability. The other specimens illustrate the younger stage, with fewer intercalatory ribs, and the end of the body chamber, respectively, the latter with almost smooth lateral tubercles, stronger peripheral clavi and stronger intercalatory ribs.

Remarks. The only related species: *M. ? laeve* RENZ has the main ribs deprived of lateral bullae.

Mathoceras leymerielliforme n. sp. AVRAM
Pl. III, Fig. 12

Holotypus: the figured example (Istocescu's coll., IG P 19186).

Derivatio nominis: the lateral shape of ornamentation, resembling that of the genus *Leymeriella*.

Locus typicus: the Pădurea Craiului Mts, in site 1 (supra).

Stratum typicum: Middle Aptian, in the Ecleja Formation.

Material: only the holotype.

Description. The holotype is small, compressed and preserves the aged one third of the body chamber. Nevertheless, it is apart from all the other species of the genus by its denser main ribs (10 on 1/3 of the last whorl), with only one, very thin intercalatory rib in between up to the end of the shell, where the last three main ribs are consecutive. The main ribs bear minute thickenings at the umbilical margin, small bullae on the external third of the sides, and strong, oblique clavi at the siphonal margin. Between the lateral bullae and the marginal clavi the main ribs enlarge and become slightly concave, like in *Leymeriella*. The whorl section preserves the flat sides, the narrow concave venter (with a siphonal sinus), the rounded umbilical margin and the steep, short umbilical wall.

Remarks. The similarity with *Leymeriella* ex gr. *tardefurcata* (LEYM. (D'ORB.)), especially of the left side of the holotype, is obvious, but the presence of the intercalatory ribs (which completely lack in *Leymeriella*) and of the lateral bullae located very high on the sides, are features typical of *Mathoceras*.

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PLATES

Plate I

Fig. 1 — *Tetragonites ?* sp. cf. *T. heterosulcatus* ANTHULA, Avram's coll. (IG P 19198), from the Hotar Valley, some 450 m below the Varu Hill summit (Site 2). Uppermost Gargasian or lowermost Clansayesian.

Fig. 2 — *Eotetragonites* sp., Preda's coll. (IG P 19219) from the Vasii Valley left side, some 250 m upstream the Măgurii Valley mouth (Site 10). Uppermost Bedoulian or lowermost Gargasian.

Figs. 3a-b — *Simionescites ?* sp., Istocescu's coll. (IG P 19172) from the right slope of the Babei Valley, some 700 m downstream its source (Site 1). Uppermost Bedoulian or lowermost Gargasian.

Fig. 4 — *Acrioceras* sp. Avram's coll. (IG P 19200) from the same site (2) and interval as Fig. 1.

Fig. 5 — *Acrioceras* sp. ex gr. *A. sarasini* SARKAR. Ştefănescu's coll. (IG P 19151), from the same site (10) and interval like Fig. 2.

Figs. 6 a-b, 7 — *Toxoceratoides stefanescui* n. sp. AVRAM: 6=paratype (Avram's coll., IG P 19214): 7=holotype (Ştefănescu's coll., IG P 19153). Both of them from the same site (10) as Fig. 2. Uppermost Bedoulian.

Fig. 8 — *Toxoceratoides* cf. *royerianus* (D'ORBIGNY), Cociuba's coll. (CU 23244) from the Rune Valley (Site 12). Uppermost Bedoulian.

Figs. 9 a-b — *Tonohamites* sp. ex gr. *T. limbatus* CASEY, Istocescu's coll. (IG P 19173), from the same site (1) and interval as Fig. 3.

Fig. 10 — *Tonohamites ?* cf. *aequicingulatus* (v. KOENEN), Avram's coll.. (IG P 19199), from the same site (2) and interval as Fig. 1.

Fig. 11 — *Hamiticeras* cf. *pilsbryi* (ANDERSON), Ştefănescu's coll. (IG P 19156), from the tributary on the left of the Măgurii Valley near its mouth (Site 11). Lowermost Gargasian.

Figs. 12, 13 — *Helicancylus aequicostatus* (GABB.), Preda's coll. (IG P 19221), from the same site (10) and interval as Fig. 2.

All the specimens are figured in natural size.

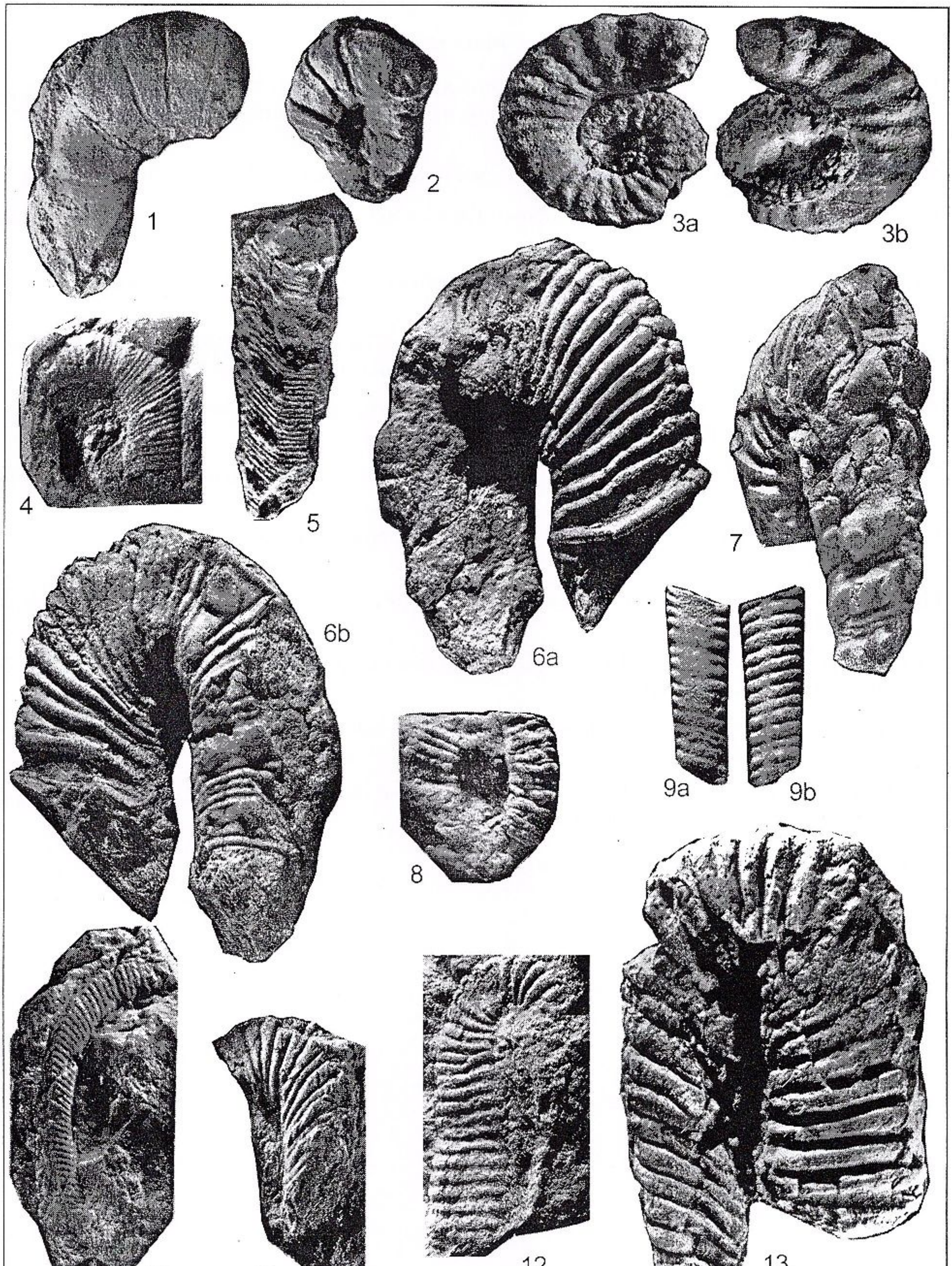
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Plate II

Fig. 1 — *Pseudocrioceratites* cf. *pseudoelegans* EGOIAN, Bordea's coll. (IG P 19166), from the right slope of the Babei Valley, some 700 m downstream its source (Site 1). Uppermost Bedoulian or lowermost Gargasian.

Fig. 2 — *Ancyloceras* sp., Istocescu's coll. (IG P 19171), from the same site (1) and interval like Fig. 1.

Fig. 3 — *Ammonitoceras* aff. *lahuseni* (SINZOW), Bordea's coll. (IG P 19167), from the same site (1) as Fig. 1. Gargasian.

Figs. 4 a-b — *Dufrenoyia* cf. *discoidalis* CASEY, Istocescu's coll. (IG P 19182), from the same site (1) as Fig. 1. Uppermost Bedoulian.

Figs. 5 a-b — *Dufrenoyia* cf. *lurensis* (KILIAN), Cociuba and Bucur's coll. (CU 23240), from the quarry in the Măgurii Valley right side (extension of Site 10). Uppermost Bedoulian.

Fig. 6 — *Dufrenoyia* cf. *notha* CASEY, Huza's coll. (CM 20764), from the Galbena Valley left side (Site 15). Uppermost Bedoulian.

Fig. 7 — *Dufrenoyia* cf. *scalata* CASEY, Huza's coll. (CM 20763), from the same site (15) and interval like Fig. 6.

Figs. 8, 9 — *Dufenoyia praedufrenoyi* CASEY, Huza's coll. (CM 20767 and 20769), from the same site (15) and interval like Fig. 6.

Figs. 10 a-b — *Dufrenoyia notha* CASEY, Istocescu's coll. (IG P 19180), from the same site (1) and interval as Fig. 4.

Figs. 11 a-b — *Dufrenoyia* sp. (cf. *D. subfurcata* KASANSKY), Istocescu's coll. (IG P 19181), from the same site (1) and interval like Fig. 4.

Figs. 12 a-b — *Cheloniceras* cf. *quadrarium* CASEY, Huza's coll. (CM 20758), from the same site (15) and interval like Fig. 6.

Fig. 13 — *Epicheloniceras* sp. (aff. *E. martini* (D'ORGIBNY)), Istocescu's coll. (IG P 19179), from the same site (1) as Fig. 1. Lowermost Gargasian.

All the specimens are figured in natural size.

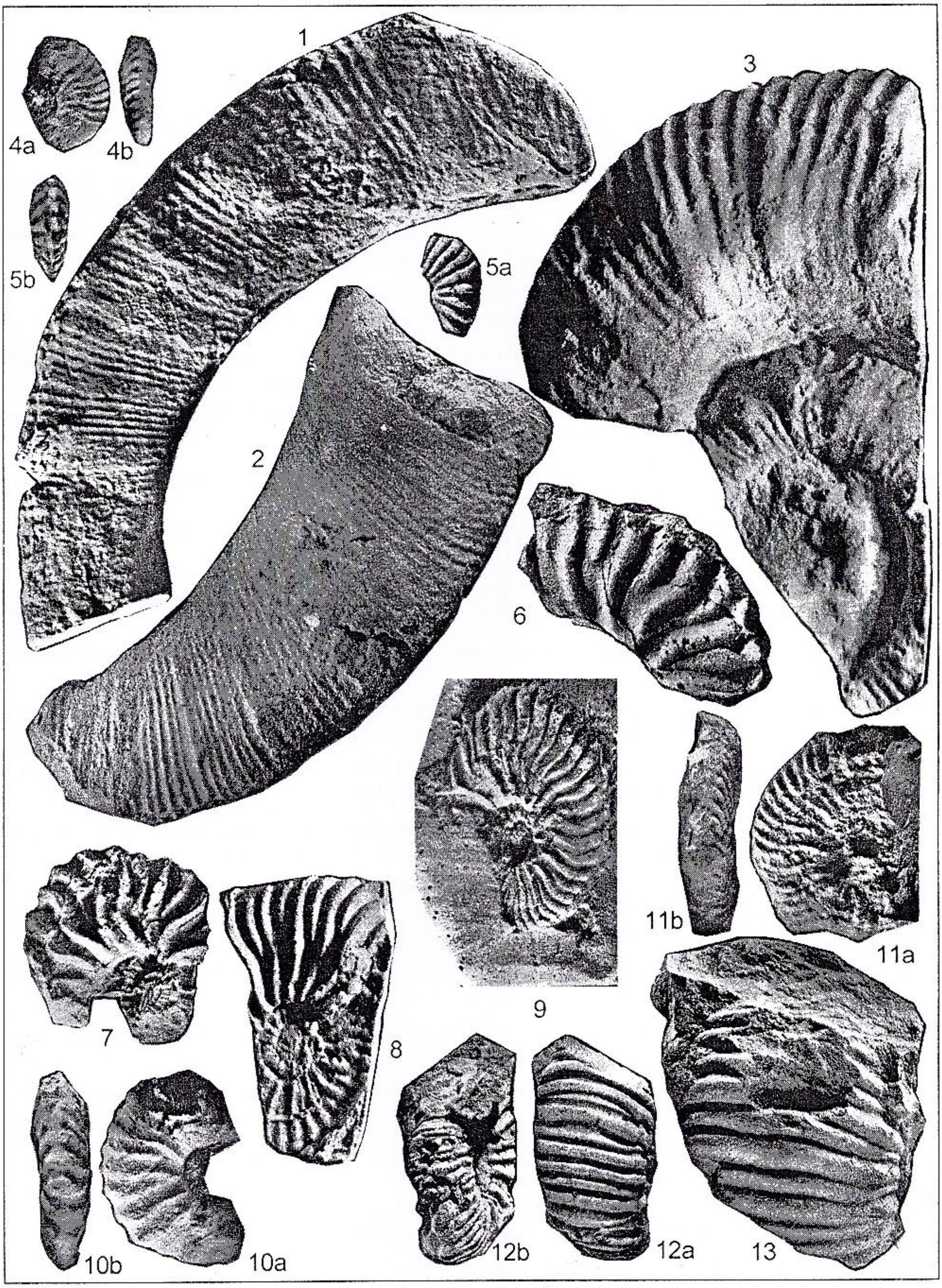
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Plate III

Figs. 1 a-b — *Cheloniceras* cf. *proteus* CASEY, Cociuba's coll. (CU 23247), from the Runc Valley (Site 12). Uppermost Bedoulian.

Fig. 2 — *Colombiceras (C.) sinzowi* (KASANSKY), Kociuba's coll. (CU 23245), from the same site (12) as Fig. 1. Gargasian.

Fig. 3 — *Colombiceras (C.)* cf. *tobleri discoidalis* (SINZOW), Bordea's coll. (IG P 19161), from the Vasii Valley, some 250 m upstream the Măgurii Valley mouth (Site 10). Lower Gargasian.

Fig. 4 — *Colombiceras (C.) tobleri discoidalis* (SINZOW), Istocescu's coll. (IG P 19176), from the right slope of the Babei Valley, some 700 m downstream its source (Site 1). Lower Gargasian.

Figs. 5, 6 — *Colombiceras (C.)* cf. *subpeltoceratoides* (SINZOW): 5=Preda's coll. (IG P 19222); 6=Bordea's coll. (IG P 19160); both of them from the same site (10) like Fig. 3. Lower Gargasian.

Figs. 7 a-b — *Colombiceras (Egoianiceras)* cf. *multicostatum* AVRAM, Huza's coll. (CM 20762), from the left slope of the Galbena Valley (Site 15). Lower Gargasian.

Fig. 8 — *Colombiceras ?* sp., Avram's coll. (IG P 19211), from the crest on the left of the Mnierii Valley (Site 8). Upper Gargasian.

Figs. 9-11 — *Mathoceras istocescui* n. sp. AVRAM, Istocescu's coll. (IG P 19184-19185); 9=holotype (IG P 19184). All of them from the same site (1) like Fig. 4. Lower Gargasian.

Fig. 12 — *Mathoceras leymerielliforme* n. sp. AVRAM, holotype. Istocescu's coll. (IG P 19186), from the same site (1) and interval like Figs. 4 and 9-11.

Fig. 13 — *Mathoceras* cf. *sucre* RENZ, Istocescu's coll. (IG P 19183), from the same site (1) and interval like Figs. 4, 9-11 and 12.

Fig. 14 — *Mathoceras ?* sp. (deprived of lateral tubercles), Istocescu's coll. (IG P 19187), from the same site (1) and interval like Figs. 4, 9-11, 12, 13.

Fig. 15 — *Hypacanthoplites* cf. *uhligi* (ANTHULA), Avram's coll. (IG P 19201), from the Hotar Valley, some 450 m below the Varu Hill summit (Site 2). Clansayesian.

Fig. 16 — *Acanthohoplites* cf. *aschiltaensis* (ANTHULA), Istocescu's coll. (IG P 19190), from the Râu Valley, 400 m downstream the Peştiş Valley mouth (Site 5). Upper Gargasian.

Figs. 17, 19 — *Acanthohoplites* sp. cf. *A. bigoti levicostatus* EGOIAN, Avram's coll.: 17 (IG P 19212), from the same site (8) and interval like Fig. 8; 19 (IG P 19205), from the right tributary of the Hotar Valley (Site 4). Upper Gargasian or Lower Clansayesian.

Fig. 18 — *Acanthohoplites* cf. *bigoureti* SINZOW non SEUNES, Avram's coll. (IG P 19204), from the same site (4) and interval like Fig. 19.

Fig. 20 — *Acanthohoplites* sp., Istocescu's coll. (IG P 19191), from the same site (5) and interval like Fig. 16.

Fig. 21 — *Sanmartinoceras ? (Sinzowia)* sp. ind., Istocescu's coll. (IG P 19193), from the same site (5) and interval like Fig. 16.

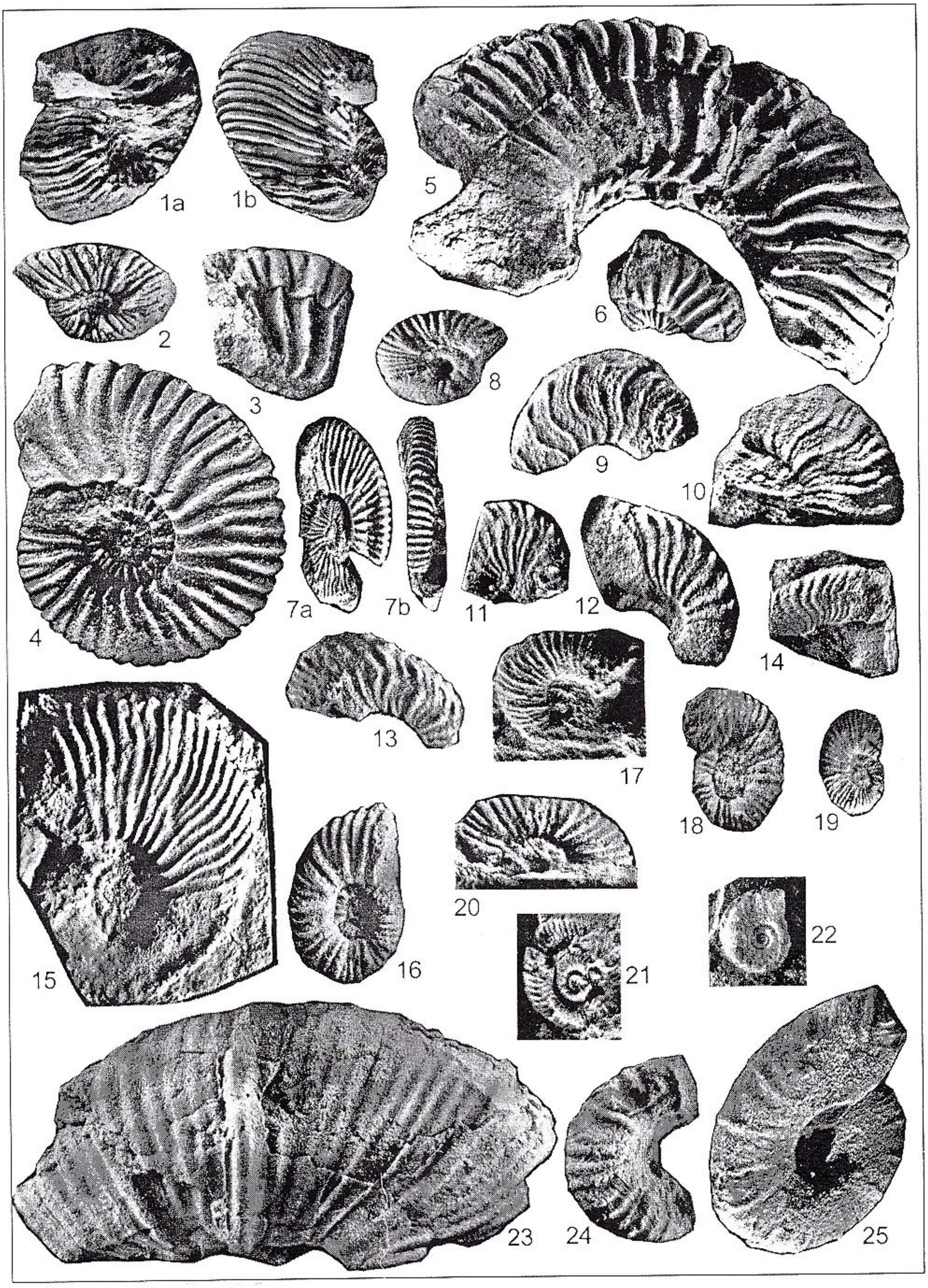
Fig. 22 — *Aconeceras* cf. *nissus* (D'ORBIGNY), Cociuba's coll. (CU 23250), from the Osiel Hill (Site 14). Uppermost Bedoulian.

Fig. 23 — *Pseudohaploceras* cf. *matheroni* (D'ORBIGNY), Avram's coll. (IG P 19217), from the same site (10) like Fig. 3. Uppermost Bedoulian or lowermost Gargasian.

Fig. 24 — *Uhligella ?* cf. *clansayense* JACOB, Istocescu's coll. (IG P 19189), from the same site (1) and interval (?) like Fig. 4.

Fig. 25 — *Uhligella ?* sp., Avram's coll. (IG P 19213), from the same site (8) and interval like Fig. 8.

All the specimens are figured in natural size, except those in Figs. 21 and 22 which are augmented 2 x.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CONTRIBUTIONS BIOSTRATIGRAPHIQUES SUR LES FORMATIONS CRÉTACÉES ET PALÉOGÈNES DE LA RÉGION VALEA PUTNEI-VALEA CAŞINULUI (MONTS DE VRANCEA)

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Abstract: *Biostratigraphic Contributions to the Cretaceous and Paleogene Formations in the Putna Valley-Caşin Valley Area (Vrancea Mountains).* The lithostratigraphic units in the external part of the Marginal Folds Nappe (namely Cozla Digitation) have been studied in micropaleontologic, palynologic and nannoplankton respectively. New data have been obtained, as follows: Streiu Beds begin in the Barremian (*Gaudryna oblonga*) and include the upper part of the Upper Albian and possibly the Lower Vraconian (*Apteodinium grande*). The base of the Tisaru Beds belongs to the Upper Vraconian (Ovoidinium verrucosum Zone); they comprise the Cenomanian. Turonian (*Praeglobotruncana* cf. *stephani, Marginotruncana marinosi, Thalmannamina meandertornata*), and their upper part, the Senonian (*Isabelidinium cooksoni*, Santonian or Campanian ?). Lepşa Beds are palynologically sterile and contain a poor microfauna, of Lower Campanian age. Caşin Beds belong to the Spiroplectammina spectabilis and Glomospira diffundens zones of the Upper Pleocene. The Upper Caşin Beds also include the Spiroplectammina spectabilis zone; at the upper part they contain the Deflandrea speciosa and Apectodinium homomorphum zones. Piatra Uscată Beds include the lower half of the Apectodinium augustum zone (terminal Paleocene/Ypresian) as well as the S. spectabilis Zone. The Apectodinium homomorphum and A. augustum zones have been identified in Romania for the first time. In the upper part of the Buciaşu Beds the Turborotalia cerroazulensis-Turborotalia pseudomayeri zones have been identified, the Lutetian/Bartonian boundary being marked by the Wilsonidium echinosuturatum Zone. Bisericani Beds contain *Turborotalia pomeroli* and *Cyclammina rotundidorsata*, of Priabonian age; the nannoplankton Isthmolithus recurvus/Sphaenolithus pseudoradians and the Kisselovia clatharata angulosa and Rhombodonium perforata zones also indicate the Priabonian age. The Eocene/Oligocene boundary is situated in the upper third of the Globigerina Marls (*Globigerina tapuriensis, G. officinalis*, the first occurrence of the species *Phtanoperidinium amoenum* and *Bohlensipollis hohli*). The slaty shales contain the Wetzeliella gochti and Helicopontosphaera reticulata/Ericsonia subdisticha zones. The lower menillites and the brown marls belong to the Rupelian (zone P17 Globigerina tapuriensis, G. selii, etc.). The lower dysodiles and the Kliwa sandstone belong to the W. gochtii Zone of the Rupelian. Regional correlations and paleogeographic considerations are made for the lithostratigraphic successions of the Marginal Folds of the Vrancea Mountains in comparison with the other lithofacies developed in the East Carpathians Flysch.

Dans les vallées de la Putna, Lepşa, Streiu et Caşin qui traversent les monts de Vrancea, affleure sur des coupes bien exposées, l'ensemble de la succession lithostratigraphique de la partie externe de la Nappe des Plis Marginaux, notamment de la Digitation de Coza. Cette succession, qui est une de plus complète de la zone du Flysch des Carpathes Orientales, comprend (Dumitrescu, 1963; Dumitrescu et al., 1971) les unités lithostratigraphiques suivantes:

a) **Couches de Streiu**, constituées par une alternance, souvent rythmique (de type flysch), d'argiles et argiles silteux noires ou noirâtres et de minces grès calcaires, souvent convolutes; des couches ou des ellipsoides de calcaires sidéritiques s'intercalent à differents niveaux. Ce développement fait partie du lithofaciès "Silésien" ou du type Schistes Noirs, largement connu dans le Crétacé inférieur de la zone du Flysch carpathiques.

b) **Couches de Tisaru**, dont l'élément spécifique sont les silicolites, noirs et vedatres dans la partie inférieure (Couches de Tisaru inférieures), rouges ou barriolés dans la partie supérieure (Couches de Tisaru supérieures). Des minces argiles noirs ou noirâtres s'intercalent parmis les radiolarites dans le tiers inférieur peuvent etre reconnus des niveaux d'argiles vert clair, bentonitiques.

c) **Couches de Lepşa**, à prédominance marno-calcaire, litées, gris clairs, gris-cendrés ou gris-verdâtres, riches en débris d'inocerames.

Les Couches de Tisaru et de Lepşa couvrent ensemble le Crétacé supérieur.

d) **Couches de Caşin**, qui se laissent diviser en trois séquences, celle inférieure, marnocalcaire, celle moyenne à calcaires marneux siltiques et celle supérieure, argilo-marneuse à intercalations minces et de microbrèches calcaires à *Lithothamnium* et briozoaires. Un "cachet" général des Couches de Caşin est un faible caractère anoxique (bitumineux). Les Couches de Caşin sont considérées d'age paléocène.

e) **Couches de Piatra Uscată**, constituées par une séquence de flysch schisteux (les pélites dominent les grès minces, quartzitiques) à argiles gris-cendrés ou verdatres, rarement rouge-griottes.

f) **Couches de Buciaş**, représentées par des marnes et marnocalcaires, lités, parfois silteux, gris ou vert clair; avec de fins "chailles" ou lits silicifiés.

g) **Couches de Bisericani**, argilo-marneuses, feuilletées ou finement litées, gris sombre, avec des très minces intercalations de grès fins. La séquence de base est souvent constituée par des argiles route-griottes (niveau de Strujinoasa).

h) **Grès de Lucăceşti et/ou Marnes à Globigérines**, avec des grès quartzeux glauconieux (Grès de Lucăceşti) qui peut admettre des intercalations de marnes, quelquefois compactes, blanchatres (Marnes à Globigérines); ces derniers rarement substituent entièrement les grès.

g) **Grès de Fierăstrău et/ou Schistes Ardésiformes**, également constitué de grès quartzeux, de type Kliwa, ayant des intercalations d'argiles silteux bitumineux (Schistes Ardésiformes), cassantes, qui peuvent substituer entièrement les grès.

i) **Menilites (silicolites bitumineux) inférieurs et Marnes Brunes** (marnocalcaires bitumineux).

j) **Schistes Dysodiliques inférieures** (argiles bitumineux, feuilletés ou finement lités, avec des très minces intercalations de grès quartzeux).

k) **Grès de Kliwa**, constitué d'une séquence massive de grès quartzeux, avec des minces intercalations de schistes dysodiliques.

La limite Paléogène/Miocène est située au-dessus du Grès de Kliwa, dans les Schistes Dysodiliques supérieures. La succession continue dans le Miocène inférieur, mais cette partie ne fait pas l'objet de cette note.

CRÉTACÉ

Les formations du Crétacé ont été étudiées sur les coupes des vallées Putna, Streiu et Lepşa.

Couches de Streiu

Nous avons échantilloné les Couches de Streiu qui affleurent dans le lit de la rivière Putna, sur le flanc oriental de l'anticlinal de Streiu depuis la voute axiale jusqu'à la limite avec les Couches de Tisaru inférieures.

Microfaune

Les associations de microforaminifères ont été identifiées dans le tiers inférieur et celui moyen des Couches de Streiu et sont constituées exclusivement de foraminifères aglutinées. Parmi les taxons importants pour la biostratigraphie on mentionne *Hippocrepina depressa*, *Verneuilinoides neocomiensis*, *V. subfiliformis*, *Trochammina vocontiana*, *Gaudryna oblonga*, *Reophax minutus*, *Pseudobolivina variabilis*. Sans pouvoir délimiter précisement des biozones définies par ces taxons, on peut neanmoins aprecier qu'ils se grouppent dans deux associations principales: (1) à *Gaudryna oblonga* et (2) à *Pseudobolivina variabilis* et *Reophax minuta* (GEROCH et NOVAK, 1984). La première correspond au Barrémien inférieur, la seconde au Barrémien supérieur-Aptien. Les FAD et LAD des taxons significatifs, que nous avons pu observer, ne coincident pas avec ceux présentés par Geroch et Novak (1984), mais les différences ne sont pas trop grandes. Elles peuvent etre dues à la fréquence différente des échantillonages et/ou aux lithofaciès plus au moins différents. Donc, selon la microfaune, les premiers deux tiers des Couches de Streiu appartiennent au Barrémien-Aptien.

Palynologie

Les données palynologiques sont assez pauvres pour la plupart des Couches de Streiu. A des nombreux niveaux échantillonées (GL 13, GL 15, GL 17, GL 27, GL 28, GL 32), comprenant la presque totalité des Couches de Streiu - hormis la partie supérieure - on à trouvé le meme palynofaciès constitué d'une masse de tissus végétaux à laquelle s'associent de très rares spores appartenant aux genres *Dictyophyllidites, Cicatricosisporites, Appendicisporites, Gleicheniidites* et de fort rares dinoflagellés, surtout des *Cribroperidinium* appartenant à *C. orthoceras* (EISENACK, 1958) DAVEY, 1969.

Cette association semble appartenir à la zone à *Cribroperidinium* spp. (Antonescu, inédit) zone d'intervalle équivalente à l'Aptien supérieur et l'Albien moyen.

Dans la partie supérieure des Couches de Streiu, sur la coupe de la rivière Putna, à 50 m en aval de l'embouchure de ruisseau Lepşa et à 30 m (15 m épaisseur stratigraphique) au-dessous de la limite avec les Couches de Tisaru s'intercallent des argiles silteux vertes ou gris-verdatres. Dans deux niveaux (GL 33, GL 33 bis) nous avons identifié dans ces argiles une microflore fort riche en individus et pauvre en genres et espèces, constituée seulement par des dinoflagellés (échantillons 4A28, 4A29) dont l'espèce *Apteodinium grande* COOKSON & HUGHES, 1964 constitue plus de 95 % de l'association. A celle-ci s'ajoutent *Cribroperidinium edwardsi* (COOKSON & EISENACK, 1958) DAVEY, 1969; *Lithosphaeridium arundum* (EISENACK & COOKSON, 1960) DAVEY, 1970 emend. LUCAS-CLARK, 1984; *Subtilisphaera* sp., *Apteodinium grande* est une espèce avec un range restreint à l'Albien supérieur et au Vraconien (Verdier, 1975; Davey & Verdier, 1973) précisant l'age des niveaux supérieurs des Couches de Streiu.

Il faut rappeler que Antonescu (dans Antonescu, Ion & Alexandrescu, 1978) a identifié dans la partie terminale des grès glauconieux qui représente la séquence supérieure des Schistes Noirs dans les Nappes d'Audia et de Tarcău, une riche association de dinoflagellés ou *A. grande* est associé avec *Ovoidinium verrucosum* (COOKSON & HUGHES, 1964) DAVEY, 1970 et *Endoceratium (=Pseudoceratium) dettmaniae* (COOKSON & HUGHES, 1964) STOVER & EVITT, 1978 qui sont caractéristiques pour le Vraconien. Au dessous de ces niveaux a été trouvé une association palynologique de l'Albien supérieur, à *Lithosphaeridium siphoniphorum* (COOKSON & EISENACK, 1958) DAVEY & WILLIAMS, 1966 emend. LUCAS-CLARK, 1984 et *Carpodinium obliquicostatum* COOKSON & HUGHES, 1964. Les associations palynologiques déterminées dans les Schistes Noirs ont d'ailleurs constitué le support fondamental pour définir la zone à *L. siphoniphorum* et *C. obliquicostatum* de l'Albien supérieur et la zone à *O. verrucosum* et *E. dettmanniae* du Vraconien (Antonescu dans Ion et al., 1987; Antonescu données inédites).

Puisque l'association à *O. verrucosum* et *E. dettmanniae* n'a pas encore été trouvée dans la partie supérieure des Couches de Streiu où on a trouvé seulement l'association à *A. grande* (qui débute dans la zone à *L. siphoniphorum* et *C. obliquicostatum*), on considère qu'elle revient à l'Albien supérieur et éventuellement au Vraconien basal?

Tenant compte des associations micropaléontologiques et palynologiques, nous estimons que les Couches de Streiu affleurant dans la vallée de la Putna couvrent l'intervalle Barrémien-Albien supérieur, le Vraconien pouvant y etre compris dans la partie terminale de la succession.

Les dinoflagellés de la partie supérieure des Couches de Streiu peuvent etre équivalents avec ceux de l'Albien supérieur (zone Inflatum) du bassin de Paris (des coupes des stratotypes - Verdier, 1975).

Deux biozones de foraminifères, à ?*Gaudryna oblonga* et *Pseudobolivina variabilis* et *Reophax minutus* pour le Barrémien-Aptien et les zones de dinoflagellée à *Cribroperidinium* spp. et *Lithosphaeridium siphoniphorum* et *Carpodinium obliquicostatum* ont été identifiées dans l'Aptien supérieur- Albien moyen et l'Albien supérieur des Couches de Streiu (tab. 1 et 2).

Couches de Tisaru

Les coupes de la vallée de la Putna et de la vallée de Streiu ont fourni autant de données micropaléontologiques que palynologiques.

Microfaune

Sur les coupes de la vallée de la Putna (tab. 3) et la vallée de Streiu (tab. 4) ont été determinées parmi d'autres espèces *Thalmannammina meandertornata*, *Reophax duplex*, *Dorothia oxycona*, ? *Rotalipora* sp., ? *Marginotruncana* sp., *Praeglobotruncana* cf. *stephani* et *Marginotruncana marinosi*. Reduite, et dans un état de conservation médiocre, cette association de foraminifères plantoniques marque un intervalle large Cénomanien-Sénonien inférieur (?). L'espèce qui a le range le plus restreint, *Marginotruncana marinosi*, débute à la partie supérieure du Turonien moyen et s'éteint à la fin du Turonien supérieur (Robaszynski et al., 1979). *Praeglobotruncana* cf. *stephani* a un range plus large et correspond à l'intervalle Albien supérieur-Turonien moyen. *Thalmannammina meandertornata* est une présence commune dans les dépôts du Turonien de la partie supérieur du Flysch Courbicortical supérieur.

Tenant compte de ces données nous estimons que la microfaune déterminée dans les couches de Tisaru peut etre rangée dans le Cénomanien et le Turonien et monte probablement aussi dans le Sénonien inférieur; elle correspond aux biozones de foraminifères aglutinées, *Bulbobaculites* et *Uvigerinammina jankoi* (après Geroch & Nowak, 1983).

Palynologie

Le niveau de base des couches de Tisaru (GL 36, échantillon 4A34) affleurant sur la coupe de la vallée de la Putna, a fourni une riche microflore composée presque uniquement de dinoflagellés. Elle comprend: *Palaeohystrichophora infusorioides* DEFLANDRE, 1935; *Ovoidinium scabrosum* (COOKSON & HUGHES, 1964) DAVEY, 1970; *Apteodinium grande*, *Cleistosphaeridium ancoriferum* (COOKSON & EISENACK, 1960) DAVEY, DOWNIE, SARJEANT & WILLIAMS, 1966 emend. COOKSON & EISENACK, 1968); *Florentinia mantelli* (DAVEY & WILLIAMS, 1966) DAVEY & VERDIER, 1973; *Florentinia deanei* (DAVEY & WILLIAMS, 1966) DAVEY & VERDIER, 1973; *Florentinia* sp., *Lithosphaeridium siphoniphorum*, *Stephodinium coronatum* DEFLANDRE, 1936; *Endoscrinium campanula* (GOCHT, 1959) VOZZHENNIKOVA, 1967; *Achomosphaera ramulifera* (DEFLANDRE, 1937) EVITT, 1963; *Xiphophoridium alatum* (COOKSON & EISENACK, 1962) SARJEANT, 1966; *Pervosphaeridium pseudhystrichodinium* (DEFLANDRE, 1937) YUN, 1981; *Leberidocysta chlamidata* (COOKSON & EISENACK, 1962) STOVER & EVITT, 1978; *Odontochitina operculata* (O. WETZEL, 1933) DEFLANDRE& COOKSON, 1953; *Cribroperidinium edwardsi*; *Cribroperidinium* spp.; *Hystricosphaeridium tubiferum* (EHRENBERG, 1838) DEFLANDRE, 1937, emend. DAVEY & WILLIAMS, 1966; *Coronifera oceanica* (COOKSON & EISENACK, 1958) MAY, 1980; *Spiniferites ramosus ramosus* (EHRENBERG, 1838) DAVEY & WILLIAMS, 1966; *S. ramosus gracilis* (DAVEY & WILLIAMS, 1966) LENTIN & WILLIAMS, 1973; *Hystrichostrogylon* sp., *Oligosphaeridium complex* (WHITE, 1842) DAVEY & WILLIAMS, 1966; *Kiokansium polypes* (COOKSON & EISENACK, 1962) BELOW, 1982.

L'association de *Ovoidinium scabrosum*, qui s'éteint à la partie terminale du Vraconien, avec *Palaeohystrichophora infusorioides*, qui débute dans la partie terminale du Vraconien indique clairement l'âge Vraconien supérieur de la partie basale des Couches de Tisaru. Il s'agit de la zone palynologique à *Ovoidinium verrucosum* et *Endoceratium dettmanniae* spécifique pour le Vraconien connue dans les nappes d'Audia, de Tarcău et du Flysch Courbicortical (Antonescu dans Ion et al., 1987)[1].

Quelques mètres plus-haut, toujours sur la coupe de la vallée de la Putna, l'association de dinoflagellés change. *A. grande* et *O. scabrosum* qui ont leur LAD à la fin du Vraconien disparaissent. A différents niveaux (GL 41, GL 45, etc.) des nouvelles espèces font leur apparition à coté des genres et espèces qui persistent de l'association vraconniene décrite à la base des Couches de Tisaru (tab. 2). L'association comprend *Epelidosphaeridia spinosa* (COOKSON & HUGHES, 1964) DAVEY, 1969; *Callaiosphaeridium asymmetricum* (DEFLANDRE & COURTEVILLE, 1939 DAVEY & WILLIAMS, 1966; *Trichodinium castanea* (DEFLANDRE, 1935) CLARKE & VERDIER, 1967; *Xenascus ceratioides* (DEFLANDRE, 1937) LENTIN & WILLIAMS, 1973; *Pterodinium cingulatum* (O. WETZEL) BELOW, 1981; *Trithirodinium suspectum* (MANUM & COOKSON, 1964) DAVEY, 1969; *Gonyaulacysta cassidata* (EISENAK & COOKSON, 1960) SARJEANT, 1966. Les espèces *L. siphoniphorum* et *C. ancoriferum*, ont été rencontrées seulement dans la moitié inférieure des Couches de Tisaru.

L'age de l'association qu'on vient de citer, suivant le "range" connu des genres et espèces semble couvrir le Cénomanien inférieur, le Cénomanien moyen et probablement, le Cénomanien supérieur et le Turonien.

[1]M. Săndulescu=levées géologiques, structure, corrélations régionales et paléogéographiques; E. Antonescu=palynologie du Crétacé et du Paléocène; D. Brotea=microfaune de l'Eocène et de l'Oligocène; E. Platon=microfaune du Crétacé et du Paléocène; A. Ionescu=palynologie de l'Eocène et de l'Oligocène; E. Strusievici=nannoplankton.

L'association du *Palaeohystrichophora infusorioides* avec *Epelidosphaeridia spinosa* précise que la moitié inférieure du Couches de Tisaru correspond à la zone à *P. infusorioides* et *E. spinosa*, d'age Cénomanien inférieur et moyen. L'age de l'association à *Xenascus ceratioides* et *Pervosphaeridium pseudhystrichodinium*, selon le "range" des genres et espèces, semble etre Cénomanien supérieur-Turonien. Cette association présente des traits particuliers, qui n'ont été pas encore rencontrés dans d'autre régions ou le Cénomanien supérieur et le Turonien ont pu être caractérisés par la microflore et ou les formations contiennent du matériel palynologique continental important (Flysch Courbicortical supérieur, Cénomanien supérieur de la Dépression de Haţeg, Couches de Dumbrăvioara à Tohan, Lăicăi, etc.; Antonescu dans Ion et al., 1987). La différence est surtout déterminée par les caractères de l'environement dans lequel se sont déposées les Couches de Tisaru, ésentiellement marin, voir océanique, sans apport de microflore continentale. L'association à *Xenascus ceratoides* et *Pervosphaeridium pseudhystrichodinium* présente des traits communs, avec l'association de dinoflagellés de la formation des argiles bigarrées et des calcaires en plaquettes des nappes d'Audia et Tarcău.

Dans la partie supérieure des Couches de Tisaru, sur la coupe de la rivière Putna, dans l'affleurement situé au bord de la route, à la limite orientale du village Lepşa (niveaux GS 1-3, échantillon 4A161), on constate un nouveau changement dans l'association des dinoflagellés. Des nouveaux genres et espèces apparaissent, tells *Isabelidinium* sp. ex gr. *I. cooksoni* (ALBERTI, 1959) LENTIN & WILLIAMS, 1977; *Ginginodinium* sp., *Florentinia* sp. cf. *F. buspina* (DAVEY & VERDIER, 1976) DUXBUTY, 1980, pollen du groupe *Normapolles* PFLUG, 1953 - *Semioculopollis* sp., à coté des genres et espèces qui persistent de l'association inférieure, *Odontochitina operculata, Leberidocysta chlamidata, Pervosphaeridium pseudhystrichodinium, Pterodinium cingulatum, Spiniferites ramosus, Endoscrinium campanula, Palaeohystrichophora infusorioides, Areoligera* sp.

L'age de l'association de la partie supérieure des Couches de Tisaru appartient au Sénonien, probablement au Santonien-Campanien inférieur?. Le genre *Isabelidinium* est connu depuis le Turonien (moyen-supérieur) et ayant son "acme- zone" au Sénonien, tout autant que *Semioculopollis*; le genre *Ginginodinium* débute au Sénonien. Quique cette association a été trouvée dans un seul niveau de la coupe de la Putna (au dessus de celui-ci il n'y a que de radiolarites rouges stériles pour la palynologie) elle est suffisante pour prouver la présence d'une partie du Sénonien inférieur dans la partie supérieure des Couches de Tisaru.

La corrélation du contenu micropaléontologique et palynologique des Couches de Streiu de la Nappe des Plis Marginaux avec celui de la Formation des Schistes Noirs, des Argiles bigarrées et des calcaires en plaquettes des nappes de Tarcău et Audia, plus internes, permet d'observer les similitudes et les différences suivantes:

- *La Formation des Schistes Noirs*, datée suivant les données palynologiques (Antonescu, Ion & Alexandrescu, 1978; Antonescu, Alexandrescu & Micu, 1982) s'étend depuis le Barrémien jusqu'au Vraconien inférieur (la présence de l' Hauterivien à la base est possible mais n'a pas été confirmé). Les grès glauconieux de la partie supérieure de la formation ont dans leurs niveaux terminaux des riches associations de dinoflagellés comprenant la zone à *L. siphoniphorum* et *C. obliquicostatum* de l'Albien supérieur et la zone à *Ovoidinium verrucosum* et *Endoceratium dettmanniae* caractéristique pour le Vraconien;

- *Les Couches de Streiu* débutent suivant les données micropaléontologiques (zones à *G. oblonga* et *P. variabilis* et zone à *R. minutus*) au Barrémien et contient aussi l'Aptien; l'intervalle Aptien supérieur-Albien moyen a été précisé par l'association palynologique correspondant à la zone à *Cribroperidinium* spp. La partie terminale des Couches de Streiu appartient à l'Albien supérieur (zone à *Lithosphaeridium siphoniphorum* et *Carpodinium obliquicostatum*). La présence du Vraconien (inférieur) au sommet des Couches de Streiu est possible mais pas encore prouvée;

- *La base des Couches de Tisaru* et celle de la *Formation des Argiles bigarrées* renferment des associations de dinoflagellés semblables (zone à *Ovoidinium verrucosum* et *Endoceratium dettmanniae*) caractérisant le Vraconien;

- *Le Couches de Tisaru* renferment donc à leur base une association vraconienne (zones à *Ovoidinium verrucosum* et *Endoceratium dettmanniae*); dans la moitie inférieure le Cénomanien inférieur et moyen (zones à *P. infusorioides* et *E. spinosa*) et dans la moitié supérieure le Cénomanien supérieur et le Turonien prouvées au moyen de la microfaune (zones à *Bulbobaculites problematicus* et *Uvigerinammina jankoi*) et de la microflore (?) - association à *X. ceratoides* et *P. pseudhystrichodinium*. La partie terminale des Couches de Tisaru renferme (un seul niveau, donc le fait est problematique?) une association à *Isabelidinium* sp. ex gr. *I. cooksoniae*, équivalente probablement du Santonien et peut-être même du Campanien inférieur?;

- *Les associations palynologiques* de la partie supérieure des Couches de Tisaru (à *X. ceratioides* et celle à *Isabelidinium* cf. *cooksoniae*) présentent quelques traits communs avec celle des calcaires en plaquettes (genres communs, etc.); pourtant la dernière monte plus-haut dans l'échelle stratigraphique;

- *Les Couches de Tisaru* et la majorité de la formation des argiles bigarrées - hormis leur moitié inférieure - sont déposées dans un milieu marin (ainsi que les calcaires en plaquettes) situé à grande distance du continent-océanique.

Couches de Lepşa

Nous avons échantilloné les Couches de Lepşa qui affleurent sur les coupes des vallées de la Lepşa et du Streiu. Jusqu'à présent les échantillons de la vallées Lepşa se sont avérés stériles pour la microflore et n'ont pas offert une microfaune détachable. Chose étonnante, quand on pense que Ion (1975, Ion dans Antonescu et al., 1988; Ion & Szasz, 1994) désigne la coupe de Lepşa comme contenant plusieurs zones de foraminifères - entre autres la seule coupe ou on a decouvert la zone à *Eugubina* du Danien.

Microfaune

Sur la coupe de la vallée de Streiu (tab. 4) les échantillons de la partie inférieure de la succession ont mis en évidence une association de foraminifères aglutinées, typiques pour le Sénonien supérieur. Pärmi les autres espèces, nous signalons la présence de *Dorothia crassa, Spiroplectammina subhaerigensis, Kalamposis grzybowski, Goesseta rugosa, Hormosina ovulum, ?Globotruncana* sp. L'association, par la présence de l'espèce index *G. rugosa* fait partie de la biozone au même nom identifiée par Geroch & Nowak (1983) et qui est équivalente au Campanien inférieur.

Palynologie

Les Couches de Lepşa se sont avérés stériles pour les palynomorphes. Un des échantillons prélèves de la partie supérieure des Couches de Lepşa contient de rares dinoflegellés - *X. ceratioides, O. operculata, Spiniferites* spp., *T. castanea*, dont l'association (selon les ranges des espèces, indique un intervalle stratigraphique plus large Cénomanien-Campanien supérieur.

Donc, d'après nos résultats les Couches de Lepşa appartiendraient au Sénonien supérieur. Fait, qui n'infirme pas les résultats de Ion (les Couches de Lepşa ont un âge compris entre le Maastrichtien supérieur zone mayaroensis - Danien inférieur zone pseudobulloides, Ion & Szasz, 1994), mais aussi ne les confirme pas.

PALÉOGÈNE

Paléocène

Couches de Caşin inférieures

Les Couches de Caşin inférieures montrent un changement de la sédimentation pendant le Paléocène: après l'épisode calcaire des Couches de Lepşa, presque/ ou totalement stérile en ce qui concerne la microflore, la reprise de l'apport détritique fait apparaitre de nouveau les palynomorphes. Les prélèves de la vallée de la Lepşa (L18/4A96 premier niveau de base et ensuite L 19/4A97; L 20/4A98; L 21/4A99; L 22/4A100) s'étalent sur une épaisseur stratigraphique d'environ 30 m à la partie inférieure de cette formation. L'association palynologique est pareille dans toute les niveaux et elle comprend: *Cerodinium (=Deflandrea=Certiopsis) diebeli* (VOZZHENNIKOVA 1963) LENTIN & WILLIAMS, 1987; *Isabelidinium* sp. cf. *I. cooksoniae*, *Isabelidinium* sp., *Trithirodinium evitti* DRUGG, 1967; *Glaphyrocysta* sp. cf. *G. pastielsi* (DEFLANDRE & COOKSON 1955) STOVER & EVITT, 1978 emend. SARJEANT, 1986; *Triblastula utinensis* (O. WETZEL, 1933) SARJEANT, 1985; *Areoligera* sp., *Turbiosphaera filosa* (WILSON, 1967) ARHANGELSKY, 1969; *Cordosphaeridium fibrospinosum* DAVEY & WILLIAMS, 1966; *Hystrichosphaeridium tubiferum, Achomosphaera* sp., *Oligosphaeridium complex*, cf. *Membranosphaera maastrichtica* SAMOILOVICH dans Samoilovich & Mtchedlişvili, 1961 ex Norris & Sarjeant, 1965; *Nudopollis* sp., *Kriegeripollenites hemiperfectus* (PFLUG, 1953) KEDVES & HERGREEN, 1980; *Dinogymnium euclaense* COOKSON & EISENACK, 1970; *Palaeotetradinium minusculum* (ALBERTI, 1961) STOVER & EVITT, 1978. L'association est dominée par les dinoflagellés, surtout les exemplaires d'*Isabelidinium*; la microflore continentale est subordonée (quelques rares exemplaires de pollen de *Normapolles*). Beaucoup d'exemplaires sont en mauvais état de conservation, brisés. Cette association pourrait appartenir au Maastrichtien selon *Triblastula utinensis* et *Dinogymnium euclaense*, espèces qui ne dépassent pas cet étage; de plus *T. utinensis* est cantonée dans les Couches de Hangu inférieures de la Nappe de Tarcău surtout au Maastrichtien supérieur, ou dans l'intervalle Maastrichtien-Danien. Rappelons que pour la partie supérieure du Maastrichtien et le Danien il y a une zone palynologique commune à *Senoniasphaera inornata* et *Thalassiphora pelagica* (ANTONESCU & ALEXANDRESCU, 1982;

Antonescu dans Ion, Antonescu & Micu, 1985; Ion, Antonescu & Alexandrescu, 1994) - le changement palynologique important étant située dans la zone Uncinata du Montien *str. s.* (ou apparaissent les genres et espèces de dinoflagellés exclusivement paléocènes - *Carpatella cornuta*, *Alisocysta circumtabulata*, etc.). Mais, tenant compte du fait que la plupart des genres et espèces de dinoflagellés au range Maastrichtien-Danien manquent (il n'y a que *T. evitti*, *C. diebeli*, *P. minusculum*), et que les espèces exclusivement maastrichtiennes sont nombreuses, il est probable que cette association appartient au Maastrichtien - et qu'elle est remaniée. D'ailleurs les remaniements sont fréquents dans le Paléocène inférieur du Flysch, et un exemple en ce genre sont les couches de Cuejdiu, ou les remaniements du Maastrchtien sont aussi fréquents. Donc, pour le moment, nous acceptons l'age Danien inférieur attribué par Ion pour les couches de Caşin inférieures (Jana Săndulescu dans Dumitrescu et al., 1971; Jana Ion Sădulescu, 1973; Jana Ion dans Antonescu et al., 1988), qui identifie dans les Couches de Caşin inférieures la zone à *Subbotina trinidadensis* (d'âge Danien inférieur, contenant *S. trinidadensis*, *Acarinina inconstans*, *Planirotalites convexus*, *Subbotina triloculinoides*, *S. pseudobulloides*, *Globigerina trivialis*, *Globoconusa daubjergensis*, *G. kozlowski*, *Globigerina edita*, *Globigerinelloides woodringina*, *Gumbelitra* et d'autres heterohelycides biseriées) et la zone à *Morozovella uncinata* (d'âge Montien, à *M. uncinata*, *M. perclara*, *M. schadagica*, *Subbotina trinidadensis*, *Acarinina inconstans*, *Planorotalites compressus*, *Subbotina pseudobulloides*, *S. triloculinoides*, *S. varianta*, *Globigerina trivialis*, *G. woodringina*, *Gumbelitra*, et les derniers exemplaires de *Globoconusa daubjergensis* de type *kozlowski*).

Couches de Caşin moyennes

Seulement la coupe de la vallée de la Lepşa a offert des données intéressantes pour ce qui concerne la microfaune; les données palynologiques, pour le momment, manquent de cette séquence.

Microfaune

L'association microfaunistiques des Couches de Caşin moyennes de la coupe de la vallée de la Lepşa comprend seulement des foraminifères agglutinees (Tab. 5): *Gyroidinoides* div. esp., *Lenticulina* div. esp., *Nodosaria* sp., *Hormosina velascoensis*, *Glomospira diffundens*, *Rzehakina epigona*, *R. fissistomata*, *Thalmannammina gerochi*, *Glomospira glomerata*, *Spiroplectamina* cf. *spectabilis*, *Karreriella horrida*, *Subreophax pseudoscalaris*, *Subreophax scalaris*, *Nothia excelsa*, *Trochamminoides irregularis*, *Glomospirella grzybowski* et *Reophax globosus* appartient à la zone à *Spiroplectammina spectabilis* et *Glomospira diffundens* Geroch & Nowak (1863) qui indique le Paléocène supérieur.

Rappelons que Ion (Jana Săndulescu, dans Dumitrescu et al., 1971; Ion dans Antonescu et al., 1988) a cité dans les Couches de Caşin moyennes deux associations de foraminifères, celle à *Acarinina primitiva*, *Morozovella uncinata*, *Subbotina triloculinoides*, *S. pseudobuloides*, *S. varianta*, *Globigerina trivialis*, *Subbotina edita* qui serrait entre les zones à *Morozovella angulata* et la zone à *M. pusilla pusilla* et l'association à *Morozovella acqua*, *M. conicontruncana*, *M. angulata*, *Globigerina velascoensis*, *G. eocaenica*, *G. linaperta*, *Subbotina varianta*, *S. triloculinoides*, *Acarinina nitida*, *A. mckannai*, *Planorotalites chapmanni* qui correspondrait à la zone à *Planorotalites pseudomenardi*.

Couches de Caşin supérieures

Micropaléontologie

Les données micropaléontologiques concernant les couches de Caşin supérieures ont été obtenues sur la coupe de la vallée de Caşin - vallée de Buciaş (Tab. 5, 9) et sur la coupe de la vallée de Lepşa (Tab. 6).

Une microfaune riche en foraminifères aglutinées a été identifiée dans les couches de Caşin supérieures, à laquelle s'ajoute quelquefois du benthos calcaire. A coté des taxons *Glomospira diffundens*, *Spiroplectammina spectabilis*, *Rzehakina inclusa*, *R. epigona*, dont les moments d'apparition, d'extinction et/ou de "acme-zone" ont leur signification stratigraphique au temps du Paléocène, il y a de nombreuses espèces caractéristiques pour la partie terminale du Crétacé supérieur, du Paléocène inférieur et du Paléocène moyen, fort probable remaniées. Nous estimons ainsi que toute l'association des Couches de Caşin supérieures (Tab. 5) ainsi que celles des Couches de Caşin moyennes (Tab. 6) appartient à la biozone à *Spiroplectammina spectabilis* Geroch & Nowak (1983) qui indique le Paléocène supérieur. Cette association nous l'avons rencontrée fréquement dans les dépots paléocènes du flysch externe (couches de Putna) et elle est en général cosmopolite. Plusieurs taxons cités par nous ont été mentionés au Trinidad (Kaminski et al., 1988) et dans le secteur central de la Mer du Nord (Charnock & Jones, 1990) ou ils indiquent le Paléocène en général.

Palynologie

Les données palynologiques concernant les Couches de Caşin supérieurs proviennent de la coupe de la vallée de Caşin - vallée de Buciaş, du tiers de la partie supérieure de ces dépôts (Tab. 7). Parmi les dinoflagellés, à coté de *Cerodinium speciosum* (ALBERTI, 1959) LENTIN & WILLIAMS, 1985, il y a *Spiniferites cornutus* (GERLACH, 1961) SARJEANT, 1970, *Deflandrea denticulata* ALBERTI, 1959, *Glaphyrocysta* sp., et des pollens, *Subtriporopollenites annulatus* (PFLUG, 1953) PFLUG, 1953, *Intratriporopollenites* sp., etc. L'association indique la zone à *Deflandrea speciosa (=Cerodinium speciosum)* caractéristique pour le Thanétien de plusieurs régions de l'Europe de l'Ouest (Powell, 1993) trouvée aussi en Roumanie (Antonescu & Alexandrescu, 1982; Ion et al., 1985; Antonescu et al., 1988) dans les Couches de Putna.

Il est utile de rappeler que Antonescu (dans Antonescu & Alexandrescu, 1982; Ion et al., 1985; Antonescu et al., 1988; Ion et al., 1994) a reconnu pour le Paléocène des Carpathes Orientales plusieurs zones palynologiques. Ce sont la zone à *Senoniasphaera inornata* et *Thalassiphora pelagica* (Maastrichtien supérieur-Danien), la zone à *Carpatella cornuta* et *Alisocysta circumtabulata* (Montien *str. s.*), la zone à *Isabelidium bakeri* et *Stephanoporopollenites hexaradiatus* (la partie moyenne du Paléocène, zone non-publiée Antonescu dans Antonescu et al., 1988) et la zone à *Deflandrea speciose* et *D. oebisfeldensis* (Thanétien).

Dans la partie terminale des Couches de Caşin supérieures de la coupe des vallées Caşin-Buciaş, à côté des genres et espèces de la zone à *Cerodinium speciosum*, apparaissent les espèces du genre *Apectodinium* (Tab. 7)- *A. homomorphum* (*Deflandre & Cookson*, 1955) *Lentin & Williams*, 1977, *A.* cf. *paniculatum* (COSTA & DOWNIE, 1976) LENTIN & WILLIAMS, 1977 emend. HARLAND, 1979 et *A. parvum* (ALBERTI, 1961) LENTIN & WILLIAMS, 1977, et d'autres dinoflagellés comme *Kleithriasphaeridium coinodes* (ALBERTI, 1961) LENTIN & WILLIAMS, 1977; *Cordosphaeridium fibrospinosum* DAVEY & WILLIAMS, 1966; *Spiniferites ramosus*.

Cette association est fort intéressante puisqu'elle met en évidence, pour la première fois en Roumanie, la zone à *Apectodinium homomorphum* (ou *hyperacanthum*, Ahy biozone de Powell, 1993) équivalente à la zone à *Wetzeliella hyperacantha* DE CARO (1973) et de la zone 5 de Heilmann-Clausen (1985); équivalente aussi à la biozone de nannoplakton NP9 (en partie), à la biozone de foraminifères P4 (de Blow, 1969 et de Bergreen, 1972). Cette zone représente un eveniment important - l'apparition des espèces du genre *Apectodinium* (Base Apectodinium Datum) et a été considerée comme équivalente par Costa & Downie (1976) avec la base de la zone de nannoplakton calcaire NP9 de Martini (1971). C'est une des zones avec le plus grand potentiel de corrélation. Ainsi le contenu palynologique de la partie terminale des Couches de Caşin peut-etre considéré comme équivalent à celui de la base de l'Illerdien de la coupe de Campo, la zone 5 du Danemark - Viborg 1 Borehole, Jutland, Holmethur Formation (en partie) et la Osterrende Clay et la Glauconitic Silt - les Woolwich Beds d'Angleterre, partie supérieure du Thanétien du Bassin de Paris, etc.

Couches de Piatra Uscată

Nos données palynologiques et micropaléontologiques se rapportent à la coupe des vallées Caşin-Buciaş.

Micropaléontologie

Selon la microfaune que nous avons trouvée dans les Couches de Piatra Uscată (Tab. 6), ces dépots peuvent étre attribués au Paléocène supérieur, la zone à *Spiroplectammina spectabilis*. Il faut noter qu'à la partie supérieure de ces couches, il y a une légère modification dans les rapports quantitatifs des genres et espèces; cet évenement pourrait-il ètre lié au passage Paléocène-Eocène?.

Palynologie

Immédiatement au-dessus de la limite Couches de Caşin supérieures/Couches de Piatra Uscată, le palynofacies devient fort riche en dinoflagellés. A côté des espèces deja présentes dans les Couches de Caşin supérieures, d'autres espèces apparaissent (tab. 7): *Apectodinium augustum* (HARLAND, 1979) LENTIN & WILLIAMS, 1981; *A. quinquelatum* (WILLIAMS & DOWNIE, 1966) COSTA & DOEMIE, 1979; *A. hyperacantha* (COOKSON & EISENACK) LENTIN & WILLIAMS, 1977; *Deflandrea cornumamillata* JAN DU CHÊNE & CHATEAUNEUFF, 1975; *Hystrichokolpoma* cf. *cinctum* KLUMPP, 1953; *Adnatosphaeridium multispinosum* WILLIAMS & DOWNIE, 1966; *Ginginodinium* sp., *Dracodinium* sp., *Caligodinium aceras* (MANUM & COOKSON, 1964) LENTIN & WILLIAMS, 1975; *Palaeocystodinium* sp.

Cette association à *Apectodinium augustum* appartient à la biozone à *Apectodinium augustum* DE POWEL (1993); elle (Aau) est équivalente à la zone 6 de Heilmann-Clausen (1985) mise en évidence dans le sondage de Viborg (Viborg 1 Borehole, Jutland et Fur Formation ainsi que Negative Series, en partie, et la Haslund Member de Danemark). Zone équivalente en partie à la zone de nannoplankton NP 9 de Martini (1971) et au zones de foraminifères P 5 (en partie et souszones P6A et P6B (en partie) de Blow (1969) et Bergreen (1972) ainsi qu'à la zone de dinoflagellés, sous zone D5a de Costa et Manum (1988). Elle a été trouvé aussi dans les Woolwich Beds et dans la Mer du Nord. Son age est Thanétien terminal et base de l'Yprésien et on remarque que la partie terminale des Couches de Caşin et des Couches de Piatra Uscată étudiées sur les coupes des vallées Caşin-Buciaş representent les dépôts qui matérialisent le mieux jusqu'à présent dans les Carpathes Orientales - en ce qui concerne les dinoflagellés - l'intervalle de temps correspondant à la fin du Paléocène et au début de l'Eocèn3, et aussi le fait que le palynofacies, les espèces, sont pareilles à celles des dépots de l'Europe de l'Ouest. Les conditions paléoécologiques gouvernant l' environement des Couches de Caşin supérieures et des Couches de Piatra Uscată et celles des formations de l'Europe de l'Ouest (formations du Danemark, de la Mer du Nord, de l'Angleterre, de Campo, etc.) devrait-etre semblables, bien que les lithofacies sont très différentes. Il est sur que dans certains moments, par exemple à l'Albien supérieur-Vraconien (Formations de Schistes Noirs/Formations des Argiles Bigarrées) ou au Paléocène supérieur-Eocène inférieur (Couches de Caşin supérieures et Couches de Piatra Uscată) les mers de l'Europe de l'Ouest et du Flysch des Carpathes devaient communiques.

A la partie supérieure des Couches de Piatra Uscată, sur la coupe des vallées Caşin-Buciaş, le palynofacies change et la plupart des genres et espèces disparaissent. Cet évenement serrait - il lié avec le passage du Paléocène à l'Eocène?. C'est une supposition seulement et la limite Paléocène/Eocène pourrait fort bien se placer plus bas dans les Couches de Piatra Uscată, puisque la biozone à *A. augustum* est à cheval sur celle-là.

ÉOCÈNE

Couches de Buciaş

Les Couches de Buciaş ont été échantillonées sur la coupe ouverte par le ruisseau Buciaş (le stratotype) ainsi que sur des coupes complémentaires, la vallée de la Putna et le versant du mont Coaşa concernant leur partie supérieure seulement.

De la partie inférieure des Couches de Buciaş (vallée de Buciaş) a été déterminée une association à *Turborotalia cerroazulensis frontosa, Acarinina bullbrooki, A. pentacamerata, A. spinoloinflata, A nitida, Globigerina velascoensis* (Tab. 9). Cette association correspond à la zone à *Turborotalia cerroazulensis/ Turborotalia pseudomayeri* (BLOW, 1979). Tenant compte de la présence dans l'association de *A. nitida* qui s'éteint à la base du Lutétien et de *G. velascoensis* qui monte du Paléocène jusqu'à l'Yprésien, on peut placer la limite Yprésien-Lutétien à ce niveau (fig. 1 et 2, tableau de corrélation). Cette conclusion est confirmée par le nannoplankton: ainsi (Tab. 8) ont été mises en évidence la biozone NP 13 à *Discoaster lodoensis*, de l'Yprésien et la biozone NP 14 à *Discoaster sublodoensis*, du Lutétien.

De la partie médiane des Couches de Buciaş, a été identifiée une association à *Nuttalides truempyi, Anomalinoides granosus, Oridorsalis umbonatus* et *Nodosarella tuberosa*, du Lutétien. Les foraminifères agglutinées dominnent l'association, mais n'apportent pas des précisions stratigraphiques suplémentaires. L'association de nannoplankton identifiée dans les mêmes niveaux correspond à la zone NP 15-Nannotetrina fulgens. Sur la coupe du mont Coaşa la biozone est précisée par la première occurènce de l'espèce *Discoaster bifax, D. saipanensis, Chiastomolithus gigas* et par la dernière apparition de *Ch. expansus* et *Ch. solitus* (Tab. 12).

La partie supérieure des Couches de Buciaş a été analysée sur trois coupes: la vallée de Buciaş, la vallée de la Putna et le versant du mont Coaşa. Les premières deux coupes ont délivré une association à *Cyclammina amplectens, Haplophragmoides walteri, Kalamopsis grzybowskii, Reophax duplex* (Tab. 9). *Cyclammina amplectens* qui est l'espèce index pour le Lutétien dans la zonation de Geroch (1983) a été signalée en Roumanie seulement dans la partie supérieure de cet étage (Bratu dans Antonescu et al., 1988).

Sur le mont Coaşa, l'âge Lutétien supérieur de la partie terminale des Couches de Buciaş est precisé par des foraminifères planctoniques - la biozone à *Turborotalia cerroazulensis possagnoensis*. Font partie aussi de l'aaociation *T. cerroazulensis frontosa, Acarinina spinulosa, A. bullbrooki* et d'autres espèces de benthos agglutiné et calcaire (Tab. 10).

Ces données peuvent être corrélés avec celles fournies par les dinoflagellés, sur la vallée de la Putna étant determinée l'association à *Wilsonidium echinosuturatum, Cordosphaeridium gracilis, Homotriblium palidum, Achomosphaera ramulifera* (Tab. 15) biozone W9 du Lutétien supérieur (fig. 1 et 2, tableau de correlation).

Couches de Bisericani

Les dépots appartenant à cette formation sont équivalents au Lutétien terminal, au Bartonien et au Priabonien.

Sur la coupe de la vallée de la Putna et sur la coupe de la vallée de Buciaş le niveau basal à argiles rouges - niveau de Strujinoasa - a été atribué au Lutétien supérieur tenant compte de la présence de l'association à *Cyclammina amplectens*. Sur la coupe du mont Coaşa, l'age Lutétien supérieur du niveau de Strujinoasa est confirmée par des foraminifères planctoniques appartenant à la zone à *Turborotalia cerroazulensis possagnoensis*. Les données du nannoplankton ont permis l'identification au même niveau, sur la coupe de la Putna, de la biozone à *Discoaster taninodifer* (Tab. 15). Les dinoflagellés trouvés sur la coupe de la vallée de Buciaş appartient à la zone à *Wilsonidium echinosuturatum* (Lutétien, Tab. 14 a, b).

Sur les coupes des vallées de la Putna et de Buciaş la limite Lutétien/Bartonien a été tracee à l'aide de la première occurence de l'espèce *Ammodiscus latus* en association avec *Cyclammina amplectens* (à une fréquence reduite, *Haplophragmoides retroseptus, Glomospira charoides, Haplophragmoides walteri, Karreriella coniformis, K. horrida* (Tab. 9, 11). Il faut préciser que l'espèce *Ammodiscus latus* a une distribution discontinue sur ces coupes, étant identifiée seulement dans quelques niveaux. Sur la coupe de la vallée de la Putna, cette espèce a une fréquence plus grande dans le Priabonien, probablement due à la résédimentation. La limite Lutétien/Bartonien est précisée aussi par l'association de dinoflagellés à *Wilsonidium echinosuturatum* correspondant à la biozone W9 qui comprend dans sa partie supérieure la partie inférieure du Bartonien. Sur la coupe de Buciaş, a été determinée *Wilsonidium echinosuturatum* en association avec *Cribroperidinium tenuitabulatum, Deflandrea phosphoritica, Thalassiphora pelagica* (Tab. 14 a).

L'association de foraminifères de la partie supérieure des Couches de Bisericani est constituée sur la coupe de Buciaş, par des espèces planctoniques dont la distribution stratigraphique ne permetent pas la separation des zones standard du Priabonien. On y a trouvé *Globigerina linaperta, G. eocaenica, G. hagni, Globigerinita pera* (Tab. 9). *Turborotalia pomeroli*, qui dans la zonation de Tourmakine et Bolli (1970) marque une zone équivalente aux zones P13-P14. est probablement ici dans sa partie supérieure de sa distribution stratigraphique. De l'association font partie aussi des espèces de foraminifères agglutinées, dont la fréquence devient plus faible vers la limite supérieure de la formation.

Au même niveau, sur la coupe de la Putna de la partie supérieure des Couches de Bisericani a été déterminée *Cyclammina rotundidorsata* en association avec *Recurvoides contortus, Glomospira serpens, Trochammina heteromorpha, Kalamopsis grzybowskii* (Tab. 11) qui temoignent l'age Priabonien de ces formations.

L'association de nannoplankton identifiée sur la coupe du Buciaş appartient au zones P18-P20-Istmolithus recurvus/Sphenolithus pseudoradians. Les espèces marker sont associaées à *Zygrhabdolithus bijurgatus, Ericsonia subdisticha, Reticulofenestra hampdenensis, R. dictyoda* (Tab. 8).

Sur la meme coupe, le Priabonien est précisé aussi par les dinoflagellés - la zone à *Kisselovia clathrata angulosa-Phtanoperidinium echinatum* (Tab. 14 a).

LIMITE ÉOCÈNE/OLIGOCÉNE

Marnes à Globigérines

La plus grande partie des dépots appartenant à la formation de Marnes à Globigérines (=Grès de Lucăceşti) affleurant sur la coupe de Buciaş contient une association de foraminifères planctoniques qui ont eu une première occurence plus bas, dans les Couches de Bisericani et qui caractérise les zones P16-17-Globigerinateka semiinvoluta et Turborotalia cerroazullensis sl., du Priabonien. On constate aussi que la fréquence du plancton est très grande et le benthos disparait. De la partie supérieure des Marnes à Globigérines ont été détérminé *Globigerina tapuriensis* en association à *G. sellii, G. officinalis, G. cryptomphala* qui marque la zone P18 de la base de l'Oligocène (Tab. 9). On arrive ainsi à confirmer que la limite Eocène/Oligocène est situé dans le tiers supérieur des Marnes à Globigérines (Micu & Gheţa, 1987; Săndulescu et al., 1987).

Les données palynologiques confirment ces données par la première occurence de l'espèce *Phtanoperidinium amoenum* et *Boehlensipollis hohli* (Tab. 14 a et b). L'association de nannoplankton de ce niveau fait partie de la zone à *Istmolithus recurvus/Sphenolithus pseudoradians* du Priabonien. Il y a donc un décalage entre le nannoplankton et les autres groupes d'organismes (Tab. 8 et tableau de corrélation).

OLIGOCÈNE

Schistes Ardésiformes (+ Grès de Fierăstrău)

L'âge de ces dépôts a été précisé, sur les coupes analysées, seulement par les dinoflagellés et le nannoplankton. Dans la vallée du Buciaş, les analyses palynologiques ont mis en évidence la zone à *Wetzeliella gochti*, d'âge Ologocène inférieur (Rupélien). L'espèce index est associée à *Tithodiscus* sp., *Thalassiphora delicata* et *Microtythodiscus* sp. (Tab. 4 a). La même position stratigraphique est précisée par l'association de nannoplankton appartenent à la zone à *Helicopontosphaera reticulata/Ericsonia subdisticha*. L'association est caractérisée par une fréquence plus grande de l'espèce *E. subdisticha* et l'absence des discoasterides.

Ménilites inférieures et Marnes Brunes

Sur la coupe de la vallée du Buciaş la microfaune identifiée à ce niveau est pauvre et mal conservée. Quelques taxons déterminés parmettent l'attribution de ces dépôts au Rupélien, zone P17: *Globigerina tapuriensis, G. ampliapertura, G. sellii, G. officinalis* (Tab. 9). On n'a pas obtenu des résultats palynologiques et du nannoplankton.

Dysodiles inférieures et Grès de Kliwa

Nous avons analysées seulement quelques échantillons isolés des Dysodiles inférieures, des intercalations pélitiques du Grès de Kliwa et de la base des Dysodiles supérieures. Les seules résultats ont été offerts par les analyses palynologiques. On a identifié une association à *Wetzeliella gochtii, W. symmetrica, Thalassiphora pelagica, Titthodiscus* sp. qui prouvent l'âge Rupélien de ces dépôts. *W. gochtii* a été identifiée aussi à la base des Dysodiles supérieures, mais à ce niveau il est possible qu'elle soit remaniée.

L'analyse de la distribution des foraminifères, des palynomorphes et du nannoplankton a permis donc d'établir une biostratigraphie assez détailée de formations appartenant à l'Eocène et à l'Oligocène. En ce qui concerne les foraminifères, la plus grande fréquence et la meilleure continuité appartient aux foraminifères agglutinèes. Les foraminifères planctoniques sont présentes seulement à quelques niveaux à la base et à la partie supérieure des Couches de Buciaş, au niveau des argiles barriolées de Strujinoasa (seulement sur la coupe de mont Coaşa) et à la partie supérieure des Couches de Bisericani et dans les Marnes à Globigérines (sur la coupe du Buciaş). Les foraminifères benthiques calcaires sont présentes seulement à la partie inférieure des Couches de Buciaş, sur la vallée de Buciaş, et à la partie supérieure de la même formation sur le mont Coaşa. L'étude des foraminifères agglutinées et benthiques calcaires ont permis de compléter l'échelle stratigraphique, là ou le plancton manque.

En général, il y a une bonne corrélation entre les biozones des trois groupes études (tableau de corrélation):

- D'après les données fournies par le plankton (dans la zone à *Turborotalia cerroazulensis frontosa*) et le nannoplakton (entre les zones à *Discoaster lodoesnsis* et *Discoaster sublodoensis*), la limite Yprésien/Lutétien se placerait près de la base des Couches de Buciaş (coupe de la vallée de Buciaş).
- La limite Lutétien/Bartonien est située dans la partie inférieure des couches de Bisericani. Elle a été étayée sur la base de l'apparition de l'espèce *Ammodiscus latus* (foraminifère agglutinée), sur les coupes de Buciaş et de la Putna. Cette zone peut-être corréllée avec en partie la zone à *Wilsonidinium echinosuturatum* trouvée sur les mêmes coupes.
- La limite Bartonien/Priabonien a été établie à l'aide des foraminifères agglutinées, des dinoflagellés et du polen; respectivement l'apparition de *Cyclammina rotundidorsata* (vallée de la Putna), de *Kisselovia clathrata angulosa* et de *Rhombodinium perforatum* (vallée de Putna).
- La limite Priabonien/Rupélien, suivant l'apparition de *Globigerina tapuriensis* et des palynomorphes *Phtanoperidinium amoenum* et *Boehlensipollis hohli* a été précisée dans le tiers supérieur de Marnes à Globigérines. Suivant le nannoplankton déterminé, la base du Rupélien serait plus-haut, à la base des Schistes ardésiformes qui contiennent la zone à *Helicopontosphaera reticulata/Ericsonia subdisticha.*
- Certaines biozones n'ont pas su être précisées, soit à cause du manque de continuité d'un groupe, soit à cause des certaines intervalles stériles, ou qui ne pouvait être échantillonnés.

CORRÉLATIONS RÉGIONALES ET PALÉOGEOGRAPHIE

Les précisions stratigraphiques qui ont été obtenues par nos recherches, permettent de mieux réaliser la corrélation des successions lithostratigraphiques des Plis Marginaux des Monts de Vrancea avec les autres lithofaciès développés dans la Zone du Flysch des Carpathes Orientales.

On constate que les précisions apportées sur l'âge de la limite supérieure des Couches de Streiu, qui est placée au niveau de la limite Albien/Vraconien souligne l'isochronisme du lithofaciès anoxique du Crétacé inférieur des Nappes d'Audia, de Tarcău et des Plis Marginaux: Serie des Schistes Noirs (Antonescu et al., 1978) = Couches de Streiu (présente note) = Couches de Sărata (Antonescu et al., 1982). Ces lithofaciès, connus sous la dénomination générale de "Lithofacies silésien", dans l'ensemble des Carpathes externes, montre certaines variations, aussi bien dans le sens transversal que longitudinal (par rapport à la direction de l'aire de dépôt), qui sont les plus sensibles au niveau de l'Albien. On constate d'abord que la séquence des "grès siliceux glauconieux" tipiquement développée dans la Nappe d'Audia et seulement en partie dans la Nappe de Tarcău, manque partout dans la Nappe des Plis Marginaux, plus externe que les deux autres. On constate que les Couches de Streiu, ayant toujours une source externe (avant-pays) pour les arénites (marquée par la présence des débris de Schistes Verts) sont dépurvus de "grès siliceux glauconitiques" (orthoquartzites glauconieux). Cette absence, qu'on constate aussi dans les Couches de Sărata (ou au niveau de l'Albien se développent des calcaires à chailles) devrait être acceptée comme dus à la distribution préférencielle du matériel orthoquartzitique toujours de source externe) suivant des "trajets" des courants de rédistribution longitudinale de ce matériel.

Les lithofaciès qui surmontent les formations anoxiques "silesiennes" montrent une forte condensation: le Vraconien-Sénonien inférieur est matérialisé dans les couches de Tisaru (monts de Vrancea) les argiles bariolés (Demi-fenêtre de Bistriţa), Couches de Lupchianu et équivalents (argiles bariolés) et Calcaire de Cârnu (Nappe de Tarău) ou argiles bariolés dans la Nappe d'Audia (Ion et al., 1994; Săndulescu et al., 1993; Micu, 1987; Micu et al., 1986; Săndulescu et al., 1987). Partout on constate que les apports arénitiques sont nuls ou fortement reduits et que la profondeur de la zone de sédimentation a auggumenté (descendant au-dessous la CCD?!). Cette situation devrait s'expliques par l'immersion totale de la source externe (avant-pays) accompagnée par l' approfondissement du bassin correspondant au lithofaciès "silésien". Il faut remarquer le large développement des silicolites dans les Couches de Tisaru qui, de ce point de vue, serait le lithofaciès le plus profond de l'ensemble des lithofaciès du Vraconien-Sénonien inférieur de la Zone du Flysch, bien qu'il est dans une position très externe, situation acceptable seulement dans les conditions de l'immersion des sources de matériel détritique mentionné ci-dessous (elle se trouvent actuellement sous-charrié au-dessous des nappes de la Zone du Flysch).

Les Couches de Lepşa sont, relativement, faciles à être mises en corrélation avec les Couches de Hangu de la Nappe de Tarcău (parties externes). Toutes les deux unités lithostratigraphiques sont cantonnées dans le Sénonien, sans pouvoir préciser la position stratigraphique de leur limite inférieure. Neanmoins on doit accepter que les deux formations ne couvrent entièrement le Sénonien puisque les formations sous-adjentes montent dans le Sénonien inférieur (Couches de Tisaru supérieures, argiles bariolés, Calcaires de Cârnu, Couches de Lupchianu). On constate que la fréquence des arénites dans les lithofaciès externes (Couches de Lepşa) est nettement inférieure à celle des lithofaciès plus internes (Couches de Hangu) bien que la source de ces arénites est externe (avant-pays). Dans ce contexte il faut admettre que les arénites des Couches de Hangu arrivaient par des cannyons sous-marins qui devraient traverser l'aire de sédimentation pélagique des faciès externes (Couches de Lepşa).

Les Couches de Caşin représentent, dans la Nappe des Plis Marginaux des monts de Vrancea, la majeure partie du Paléocène. De ce point de vue elles peuvent êtres très bien équivalentes avec les Couches de Putna de la partie externe de la Nappe de Tarcău, qui couvrent le même intervale stratigraphique (Antonescu et al., 1989; Ion et al., 1994). Sur plusieures coupes la limite Sénonien/Paléocène serait commune avec celle de Couches de Hangu ou Couches de Lepşa ou Couches de Caşin (Săndulescu et al., 1987, et note présente). Deux situations "non conformistes" sont à retenir: les derniers niveaux de Couches de Lepşa dans la vallée du même nom (Ion, 1975) et les Couches de Hangu dans la vallée de Oanţu (Ion et al., 1994). Ces exceptions "aberrantes" sont encore sujet d' étude et il est encore difficile d'avancer une conclusion.

Du point de vue paléogeographique le Paléocène montre les mêmes traits que les formations plus anciennes: faciès typiquement turbiditiques (flysch) sont plus internes que ceux pélitosiltiques externes (Couches de Caşin). C'est le même modèle qui doit permettre aux arénites des flysch (Couches de Putna) de traverser les aires ou la sédimentation est de type pélagique mais pas turbiditique (Couches de Caşin). Pourtant à certains niveaux (séquences supérieures des Couches de Caşin) des intercalations de microbrèches et conglomérats à débris de Schstes Verts sont les temoins de périodes accentuées de l'activité de la source externe.

Dans la Nappe des Plis Marginaux et dans la partie externe de la Nappe de Tarcău se développent, à cheval par rapport à la limite Paléocène/Eocène une séquence de flysch de type "couches à hiéroglyphes", les Couches de Piatra Uscată dans la première et les Couches de Straja dans la deuxième. Les deux séquences turbiditiques sont tributaires à la source externe d'arénites. Pour le domaine des Plis Marginaux cette séquence typiquement flysch (les Couches de Piatra Uscată) suit et précède des développements faiblement turbiditiques: les Couches de Caşin et, respectivement, les Couches de Buciaşu. La présence des niveaux paléocènes à la base des Couches de Piatra Uscată, ainsi que des Couches de Straja (Bratu & Alexandrescu, 1970; Dumitrescu et al., 1971) souligne le vraisemblable synchronisme de la limite inférieure de ces deux unités lithostratigraphiques. La présence des niveaux yprésiens dans la moitié supérieure des Couches de Piatra Uscată et de Straja (Dumitrescu et al., 1971) soutient l'affirmation faite ci-dessus concernant la position de la limite Paléocène/ Eocène. En ce qui concerne la corrélation de la limite supérieure des Couches de Piatra Uscată et de Couches de Straja il semble qu'elle se place en général au même niveau stratigraphique, sans pouvoir apporter plus de details.

Les Couches de Buciaş, situées entre les Couches de Piatra Uscată (au m ur) et les Couches de Bisericani (*s.l.* - comprenant les équivalents des Couches de Strujinoasa à leurs extrèmes base au toit) peuvent être facilement équivalentes dans les monts de Vrancea avec les Couches de Greşu (Dumitrescu, 1963) situées dans les parties internes des Plis Marginaux, ainsi qu'avec les Couches de Leşunţ et les Couches de Piepturi de la partie externe de la Nappe de Tarcău. En ce qui concerne la corrélation avec les Plis Marginaux de la Demi-fenetre de Bistriţa et la partie externe de la Nappe de Tarcău au nord de celle-ci, on constate que les Couches de Buciaşu correspondent à deux unites lithostratigraphiques: les Couches de Jghiabu Mare et le Calcaire de Doamna, respectivement aux Couches de Suceviţa et au Calcaire de Paszieczna. Les aspects paléogeographiques de la séquence yprésien-lutétienne sont intéresantes à etre analysés. On constate:

- l'absence, ou le très faible développment, des turbidites dans le lithofaciès de Buciaşu, pélagique, partiellement silicifié, ayant la position paléogeographique la plus externe (au moins sur la transversale des monts de Vrancea);

- le développment inégal, sur différentes transversales, du nord vers le sud, des calcaires à "chailles" (Paszieczna/Doamna) en tant qu'unité lithostratigraphique, dans le sens qu'ils sont développés sur touts la largeur (visible) de la zone correspondant à la partie externe de la Nappe de Tarcău et à l'ensemble des Plis Marginaux (au Nord de Trotuş); étant restreintes sur une petite aire (affleurements de la vallée de Doftana - Săndulescu et al., 1987) à la partie interne des Plis Marginaux (au sud de Trotuş; cette situation est accompagnée du fait que dans la partie externe de la Nappe de Tarcău (lithofaciès de Leşunţ) ainsi que dans la partie interne des Plis Marginaux (lithofaciès de Greşu - exception la vallée de Doftana), des calcaires à "chailles" (de type Pasieczna/Doamna) constituent des intercalations discontinues dans des séquences flysch;

- comme pour les séquences albiennes et sénoniennes, les faciès turbiditiques typiques sont situés à l'ouest (plus internes) (Greşu, Leşunţ, Suceviţa) que les successions pélagiques (Buciaşu, partiellement Jghiabu Mare);

- le développement au Lutétien supérieur de la séquence des calcaires à "chailles" sur une large surface du bassin tributaire au matériel détritique à la source externe (avant-pays) montre une immersion importante de cette source au moins dans les parties corréspondant aux secteurs centrales et septentrionales. En effet des calcaires à nummulites lutétiens sont très abbondantes dans les éléments des conglomérats oligocènes et miocènes, dont la source est toujours l'avant-pays, ce qui prouve un large recouvrement par les calcaires lutétiens de l'avant-pays.

La corrélation de la limite inférieure des Couches de Bisericani avec celle des Couches de Plopu (de la Nappe de Tarcău) est bien documenté (Săndulescu et al., 1987; Antonescu et al., 1989). Du point de vue paléogéographique on doit remarquer, encore une fois que vers l'exterieur du bassin se développent un lithofaciès pélagique, à faibles apports turbiditiques (Couches de Bisericani) tandis que les faciès flysch sont situés plus loins vers l'exterieur (Couches de Plopu et Couches de Podu Secu).

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Sousdivision stratigraphiques	Unités lith.	Foraminifères	Dinoflagellés	N-plct		
EOCENE		?	?	NP 9		
PALÉOCÈNE supérieur	Couches de Casin	*Spiroplectammina spectabilis*; *Glomospira difundens*	*Apectodinium augustum*; *Apectodinium homoomorphum*; *Cerodinium speciosum*	NP 9		
PALÉOCÈNE inférieur	Couches de Casin ---?--- Couches de Lepşa	?	?	?	Casin inf.	*Glob. compressa*, *G. triloculinoides*, *G. psedobulloides*
MAASTRICHTIEN	Couches de Lepşa	?	?	?	Couches de Lepşa	*Petites glob, Heterohelix*; *Gl-ana stuarti*, *Gl-ana citae*
CAMPANIEN	Couches de Lepşa	*Goesella rugosa*	?	?	Couches de Lepşa	*Goesella carpathica, H. suborb, Spira. subhaering.*
SANTONIEN	Couches de Lepşa	*Uvigerinammina jankoi*	?	?	Couches de Lepşa	
CONIACIEN	Couches de Tisaru	*Uvigerinammina jankoi*	?	?	Tisaru sup.	
TURONIEN	Couches de Tisaru	*M. schn.*; *P. helv.*; *Uvigerinammina jankoi*	*Xenascus ceratioides*	?	Tisaru inf.	*G. ex. gr. lapparanti, N. velascoense, H. ov. gigant*
CÉNOMANIEN	Couches de Tisaru	*W. arch.*; *R. cush.*; *Bulbobaculites problematicus*	*Pervosphaeridium pseudohystricodinium*	?	Couches de Streiu	*Haplophrag. gigas miner*; *Gr. irregularis*; *R. imperfectus*
VRACONIEN	Couches de Tisaru	?	*Ovoidinium scabrosum*; *Paleohystrichophora infusorioides*	NC 9	Couches de Streiu	*Psamoosphaera*
ALBIEN S.S.	Couches de Tisaru; Couches de Streiu	?	*Apteodinium grande*; *Cribroperidinium orthoceras*	NC 9 ---?---	Couches de Streiu	*Hedbergella trechoidea*, *H. planispira*, *Ticinella*, *H-ides concavus*
APTIEN	Couches de Streiu	*Pseudobolivina variabilis*; *Reophax minutus*	*Cribroperidinium orthoceras*		Couches de Streiu	*Verneuilinoides subfiliformis*; *Hipocroppina depressa*
BARRÉMIEN	Couches de Streiu	*Reophax minutus*; *Gaudryna oblonga*	*Cribroperidinium orthoceras*		Couches de Streiu	
HAUTERIVIEN					Couches de Streiu	

Fig. 1 – Tableau de corrélation. Biostratigraphie des formations du Crétacé et Paléogène inferieure.

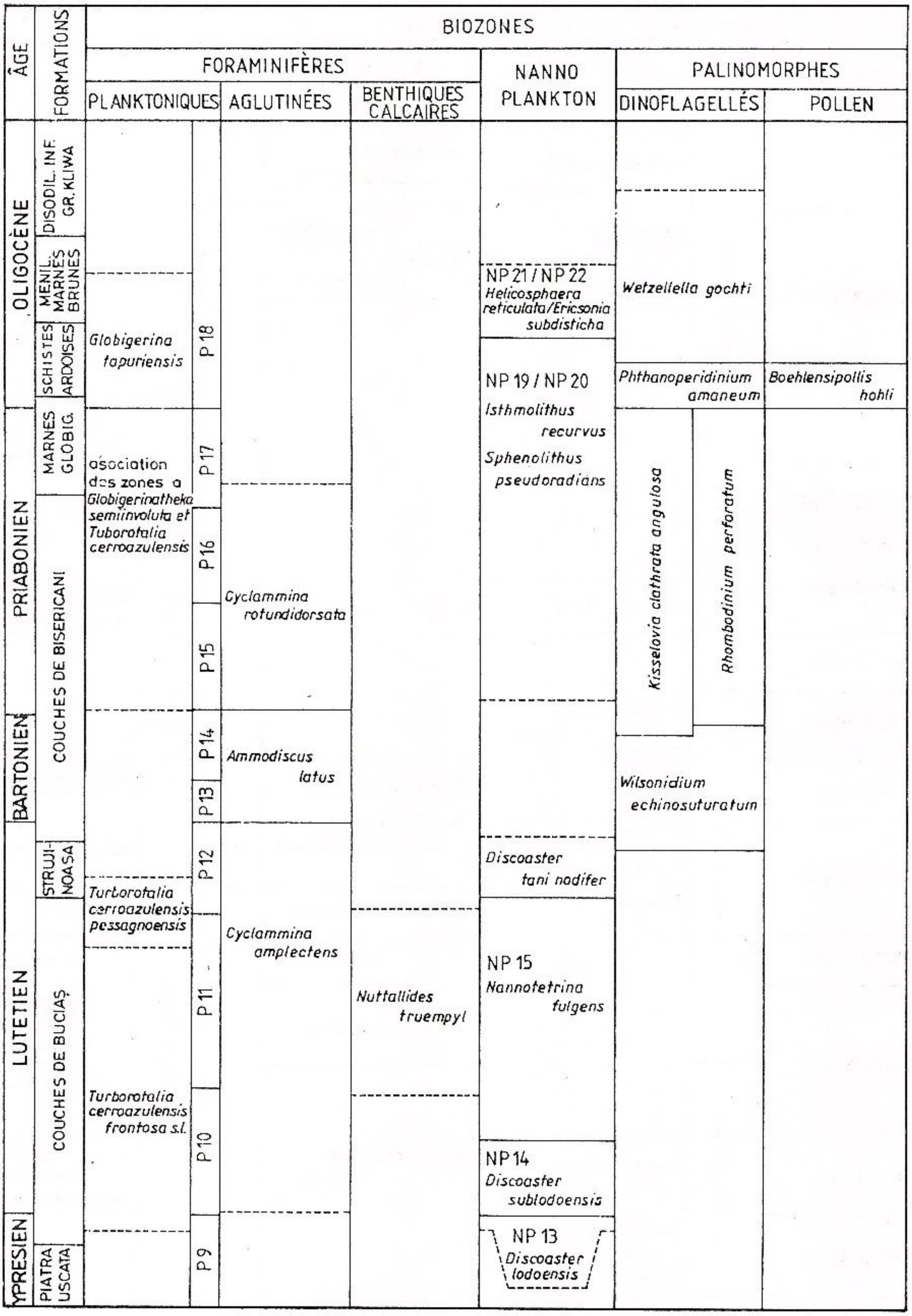


Fig. 2 - Tableau de corrélation. Paléogène moyen et supérieur.

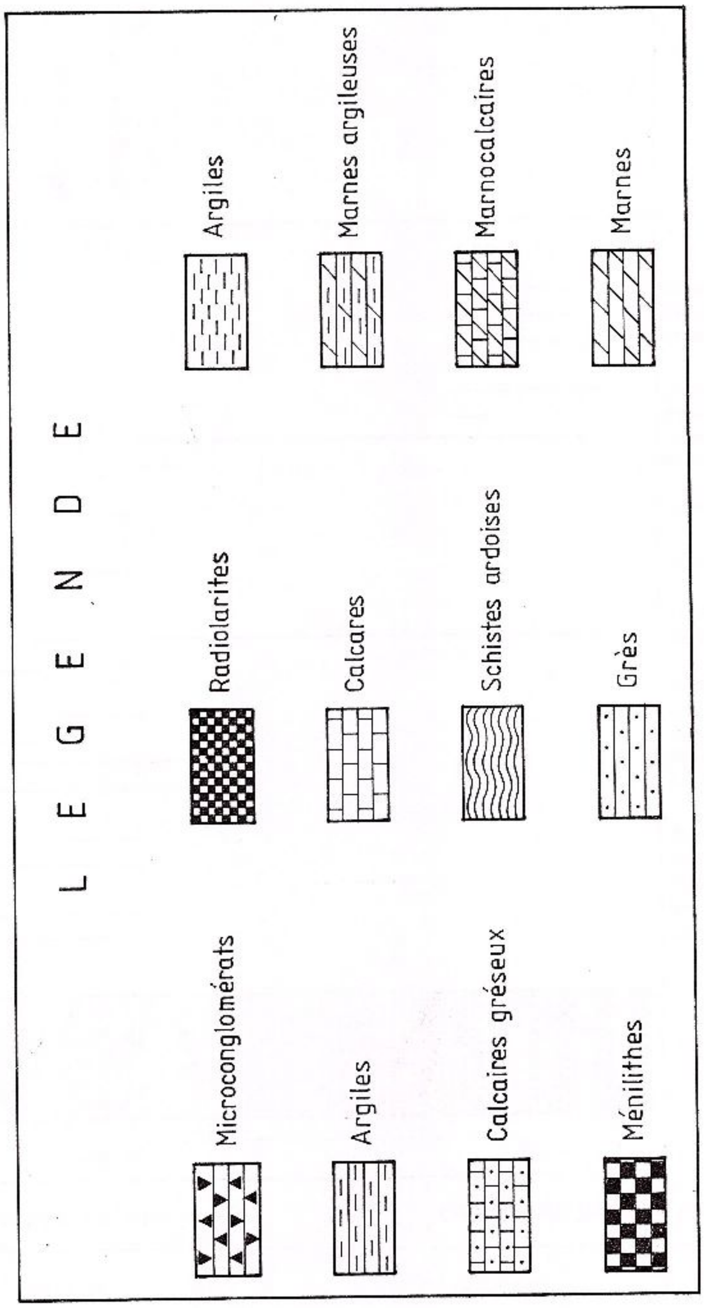

L E G E N D E
Microconglomérats
Argiles
Calcaires gréseux
Ménilithes
Radiolarites
Calcares
Schistes ardoises
Grès
Argiles
Marnes argileuses
Marnocalcaires
Marnes

Fig. 3

Tableau 1 - COUPE DE LA VALLÉE DE PUTNA (GURA LEPSA - GURA STREIU)

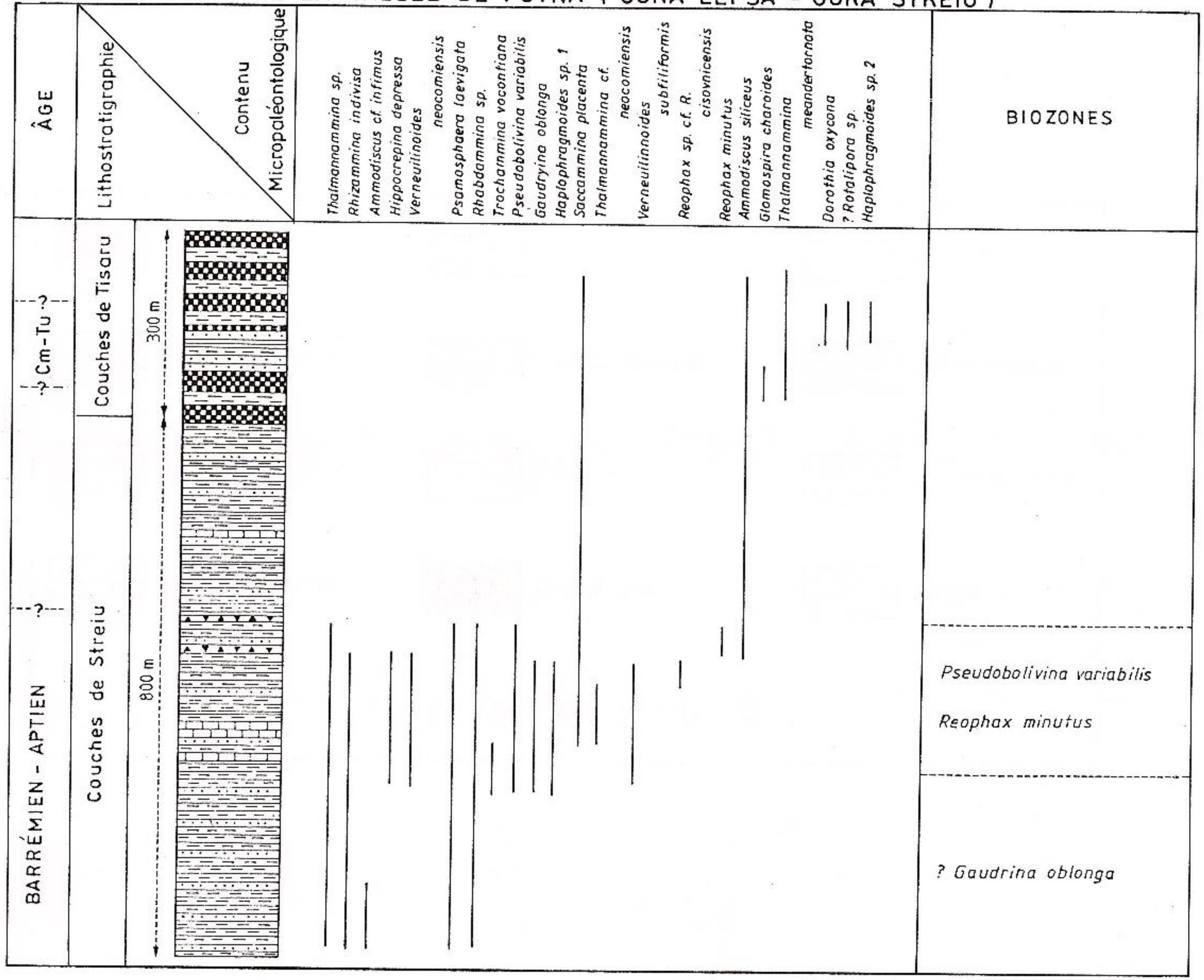


Tableau 2 - COUPE DE LA VALLÉE PUTNA (GURA LEPȘA - GURA STREIU) DÉTAIL

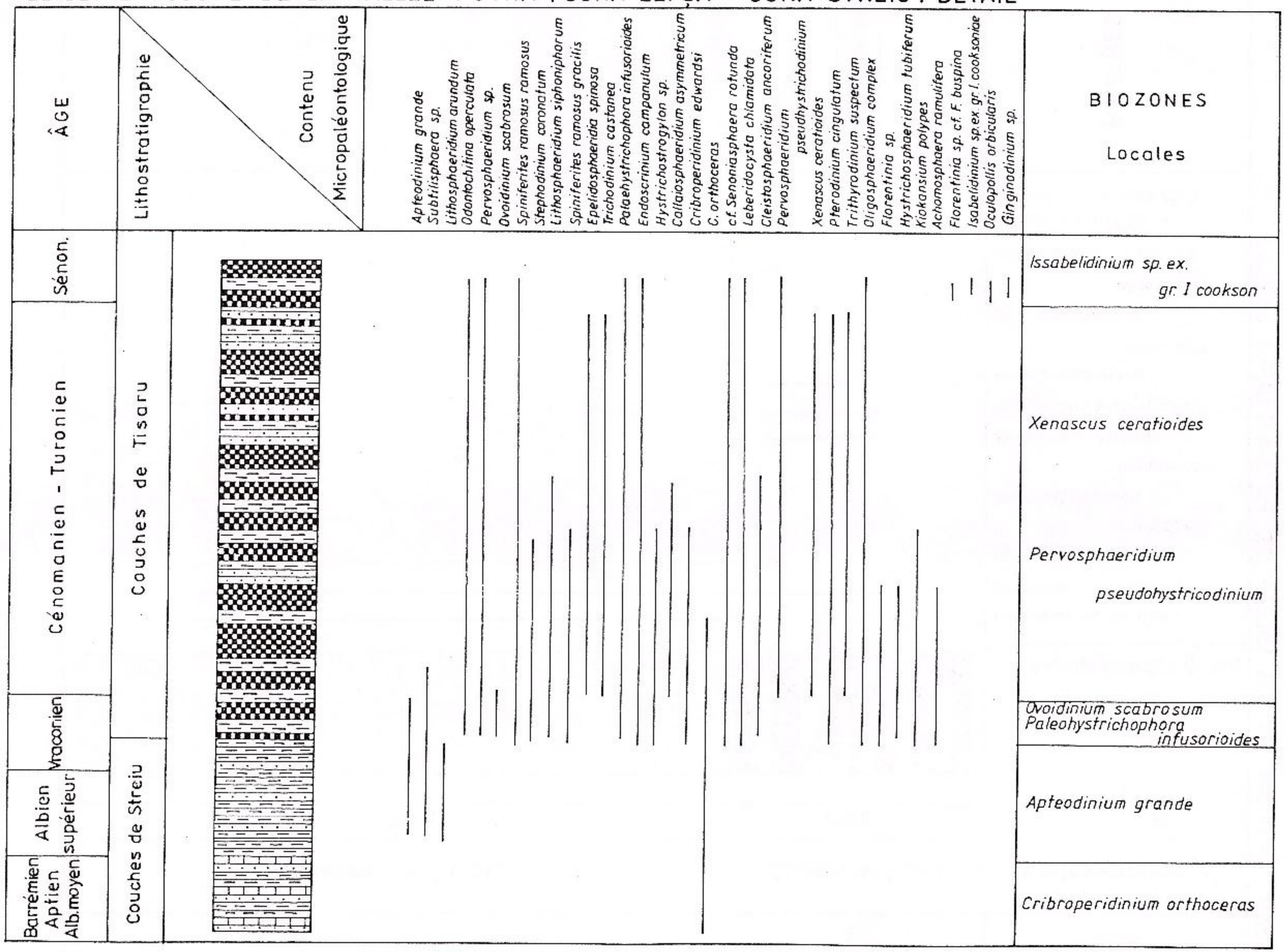


Tableau 3 - COUPE DE LA VALLEÉ DE PUTNA (GURA LEPȘA)

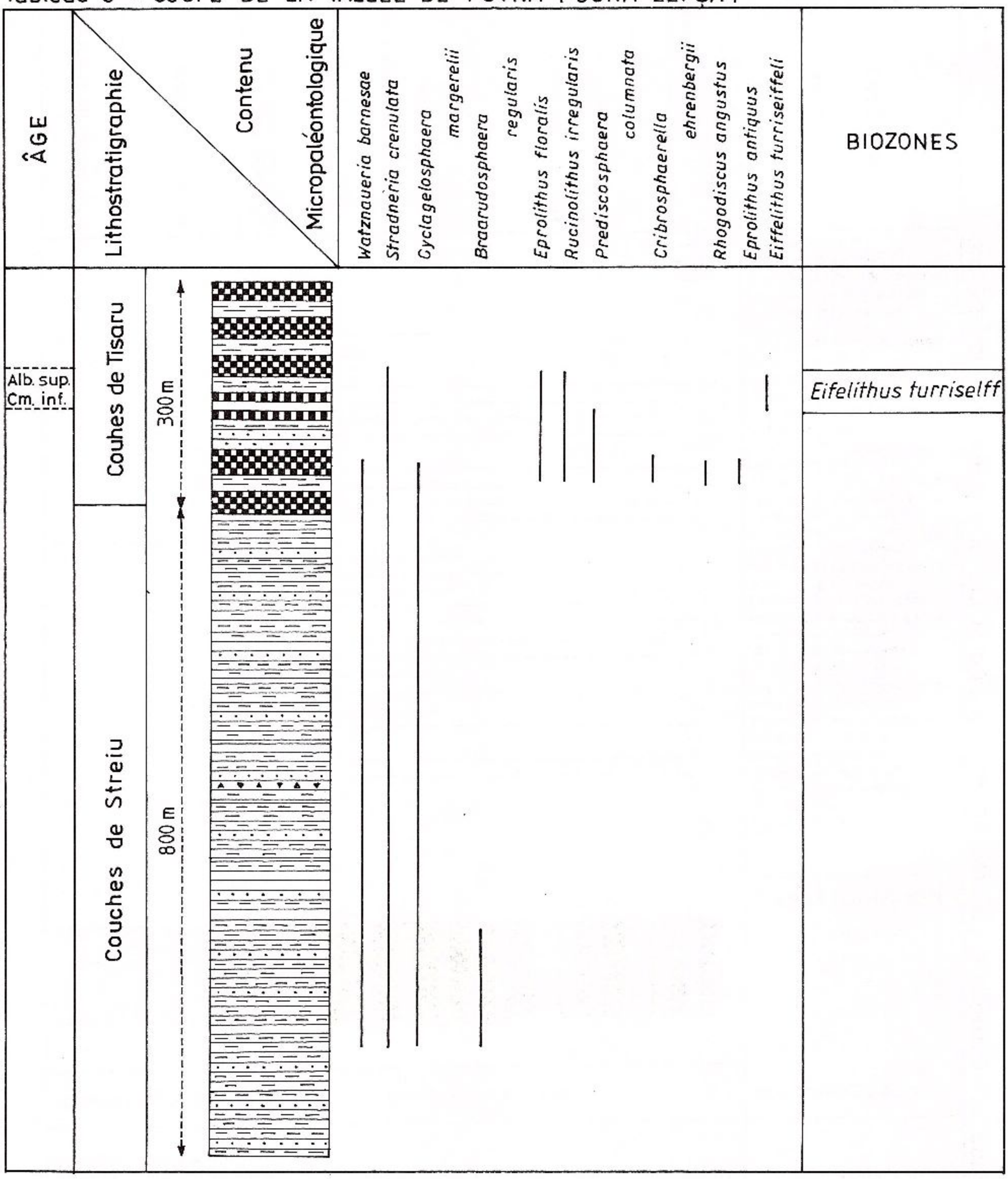


Tableau 4 - COUPE DE LA VALLÉE DE STREIU

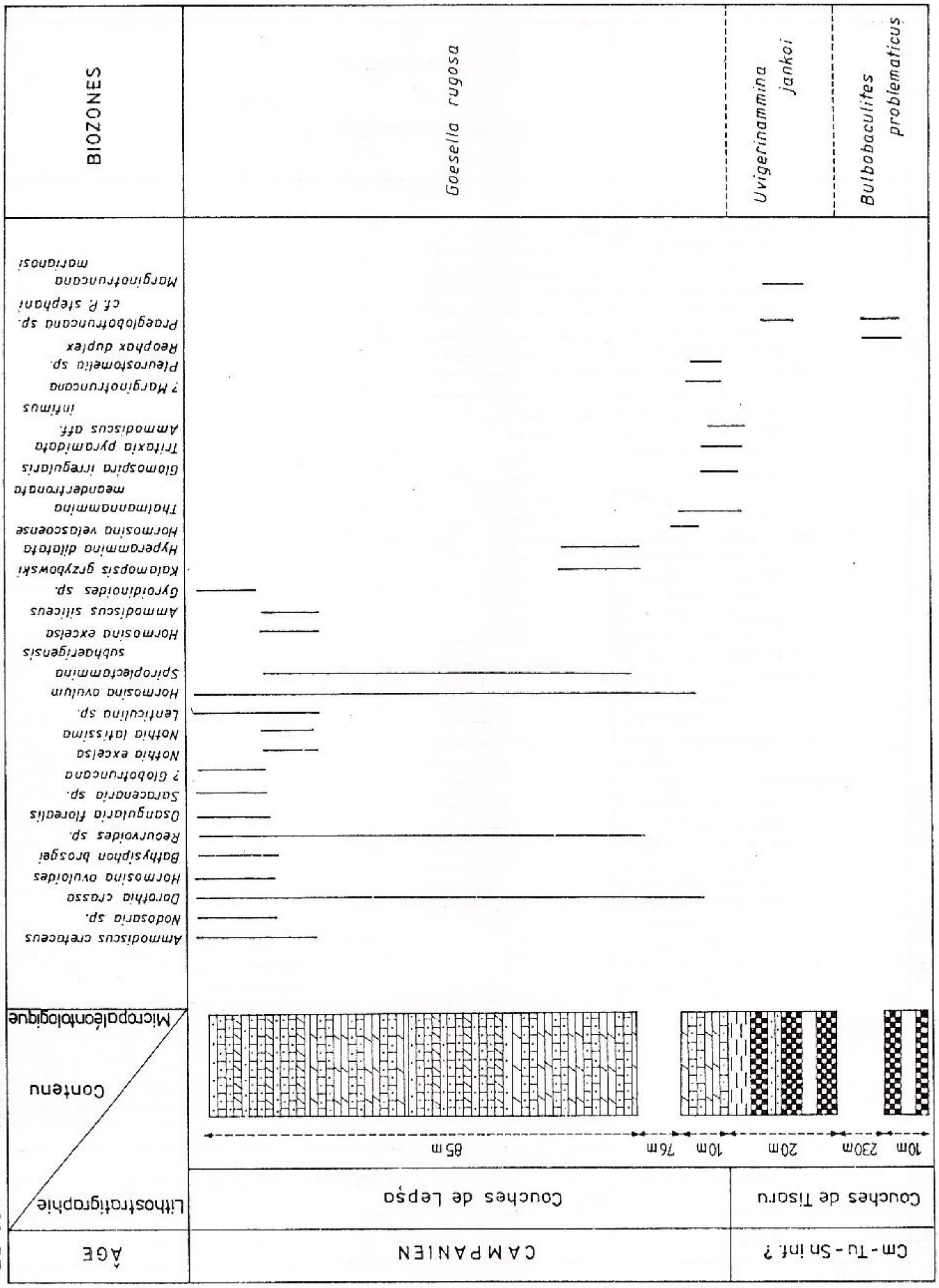
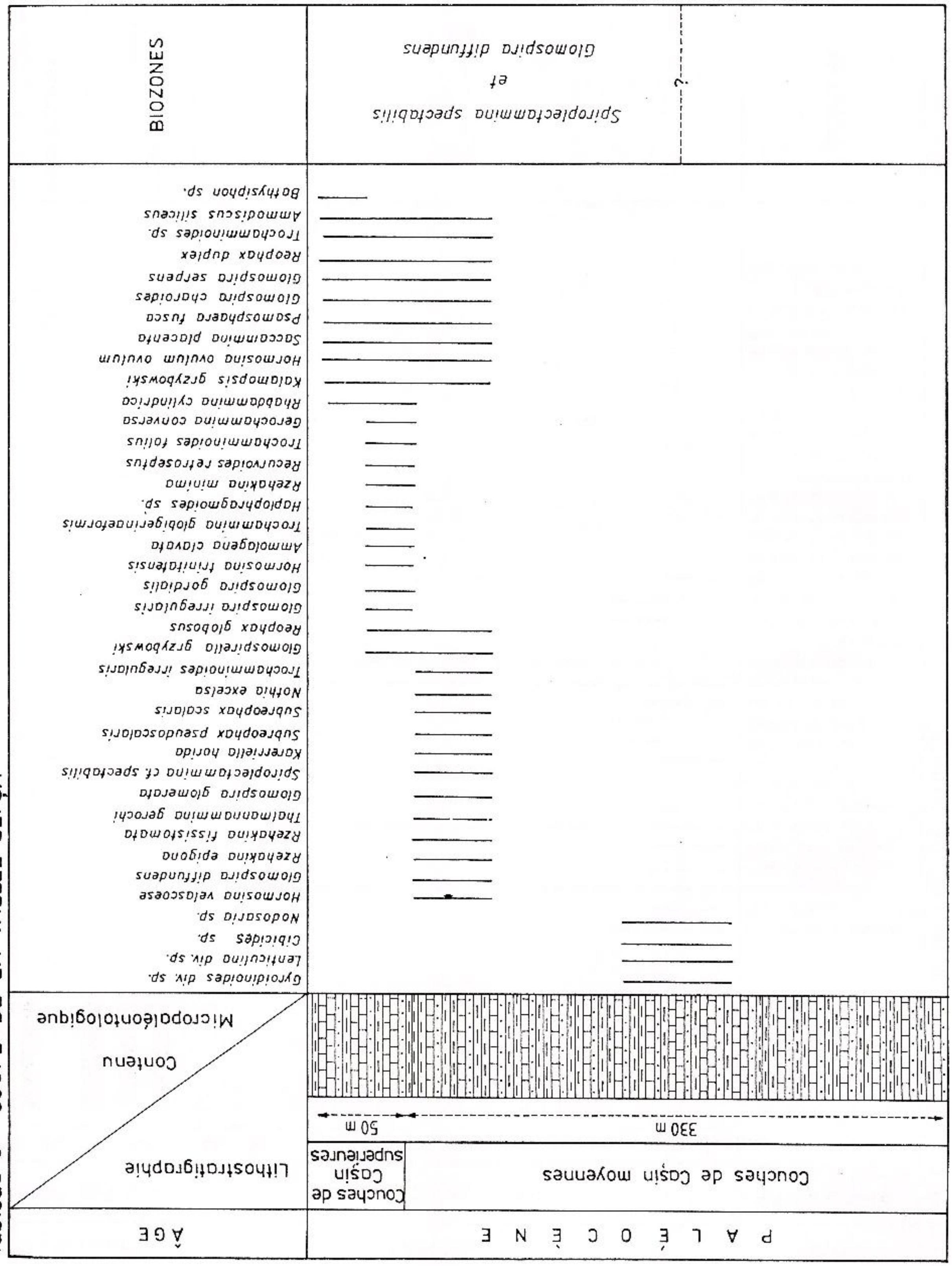


Tableau 5 - COUPE DE LA VALLÉE LEPŞA



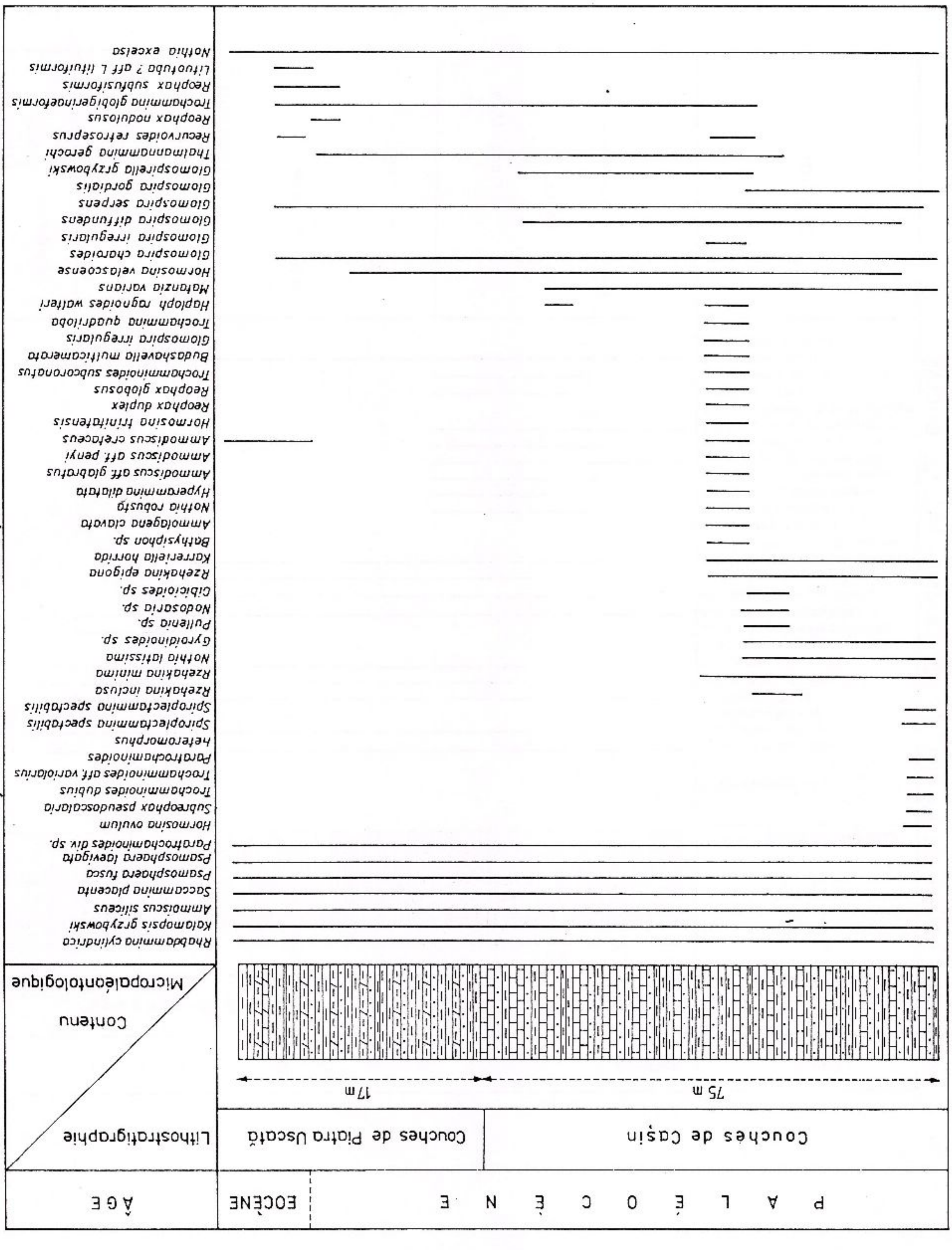
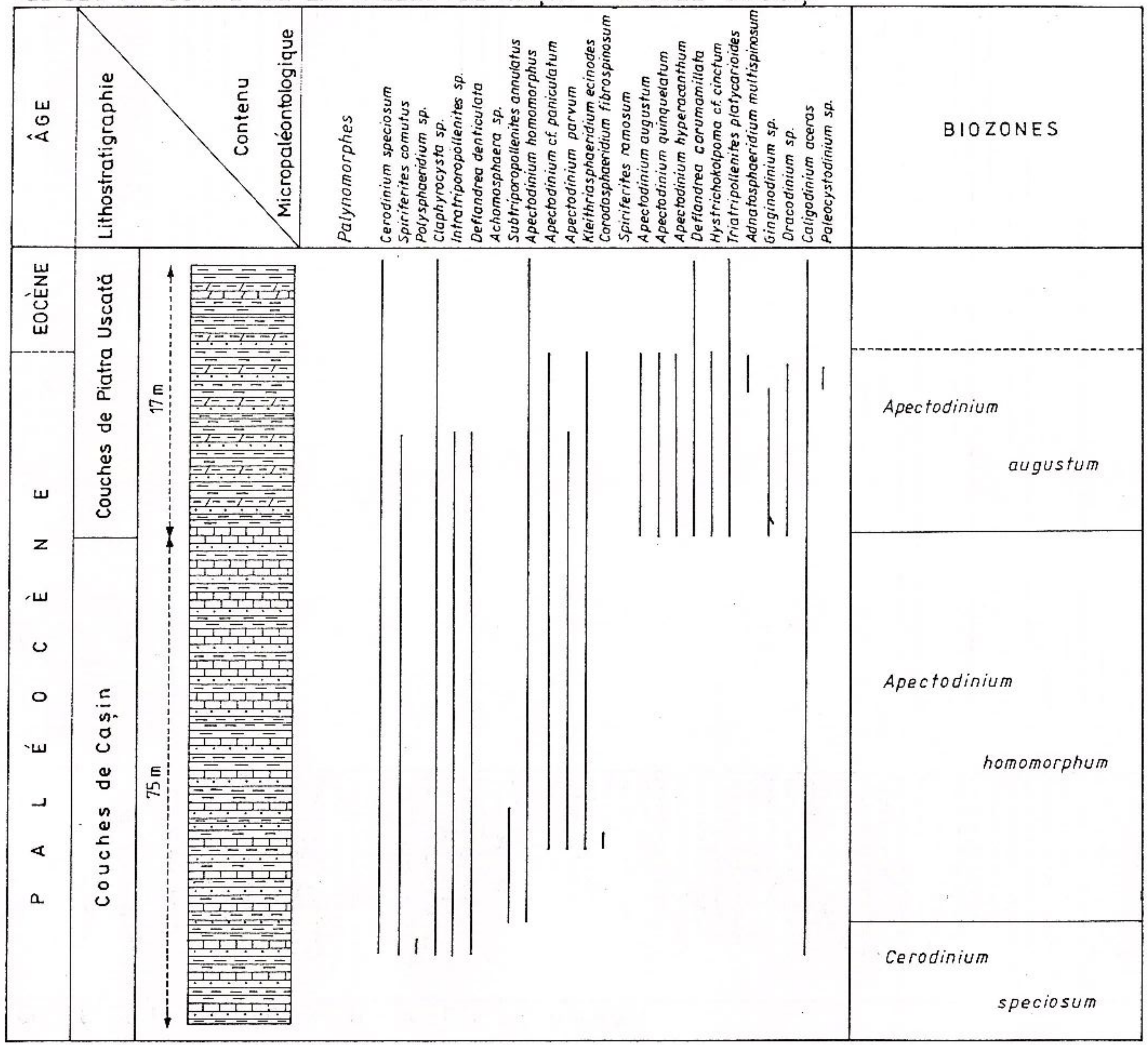


Tableau 6 - COUPE DE LA VALLÉE CASIN - VALLÉE DE BUCIAS

Tableau 7 - COUPE DE LA VALLÉE DE CAȘIN - VALLÉE BUCIAȘ



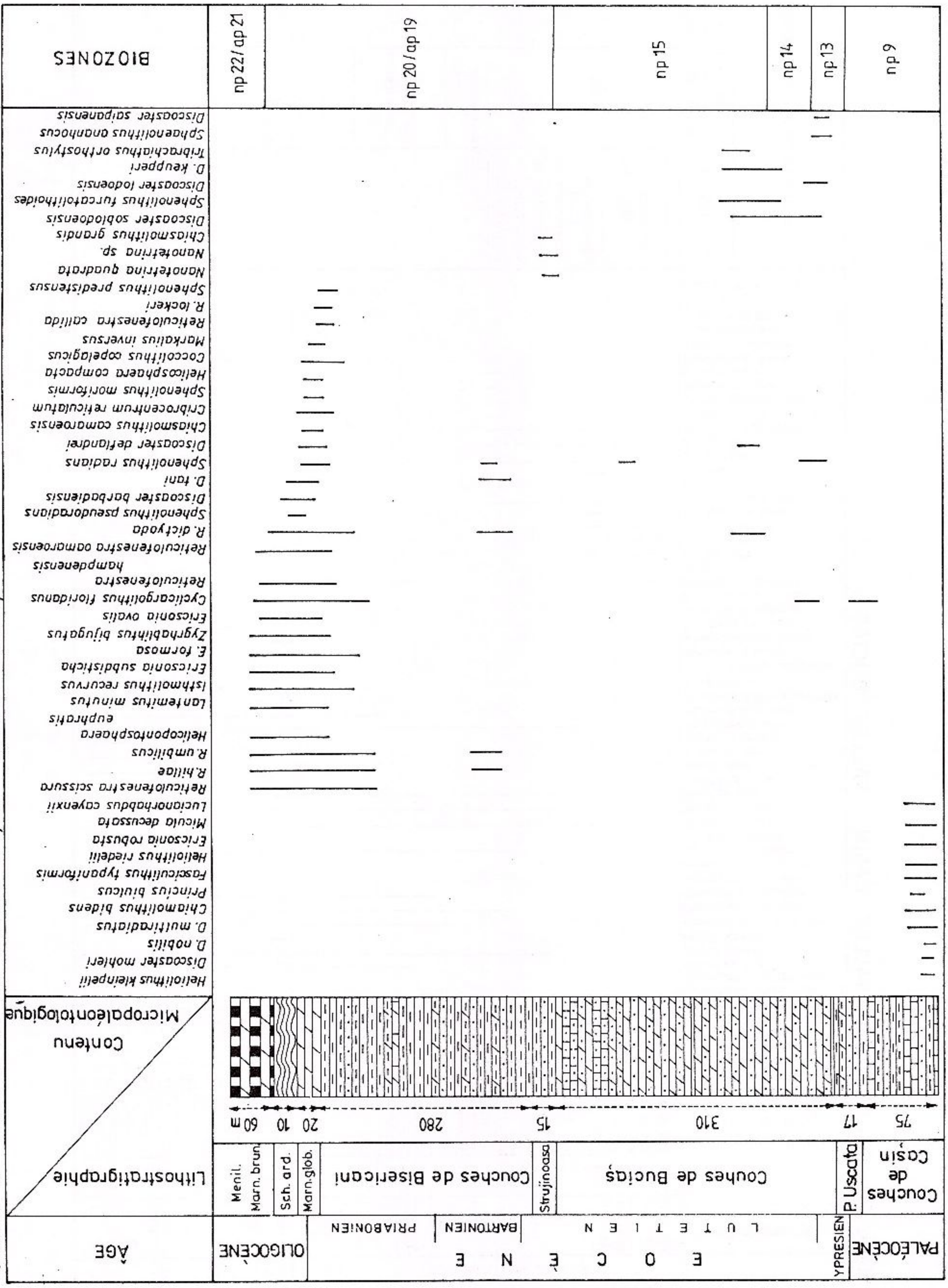
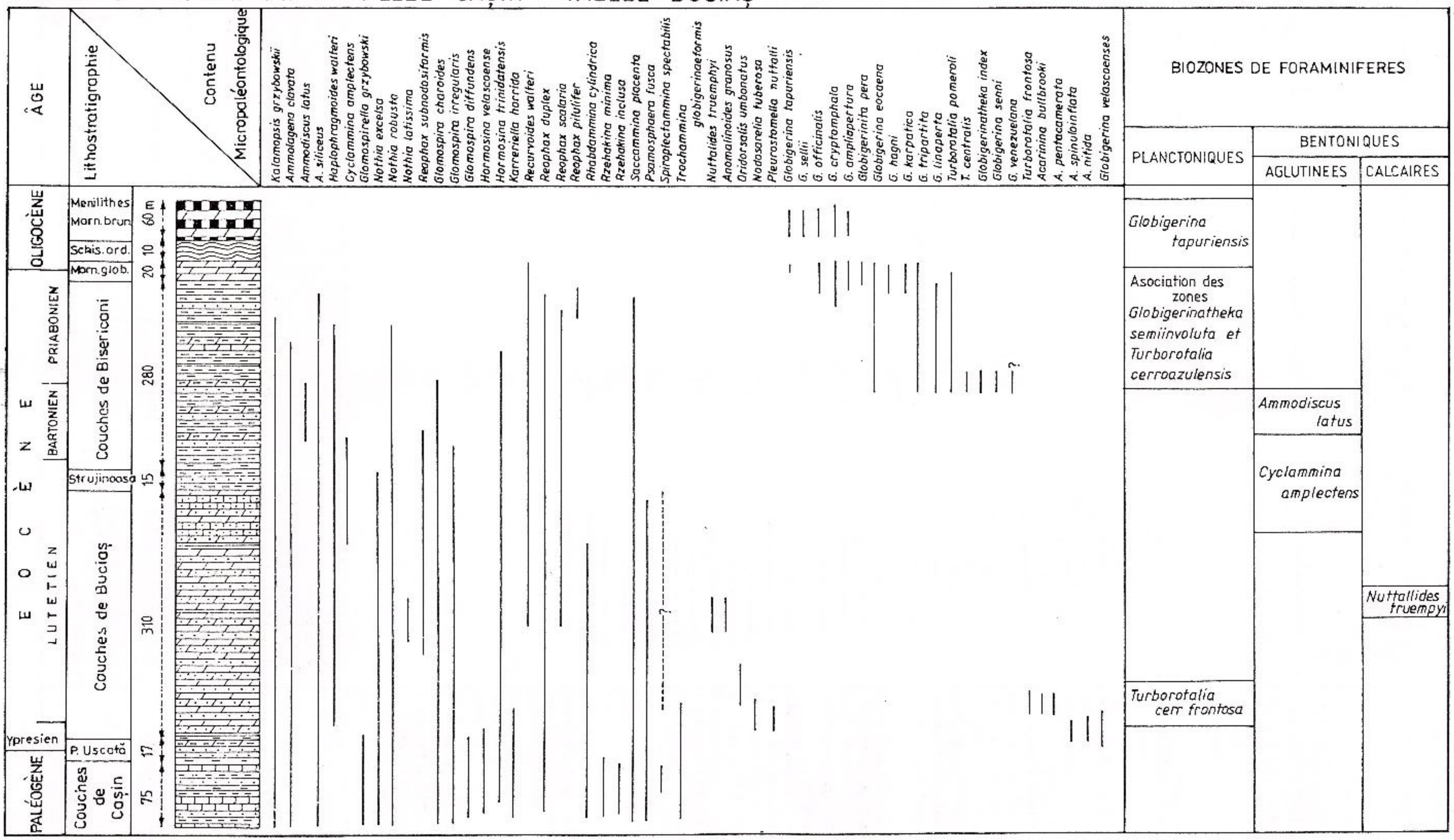


Tableau 8 - COUPE DE LA VALLÉE CASIN - VALLÉE BUCIAS

Tableau 9 - COUPE DE LA VALLÉE CAȘIN - VALLÉE BUCIAȘ



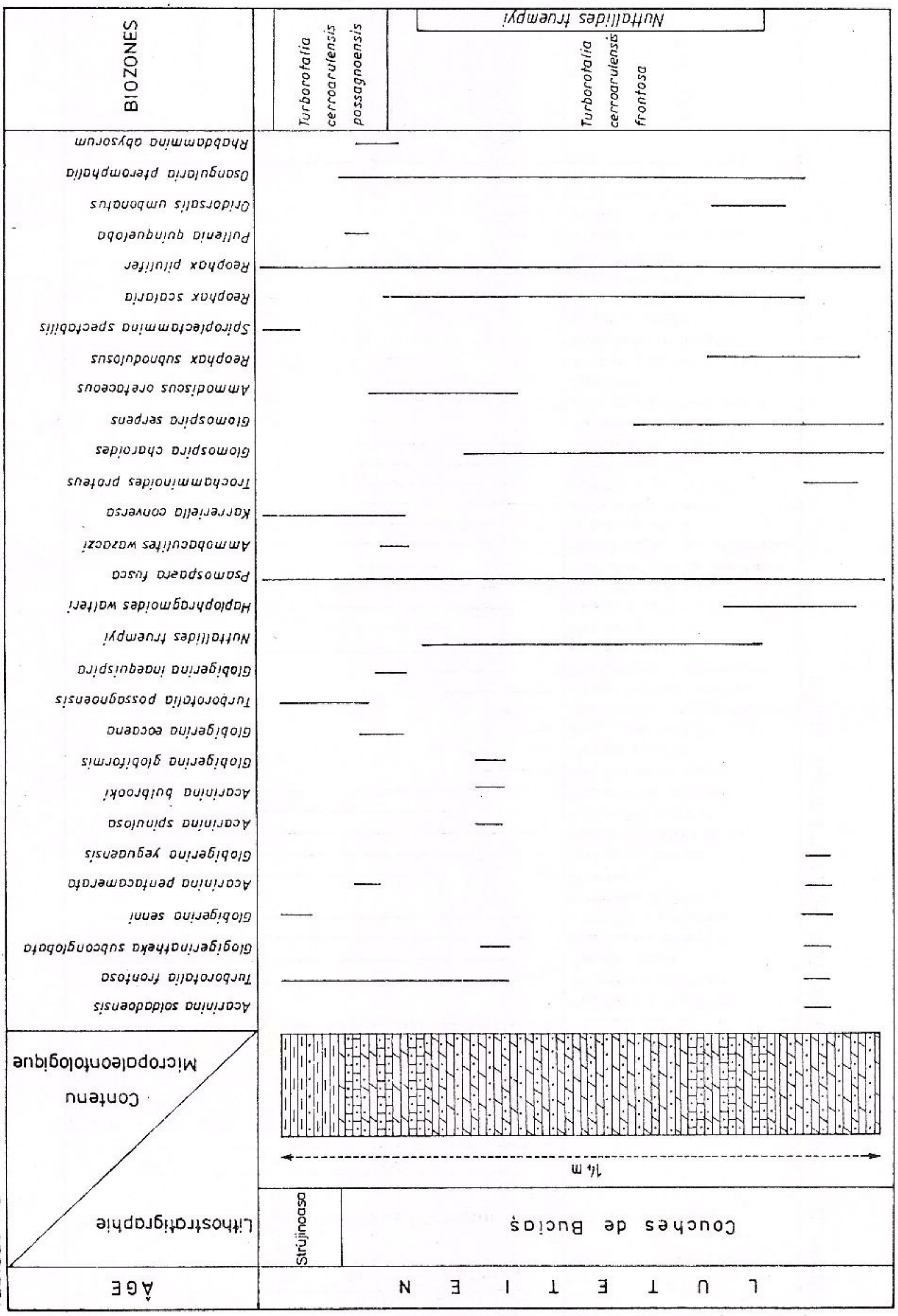


Tableau 10 - COUPE COASA

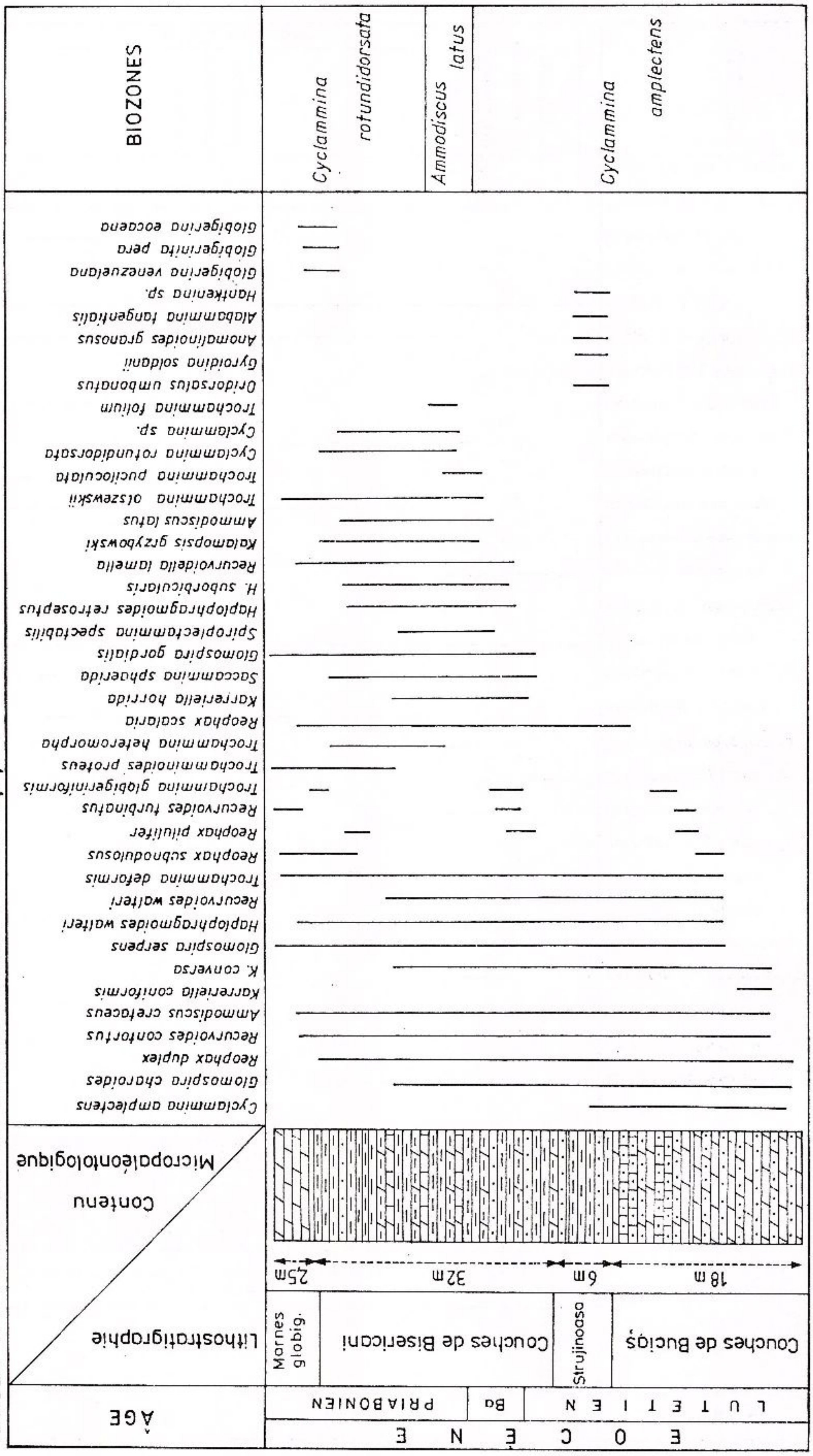


Tableau 11 - COUPE DE LA VALLÉE PUTNA (GURA TIŞIŢA)

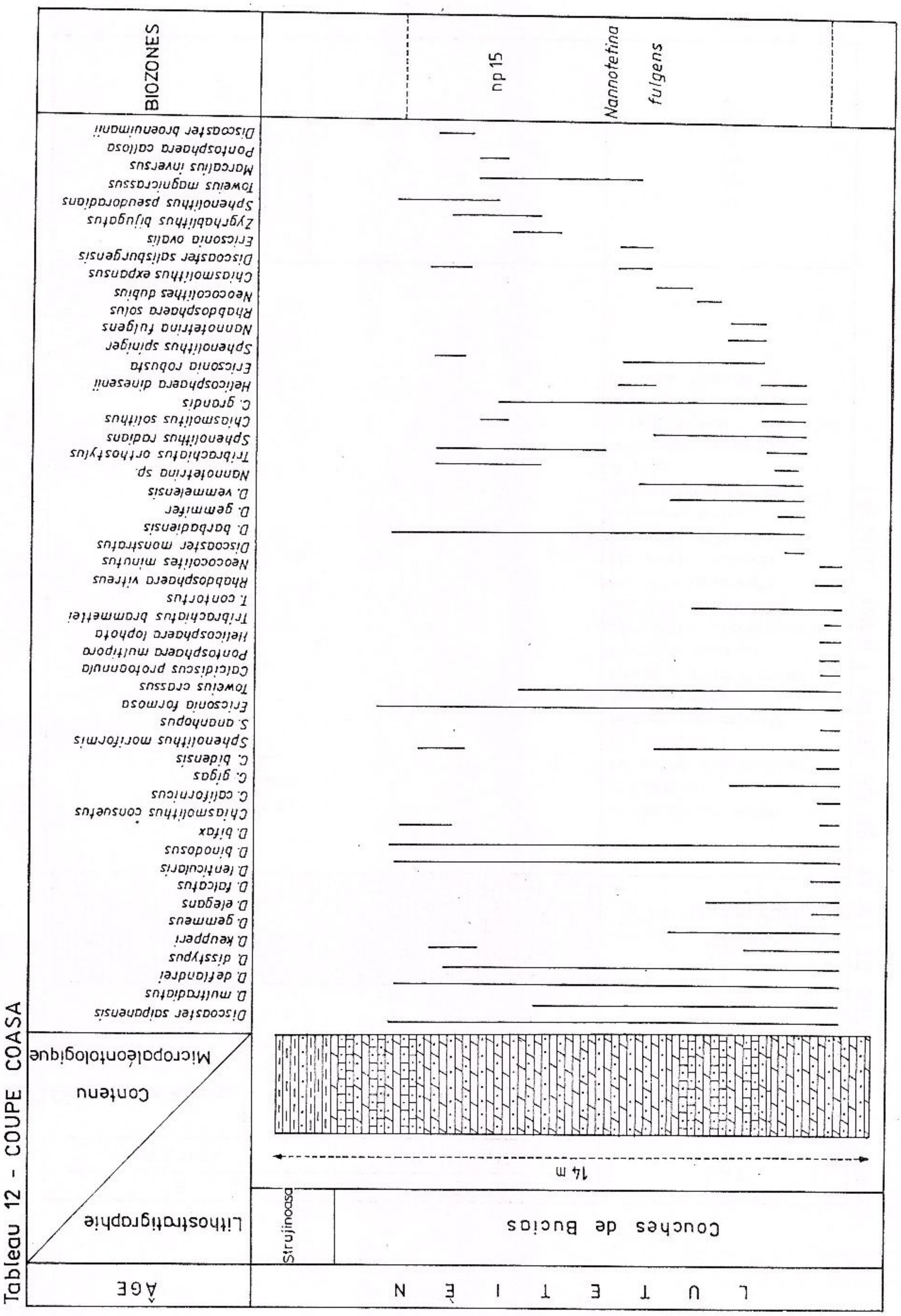


Tableau 12 - COUPE COASA
BIOZONES
np 15
Nannotetina fulgens
Discoaster broenimanii
Pontosphaera callosa
Marcalius inversus
Toweius magnicrassus
Sphenolithus pseudoradians
Zygrhablithus bijugatus
Ericsonia ovalis
Discoaster salisburgensis
Chiasmolithus expansus
Neococcolithes dubius
Rhabdosphaera solus
Nannotetina fulgens
Sphenolithus spiniger
Ericsonia robusta
Helicosphaera dinesenii
C. grandis
Chiasmolithus solitus
Sphenolithus radians
Tribrachiatus orthostylus
Nannotetina sp.
D. vemmelensis
D. gemmifer
D. barbadiensis
Discoaster monstratus
Neococolites minutus
Rhabdosphaera vitreus
T. contortus
Tribrachiatus bramlettei
Helicosphaera lophota
Pontosphaera multipora
Calcidiscus protoannula
Toweius crassus
Ericsonia formosa
S. anarhopus
Sphenolithus moriformis
C. bidensis
C. gigas
C. californicus
Chiasmolithus consuetus
D. bifax
D. binodosus
D. lenticularis
D. falcatus
D. elegans
D. gemmeus
D. keupperi
D. diastypus
D. deflandrei
D. multiradiatus
Discoaster saipanensis
Contenu Micropaléontologique
14 m
Lithostratigraphie
Strujinoasa
Couches de Bucios
ÂGE
L U T E T I E N

Tableau 13 - COUPE DE LA VALLÉE DE PUTNA (GURA TIȘIȚEI)

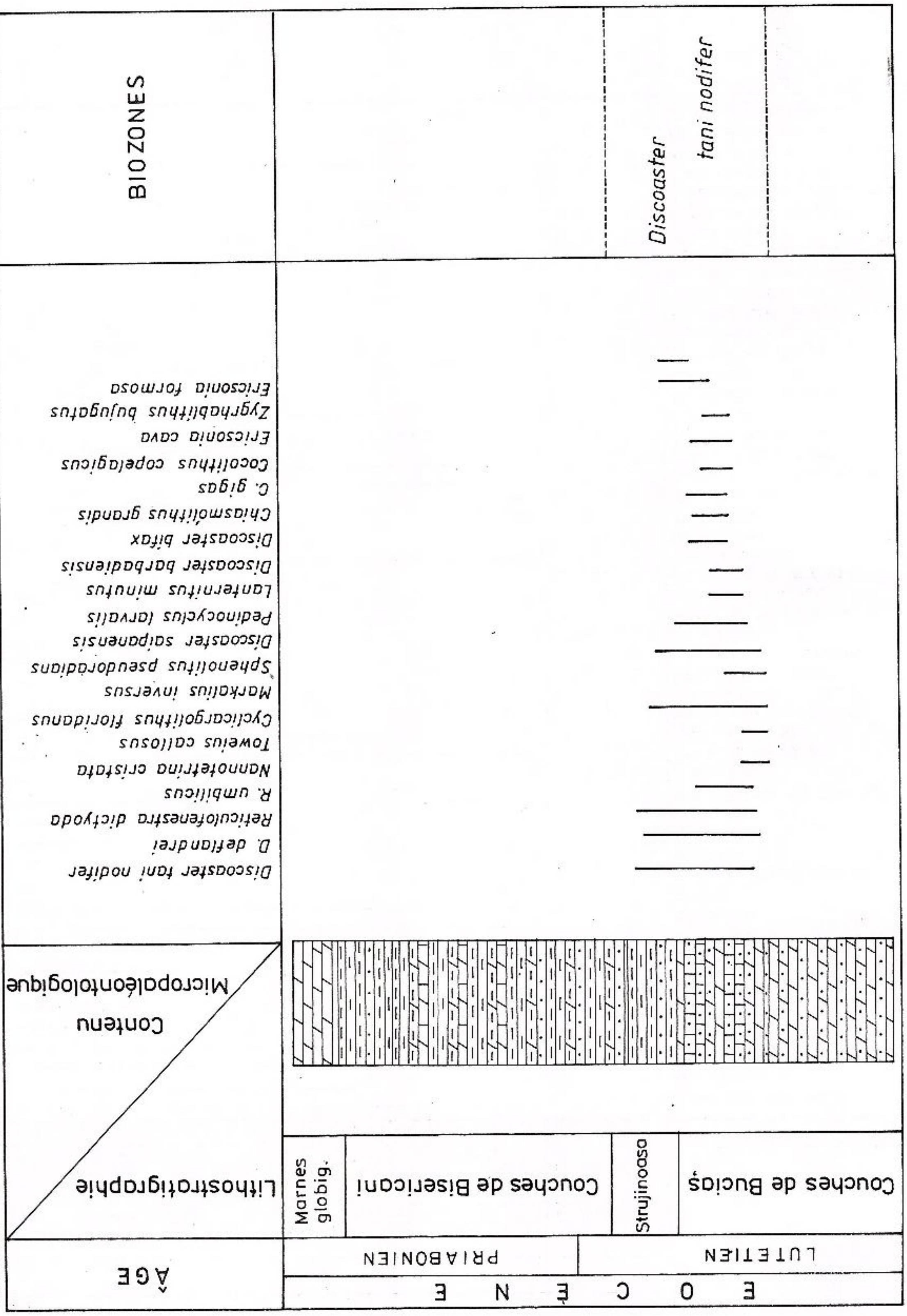


Tableau 14 a - COUPE DE LA VALLÉE CASIN - VALLÉE BUCIAS

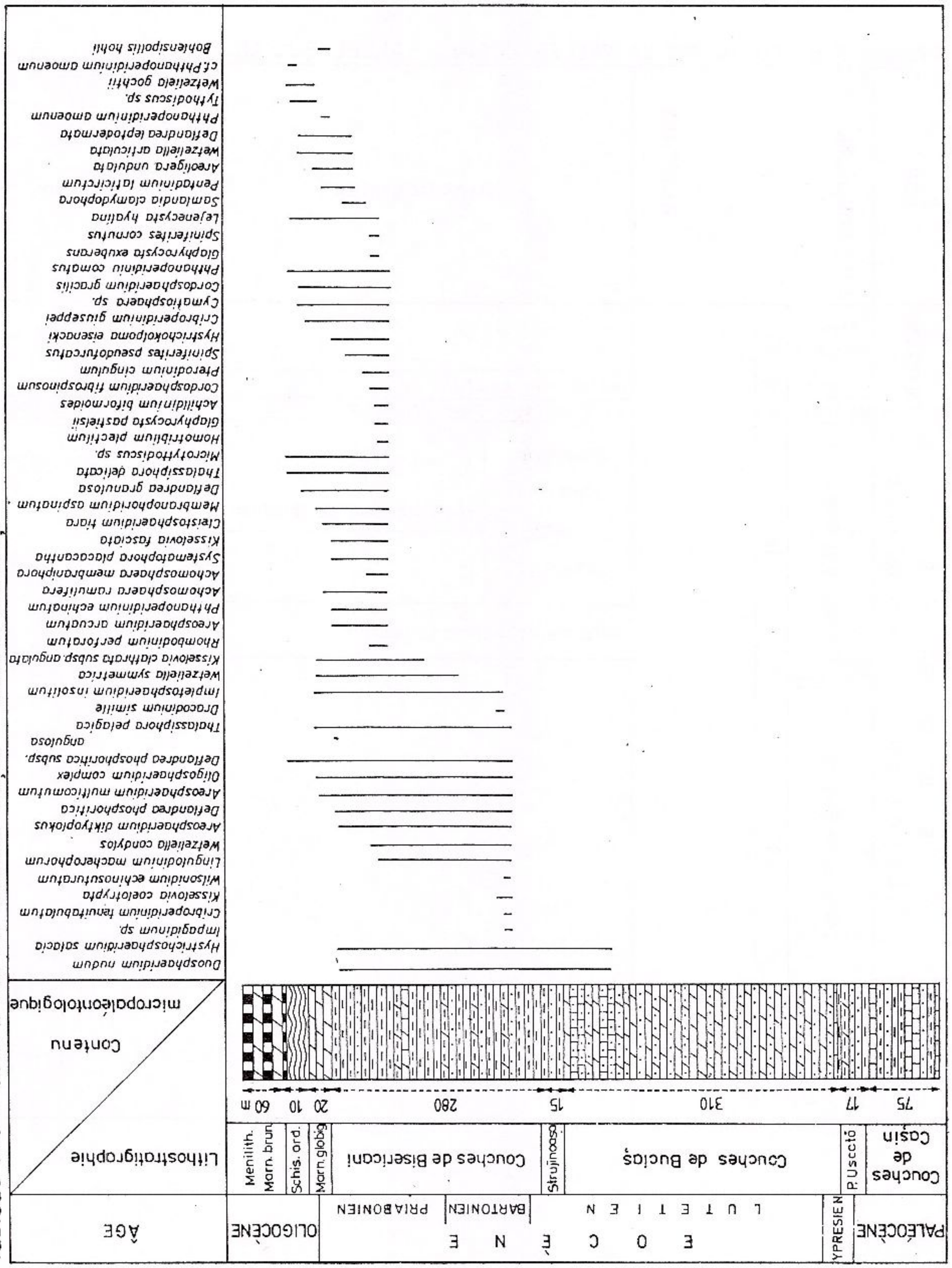


Tableau 14 b - COUPE DE LA VALLÉE CAȘIN - VALLÉE BUCIAȘ

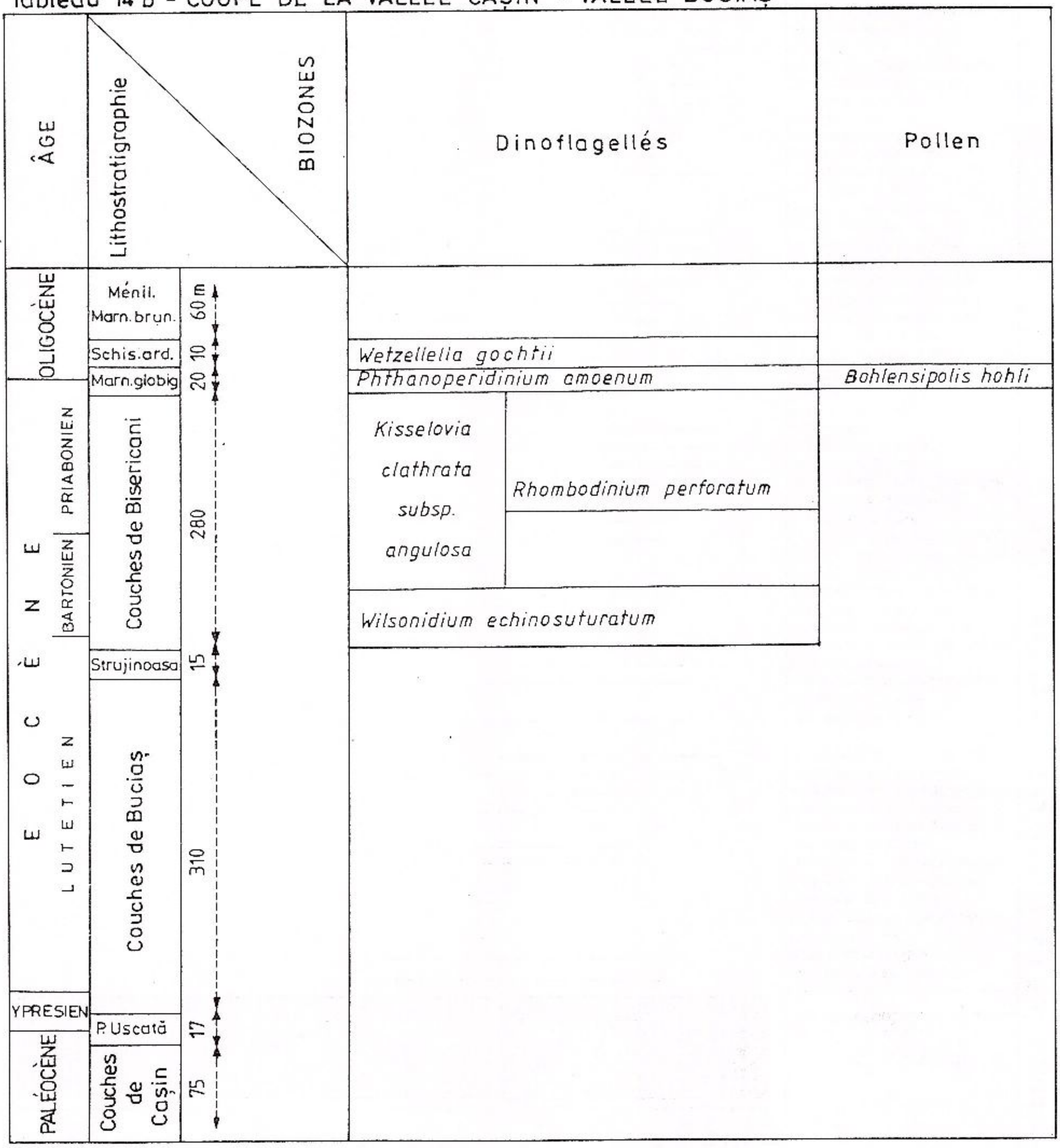


Tableau 15 - COUPE DE LA VALLÉE PUTNA

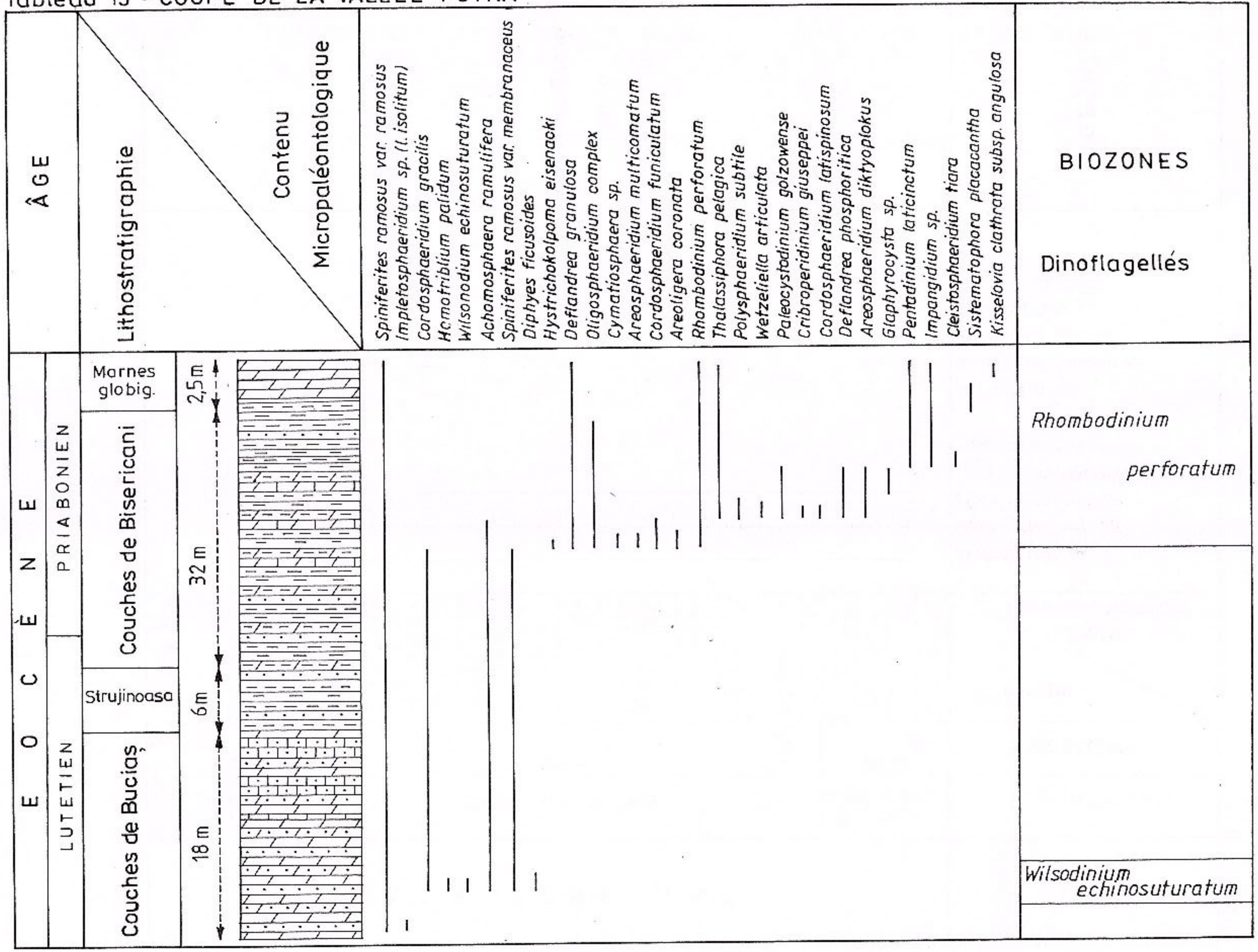
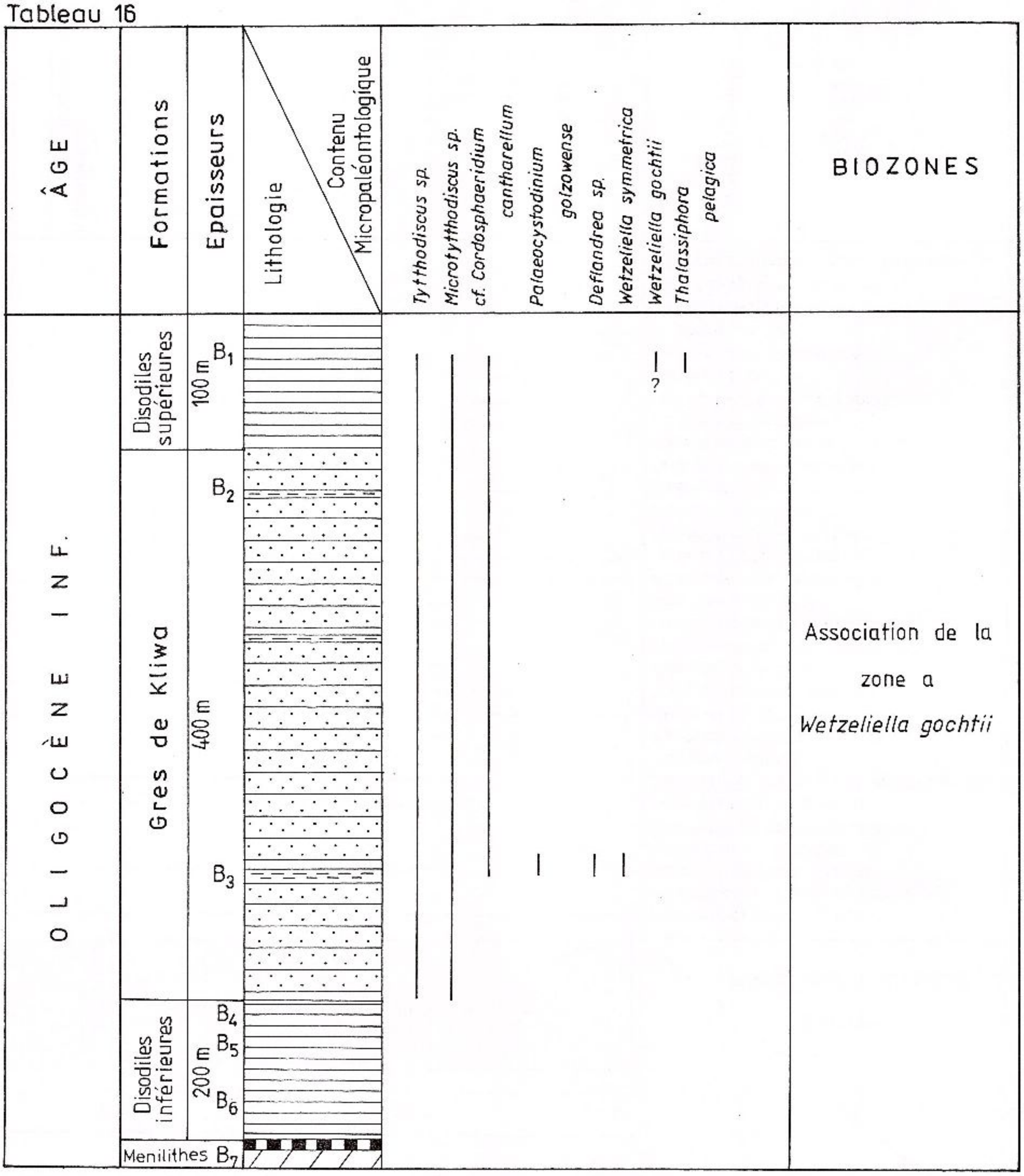


Tableau 16


ÂGE
Formations
Epaisseurs
Lithologie
Contenu Micropaléontologique
Tytthodiscus sp.
Microtytthodiscus sp.
cf. Cordosphaeridium cantharellum
Palaeocystodinium golzowense
Deflandrea sp.
Wetzeliella symmetrica
Wetzeliella gochtii
Thalassiphora pelagica
BIOZONES
OLIGOCÈNE INF.
Disodiles supérieures
100 m
B_1
?
Gres de Kliwa
B_2
400 m
B_3
Disodiles inférieures
200 m
B_4
B_5
B_6
Menilithes B_7
Association de la zone a *Wetzeliella gochtii*

P L A T E S

Planche I

Eléments de la microflore des Couches de Streiu (Barrémien-Vraconien inférieur).
Partie terminale des Couches de Streiu, Albien supérieur-Vraconien inférieur?. Gl 33, GL 33 bis.

Figs. 1-2 — *Apteodinium grande* COOKSON & HUNGHES, 1964 Fig. 1, lame 4A28/1; 104/58,5; 113μ f. 3A86. Fig. 2, lame 4A28/1; 114,6/60,9; 110μ f. 3A86.

Fig. 3 — *Cribroperidinium edwardsi* (COOKSON & EISENACK, 1958) DAVEY, 1969
Lame 4A28/1; 123,5/60,8; 100μ f. 3A86.

Eléments de la microflore des Couches de Tisaru (Vraconien supérieur-Campanien inférieur).
Partie basale, Vraconien-Cénomanien inférieur, GL 39.

Fig. 4 — *Ovoidinium scabrosum* (sc Cookson & EISENACK, 1964) DAVEY, 1970
Lame 4A34/18; 117,9/57,6; 63μ f. 3A86.

Fig. 5 — *Apteodinium grande* COOKSON & HUGHES, 1964
Lame 4A34/2; 110,3/44; 128μ f. 3A86.

Fig. 6 — *Lithisphaeridium siphoniphorum* (COOKSON & EISENACK, 1958) DAVEY & WILLIAMS, 1966
Lame 4A34/1; 109,7/49,5; 60μ f. 3A87.

Fig. 7 — *Stephodinium coronatum* DEFLANDRE, 1936
Lame 4A34/17; 107,2/48; 90μ f. 3A88.

Fig. 8 — *Florentinia* sp.
Lame 4A34/4; 110/61; 78μ f. 3A88.

Fig. 9 — *Histrichostrogylon* sp.
Lame 4A34/3; 124,9/59; 75μ f. 3A88.

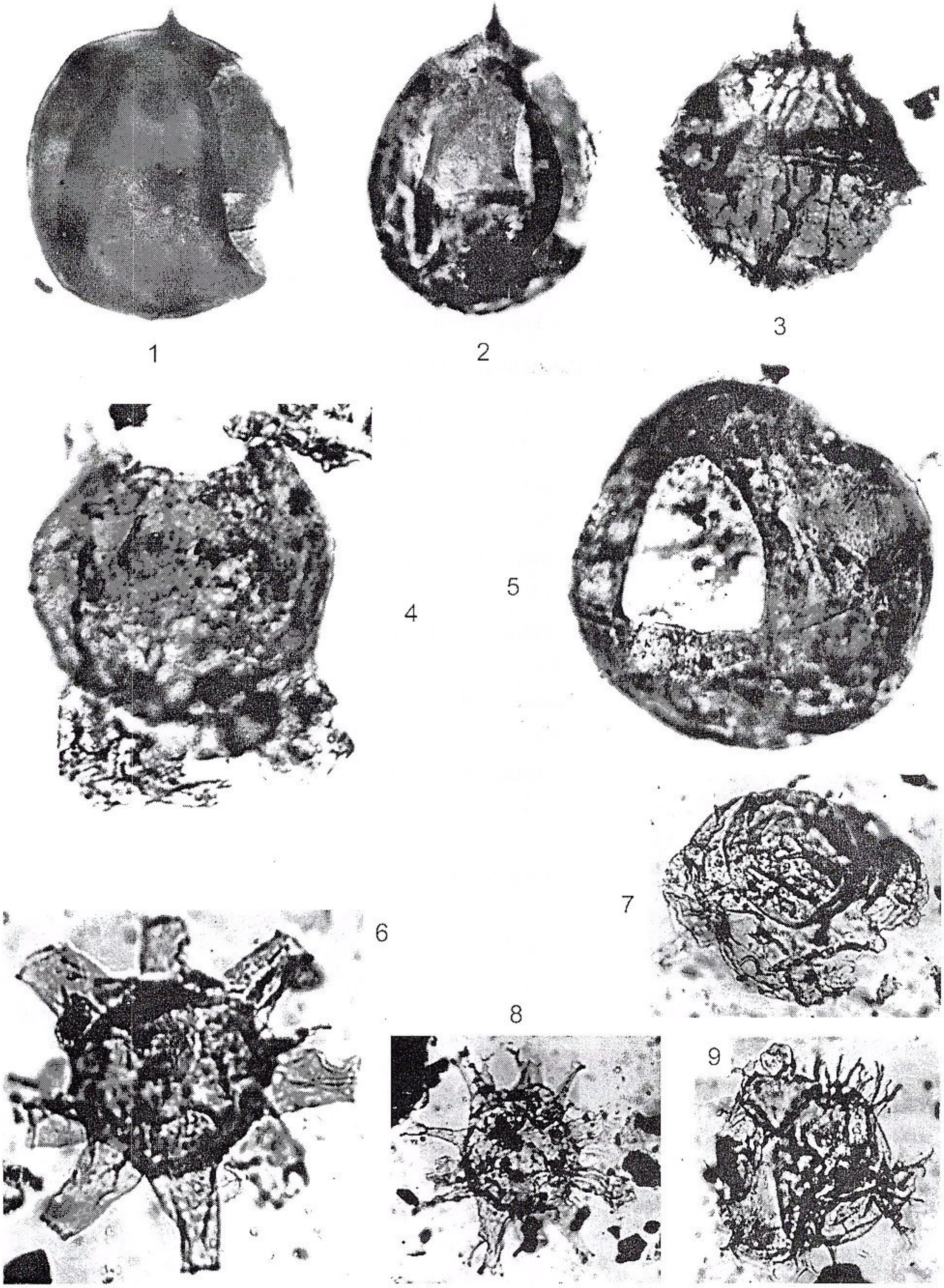

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Planche II

Eléments de la microflore des Couches de Tisaru (Vraconien supérieur - Campanien inférieur).
Vallée de Putna.
Partie basale, Vraconien-Cénomanien inférieur (GL 39).

Fig. 1 — *Spiniferites ramosus ramosus* (EHRENBERG, 1838) DAVEY & WILLIAMS, 1966.
Lame 4A34/3; 124/59; 75μ f. 3A88.

Partie inférieur-moyenne des Couches de Tisaru, Cénomanien inférieur-Turonien (GL 36-GL 45).

Fig. 2 — *Palaeohystrichophora infusorioides* DEFLANDRE, 1935.
Lame 4A36/13; 119,5/62,5; 55μ f. 3A86.

Fig. 3 — *Epelidosphaeridia spinosa* (COOKSON & HUGHES, 1964) DAVEY, 1969.
Lame 4A45/3; 121,9/58; 55μ f. 3A86.

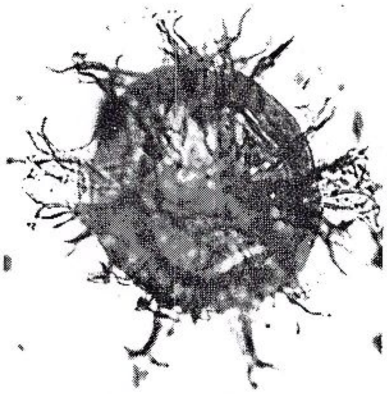
Fig. 4 — *Trichodinium castaneum* (DEFLANDRE, 1935) CLARKE & VERDIER, 1967.
Lame 4A36/11; 114,8/54,3; 60μ f. 3A88.

Fig. 5 — *Pervosphaeridium pseudhystrichodinium* (*Deflandre* 1937) *Yun*, 1981.
Lame 4A36/11; 114,8/54,3; 60μ f. 3A88.

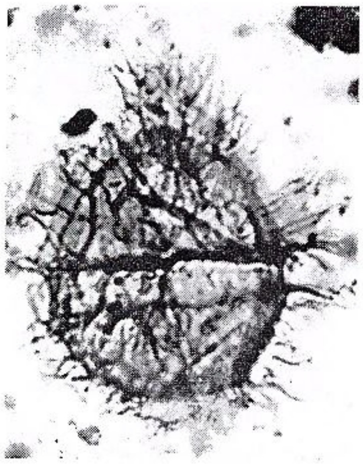
Fig. 6 — *Hystrichosphaeridium tubiferum* (EHRENBERG, 1838) DEFLANDRE, 1937, emend. DAVEY & WILLIAMS, 1966.
Lame 4A36/1; 123,2/59,3; 88μ f. 3A85.

Fig. 7 — *Gonyaulacysta cassidata* (EISENACK & COOKSON, 1960) SARJEANT, 1966.
Lame 4A36/3; 121/59,1; 68μ f. 3A88.

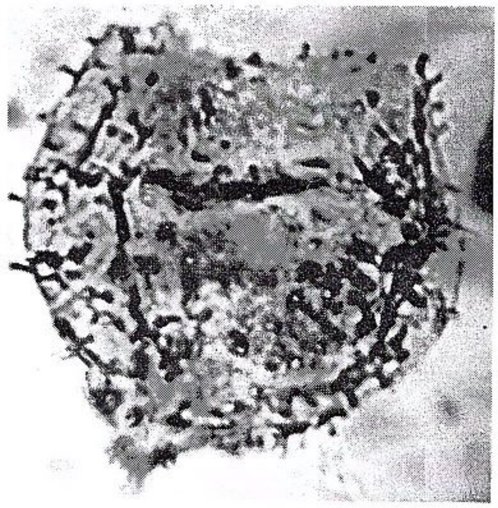
Fig. 8 — *Endoscrinium campanulum* (GOCHT, 1959) VOZZHENNIKOVA, 1967.
Lame 4A36/4; 116/46; 78μ f. 3A86.



1



2



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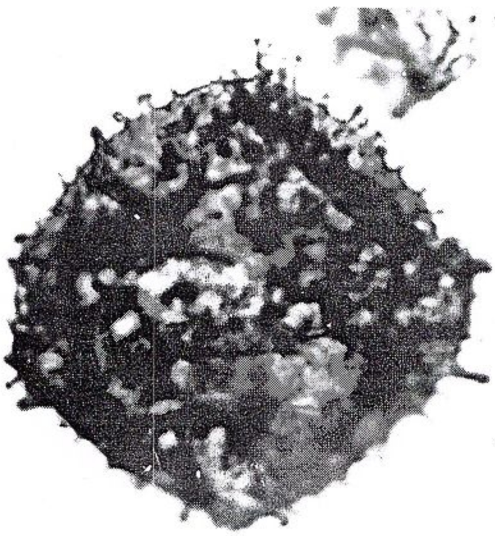
4



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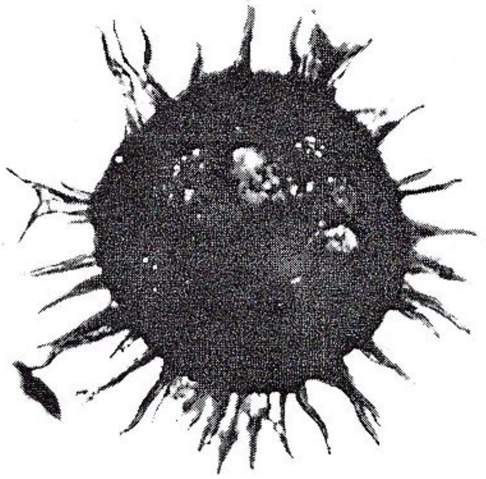

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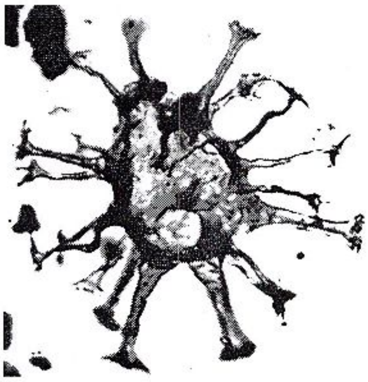

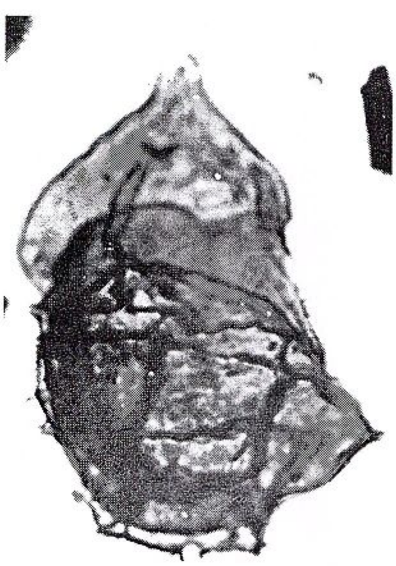

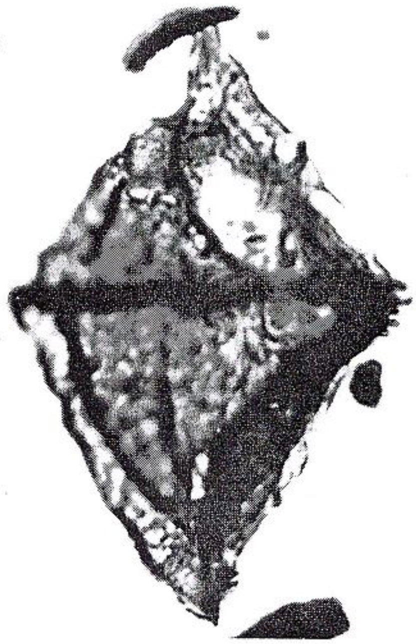

Planche III

Eléments de la microflore des Couches de Tisaru (Vraconien supérieur-Campanien inférieur).
Vallée de Putna.
Partie inférieur-moyenne des Couches de Tisaru. Cénomanien inférieur-Turonien (GL 36-GL 45).

Fig. 1 — *Pterodinium cingulatum* (O. WETZEL, 1933) BELOW, 1981.
Lame 4A36/11; 105,5/52,6; 60μ f. 3A88.

Fig. 2 — *Xenascus ceratioides* (DEFLANDRE, 1937) LENTIN & WILLIAMS, 1937.
Lame 4A35/8; 103,1/40; 80μ f. 3A85.

Fig. 3 - cf. *Leberidocysta chlamidata* (COOKSON & EISENACK, 1962) STOVER & EVITT, 1978.
Lame 4A36/1; 120,3/60,9; 73μ f. 3A83.

Fig. 4 — *Spiniferites* sp. cf. *S. crassimuratus* (DAVEY & WILLIAMS, 1966) SARJEANT, 1970.
Lame 4A36/3; 117/45,6; 55μ f. 3A88.

Fig. 5 — *Callaiosphaeridium asymmetricum* (DEFLANDRE & WILLIAMS, 1966) SARJEANT, 1970.
Lame 4A36/18; 122/55,5; 90μ f. 3A85.

Fig. 6 — *Trithirodinium suspectum* (MANUM & COOKSON, 1964) DAVEY, 1969.
Lame 4A36/11; 110/46,5; 63μ f. 3A88.

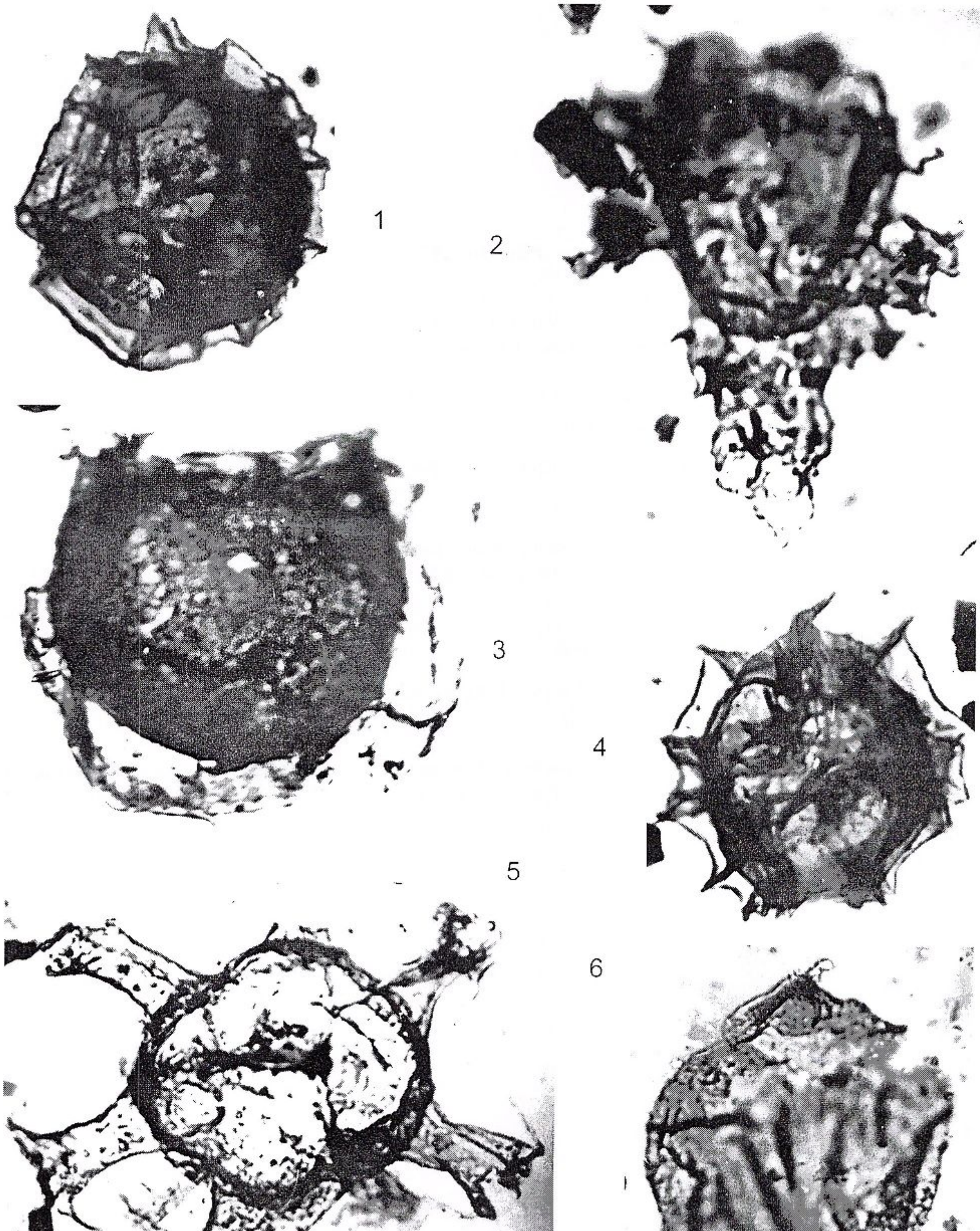
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Planche IV

Eléments de la microflore des Couches de Tisaru (Vraconien supérieur-Campanien inférieur).
Vallée de Putna.
Partie inférieur-Turonien (GL 36-GL45).

Fig. 1 — *Xiphophoridium alatum* (COOKSON & EISENACK, 1962) SARJEANT, 1966.
Lame 4A36/12; 105/63,5; 100μ f. 3A86.

Fig. 4 — *Coronifera oceanica* (COOKSON & EISENACK, 1958) MAY, 1980.
Lame 4A36/11; 125/50; 71μ f. 3A86.

Partie supérieur des Couches de Tisaru, affleurement sur la chaussée Sud-Est de Lepşa.
Santonien-Campanien inférieur (Gs1-Gs3).

Fig. 2 — *Semioculopollis* sp.
Lame 4A161/9; 116,4/55,8; 28μ f. 3A86.

Fig. 3 — *Odontochitina operculata* (O. WETZEL, 1933) DEFLANDRE & COOKSON, 1955.
Lame 4A 159/4; 120/50,9; 261μ f. 3A86.

Figs. 5, 7 — *Isabelidinium* sp. ex gr. *I. cooksoni* (ALBERTI, 1959) LENTIN & WILLIAMS, 1977. Fig. 5, lame 4A161/12; 108,3/42; 88μ f. 3A86; Fig. 7, lame 4A161/20; 113/63; 88μ f. 3A85.

Fig. 6 — *Ginginodinium* sp.
Lame 4A161/7; 111/52,5; 65μ f. 3A86.

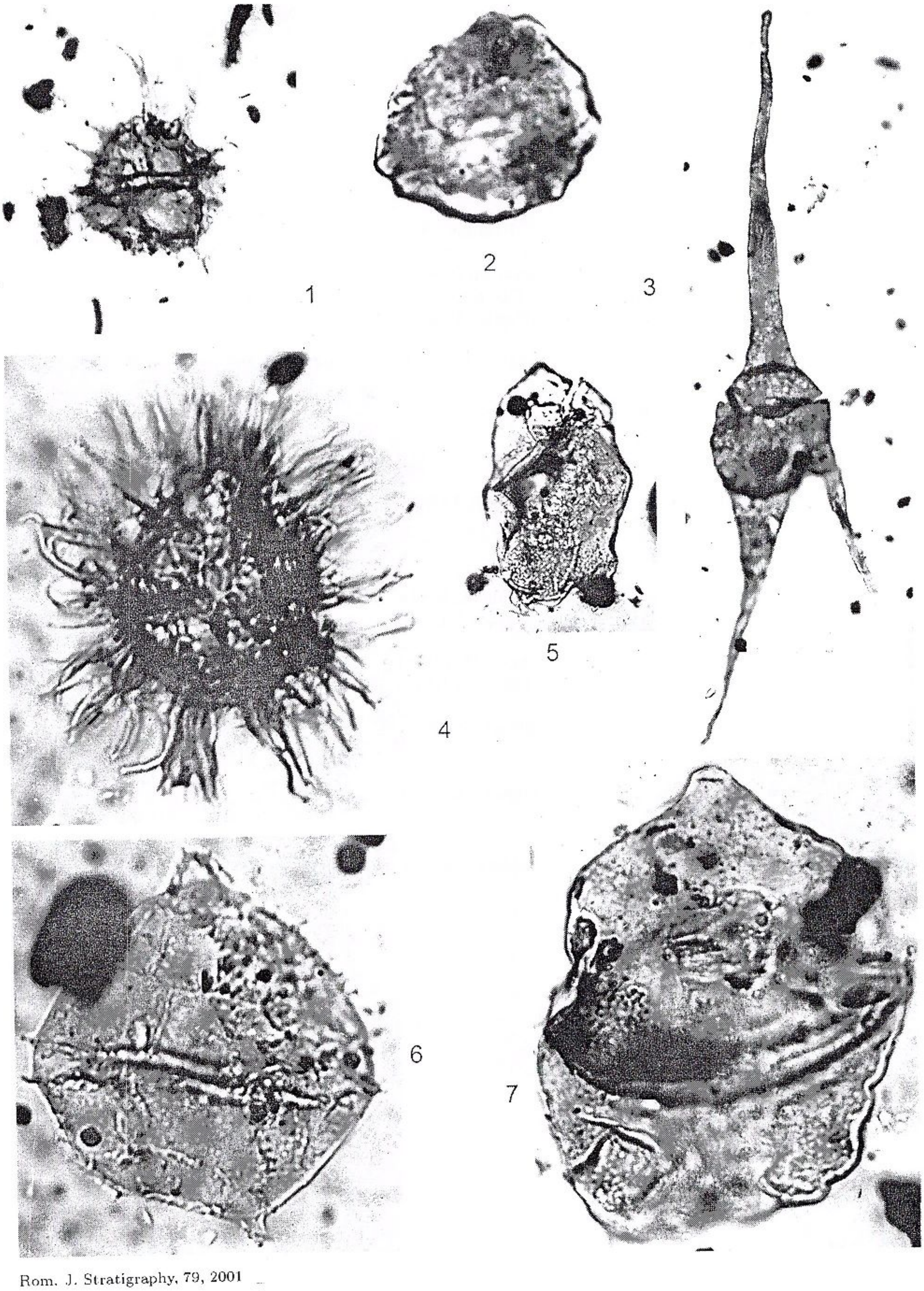
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Planche V

Eléments de la microflore des Couches de Caşin supérieures. Paléocène supérieur, zone à *Apectodinium homomorphum* (DEFLANDRE & COOKSON, 1955) LENTIN & WILLIAMS, 1977.
Vallée de Caşin - Vallée de Buciaş.

Fig. 1 — *Cerodinium speciosum* (ALBERTI, 1959) LENTIN & WILLIAMS, 1987.
Lame 4A126/4; 119/43,7; 138μ f. 3A83.

Fig. 2 — *Caligodinium* sp.
Lame 4A126/4; 105/54; 55μ f. 3A83.

Fig. 3 — *Apectodinium* cf. *paniculatum* (COSTA & DOWNIE, 1976) LENTIN & WILLIAMS, 1977, emend HARLAND, 1979.
Lame 4A128/7; 115,9/52,8; 80μ f. 3A82.

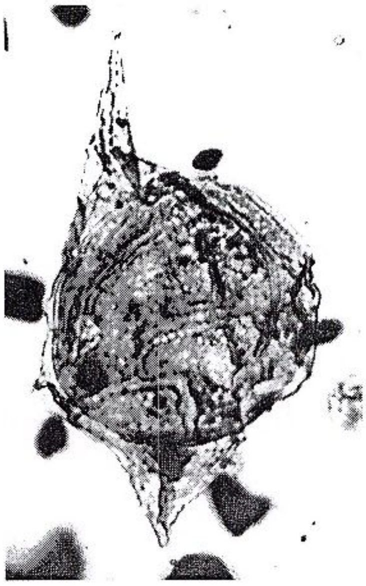
Eléments de la microflore des Couches de Piatra Uscată. Paléocène supérieur- Eocène inférieur.
Vallée du Caşin - Vallée de Buciaş.

Figs. 4-5 — *Apectodinium augustum* (HARLAND, 1979) LENTIN & WILLIAMS, 1981. Fig. 4, lame 4A137/8; 105/58,5; 158μ f. 3A81; Fig. 5, lame 4A141/3; 112/51; 128μ f. 3A83.

Fig. 6 — *Apectodinium parvum* (ALBERTI, 1961) LENTIN & WILLIAMS, 1977.
Lame 4A141/6; 116,7/52; 88μ f. 3A82.

Fig. 7 — *Apectodinium* cf. *homomorphum* (DEFLANDRE & COOKSON, 1955) LENTIN & WILLIAMS, 1977.
Lame 4A137/25; 122/51,3; 93μ f. 3A82.

Fig. 8 — *Apectodinium homomorphum* (DEFLANDRE & COOKSON, 1955) LENTIN & WILLIAMS, 1977.
Lame 4A137; 114,5/43,1; 85μ f. 3A85.



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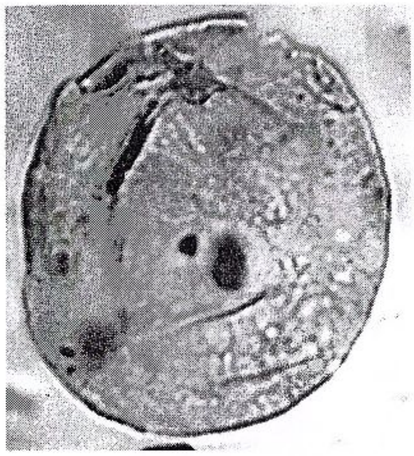

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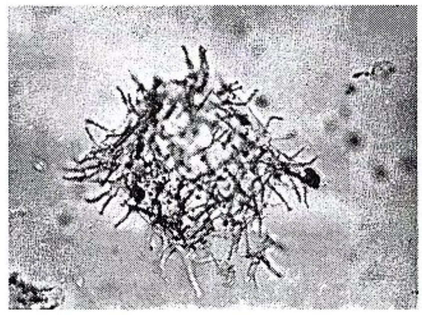

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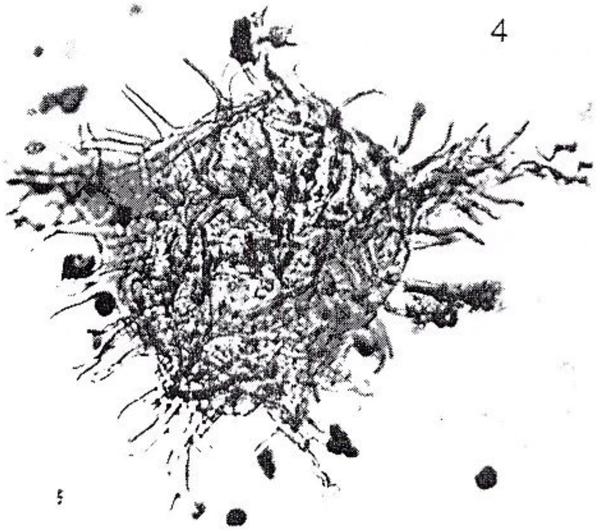

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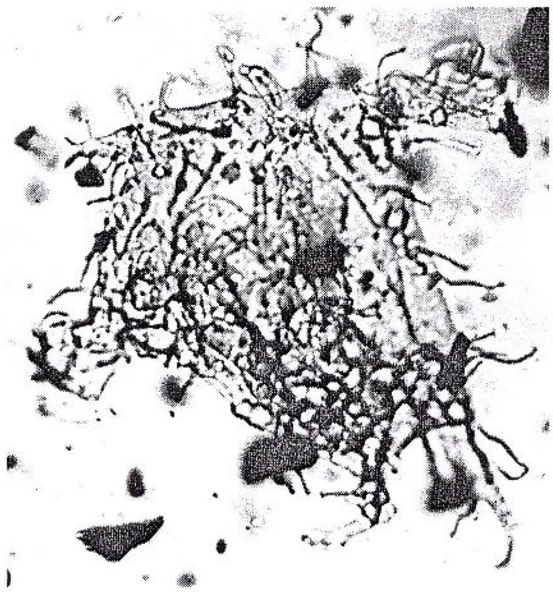

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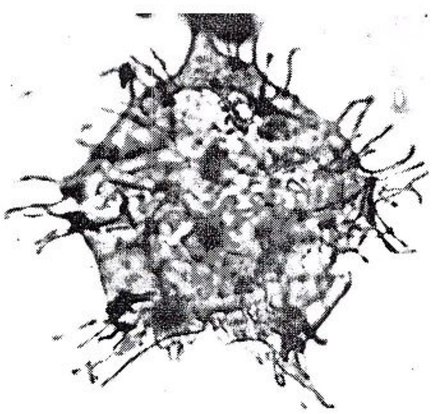

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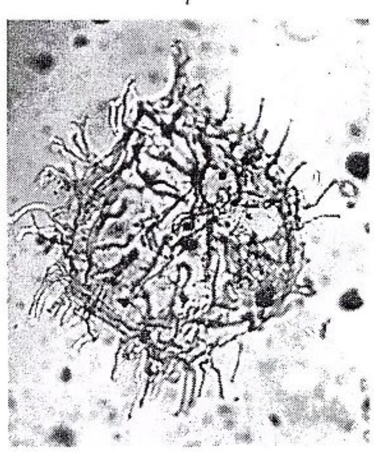

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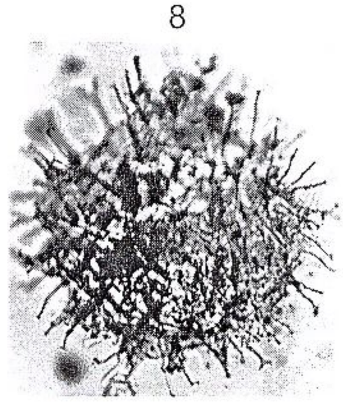

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Planche VI

Eléments de la microflore des couches de Piatra Uscată. Limite Paléocène/Eocène - Eocène inférieur? Zone à *Apectodinium augustum*. Vallée de Caşin - Vallée de Buciaş.

Figs. 1, 3 — *Apectodinium augustum* (HARLAND, 1979) LENTIN & WILLIAMS, 1977. Fig. 1, lame 4A137/1; 112/56,7; 153μ f. 3A86; Fig. 3, lame 4A137/3; 124,8/50; 150μ f. 3A80.

Fig. 2 — *Apectodinium* sp. cf. *A. hyperacanthum* (COOKSON & EISENACK, 1965) LENTIN & WILLIAMS, 1977.

Lame 4A137/21; 120,8/63,8; 118μ 3A80.

Fig. 4 — *Apectodinium quinquelatum* (WILLIAMS & DOWNIE, 1966) COSTA & DOWNIE, 1979.

Lame 4A137/4; 109/49,5; 100μ f. 3A80.

Fig. 5 — *Apectodinium hyperacanthum* (COOKSON & EISENACK, 1965) LENTIN & WILLIAMS, 1977.

Lame 4A144/1; 123,9/62,1; 100μ 3A83.

Fig. 6 — *Apectodinium parvum* (ALBERTI, 1961) LENTIN & WILLIAMS, 1977.

Lame 4A137/23; 124,5/51,5; 83μ f. 3A81.

Fig. 7 — *Apectodinium homomorphum* (DEFLANDRE & COOKSON, 1955) LENTIN & WILLIAMS, 1977.

Lame 4A144/4; 115/50,1; 100μ f. 3A83.

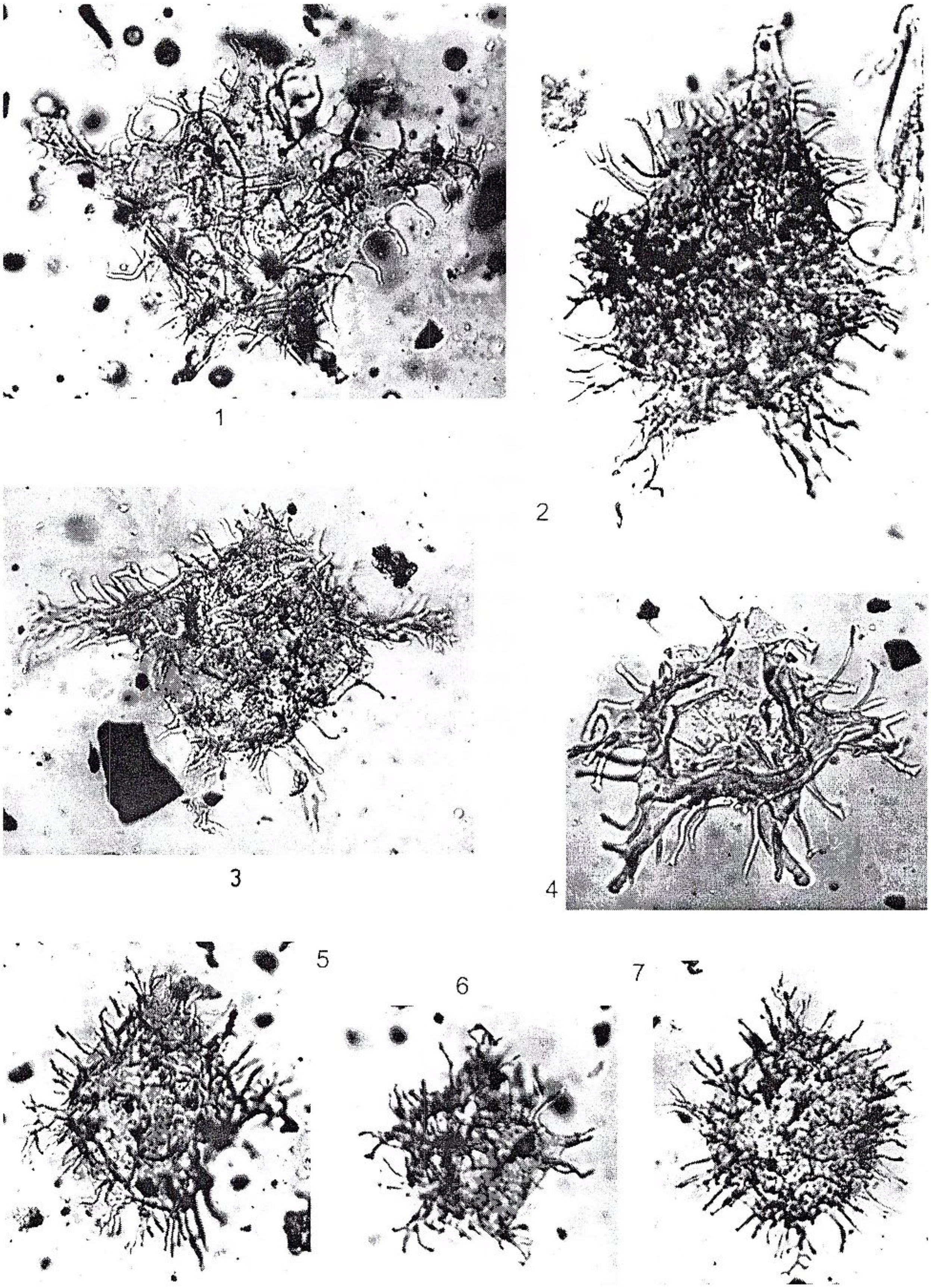
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Planche VII

Eléments de la microfaune du Crétace et du Paléocène.

Fig. 1 — *Hippocrepina depressa* VASICEK.

Fig. 2-3 — *Gaudryina oblonga* ZASPELOVA.

Fig. 4 — *Hormosina ovulum ovulum* (GRZYB.).

Figs. 5-6 — *Verneuilinioides neocomiensis* MJATLIUK.

Fig. 7 — *Glomospira diffundens* CUSHMAN & RENZ.

Fig. 8 — *Verneuilinioides subfiliformis* BARTENSTEIN.

Fig. 9 — *Haplophragmoides* sp.

Figs. 10-11 — *Pseudobolivina variabilis* (VASICEK).

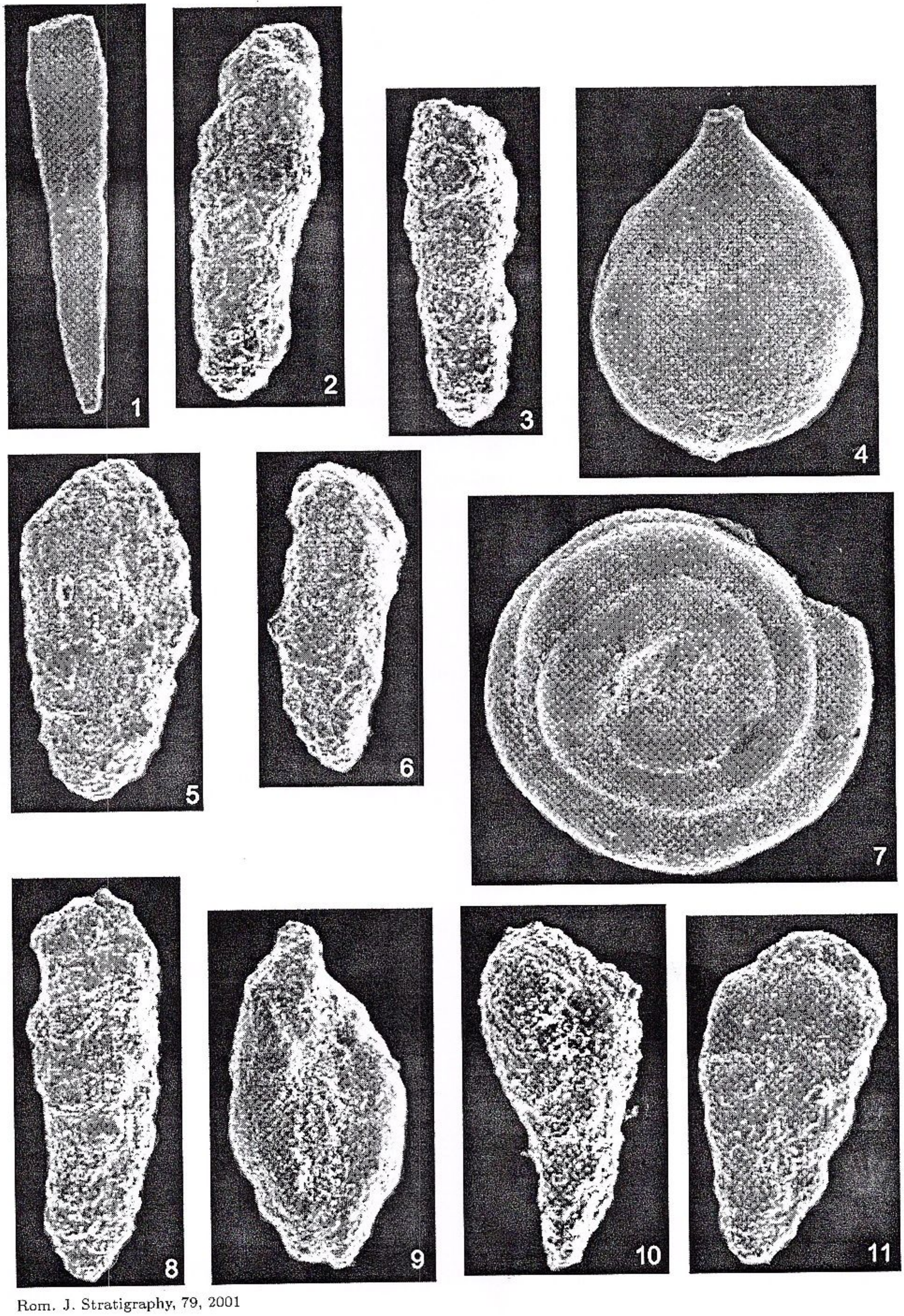
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Planche VIII

Eléments de la microfaune du Crétacè et du Paléocène.

Fig. 1 — *Pseudobolivina variabilis* (VASICEK).

Figs. 2, 5 — *Rzehakina epigona epigona* (RZEHAK).

Fig. 3 — *Reophax duplex* GRZYB..

Figs. 4, 6, 7 — *Reophax minutus* TAPPAN.

Fig. 8 — *Spiroplectammina spectabilis* (GRZYB.).

Fig. 9 — *Trochammina vocontiana* MOULLADE.

Fig. 10 — *Spiroplectammina subhaerigensis* (GRZYB.).

Fig. 11 — *Goesella rugosa* (*Hanzlikova*).

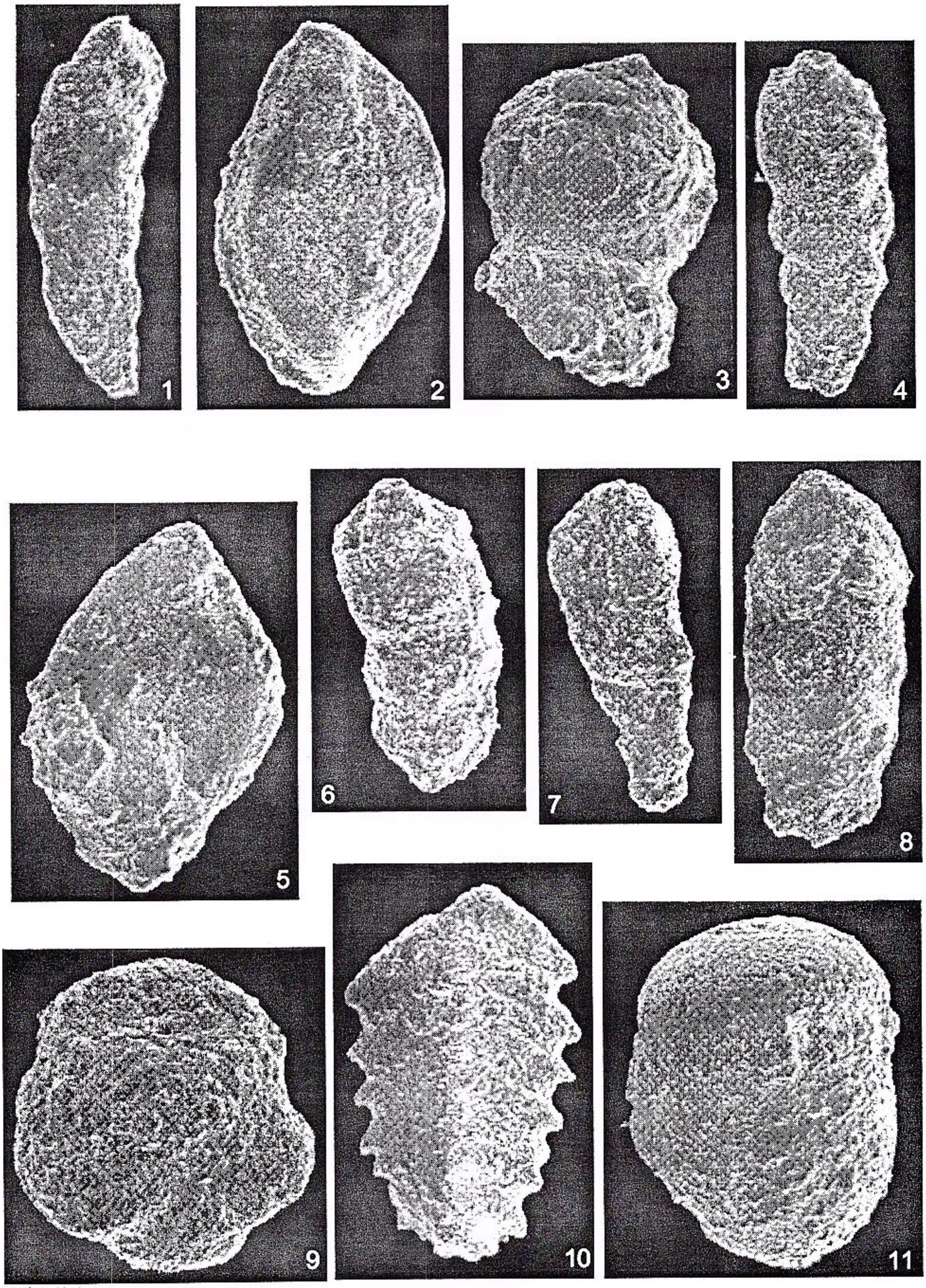
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Planche IX

Eléments de la microfaune du Crétacé et du Paléocène.

Figs. 1-4 — *Marginotruncana marianosi* DOUGLAS.

Fig. 5 — *Marginotruncana* sp.

Fig. 6 — *Thalmannammina meandertornata* NEAGU & TOCORJESCU.

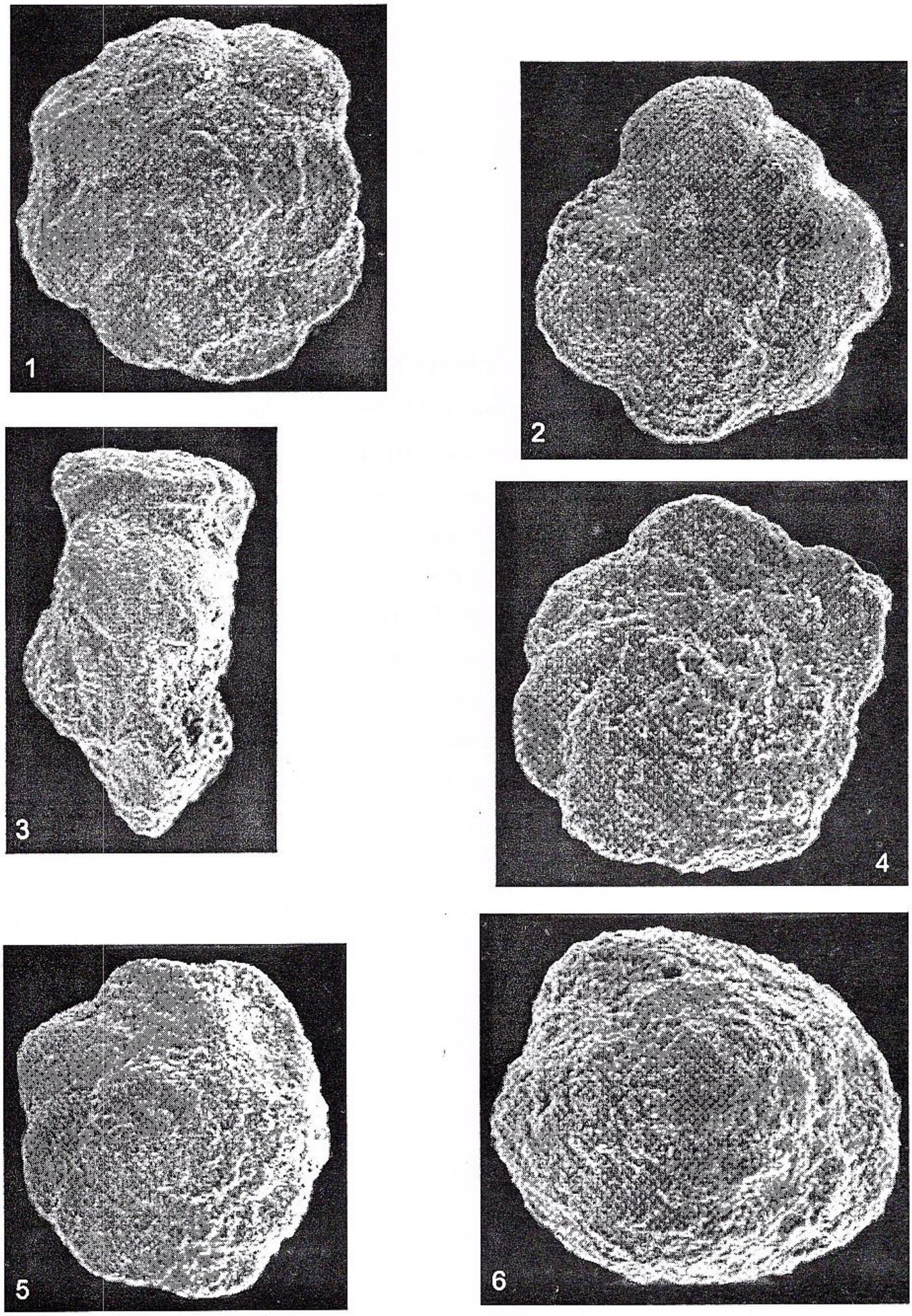
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Planche X

Eléments de la microfaune de l'Eocène et de l'Oligocène.

Fig. 1 — *Acarinina nitida* MARTIN, x 200. Couches de Buciaş.

Fig. 2 — *Globigerina prasaepsis* BOW, x 125. Couches de Buciaş.

Fig. 3 — *Globigerina linnaperta* EINLAG, x 140. Couches de Bisericani.

Fig. 4 — *Globigerina inaequispira* SUBBOTINA, x 150. Couches de Buciaş.

Fig. 5 — *Acarinina spinuloinflata* (BONDY), x 170. Couches de Buciaş.

Fig. 6 — *Acarinina bullbroki* BOLLI, x 150. Couches de Buciaş.

Fig. 7 — *Haplophragmoides walteri* (GRZYBOWSKI), x 125. Couches de Buciaş et Couches de Bisericani.

Fig. 8 — *Turborotalia cerroazullensis frontosa* SUBBOTINA, x 150. Couches de Buciaş.

Fig. 9 — *Globigerina ampliapertura* BOLLI, x 150. Couches de Bisericani, Menilithes.

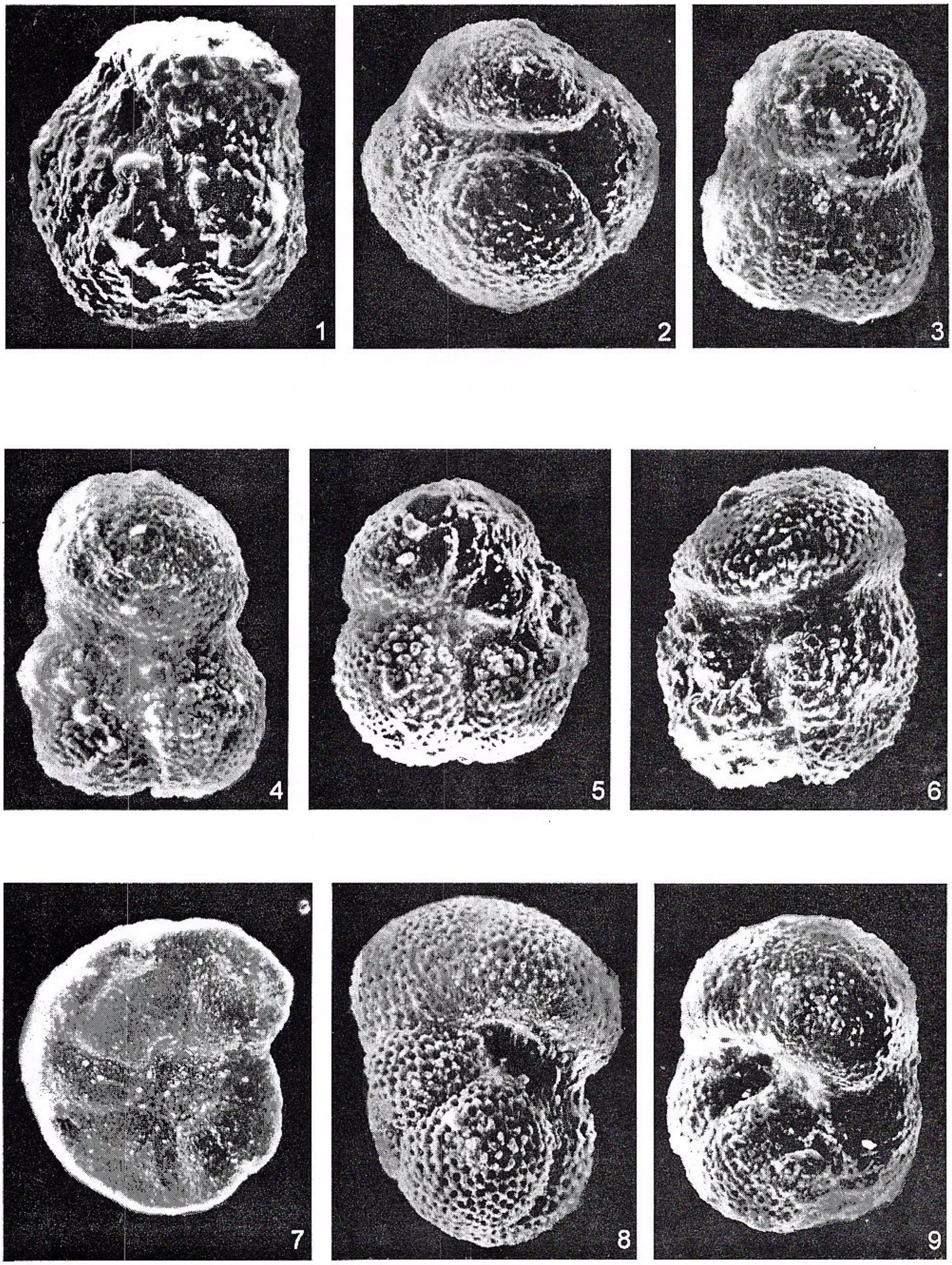

Planche XI

Eléments de la microflore des dépôts de l'Eocène inférieur et de l'Eocène moyene.

Fig. 1 — *Homotryblium tenuispinosum* DAVEY & WILLIAMS, 1966, Couches de Buciaş, 1F 581/1 (GT 15), 101,2/18,9, 102,5 μ.

Fig. 2 — *Impagidinium* sp., Couches de Buciaş, 1F 581/2 (GT 15), 116/24, 65 μ.

Figs. 3, 6 — *Cordosphaeridium gracilis* (EISENACK, 1954) DAVEY & WILLIAMS, 1966. Fig. 3, Couches de Buciaş, 1F 581/2 (GT 15), 102,9/17,0, 90 u; Fig. 6, Couches de Caşin, 1F 669/5 (GT 4), 120,5/8,1, 95 μ.

Fig. 4 — *Systematophora placantha* (DEFLANDRE & COOCKSON, 1955) DAVEY, DOWNIE, SARJEANT & WILLIAMS, 1969, Couches de Caşin, 1F 669/5 (GT 4), 105,0/5,0, 75 μ.

Fig. 5 — *Deflandrea eocenica* BALTES, 1969, Couches de Buciaş, 1F 574/4 (GT 8), 108,1/24,1, 100 μ.

Fig. 7 — *Wilsonidium* sp., Couches de Buciaş, 1F 581 (GT 15), 101,9/26,2, 115 μ.

Fig. 8 — *Cordosphaeridium latispinosum* DAVEY & WILLIAMS, 1966, Couches de Bisericani, 1F 665/3 (GT 29), 112,1/22,0, 80 μ.

Fig. 9 — *Spiniferites cornutus* (GERLACH) SARJEANT, 1970, Couches de Bisericani, 1F 655/3 (GT 39), 104,3/20, 117,5 μ.

Fig. 10 — *Deflandrea speciosa* (ALBERTI, 1959) LENTIN & WILLIAMS, 1977, Couches de Caşin, 1F 669 (G), 111,9/23,5, 80 μ.

Fig. 11 — *Diphyes colligerum* (DEFLANDRE COOCKSON) COOCKSON, 1965, Couches de Buciaş, 1F 581 (GT 15), 114,0/7,9, 110 μ.

Fig. 12 — *Duosphaeridium nudum* (COOCKSON) LOEBLICH & LOEBLICH, Couches de Caşin, 1F 670/1 (GT 5), 100,1/14,1, 92,5 μ.

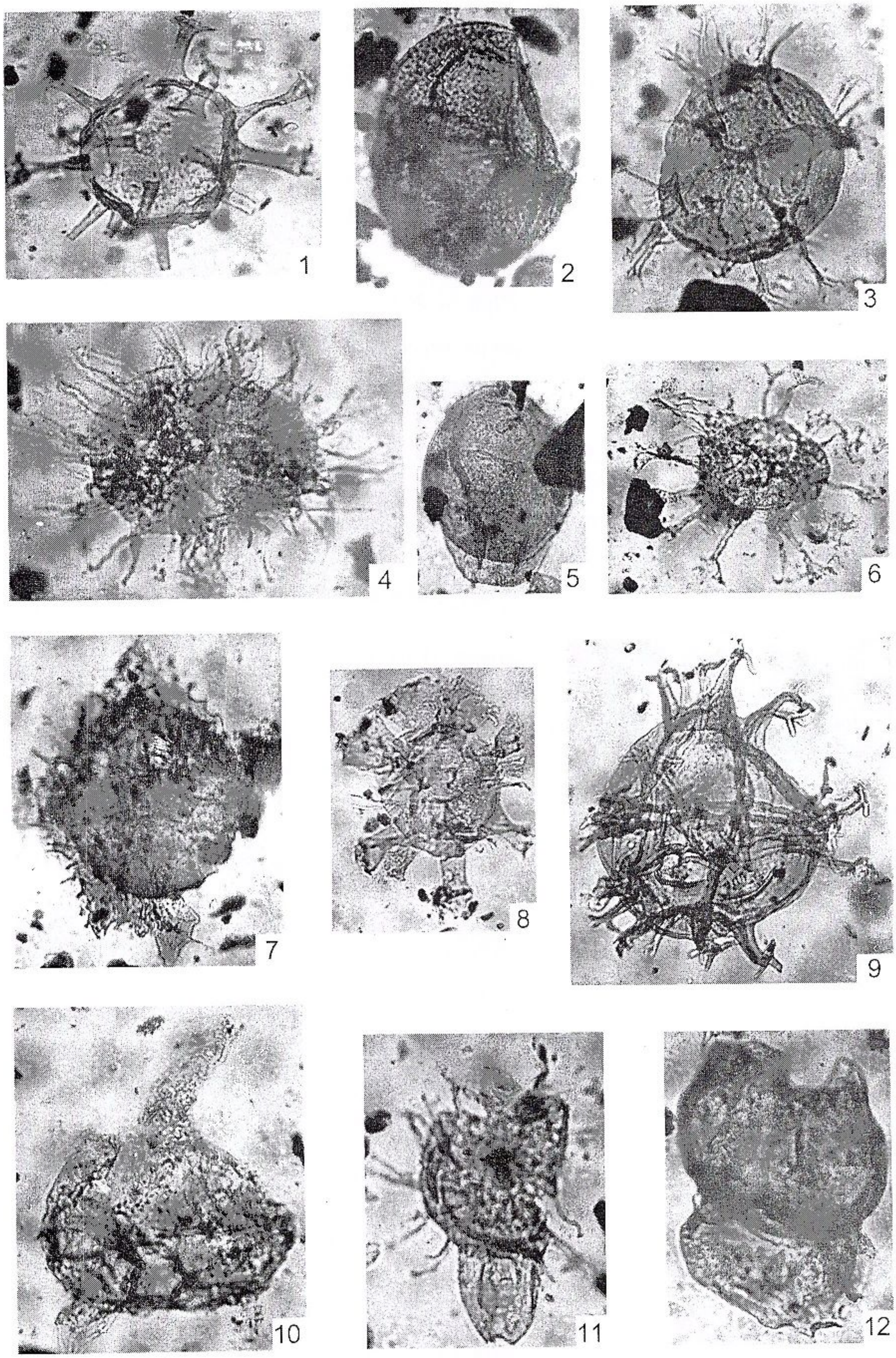
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Planche XII

Eléments de la microflore des dépôts de l'Eocène supérieur.

Fig. 1 — *Charlesdowniea fasciata* LENTIN & VOZZHENNIKOVA, 1989, Couches de Bisericani, 1F 714/3 (B 39), 112,1/12,1, 125 μ.

Fig. 2 — *Deflandrea leptodermata* COOCKSON & EISENACK, 1965, Marnes à Globigérines, 1F 635/2 (B 17), 115,0/16,4, 124 μ.

Fig. 3 — *Oligosphaeridium complex* (WHITE, 1842) DAVEY & WILLIAMS, 1966, Couches de Bisericani, 1F 714/3 (B 39), 109,6/17,0, 105 μ.

Fig. 4 — *Wetzeliella symmetrica* WEILLER, 1956, Marnes à Globigérines, 1F 641,0/8,0 (B 23), 120,2/11,1, 130 μ.

Fig. 5 — *Operculodinium* sp., Marnes à Globigérines, 1F 641,0/12,0 (B 23), 105,0/21,0, 120 μ.

Fig. 6 — *Wilsonidium baltesii* n. sp., Couches de Bisericani, 1F 725,0/1,0, 107,6/3,9, 150 μ.

Fig. 7 — *Areosphaeridium diktyoplokus* (KLUMPP, 1953) EATON, 1971, Marnes à Globigérines, 1F 645,0/4,0 (B 27), 103,2/25,0, 135 μ.

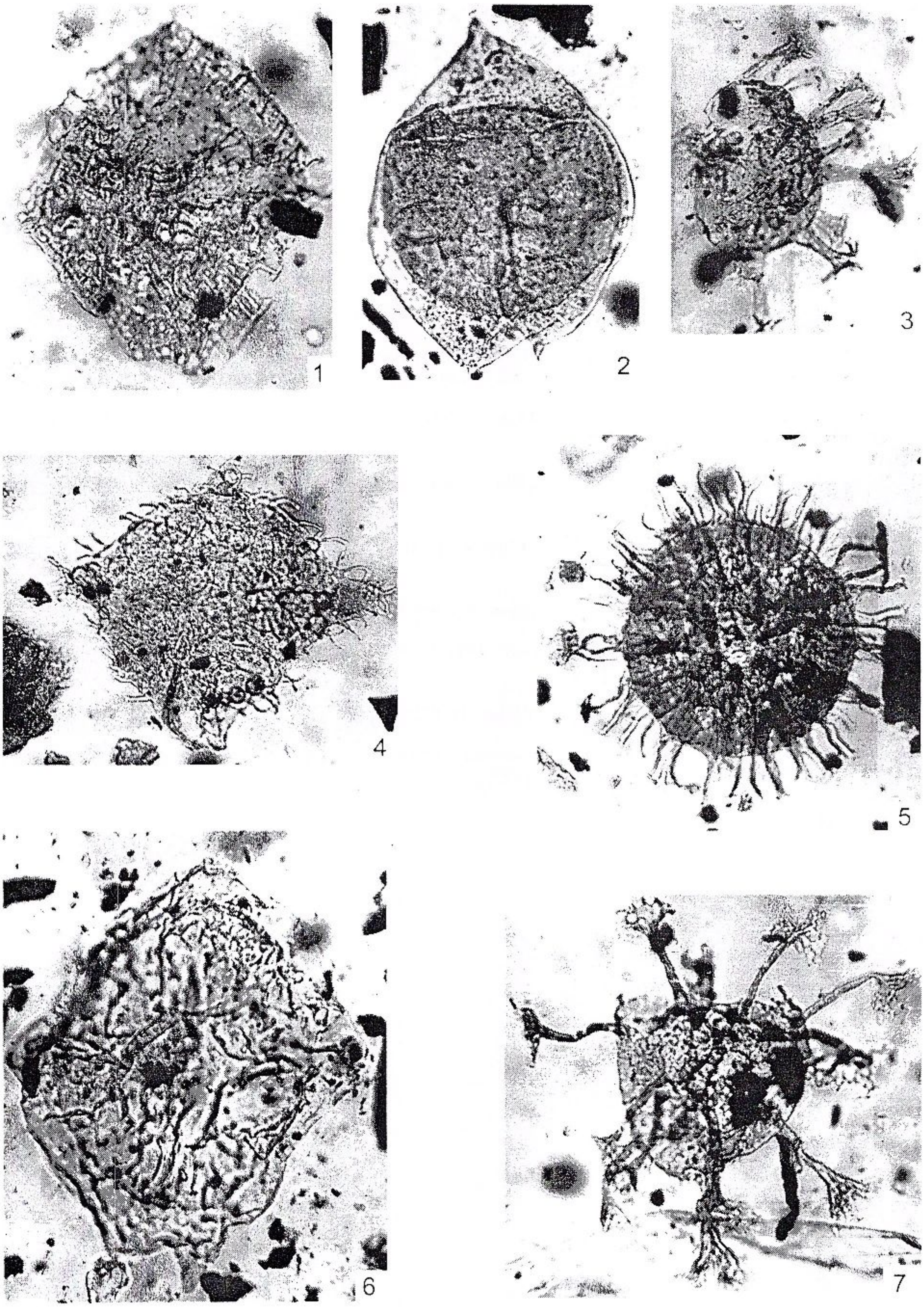
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Planche XIII

Eléments de la microflore des dépôts de l'Eocène supérieur et de l'Oligocène inférieur.

Fig. 1 — *Lejeunecysta hyalina* (GERLACH, 1961) emend. KJELSTROM, 1972, ARTZNER & DORHOFER, 1978, Schistes ardesiformes, 1F 712/2 (MO 8), 107,0/23,0, 85 μ.

Fig. 2 — *Hystrichokolpoma eisenacki*, WILLIAMS & DOWNIE, 1966, Couches de Bisericani, 1F 715,0/2,0 (B 40), 99,1/17,1, 67,5 μ.

Fig. 3 — *Deflandrea granulosa* MENENDEZ, 1965, Couches de Bisericani, 1F 616/6 (B 31), 111,1/4,9, 135 μ.

Fig. 4 — *Thalassiphora pelagica* (EISENACK, 1954) EISENACK & GOCHT, 1960, Couches de Bisericani, 1F 620,0/1,0 (B 35), 104,5/20,9, 132,5 μ.

Fig. 5 — *Wilsonidium antonescui* n. sp., Couches de Bisericani, 1F 725/4 (B 50), 102,1/9,1, 120 μ.

Fig. 6 — *Areosphaeridium multicornutum* EATON, 1971, Couches de Bisericani, 1F 622/2 (B 37), 102,0/6,8, 55 μ.

Fig. 7 — *Glaphyrocysta* sp., Marnes à Globigérines, 1F 632/7 (B 14), 104,2/12,1, 75 μ.

Fig. 8 — *Charlesdowniea clathrata* subsp. *angulosa* LENTIN & VOZZHENNIKOVA, 1989, Marnes à Globigérines, 1F 632/10 (B 10), 120,9/23,0, 162,5 μ.

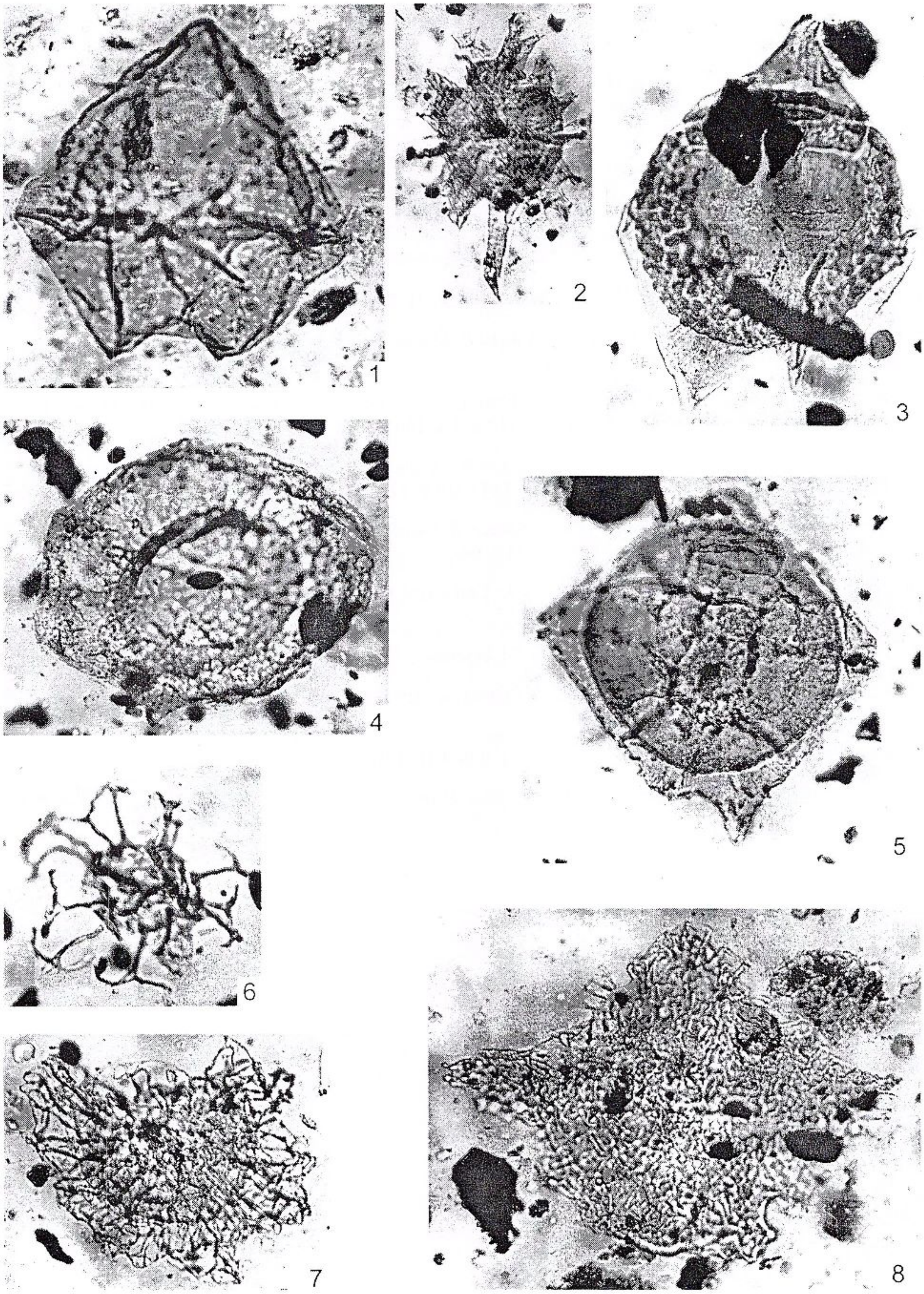
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Planche XIV

Eléments de la microflore des dépôts l'Eocène supérieur/l'Oligocène inférieur.

Fig. 1 — *Impletosphaeridium* sp., Couches de Bisericani, 1F 655/3 (GT 29), 114,9/23,8, 100 μ.

Figs. 2, 3, 6 — *Boehlensipollis hohli* KRUTZSCH, 1962, Marnes á Globigérines, 1F 635/1 (GT 17), 105,2/5,7, 30 u; 110,0/10,9, 27,5 u; 120,0/8,9, 30 μ.

Fig. 4 — *Rhombodinium perforatum* (JAN DU CHENE & CHATEUNEUF, 1975) LENTIN, WILLIAMS, 1977, Couches de Bisericani, 1F 654/1 (GT 28), 118,2/4,3, 100 μ.

Fig. 5, 8 — *Phthanoperidinium amoenum,* DRUGG & LOEBLICH, 1967, Marnes à Globigérines, 1F 635/2 (B 17), 1052/13,5, 110 u; 1F 635/8 (B 17), 114,0/10,9, 37,5 μ.

Fig. 9 — *Systematophora placacantha* (DEFLANDREA & COOKSON, 1955) DAVEY, DOWNIE, SARJEANT & WILLIAMS, 1969, 1F 635/5 (B 17), 105,0/23,1, 50 μ.

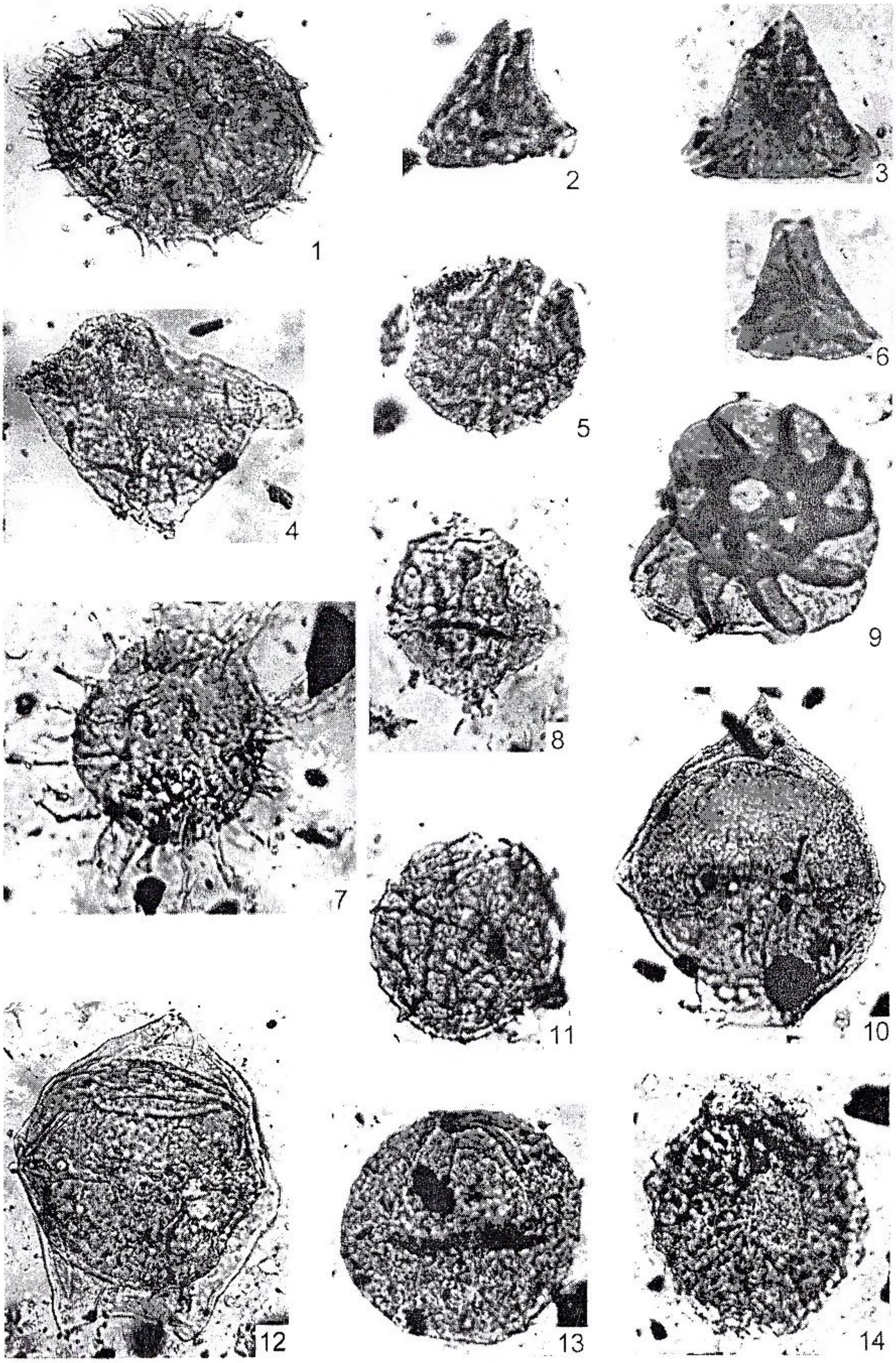
Fig. 10 — *Deflandrea leptodermata* COOKSON & EISENACK, 1965, Marnes à Globigérines, 1F 635/4 (B 17), 114,9/13,1, 45 μ.

Fig. 11 - cf. *Phthanoperidinium* sp., Marnes à Globigérines, 1F 635/2 (B 17), 99,0/25,2, 112,5 μ.

Fig. 12 — *Deflandrea granulosa* COOKSON & EISENACK, 1965, Marnes à Globigérines, 1F 635/1 (B 17), 99,0/25,2, 112,5 μ.

Fig. 13 - Non identifié, Marnes à Globigérines, 1F 635/1 (B 17), 107,8/7,2, 80 μ.

Fig. 14 — *Samlandia clamydophora* EISENACK, 1954, Marnes à Globigérines, 1F 635/1 (B 17), 98,6/25,1, 75 μ.



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INFLUENCE OF THE NEOGENE TECTONIC ON THE ENTRAPMENT OF THE OIL RESERVES IN THE ZAGROS OROGENIC BELT OF IRAN

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Key words: Neogene. Tectogenesis. Oil reserves. Thermal modelling. Zagros. Iran

Abstract: The relative chronology of the orogenic events related to the closing of the South Tethys, during the Neogene, significantly influenced the generation, migration and entrapment of petroleum in one of the world's richest province, the Dezful Embayment of the Zagros orogenic belt of Iran, which contains some 8 % of the global reserves in an area of only 40,000 sq km. Oil and associated gas are trapped in large-sized whaleback anticlines, which resulted from the main Zagros compressive phase. This phase began towards the end of the Middle Miocene at around 10 Ma and continued throughout the Late Miocene and Pliocene. Oil and gas accumulated in two calcareous reservoirs, the Sarvak Fm of Cenomanian/Turonian age and the extremely prolific Asmari Fm. of the Early Miocene, well protected by the evaporites of the Gachsaran Fm. Three excellent source rocks, i.e. the Neocomian part of the Garau Fm, the Kazhdumi of Albian age and the Padbeh of the Middle Eocene to Early Oligocene, were deposited in intracratonic depressions located in the Arabian Platform, when anoxic conditions prevailed. The Arabian Platform remained stable, with a low subsidence during the Late Cretaceous and part of the Cenozoic, up to the Early Miocene, circa 30 Ma. The subsidence was slightly higher during the deposition of the Asmari, between 30 and 20 Ma, then a sudden increase, between 20 and 3 Ma, corresponded to the deposition of the Gachsaran evaporites, the Mishan Fm., and the extremely thick, up to 4000 m, molasse-type sediments of the Agha Jari Fm. Modelling, using burial profiles and heat flow assumptions based upon maturity indices, such as the actual geothermal gradient, the vitrinite reflectance and rock- eval parameters, showed that the Kahdumi and the Pabdeh source rocks attained the onset of the oil expulsion window, during the deposition of the Agha Jari Fm between 8 and 3 Ma. According to this chronology, oil was expelled from source rocks recently when the Zagros structures already existed. Therefore, oil migrated over short distances to the next trap within an already well-characterized system of drainage areas, the geometry of which could be deduced from seismic data. Moreover, the Zagros folding induced an important fracturing observed at the outcrops and in the wells. This fracturing, which affects the limestones as well as the marls, enhanced a subvertical transfer of generated hydrocarbons towards the reservoirs, and especially towards the well capped Asmari Limestone, in which 75 % of the oil accumulated. Due to such short distances of migration, the oils could be directly related to the source rocks which generated them. These relations are confirmed by oil-to-source rock correlations based upon stable isotopes (^{13}C, ^{34}S) and biomarkers. Modelling applied for each drainage area provides estimates of the amount of the oil generated by each source rock. The calculated estimate was then compared to the oil-in-place really accumulated in the corresponding field. Therefore, the use of such a modelling technique, which could provide before drilling an estimate of the amount of the oil to be found, was only possible because the Zagros structures already existed at the time of the oil migration.

Introduction

In the Zagros Foldbelt of Iran, a small-sized depressed area, the 40 000 Sq km Dezful Embayment, part of the Khuzestan province, corresponds to an impressive gathering of 45 oilfields which contain more than 330 billions barrels of oil in place, accounting for 8 % of the global reserves (Figs. 1 and 2).

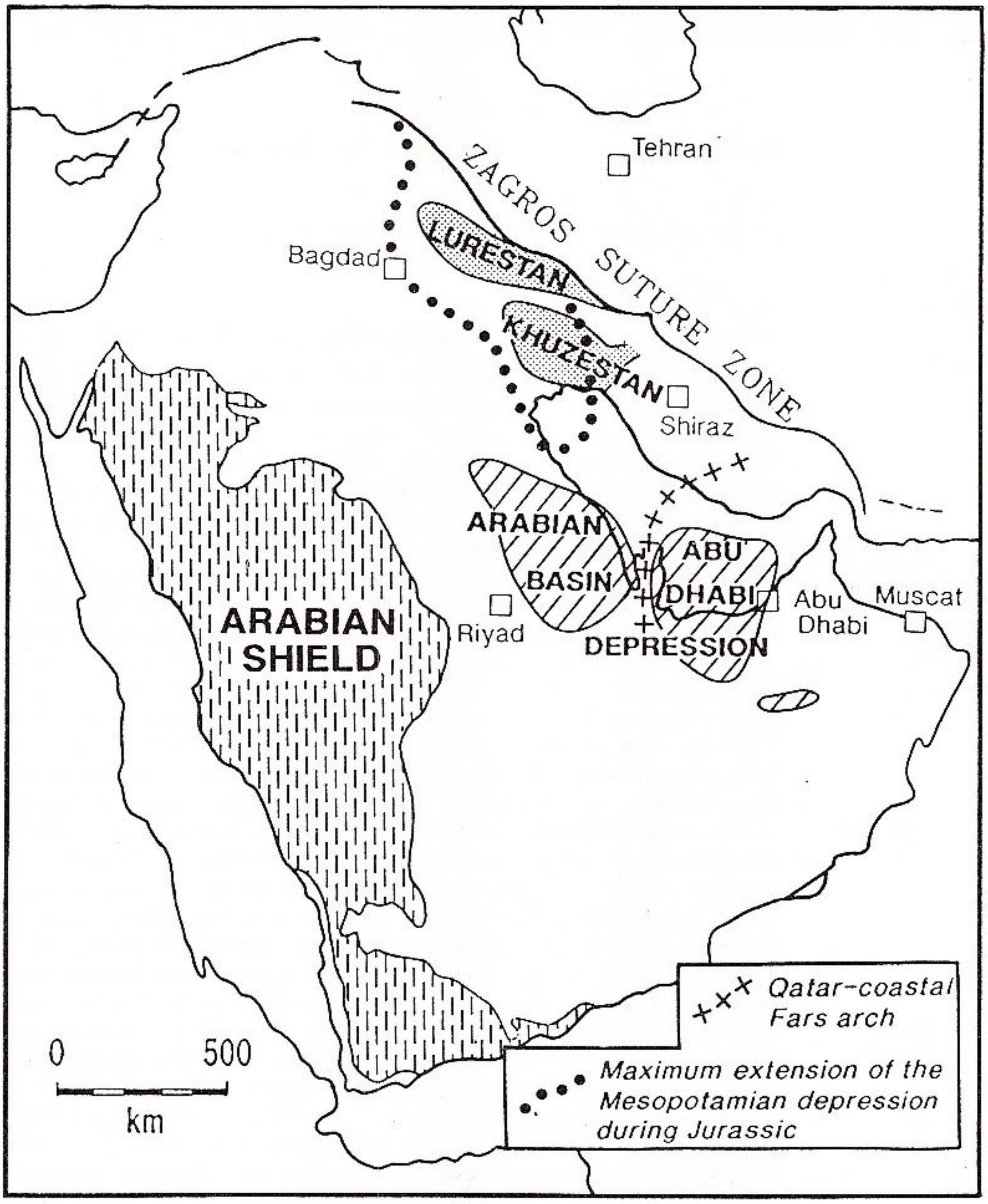


Fig. 1 – Middle East sketch map

The coexistence of favorable factors, such as extremely rich source rocks, excellent reservoirs, a widespread efficient evaporitic cap rock, large-sized anticlines and rock fracturing, resulting from the Neogene folding, explained in part this prolific oil province. Another reason which is developed in this paper is **the relative chronology of the oil expulsion from the source rocks as compared to those of the formation of the huge Zagros anticlines.** The two aspects of this relative chronology are discussed, i.e. the history of the Zagros orogenic phases, mainly during the Neogene, and the timing of oil generation, expulsion and migration which is related to the subsidence history of the study area. The timing of the oil expulsion was estimated by thermal modelling using the Arrhenius equation and the GENEX program designed by Institut Français du Pétrole (IFP) and BEICIP.

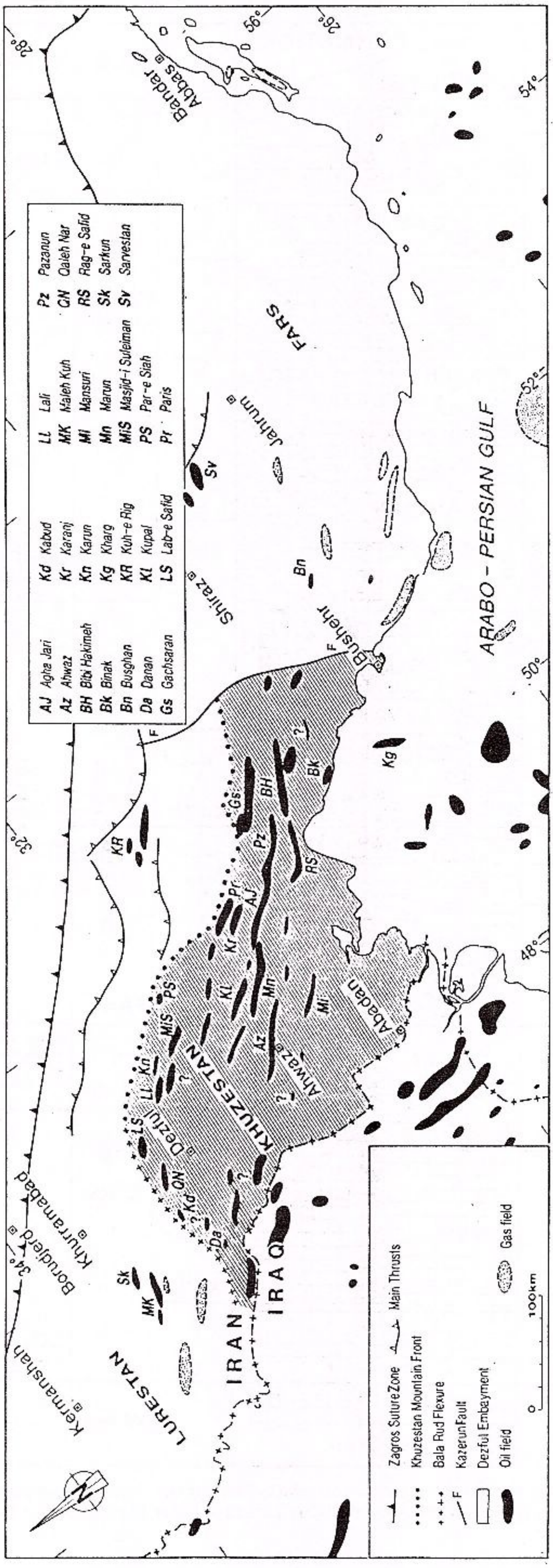


Fig. 2 – Location map of the Zagros Foothills, Dezful Embayment and northeastern part of the Arabo-Persian Gulf.

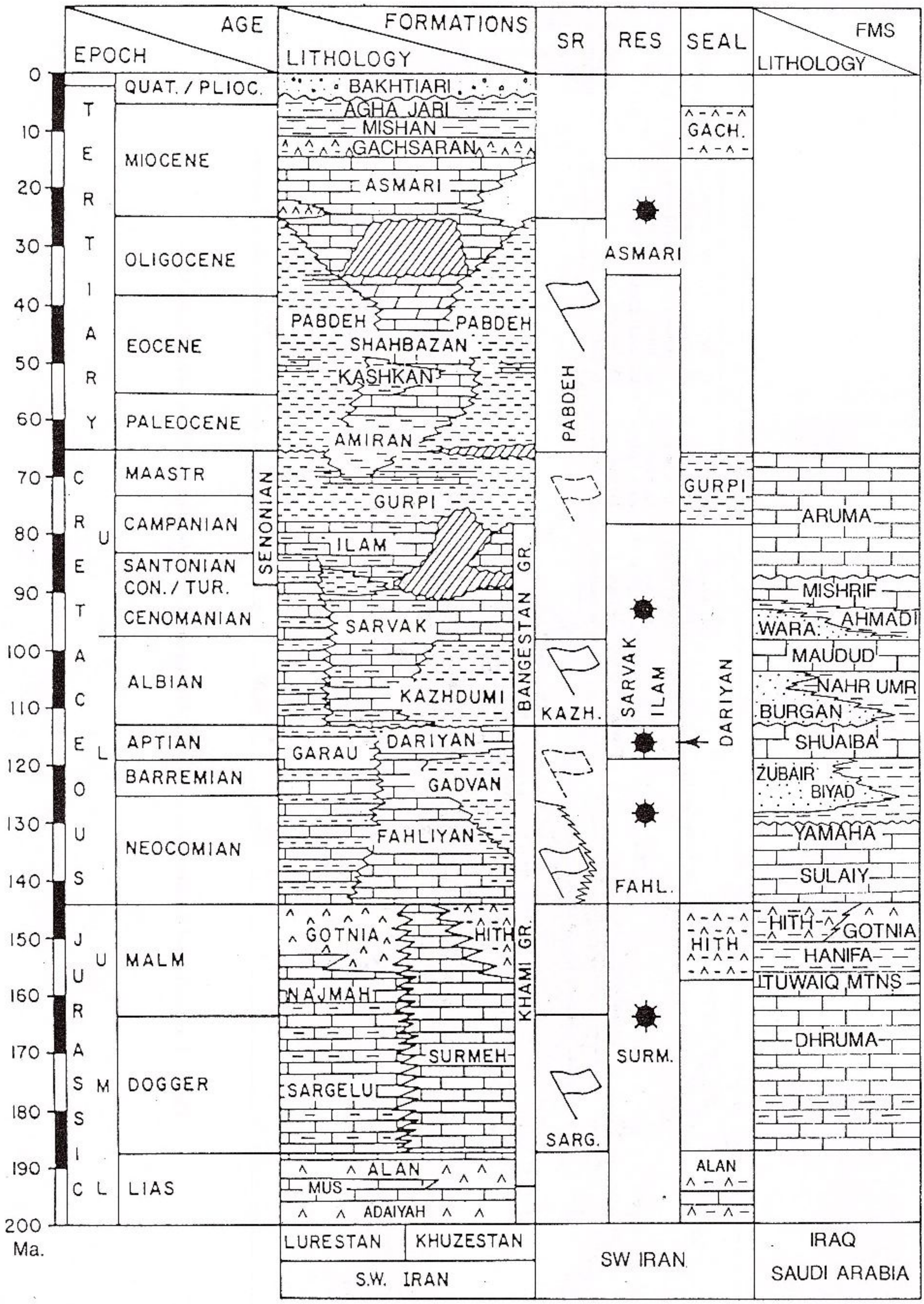


Fig. 3 – Schematic stratigraphy and source-reservoir-seal relationship for the Dezful Embayment and neighbouring areas. Main source rocks (SR), reservoirs (Res) and seals are indicated (after Bordenave and Burwood, 1995).

ZAGROS FOOT HILLS - CANDIDATE SOURCES

AGE	FORMATION	LITHOLOGY	MAXIMUM EFFECTIVE SOURCE THICKNESS (M)	SOURCE ROCK QUALITY			SOURCE SIGNATURE KEROGEN ◇/ KEROGEN PYROLYSATE ◆ $\delta^{13}C$ (ppt)
				KEROGEN TYPE	TOC (% WT)	S2 (KG / TN)	
L.TERT.	PABDEH	LST. ARG	230 +	II_S	1.0-12.0	7 - 40	
LOWER - CRETACEOUS - UPPER	GURPI	LST.V . ARG	150 (N.FARS)	II / III	0.5-2.0	2 - 10	
	SURGAH/ LAFFAN	SH. CALC.	150 (S.LURESTAN)	III	1.5-3.0	3 - 8	
	KAZHDUMI	SH.V. ARG LST. ARG	300 + (DEZFUL)	II_S	3.1-12.0	17 - 40	
	GADVAN	LST. ARG	100 (N.W. FARS)	II_S	1.0	5	
	GARAU	LST.V . ARG	300 + (LURESTAN)	II_S	1.5-10.0	ND	
M. JUR.	SARGELU	LST. ARG	150 + (LURESTAN)	II_S	3.1-4.4	18	
L.PALZ	SILURIAN	SH.	70 + (S.FARS)	II	2.5-4.3	ND	-30.8

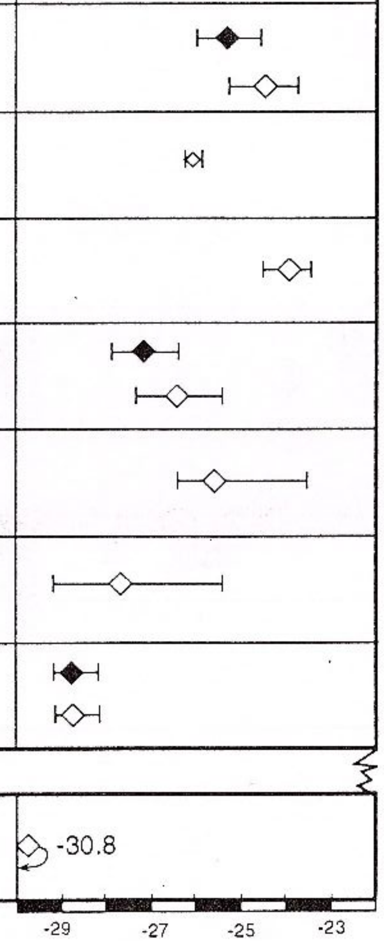


Fig. 4 – Iranian source rock characteristics (after Bordenave and Burwood, 1990).

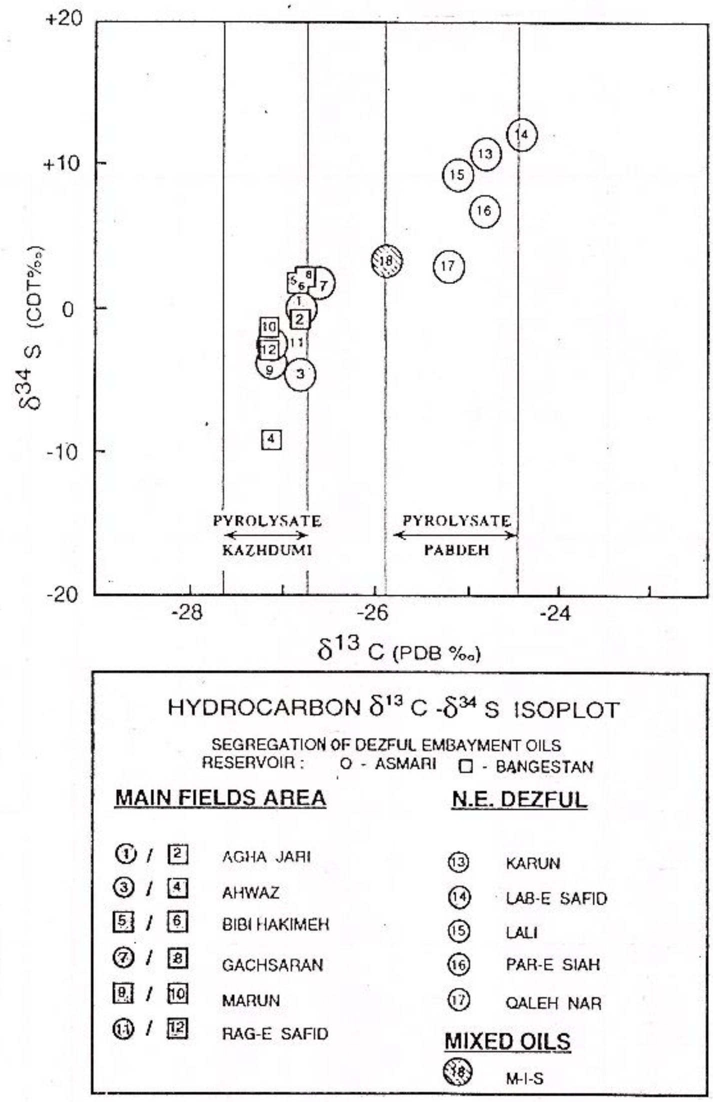


Fig. 5 – Carbon and Sulfur isotope cross-plot of Dezful Embayment oils (after Bordenave and Burwood, 1990).

Geological setting

The stratigraphic column is summarized in Figure 3. Sedimentation remained dominantly calcareous, consisting of areobic high-energy limestones, with breaks of sedimentation on the highs and low-energy argillaceous limestones and marls in the depressions (James and Wynd, 1965, Murris, 1980). The calcareous sedimentation was temporarily interrupted, either by evaporitic episodes, or by sudden influxes of clastics. Evaporites developed as a result of arid climate episodes during the Late Proterozoic with the deposition of the thick Hormuz salt, during the Triassic (Dashtak Fm), at the end of the Jurassic (Gotnia and Hith Fms), and at the end of the Early Miocene (Gachsaran Fm). Large amounts of clastics, resulting from the erosion of the Arabian Shield, likely after uplift phases and under humid climates, invaded the shallow marine habitat during the Rhaetian, Middle to Late Barremian (Zubair Sands), Albian (Burgan Sands), Cenomanian (Wara Fm) and Early Miocene (Ahwaz Sands).

Excellent source rocks were deposited during Silurian, Middle Jurassic, Neocomian, Albian, Early Cenomanian and Middle-Late Eocene. Several papers have already described these source rocks, and in particular their depositional environment, areal extent, thickness and geochemical characteristics. The prolific **Kazdhumi source rocks** of Albian age were described in some detail (Bordenave and Burwood, 1995). Another paper was devoted to the Cretaceous source rocks including, in addition to the Kazhdumi, the **Garau Fm** of the Neocomian and two layers which were found to have only a marginal potential, namely the Gadvan Formation of Barremian and the Gurpi marls of Senonian (Bordenave and Huc, 1995). An overall picture of the source rocks, also including the rich **Silurian Shales**, the **Sargelu Fm** of the Middle Jurassic and the **Pabdeh Fm** of the Middle Eocene to Early Oligocene was given in another paper (Bordenave, in press).

These source rocks were deposited in intracratonic depressions when anoxic conditions prevailed as a result of water stratification, in high-stand waters, associated to either humid climates or oceanic upwellings.

With the exception of the Silurian shales, deposited in a polar environment and related to a marked transgression resulting from the melting of Ordovician glaciers, the other source rocks were deposited in tropical or equatorial dominantly calcareous environments. They contain at least 70 % carbonates and should be considered as marls or even as argillaceous limestones. All these source rocks have excellent characteristics with TOC values reaching 4 and 12 % and Hydrogen Indices in the 650 g HC/kg C range. They are generally thicker than 250 m in the center of the depressions (Garau, Kazhdumi and Pabdeh). While the Silurian shales contain little sulfur, Jurassic and younger source rocks, deposited in euxinic environments, contain up to 7 to 9 % organic sulfur.

Intracratonic depressions, in which Jurassic and Cretaceous source rocks were deposited, were surrounded by shallow platforms and carbonate shoals. Their communication with the Neo-Thethys was hampered by the existence of shallow water sills. On the contrary, the Pabdeh source rocks were deposited in a NE-SW elongated trough parallel to the Zagros suture. Anoxic conditions during the Eocene were probably related to upwelling currents in the Western Mediterranean Sea, as suggested by the phosphatic and cherty character of the bituminous argillaceous carbonates deposited in Jordan from the Maestrichtian to the Eocene (Beydoun et al., 1974).

The Silurian source rocks caused the accumulation, before the Zagros folding, of huge quantities of gas in the Permian limestones capped by Triassic evaporites on **early regional highs** and salt-related structures, emplaced during Cenomanian. The influence of the Zagros folding was limited to the relocation of the gas in newly formed anticlines such as Aghar, Dalan, Kangan and Nar in Fars, and Samand and Kabir Kuh in Lurestan. The Paleozoic Petroleum System, related to the Silurian source rocks, is not studied here (see Bordenave, in press). By contrast, **the generation of oil from Cretaceous and Tertiary source rocks, its expulsion, migration and entrapment resulted from Neogene events.**

The Cretaceous Tertiary Petroleum complex in the Dezful Embayment

Three reservoirs were potential candidates to accommodate oil and gas accumulations in the Dezful Embayment: the extremely prolific **Asmari limestone** of Early Miocene which contains about 75 % of Iran onshore oil reserves, the Bangestan Group, which includes the Sarvak Formation of Cenomano-Turonian and the Ilam Formation (Santonian), accounts for 23 % of the reserves, and the **Khami Group** of Neocomian to Aptian age which would contain only 2 % of the total (Fig. 3).

The 250 to 500 m thick Asmari reservoir is a high energy limestone rich in large foraminifers (Nummulites, Neoalveolins, Myogypsins...) which retains excellent reservoir characteristics over most of the study area. Its reservoir quality is generally enhanced, close to the tops of anticlines, by a prominent system of fracturing. In SW Khuzestan, the basal part of the Asmari becomes sandy (Ahwaz Sandstone Member), increasing its porosity even more. The Asmari is capped by thick evaporites of the Gachsaran Formation which provide an efficient seal.

The Bangestan Group includes the very thick (300 to 1000 m) Sarvak limestone and the thinner Ilam (50 to 200 m). The two reservoirs are separated in Lurestan by the Surgah marls. The Bangestan is known under two facies: a high energy neritic facies with massive limestones, rich in gastropods, pelecypods, large forams and rudist debris, which has often fair reservoir characteristics, and a basinal equivalent, tight, micritic, often argillaceous limestone, known as the "Oligostegina facies". Basinal facies extend over Central Lurestan and in NW-SE trough in Khuzestan, from MIS to Karanj. The Pabdeh and Gurpi marls are most of the time, heavily fractured, as observed at surface outcrops as well as in well cores. Therefore, the Asmari and Bangestan are often connected.

The Khami Group includes two high energy limestones of the Fahliyan Formation of Neocomian and the Dariyan of Aptian separated by the more argillaceous lower energy Gadvan Formation. The Dariyan is capped by the Kazhdumi marls. Reservoir characteristics of the Khami are developed in Fars, where it is oolitic and pellety, but not associated there with Cretaceous or Tertiary source rocks. Its reservoir quality becomes poor in Khuzestan and disappears in NE and Central Lurestan, where it is replaced by its basinal time-equivalent, the Garau facies. Up to now, the exploration of the Khami reservoir has been disappointing with only limited amounts of oil found (Ahwaz, Chillingar) and marginal gas reserves (Agha Jari, Gachsaran). However, the Khami targets were often not properly tested due to technical problems, when drilling deeper than 4500/5000 m.

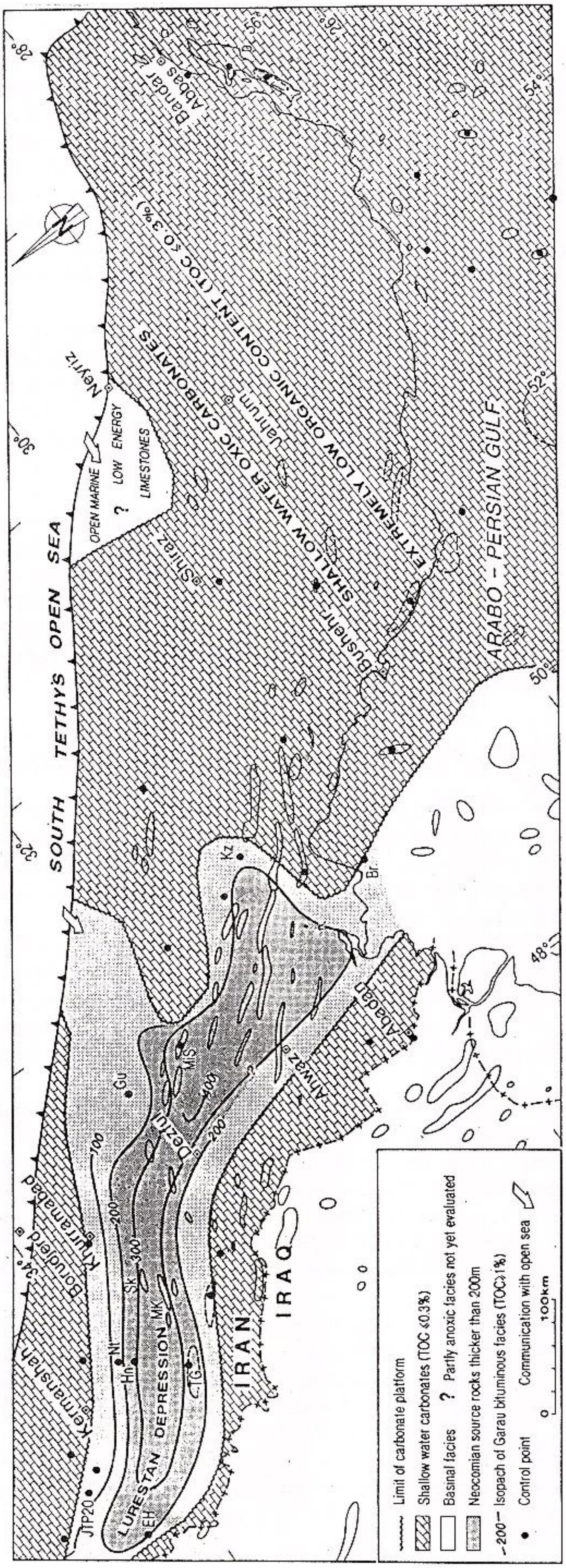


Fig. 6 – Isopachs of the organic-rich basal Garau interval (After Bordenave and Huc, 1995).

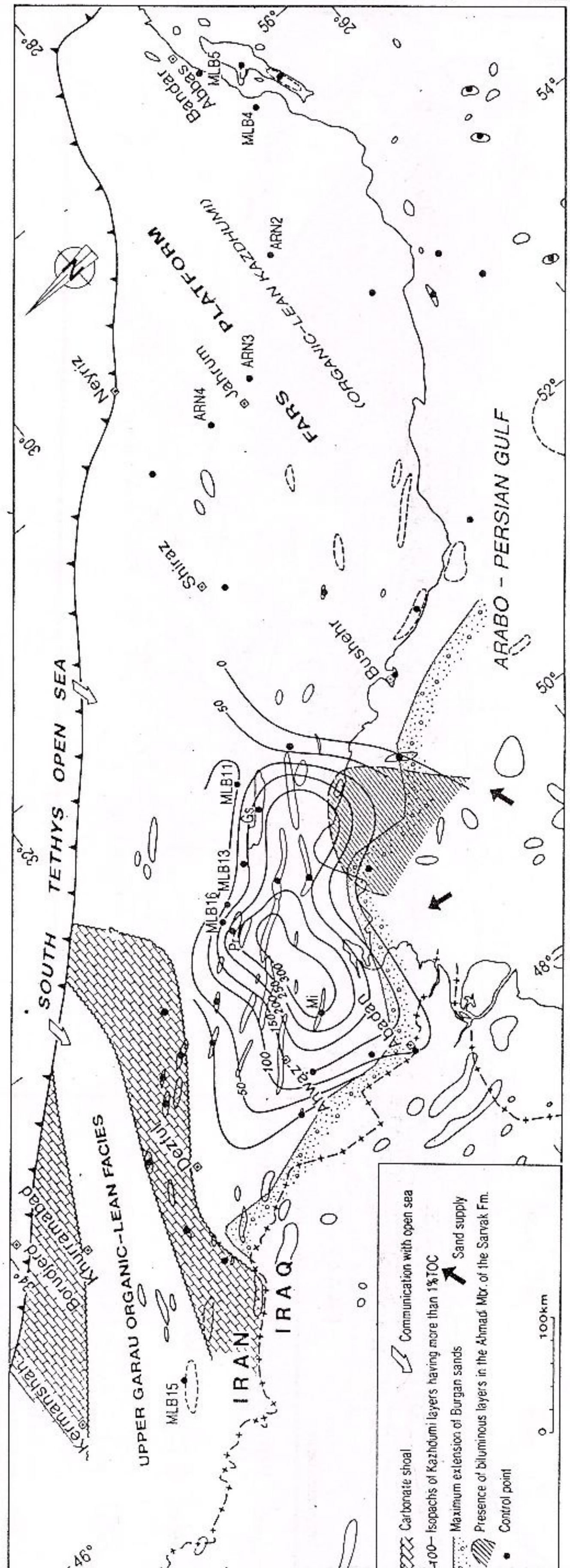


Fig. 7 – Isopachs of the organic-rich layers (TOC $\geq$ 1 %) of the Kazhdumi Formation (after Bordenave and Burwood, 1995).

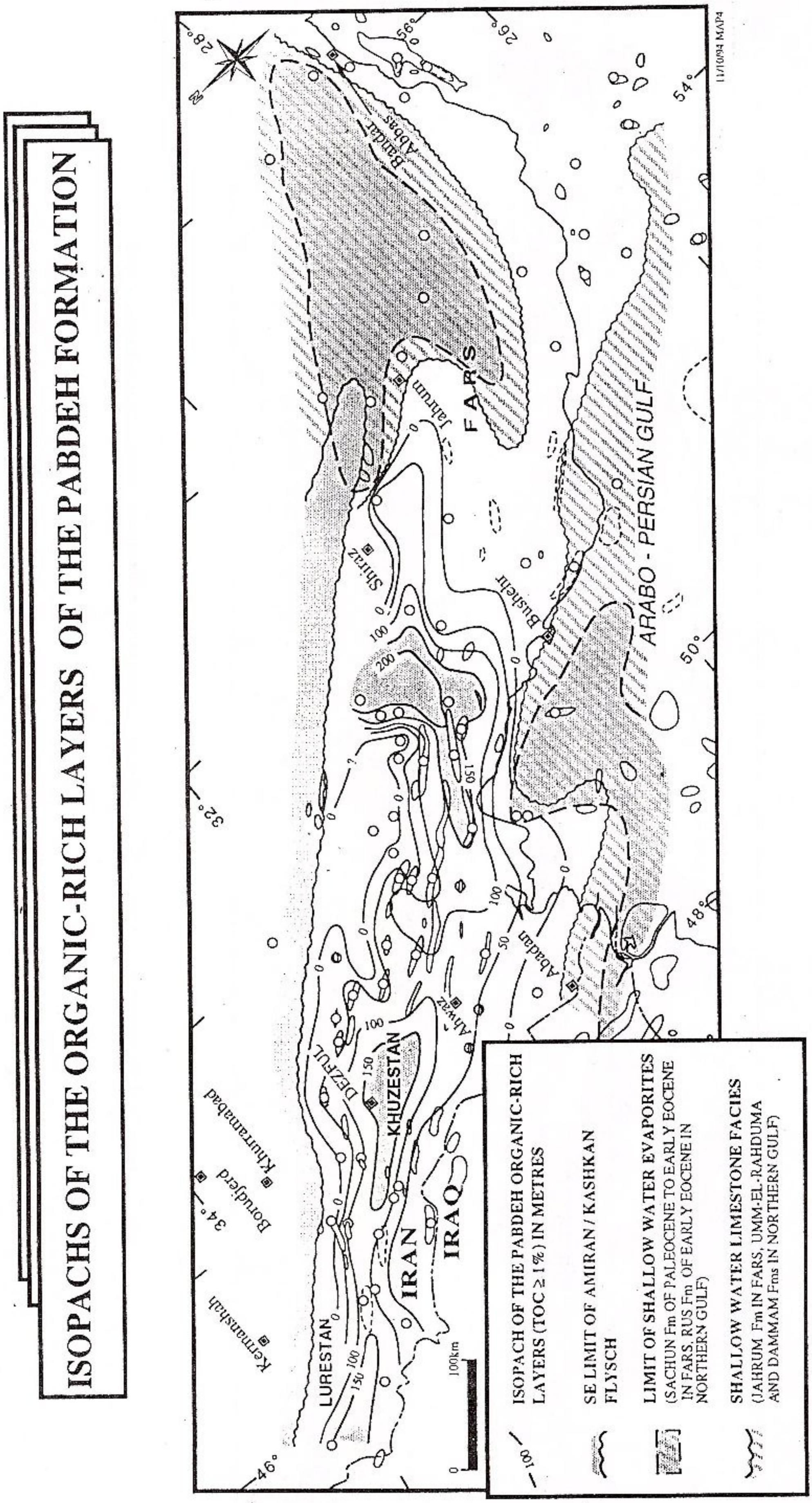


Fig. 8 – Isopach of the organic-rich layers (TOC ≥ 1 %) of the Pabdeh Formation (after Bordenave, in press).

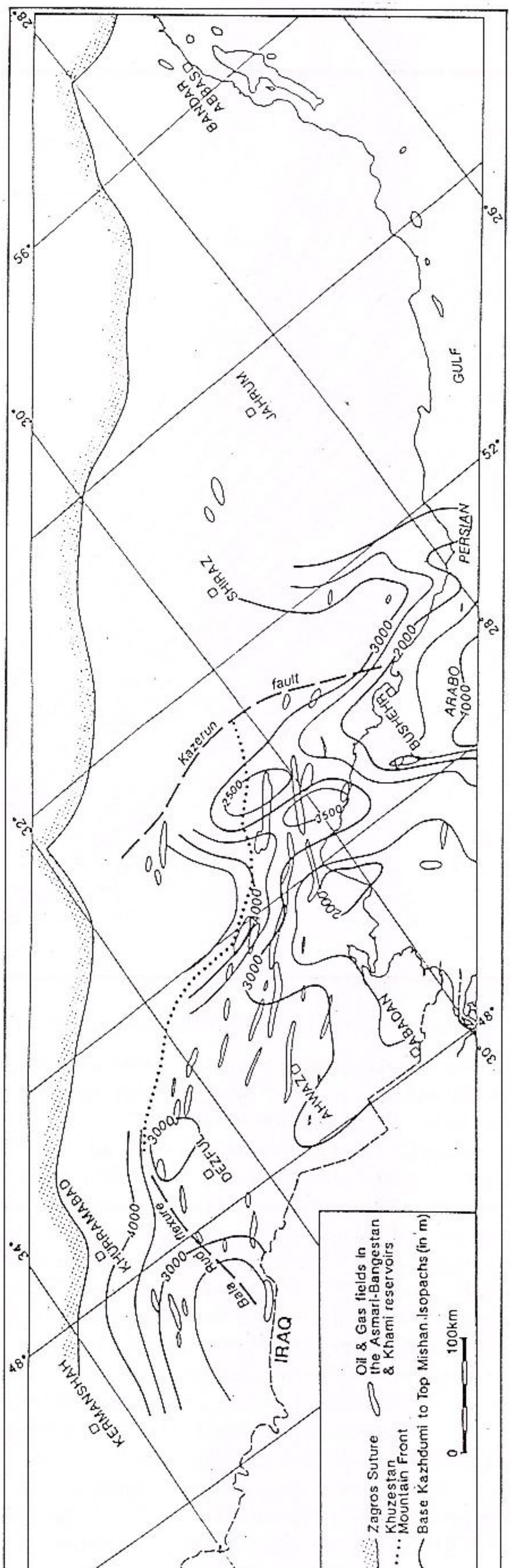


Fig. 9 – Isopach map of the base Kazhdumi - top Mishan (Bordenave and Burwood, 1995).

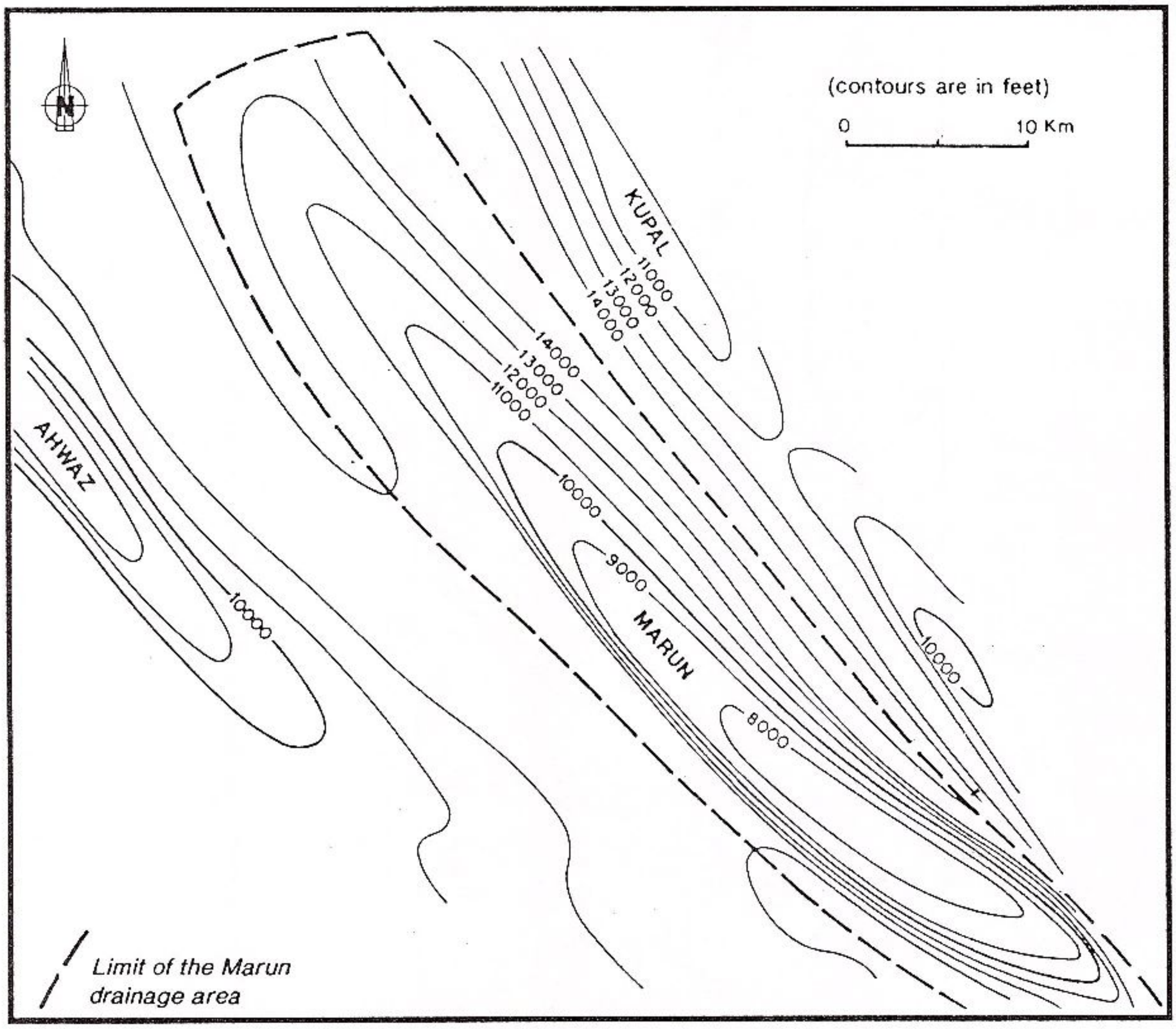


Fig. 10 – Area of drainage: the Marun field (after Bordenave and Nili, 1973).

A question to be answered is what source rocks really charged the three reservoirs available in the Dezful Embayment i.e. Asmari, Bangestan and Khami, or more precisely have the older source rocks: Silurian shales and Sargelu marls, participated in the charge of these reservoirs?

The Paleozoic Petroleum System, related to Silurian shales, does not interfer with younger systems. In Coastal and Central Fars and offshore, no post-Paleozoic source rocks are known, while in Lurestan and perhaps Khuzestan, thick basinal evaporites of the Adaiyah, Alan and Gotnia Formations (Fig. 3) prevent Paleozoic hydrocarbons to reach Cretaceous or younger reservoirs. Moreover, the $\delta^{13}C$ of the Silurian shales kerogen is about -31 $^0/_{00}$ (Fig. 4) as compared to -27.2 to -24.5 $^0/_{00}$ for the Dezful Embayment and Zagros Foothills oils (Fig. 5) (Bordenave and Burwood, 1990, Brosse and Bordenave, 1993).

The Sargelu organic-rich marls (initial TOC in the 6 to 10 % range) of Middle Jurassic are developed in Lurestan and Khuzestan where Gotnia evaporites prevent hydrocarbons originating in the Sargelu to reach the overlying Bangestan and Asmari reservoirs. Moreover, the efficiency of the 100 to 200 m Gotnia seal is confirmed by a well marked difference between the isotopic composition of the Sargelu pyrolysate, i.e. ($\delta^{13}C$=28.7±0.5$^0/_{00}$ (Fig. 4) and oils accumulated in Asmari and Bangestan reservoirs.

A Cretaceous/Tertiary Petroleum Complex, independent from older Petroleum Systems, can be defined in the Dezful Embayment. It is formed of three reservoirs, Asmari, Bangestan and Khami mainly charged by three source rocks, the Valanginian part of the Garau Formation (Fig. 6), the Albian Kazhdumi (Fig. 7) and the Pabdeh of Middle Eocene to Early Oligocene (Fig. 8). Another source rock, the Ahmadi Member

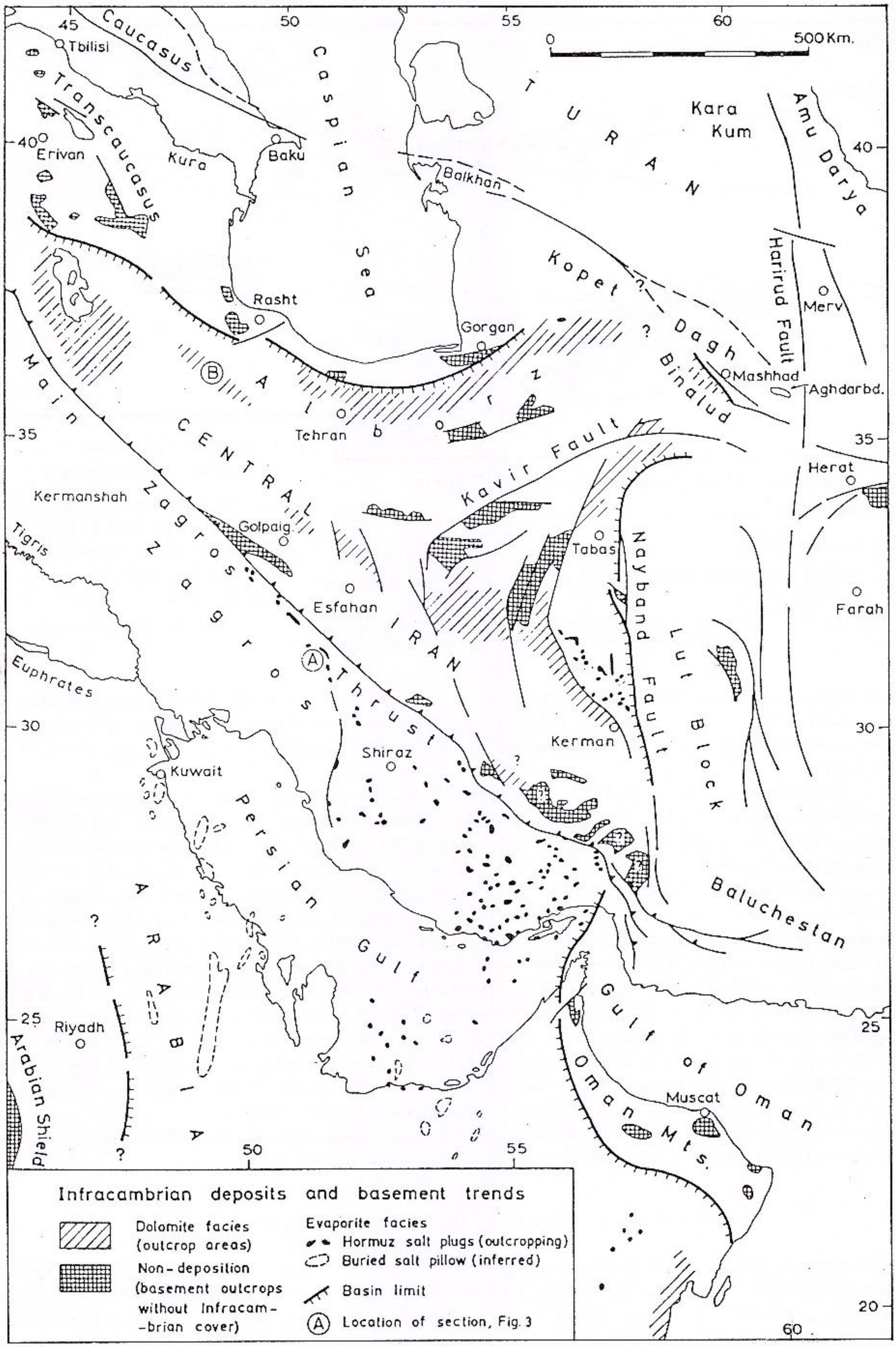


Fig. 11 – Iran structure map, showing the locations where Proterozoic to Cambrian successions were studied (A) in the High Zagros, (B) in Central Iran block (after Stöcklin, 1974).

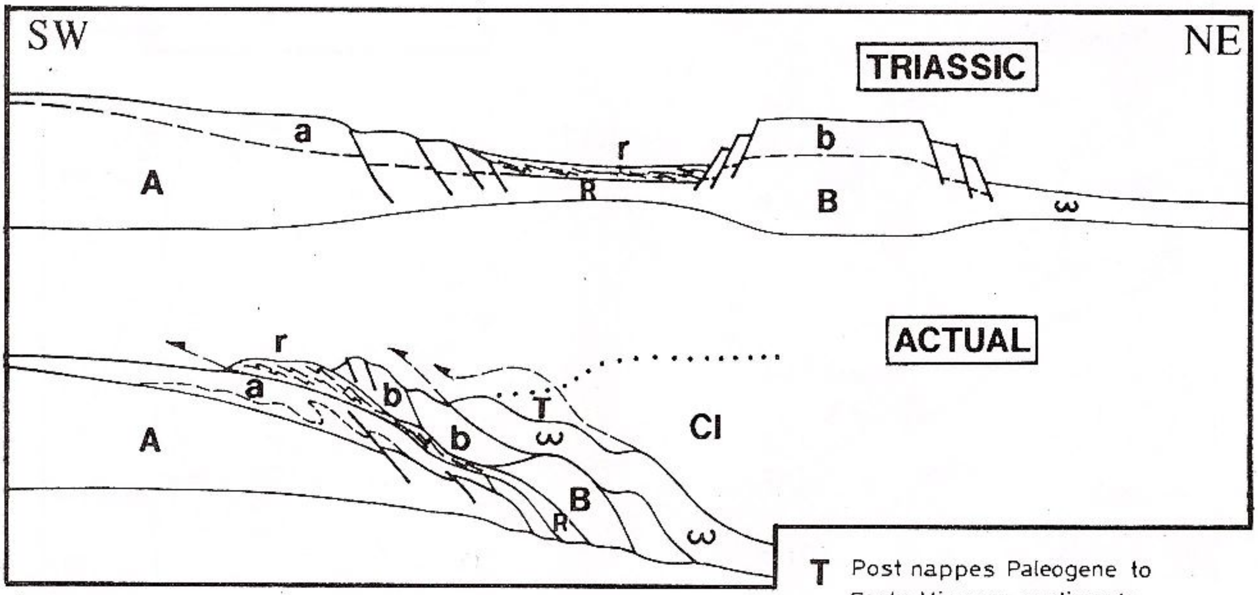


Fig. 12 – Sketch showing in Lurestan the relationship between the Arabian Platform and the Neo-Tethys during in the Triassic and after the pre-Maestrichtian phase (after Braud, 1987).

of the Sarvak Formation, which is bituminous in the northern Arabo-Persian Gulf and in the Binak area, accumulated in front of the prograding wedge of the Wara Formation of Late Cenomanian age. The impact of the Ahmadi Member, as far as oil generation is concerned, is limited to the southern top of the Dezful Embayment.

The Garau source rocks are overlain, in the Dezful Embayment, by thick basinal argillaceous limestones of the Fahliyan and the Gadvan Formations, then by a poor characteristic reservoir formed of the Dariyan Formation (Fig. 3). Some oil generated by the Garau may have reached the Dariyan together with some oil migrated per descensum from the Kazhdumi. However, as the amount of oil trapped in the Dariyan represents less than 2 % of the total of the oil accumulated in the Dezful Embayment, the contribution of the Garau is considered as negligible. **Only two source rocks, Kazhdumi and Pabdeh are assumed to have charged the Asmari and Bangestan reservoirs** and are therefore considered for modelling.

If oil expulsion from Kazhdumi and Pabdeh source rocks have occurred prior to the Zagros folding, oil would have then migrated along a low-angle dipping ramp, toward the Arabo-Persian Gulf, northern part of Saudi Arabia and Southern Iraq (Fig. 9). With such a scenario, oil may have been trapped far away from the kitchen area where it was formed. On the contrary, if the oil expulsion happened when the Zagros folds existed, or at least have commenced to grow, **the oil generated would have moved toward the closest anticlines** (Fig. 10). In this case, seismic data would allow to define, for each anticline, an area of drainage from which all the oil formed would charge the anticline.

In an area of drainage situation, **oil-to-source rock correlations become possible and a modelling technique can be applied as a prevision tool** (Bordenave and Nili, 1973).

The Tectonic History of the present Zagros Foothills

The orogenic phases, observed near the edge of the Arabian Platform, close to the margin of the Neo-Tethys, were reviewed together with their impact on the Platform itself, in terms of unconformities, readjustment of the previous fault network, movement of the Late Proterozoic Hormuz salt, erosion and subsidence.

The relationship between the subsidence of the Platform and the orogenic phases are discussed with some details, as the source rock burial and the timing of oil and gas generation depend upon subsidence. The entrapment of the generated oil and gas, if it happened before the Zagros folding, is rules by the geometry of drains and reservoirs, at the time of migration, which is related to the differential subsidence.

The tectonic history may be summarized, as follows:

Paleozoic. Available data suggest that during Precambrian and Paleozoic, the Central Iran block was part of the Arabian Platform and thus belonged to Gondwanaland (Fig. 11). On the Arabian Platform, a "Hercynian" phase caused the formation of a system of N-S oriented horsts and tilted blocks. This phase was well observed in Saudi Arabia where exploration wells have shown that more than 1000 m of sediments were removed on some horsts (Mc Gillivray, 1992). Similar conclusions were made in Iran from outcrops in the High Zagros (Szabo and Kheradpir, 1978).

The Triassic pull-apart and the formation of the Neo-Tethys. From the Late Triassic, the similarity between the Arabian Platform and Central Iran ceased abruptly. While the shallow water, mostly calcareous sedimentation continued on the Platform, a sudden change into coal-bearing paralic and continental clastic deposition attested the uplift of Central Iran and its fragmentation into a horst-and-graben system, indicated by abrupt changes of thickness of Upper Triassic and Liassic deposits across fault block boundaries (Stöcklin, 1968 and 1974).

As part of the rifting, a deep trough was formed on the NE edge of the Arabian Platform new margin. Deep marine sediments, i.e. radiolarites and siliceous micritic limestones were associated with turbiditic deposits originating in the adjacent platforms. The lower layers of these sediments were dated as Late Triassic. NE of the Radiolarite Trough, a narrow continental slab, parallel to the newly-formed margin, extending over 400 km from Lurestan to Iraq was individualized. This slab, known as the Bisotun Shoal, acted as a barrier reef on which more than 3000 m of high energy limestones were deposited almost continuously from Late Triassic to Cenomanian (Braud, 1987).

To the NE of the Bisotun Shoal, a new ocean or Neo-Tethys resulted from the pull- apart between the Arabian Platform and Central Iran (Fig. 12).

The first period of stability of the Arabian Platform. The formation isopachs, the facies distribution and the absence of marked unconformities show the **remarkable stability** of the Arabian Platform during 180 Ma, from the Permian transgression to the end of Albian. A monotonous shallow-water dominantly calcareous sedimentation continued in Fars, while more argillaceous basinal facies and relatively deep-water evaporites accumulated in the Lurestan depression.

Contrarily to what could have been expected, the breaking between the Arabian Platform and Central Iran did not correspond to any anomaly on the Arabian Platform, even close to its NE edge. As shown on Figure 13, subsidence remained moderate and almost constant during this long period. Apparent subsidence (ratio between a formation thickness and the duration of its deposition) remained in the 20 m/Ma range.

During this period, the Neo-Tethys began to widen with formation of oceanic crust, as the result of the northward drift of Central Iran, while a Paleo-Tethys, which existed in Paleozoic at the present northern foot of the Elbourz range, progressively narrowed and was closed by the end of Lias (Stöcklin, 1974). After the suture of the Paleo-Tethys, the African-Arabian plate continued to move northwards (Dercourt et al., 1985, Scotese and Golenka, 1993). The Neo-Tethys progressively narrowed as a result of its subduction underneath Central Iran, whose SE edge was an active margin, as shown by tectonic activity and metamorphism which ceased at the end of Jurassic. Later, it is likely that this subduction was relayed by an intra-oceanic arc with local duplication of oceanic crust and isostatic compensation (Desmond and Beccaluva, 1983, Braud, 1987).

The Middle Cretaceous instability. Epeirogenic movements induced during Cenomanian a partial readjustment of the "Hercynian" fault network, together with the mobilization of the Hormuz salt, i.e. the piercement of a large number of salt diapirs and the formation of salt domes and NS elongated salt swells (Kent, 1979). These salt-related structures caused a marked thinning of the Upper Cretaceous sediments (Koop and Stoneley, 1982).

A sudden and simultaneous increase of subsidence was observed everywhere on the present Zagros Foothills, during Cenomanian. The apparent subsidence, during the deposition of the Sarvak Formation, increased almost ten times as compared to those of the previous period, to reach 180±50 m/Ma (Fig. 13). This downwarping of the Arabian Platform was synchronous to the very beginning of a pre-Maestrichtian orogenic phase marked, near the NE edge of the Arabian Platform, by the appearance of brecciated facies in the Sarvak Formation, indicating a growing instability.

The pre-Maestrichtian collision. The uplift of the former Radiolarite Trough and its erosion were observed along the NE edge of the Arabian Platform, both in Fars, around Neyriz (Ricou, 1974) and in Lurestan (Braud, 1987).

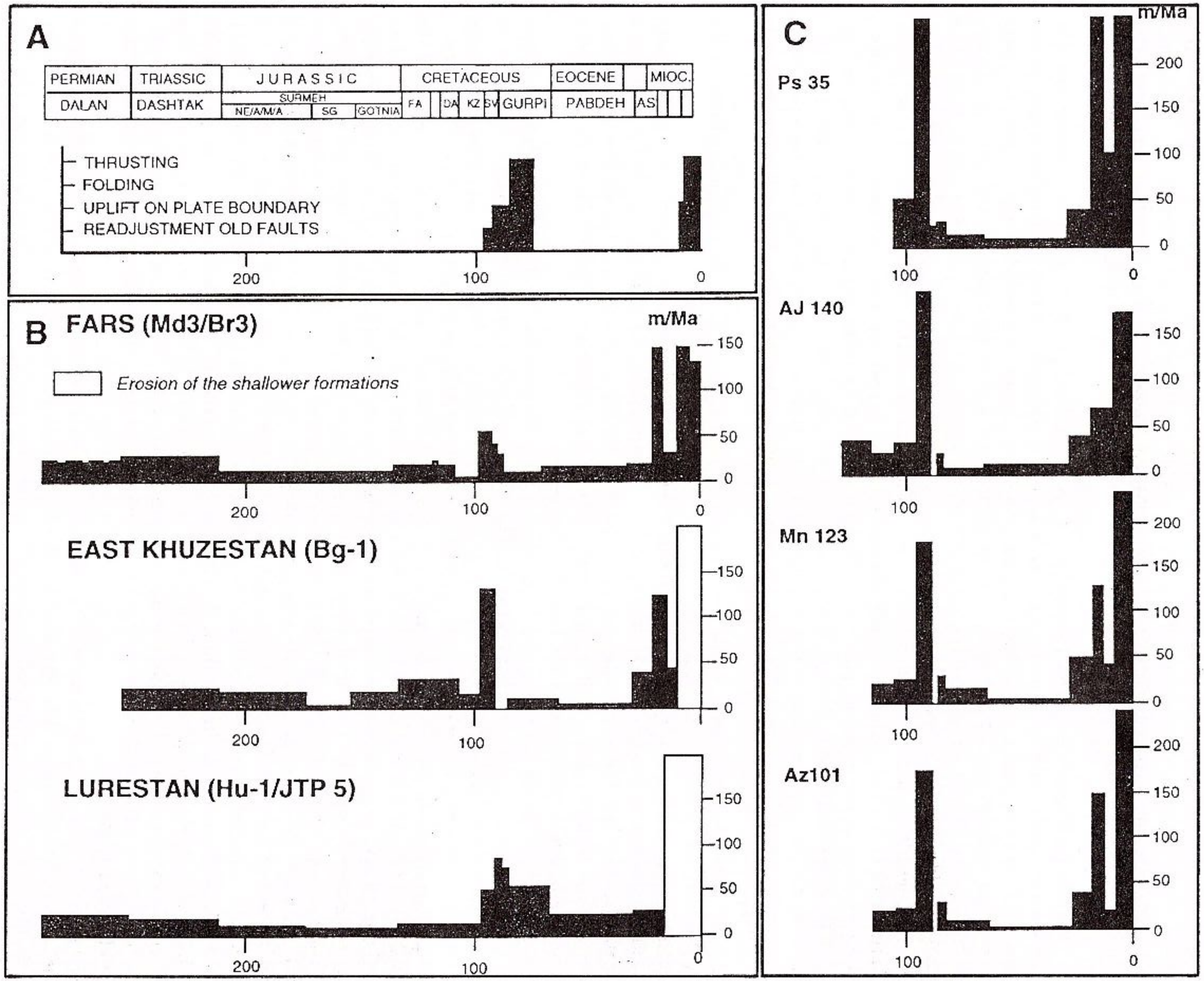
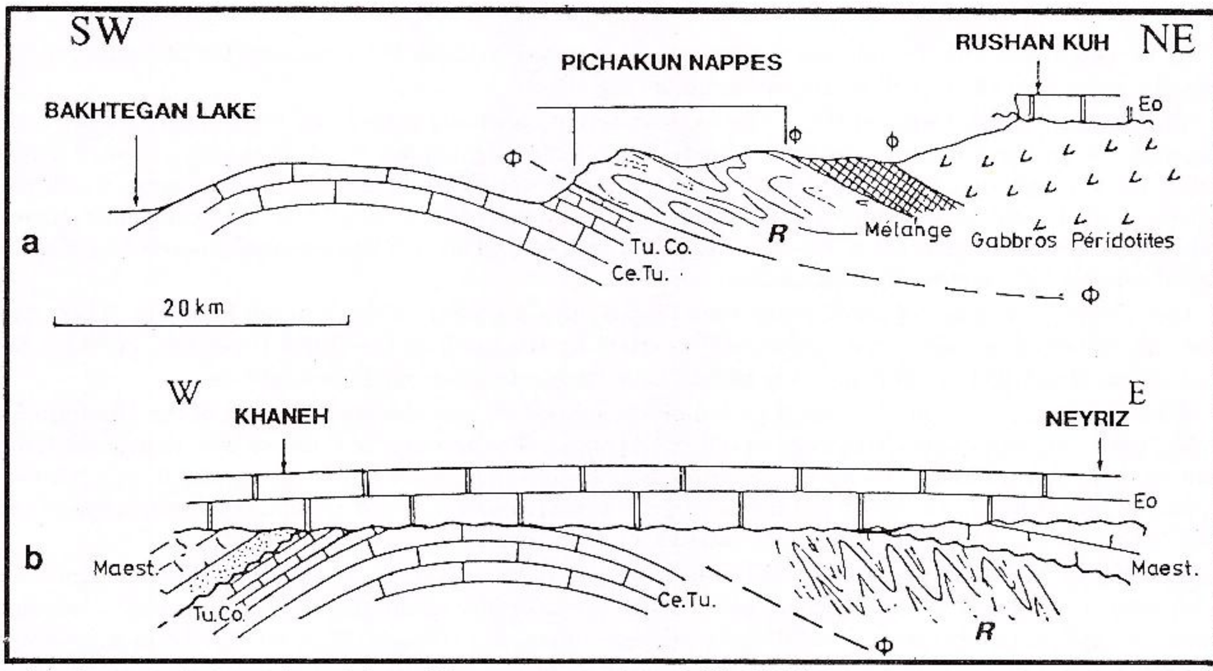


Fig. 13 – Relationship between subsidence variations in the Iranian part of the Arabian Platform, corresponding to the present Zagros Fold belt and orogenic events observed on the NE limit of the Platform: A. Apparent subsidence (Formation thickness/time of deposition ratio) observed from deep wells and surface sections from Fars to Lurestan. B. Apparent subsidence in Dezful Embayment wells from west to east. C. Orogenic events observed close to Platform edge.



a – Section from Bakhtegan Lake to Rushan Kuh.

b – Section on the southern edge of the Bakhtegan Lake showing the transgression of the Tarbur Limestones (Maestrichtian) on the radiolarite nappes (R), then the transgression of Eocene limestones (Eo) over all the older formations, nappes and autochthonous (Cenomanian = Ce, Turonian = Tu and Coniacian = Co.

Fig. 14 – Schematic section of the pre-Maestrichtian nappes NW of Neyriz (after Ricou, 1974).

In the Neyriz area, Coniacian globotruncana limestones contain thin beds of conglomerates consisting of radiolarite microclasts and debris of turbiditic shallow water limestones. NW of Neyriz, at Rushan Kuh and Kuh-e Dalneshin (Ricou, 1974, Stoneley, 1982) these beds are replaced by a "flysch" made of radiolarite material (Arsenjan "Flysch"). In Lurestan, the first radiolarite micro-elements were found in Campanian marls. At Kuh-e Zangalian, 10 km NW of Kermanshah, in Lurestan, marls dated as Campanian become chaotic at their top and contain unsorted blocks from pebble-size to a few cubic meters of contorted radiolarite and Cretaceous limestones interpreted as an olistostrome, originating in a very close uplifted area.

Always in the Neyriz area, at least five low-angle pellicular units, thrust one over the other (Pichakun nappes), were emplaced upon Cenomanian to Coniacian autochthonous sediments without any visible truncation at the tectonic contact. Each of these units contain a similar succession of well-dated sediments, i.e. pre-rift Triassic marls, associated with Ladinian-Norian reefal facies, and post-rift sediments: siliceous limestones with few radiolarite beds and turbidites (Lias-Dogger), thick radiolarite (Late Jurassic to Aptian) and Cenomanian conglomeratic limestones. The contact between units correspond, most of the time, to plastic Triassic marls which acted as detachment surfaces. These nappes were probably emplaced by gravity and not as the result of compression (Ricou, 1974). Extremely thick, up to 7000 m, massive basic volcanic rocks made of gabbro and peridotite, interpreted as ophiolite are thrust over the Pichakun nappes. However, between the radiolarites and the ophiolites, a chaotic assemblage of polymorphic blocks of various size, mapped as a "Coloured Melange", include huge slabs of Megalodon limestones dated as Triassic, up to 7 km long, 2 km wide and 300 m thick. This "Coloured Melange" was interpreted by Ricou, as a tectonic melange scrapped by the ophiolite nappes, when emplaced as a result of an extremely strong compression, and now accumulated in front of the nappes (Fig. 14).

A similar history can be reconstructed in Lurestan, but folded radiolarite nappes are thrust by the massive limestones of the Bisotun Shoal, which are in turn thrust by at least two ophiolite units. The first one, according to Braud (1987) which includes gabbros, harzburgite and diorite could have been derived from the tholeitic magma of an intracratonic island arc. This arc would have corresponded to the subduction of the SW oceanic crust associated with the Arabian-African plate below the NE oceanic crust associated to Central Iran. A short (10 Ma) but violent compression may have resulted from the collision between the descending Bisotun Slab and the NE oceanic crust. This compression would have provoked the aggradation of the former Radiolarite Trough and the Bisotun Slab to the Arabian Platform, and the obduction of the ophiolite nappes on the top of the two other units (Fig. 12).

The second period of stability. The tectonic activity abruptly ceased and thick shallow-water limestones of the Tarbur Formation dated as Maestrichtian sealed the nappes in NE Fars and remained undisturbed for a long period (Fig. 14). In Lurestan, the thrust units still formed a high-relief zone, which was actively eroded with the accumulation of thick detrital sediments (Amiran "Flysch" and conglomeratic red beds of the Kashkan Formation of Paleocene to Middle Eocene). The area was eventually covered by slightly unconformable Late Eocene Shahbazan limestones.

Unexpectedly, no trace of pre-Maestrichtian orogeny was observed on the Arabian Platform. There, the Ilam limestones of Santonian are conformably overlain by the marls of the Gurpi Formation in which the Campanian-Maestrichtian limit was only recognizable from micro-paleontologic evidence.

After the Coniacian uplift, a second period of low subsidence was observed on most of the Platform for 67 Ma, extending from Late Cretaceous to end of Oligocene. The apparent subsidence was, in general, lower than 10 m/Ma. A noticeable exception corresponds to a subsiding depression which extended over NE Fars (apparent subsidence in excess of 100 m/Ma) where Tarbur limestones and Sebkha-type anhydrites of the Sachun Formation accumulated during Maestrichtian and Early Paleocene.

The Early Miocene instability. With the exception of the low angle unconformity already mentioned in Lurestan towards the Middle Eocene, and a slight unconformity in the Rushan Kuh, between alveolinid limestones of Middle Eocene and reefal Early Miocene limestones (Ricou, 1974), no significant movements were observed on the Platform NE edge until the end of Early Miocene. However conglomerates in the Rezak marls of the Neyriz area and a 1000 m thick conglomeratic unit close to Firuzabad in Lurestan (Braud, 1987), both dated as Early Miocene, are evidence of another growing instability marked by the uplift and the active erosion of the NE edge of the Arabian Platform including the radiolarites and the pre-Maestrichtian ophiolite nappes.

On the Platform, the subsidence increased moderatly during the deposition of Asmari (20 to 50 m/Ma), but became extremely important with the deposition of the Gachsaran evaporites (70 to 500 m/Ma according to places). The low subsidence corresponding to the deposition of the Mishan Formation, toward the limit of Early to Middle Miocene may be due, at least in part, to a period of deposition shorter than estimated and may not be significant (Fig. 13).

As for the pre-Maestrichtian phase, the downwarping of the platform, which began during the deposition of Asmari and Gachsaran, preceded the main phase of the Zagros orogeny.

The main Zagros orogenic phases. The first evidence of the closure of the Neo Tethys was probably observed at Rushan Kuh, NW of Neyriz. There, undisturbed reefal Early Miocene limestones are overlain by 500 m of "flysch" becoming conglomeratic towards the top. **These conglomerates contain elements of quartz and biotite which proved the vicinity of the metamorphosed Central Iran block** (Ricou, 1974). Later, locally huge developments of syntectonic conglomerates, grouped into a convenient, but vague, Bakhtiari Formation remained undated and did not provide any age-dating of the main Zagros orogenic phases. **The younger undisturbed sediment on the NE edge of the Platform has been dated as Burdigalian** (Ricou, 1974).

Large-sized nappes were thrust, as a result of the collision with Central Iran, over an already eroded surface including Miocene "flysch", various older Tertiary layers of the extended Arabian Platform and even on Eocene volcanics which were interpreted by Braud (1987) as a former Neo-Tethys residual oceanic crust. The nappes were emplaced through almost horizontal thrust planes.

In Fars, the front of the Khounsar thrust unit extends on more than 100 km following a NNE-SSW direction. The nappes include from bottom to top (Ricou, 1974).

- highly metamorphosed ante-Triassic sediments, including gneiss, amphibolites, massive limestones in which fusulins are still visible and chloritoschists,
- weakly metamorphosed massive shallow-water with few slightly schistose argillaceous intervals, dated as Early to Middle Cretaceous;
- Eocene conglomerates, with Nummulites and Alveolins.

These sediments are similar to those described in Central Iran, in the Sirjan-Senandaj zone defined by Stöcklin (1968). In Lurestan, the Kuh-e Garun nappe (Braud and Ricou, 1971) included a metamorphic unit with schists, marbles and quartzites overlain by 1200 m of massive limestones dated from Barremian to Senonian.

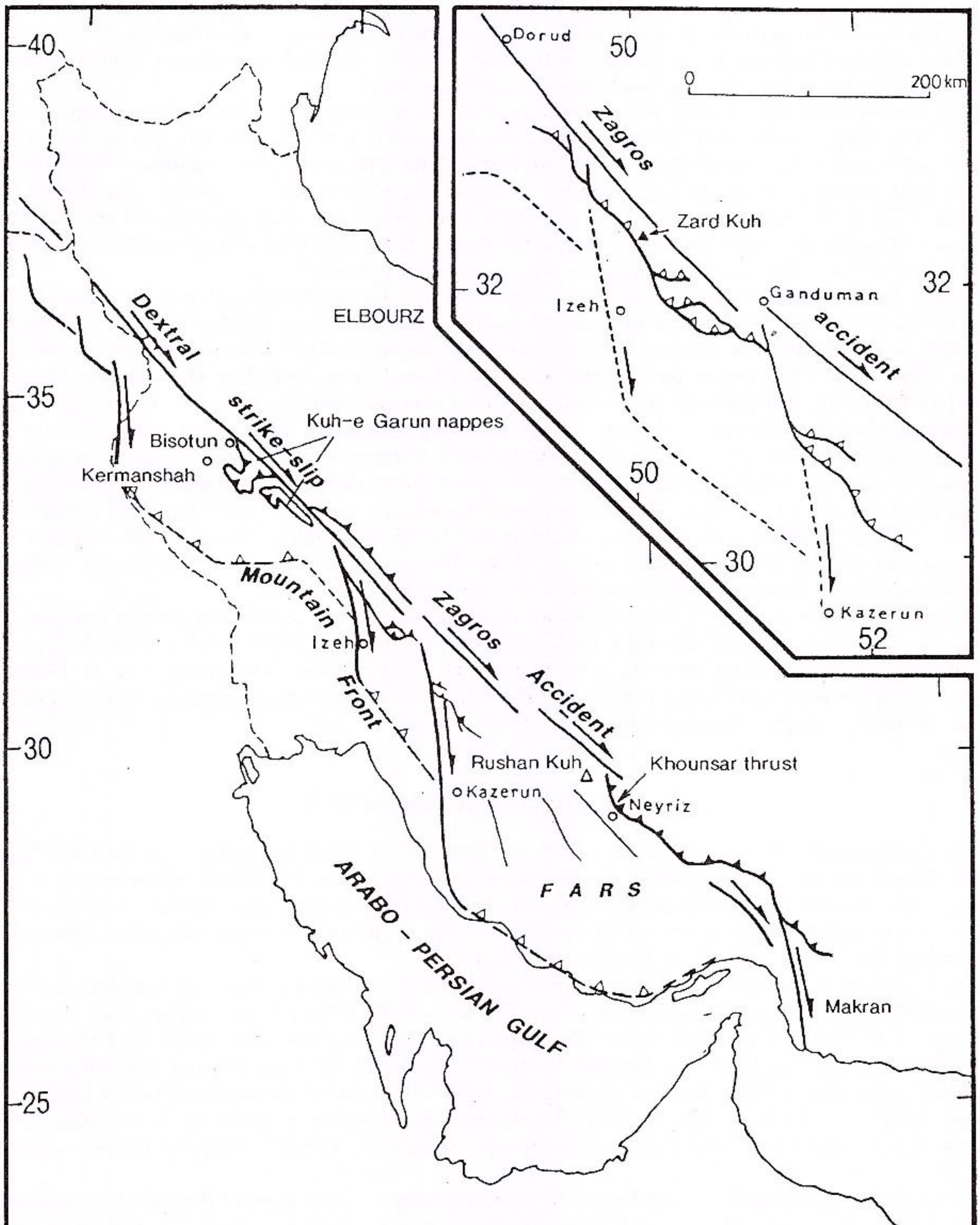


Fig. 15 – Location of the Khounsar and Kuh-e Garun Pliocene nappes and relationship with the NE-SW dextral strike slip Zagros main accident (modified from Braud and Ricou, 1971).

NE of the front of the Khounsar and Kuh-e Garun nappes (Fig. 15), a NW-SE almost linear accident, subvertical to a 45^0-dipping thrust plane, made of a succession of en-echelon dextral faults seems to limit to the SW the Central Iran Block metamorphic sediments. This accident would extend over 1700 km from the Van Lake in Turkey to the Hormuz Strait (Braud and Ricou, 1971).

This dextral strike-slip accident shifted the nappes and would be younger than the emplacement of the nappes. According to Ricou (1974), the first phase corresponded to a NW-SE shortening of the Zagros area and to the formation of the large amplitude NW-SE Zagros folds. The second phase provoked a NS shortening with dextral movement along the Zagros suture. It would also have reactivated the old "Hercynian" fault network, such as the Kazerun fault. It is also likely that the second phase accentuated the folds with formation of assymetric and even slightly thrust SW flanks. Some anticlinal axis anomalies such as Rag-e Sefid and several structures in Fars may be related to the second phase.

On the Arabian Platform, the subsidence remained very high during the deposition of the Agha Jari Fm, from 100 to 300 m/Ma on anticlinal tops and up to 500 m/Ma in deepest synclines (Fig. 13). However, according to seismic data, the Mishan Formation and the lower part of the Agha Jari Formation remained almost isopach, while **the upper part of the Agha Jari Formation, including the Lahbari Member dated as Pliocene, pinched out as a result of the structure growth.** Therefore, the main structure growth would be Late Miocene to Pliocene, mainly occurring between 10 and perhaps 3 Ma. However, the Zagros belt is still tectonically active as shown by the gentle warping of the Bakhtiari conglomerates and by Quaternary terraces affected by strike-slip faults. Moreover recent earthquake epicenters are located in the Zagros Foothills, i.e. in the Arabian Platform, while the former metamorphic Central Iran zone is seismically inactive. A large majority of the epicenters corresponded to NS major dextral faults which are still active, and are rather shallow, 0 to 20 km (Niazi et al., 1978). They seem to be related to a continuing folding of sediments above the Hormuz salt which acts as a detachment level.

It is also interesting to note that the Bakhtiari conglomerates do not contain any pebbles from the Fars Group i.e. Gachsaran, Mishan and Agha Jari Formations, but rounded pebbles and cobbles of Oligocene, Eocene and Cretaceous limestones and cherts (James and Wynd, 1965). This shows that the Bakhtiari conglomerates are not formed from the erosion of the higher parts of growing anticlines, but rather from the erosion of the High Zagros, NE of the Mountains Front.

Characteristics of the Zagros folds

The amplitude of folding progressively decreased from NE to SW, when going away from the Zagros suture, from assymetric, often thrust, high-relief large-sized anticlines to low-relief structures such as Ab-e Teymur, Mansuri and Binak. SW of these structure, the influence of the Zagros orogeny is hardly visible (Falcon, 1961), and structures are rather NS oriented, i.e. they are related to salt pillowing along "Hercynian" faults which were reactivated during Middle Cretaceous.

Typical Zagros anticlines formed impressive large-amplitude whale-back shaped mountains, sometimes deeply dissected by erosion. These anticlines, 5 to 15 km wide and 50 to 100 km long are often assymetric with a steeper or even reverse SW flank. The large size of the anticlines is explained by a thick pile of competent sediments, up to 10 km, bounded by detachment layers, the thick Hormuz salt below and the Gachsaran anhydrite and salt above (Colman-Sadd, 1978). However in SW Fars and Dezful Embayment another detachment level, the Triassic Dashtak anhydrite should also be considered. In Lurestan, several other potential detachment layers exist in the Jurassic, namely the Adaiyah, Alan and Gotnia evaporites (Fig. 3).

The anticlines are formed by parallel folding resulting from a combination of flexural slip and neutral surface folding (Fig. 16). In most of the anticlines, the Asmari limestones show normal faults or small tensional grabens at their top (Mc Quillan, 1973 and 1974). Tensional conditions also provoked an intense fracturing (Fig. 17). As mentioned by Mc Quillan, anticlinal tops provide "a text book of structure related tension fracture sets". In a parallel neutral surface folding, fractures theoretically extend down to the plane of no strain which limits the tensional zone above from the compressional one below.

Surface observations, made in Fars, Lurestan and NE Khuzestan Mountain Front show that fractures generally extend at least down to the Sarvak Formation, i.e. that the Asmari and Sarvak limestones, as well as the thick Pabdeh-Gurpi marls are fractured. A similar conclusion can be drawn from the study of the numerous cores taken in exploration and production wells.

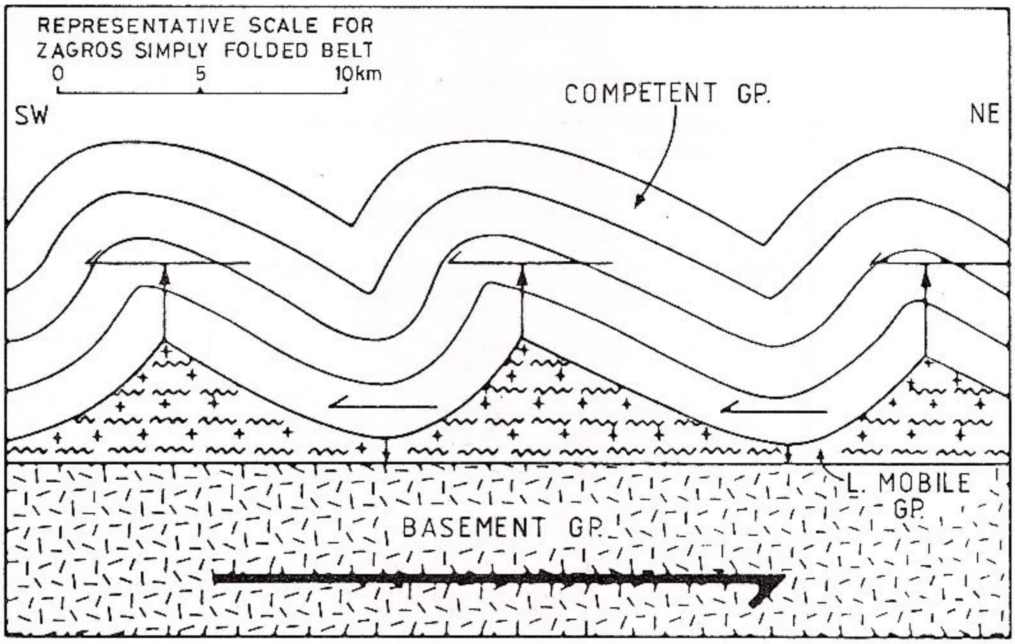


Fig. 16 – Assymetric folds caused by shearing on detachment zone (after Colman-Sadd, 1978).

Tectonic history and geochemical modelling

Few points, which resulted from the tectonic history of the present Zagros Foldbelt are critical for geochemical modelling:

1. Subsidence

The subsidence in the Zagros domain remained moderate for 250 Ma, from the Permian transgression, i.e. 280 Ma, to the end of the Asmari deposition, i.e. 30 Ma, excluding the short increase of subsidence during the deposition of the Sarvak Formation. During this long period, the average apparent subsidence varied little, around 20 m/Ma (17 to 20 m/Ma in Fars, and 20/23 m/Ma in the Dezful Embayment and Lurestan).

Then, the subsidence dramatically increased from 20 to 3 Ma, averaging up to 350 m/Ma. As a consequence, oil generation occurred very recently. The burial of the Kazhdumi source rocks was not sufficient for the oil window to be reached by the end of the Asmari deposition. At the end of the Mishan deposition (10 Ma) the oil expulsion window was not reached, except perhaps in NE Khuzestan in the Paris area.

2. Age-dating of the Zagros folding

As previously mentioned, the last well-dated undisturbed sediments on the NE edge of the Arabian Platform belong to Burdigalian. The thrusting of the Kuh-e Garun and Khounsar nappes is younger than 15 Ma. According to seismic, the folding would commence not too long after the beginning of the deposition of the Agha Jari Formation, perhaps around 10 Ma ago. The period of structural growth is likely to have continued from 10 to 3 Ma, later the tectonic activity continued, but at an extremely low pace.

3. Fracturing

Extensive fracturing developed in high relief anticlines. This probably explains why in most of the Dezful Embayment major oil and gas fields, the Asmari and Sarvak reservoirs have some degree of connection, from a complete interconnection with the same oil-water contact, as for Agha Jari, Bibi Hakimeh, Gachsaran, Pazanan and Rag-e Sefid fields, to a partial connection, in which the lighter fractions of oil and gas moved with difficulty through a poorly developed fracture network (Ahwaz, Mansuri). The fracturing is extremely important, as it allowed easy subvertical migrations of the Kazhdumi-originated oil up to the Asmari reservoir, well capped by the Gachsaran evaporites. It is also possible that the fracturing of the Kazhdumi marls facilitated the oil expulsion process.

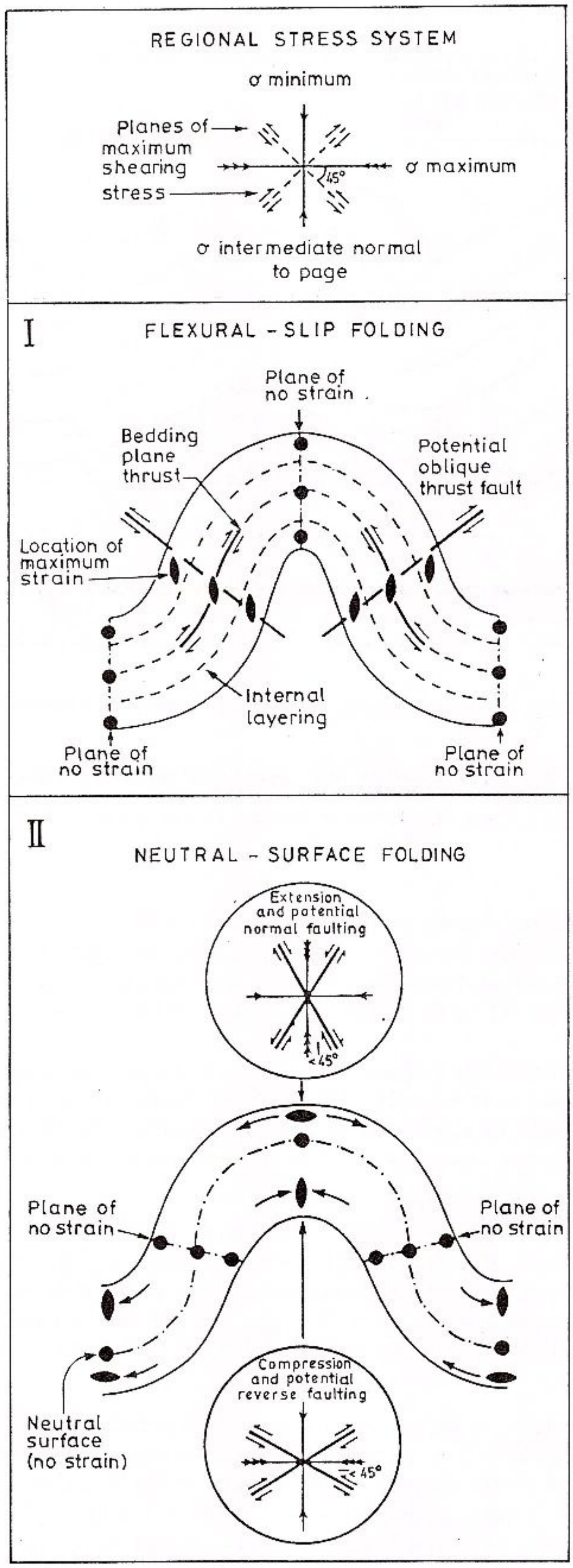


Fig. 17 – Parallel folds developed by (I) flexural slip, II neutral-surface folding (after Colmaan-Sadd, 1978).

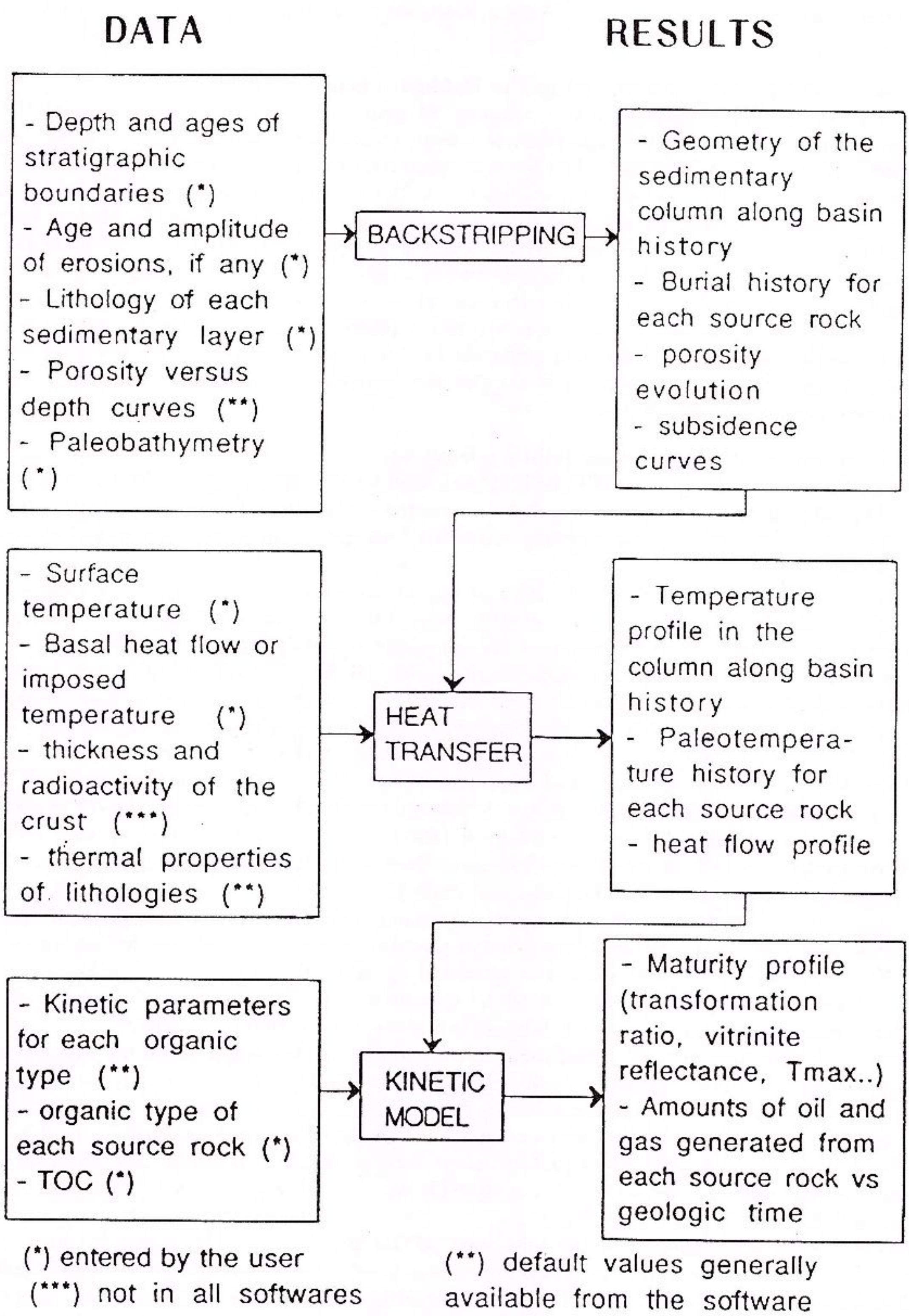


Fig. 18 – General features of classical 1D model of burial reconstruction, heat transfer and petroleum generation (after Ungerer, 1993).

By contract no fracturing is observed in some low-curvature deeply buried anticlines no fracturing is observed and Asmari and Sarvak appear to be independent reservoirs as at Ab-e Teymur. Oil generated by the Kazhdumi would remain trapped in the Sarvak, while oil originating in the Pabdeh would accumulate in the Asmari.

4. Absence of Fars Group material in the Bakhtiari conglomerates

The absence of material derived from the Gachsaran, Mishan and Agha Jari Formations in the Bakhtiari conglomerates seems to discount the assumption of a deep erosion on the top of Zagros growing structures in **the Dezful Embayment**. The Agha Jari Fm was measured to be 3000 m thick in the syncline between Gachsaran and Agha Jari-Pazanan structures (James and Wynd, 1965), while it is only between 700 and 1200 m on Agha Jari and Pazanan tops, and it does not exist at Gachsaran, where the Mishan Formation outcrops. The thinning of the Agha Jari Formation seems to result more from a depositional pinch out rather than the deposition of a thick Agha Jari followed by a deep erosion. Seismic observations corroborate this assumption. The evaluation of the erosion is a critical factor in modelling.

The assumption of a relatively modest erosion is not applicable, however, to the Kuzhestan Mountain Front and to the High Zagros anticlines from where the Bakhtiari material was derived. For instance, at the Bangestan well, the erosion attained the top of the Dariyan Formation (Aptian) and thickness of the eroded section was estimated to almost 3000 m.

5. Determination of the main modelling parameters

The GENEX program designed by IFP/BEICIP was used for this modelling exercise (see Ungerer, 1993 and Fig. 18). Burial profiles were constructed for selected wells, located close to the top of the main field anticlines (Fig. 19). Erosion was carefully estimated from geological reconstruction and extrapolated vitrinite profiles.

In order to determine a thermal history, different hypotheses using either a variable or a constant heat flow through geological time were tested. However, because of the remarkable stability of the Platform, from the Permian transgresssion to the beginning of the deposition of the Gachsaran Fm, the assumption of a constant heath flow was used. The actual temperature profiles and the values calculated for each burial profile were checked (Fig. 20), keeping in mind that BHT data are often 10 to 15 C lower than true temperatures. Vitrinite profiles from 3 wells: Agha Jari 123, Bangestan 1 and Paris 35 (Burwood, unpublished company reports cited by Khosrovi Said, 1987) were also used for calibration (Fig. 21 a, b, c). A reasonable fit was obtained for both measured temperature and vitrinite profiles with a constand heat flow of 36 mW/m^2.

Kinetic parameters determined by pyrolysis in the lab and the OPTKIN program of the IFP for immature samples from the Kazhdumi and Pabdeh source rocks (Espitali, personal communication) were used in the model. These parameters were calibrated with measured Rock Eval data, Tmax (Fig. 22 a, b) and Hydrogen Indices (Fig. 23 a, b) from two wells, Mansuri 1 and Paris 35.

The onset of the oil window was then determined by using both the calibrated heat flow and the kinetic parameters of each source rock. The expulsion window was also determined, assuming that expulsion started when 30 % of the source rock pore space was saturated by generated petroleum. The beginning of the expulsion window was compared to the time of the structural growth of the Zagros anticlines which was assumed to have commenced 8 Ma ago, much later than previously assumed (Ala, 1982).

In the example presented here for a **well located on the top of the Marus** structure, the beginning of the oil window (Transformation Ratio, TR=10 %) was dated as 15 Ma (Figs. 19 and 24) for the Kazhdumi and 3 Ma for the Pabdeh, both corresponding to the time when source rocks were buried around 3000 m. However, the **expulsion window was attained 5 Ma ago for the Kazhdumi** and it is not yet reached for the Pabdeh (Fig. 24). The expulsion of oil from the Kazhdumi would have commenced when the Transformation Ratio reached 24 %. The actual TR would reach respectively 56 % and 17 % for the Kazhdumi and the Pabdeh (Figs. 25 and 26).

The same exercise was made for the **deepest part of the synclinal** area bordering Marun to the NE, where according to seismic data, the Agha Jari Formation, together with thin Bakhtiari conglomerates was assumed to be 3800 m thick (Figs 27 a, b). The onset of the **Kazhdumi** oil window also happened 15 Ma ago, which is obvious and no difference of burial is expected, before the deposition of the Agha Jari Fm and before the structural growth, between a well located on the Marun field and in the present deep syncline before the deposition of the Agha Jari Formation and before the structural growth. The **expulsion commenced 8 Ma ago**, about synchronous with the structural growth. The actual TR would be as high as 97 % (Figs. 28 a, b). At that depth, the gas window was reached. For the **Pabdeh source rocks,**

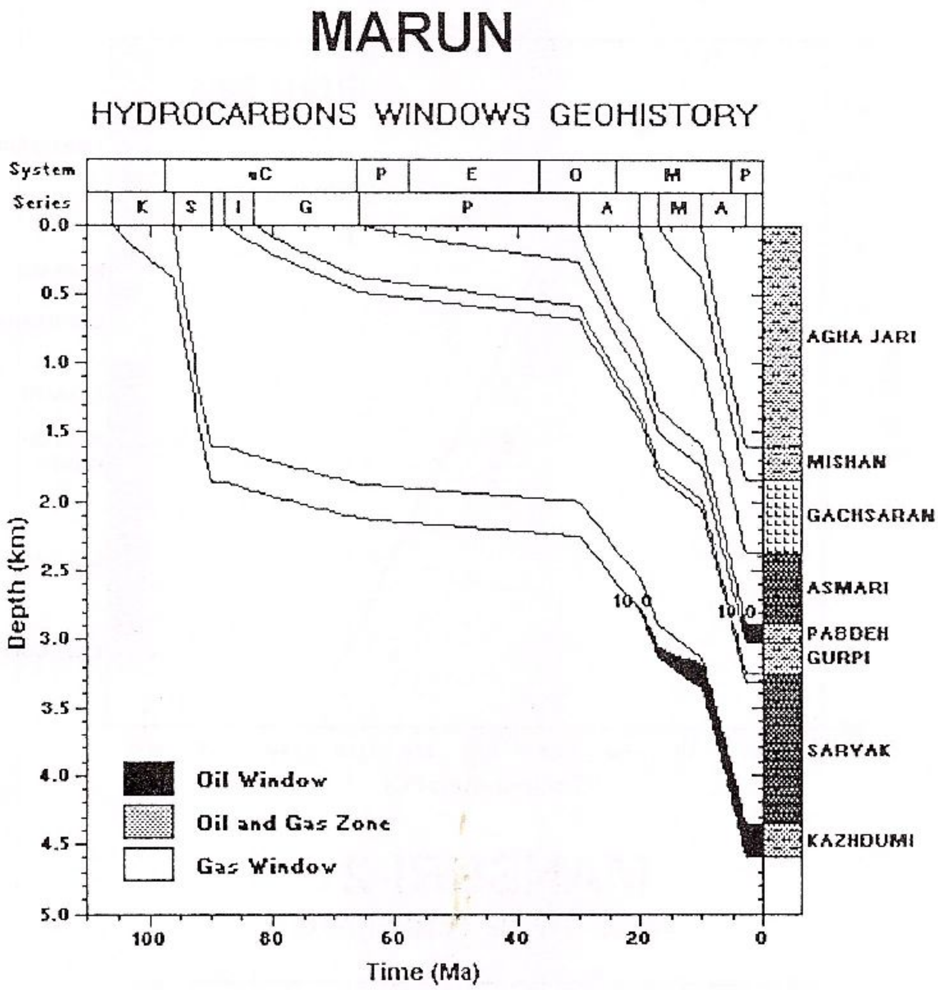


Fig. 19 – Burial profile and beginning of oil generation (hydrocarbon windows) for a well located close to the top of the Marun anticline.

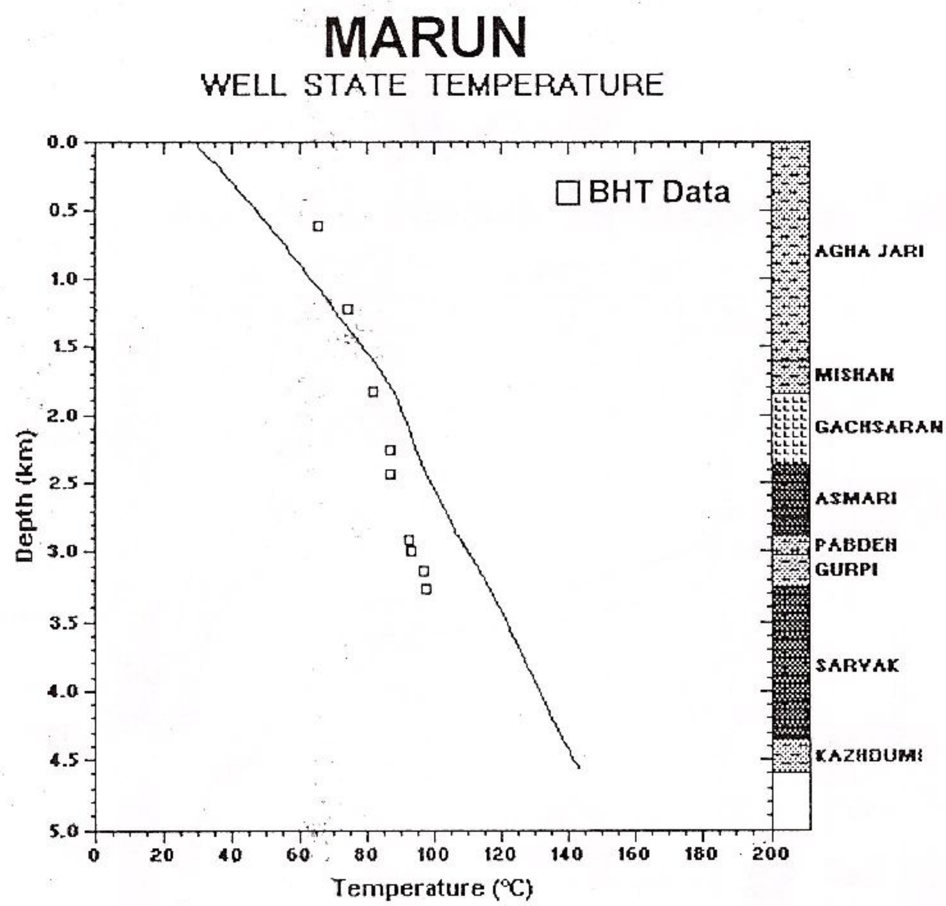
expulsion began 3.5 Ma ago, while the actual TR reached 66 % (Figs. 29 a, b).

6. The concept of area of drainage and calculation of the amount of oil generated

In the example of Marun, oil began to be expelled from the Kazhdumi 8 to 5 Ma ago, when the structure growth has already begun. Therefore, the drainage area concept could be applied and short range, almost vertical migration enhanced by fracturing of the Bangestan to Asmari interval, allowed to closely relate oils to the source rocks they were derived from. **Similar conclusions were obtained for other fields for which oil expulsion began, at the earliest 8 Ma ago, in the deeper synclinorial area, but was in general younger.**

The geometry of the drainage area for each field was deduced from seismic data. Each drainage area was then cut into slices of 1000 feet (304.8 m) thick (Fig. 10). A burial profile was reconstructed for each of these slices, which allowed to calculate for each source rock, i.e. Kazhdumi and Pabdeh, the actual Transformation Ratio (TR), and the amount of oil generated and expelled. The calculated amount of oil generated by a source rock for the whole area of drainage, Qc, would be:

a.



b.

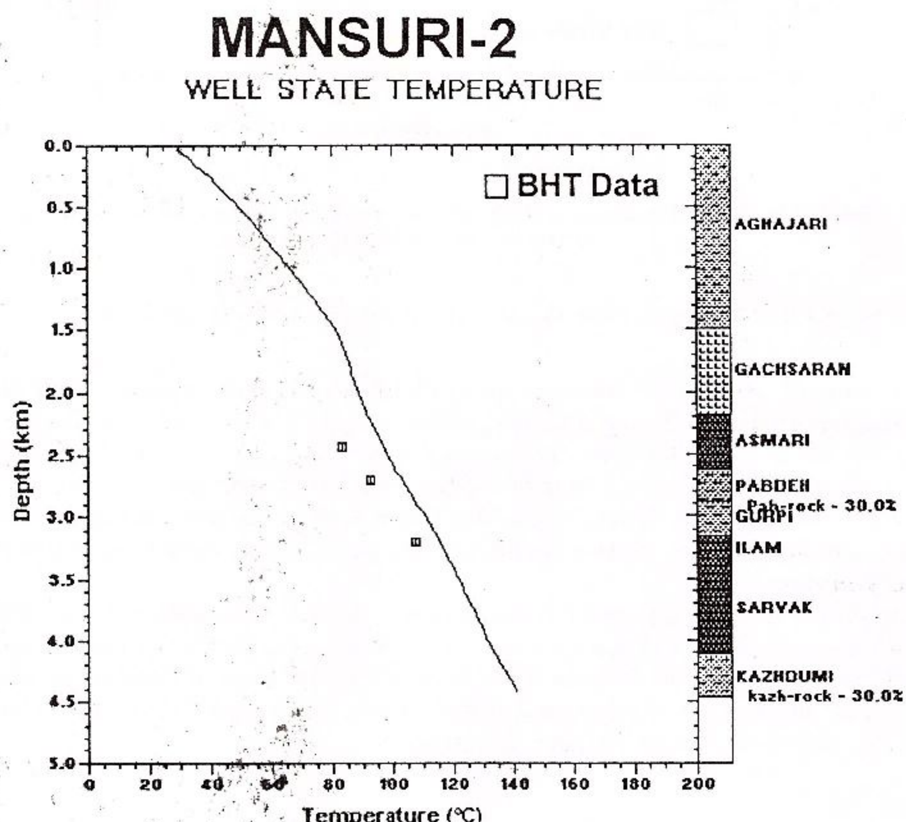


Fig. 20 – Calculated and observed present temperature profiles.

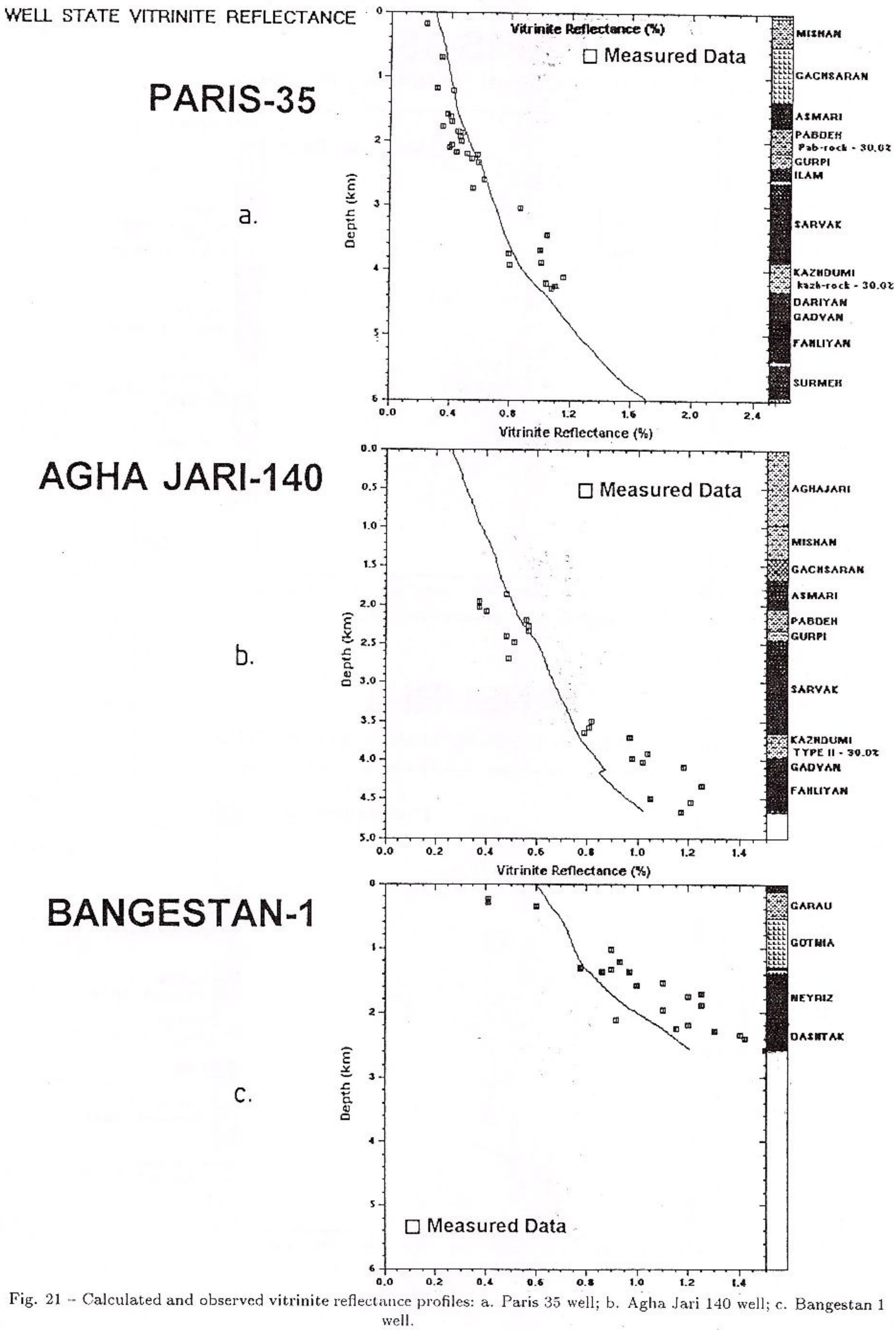


Fig. 21 – Calculated and observed vitrinite reflectance profiles: a. Paris 35 well; b. Agha Jari 140 well; c. Bangestan 1 well.

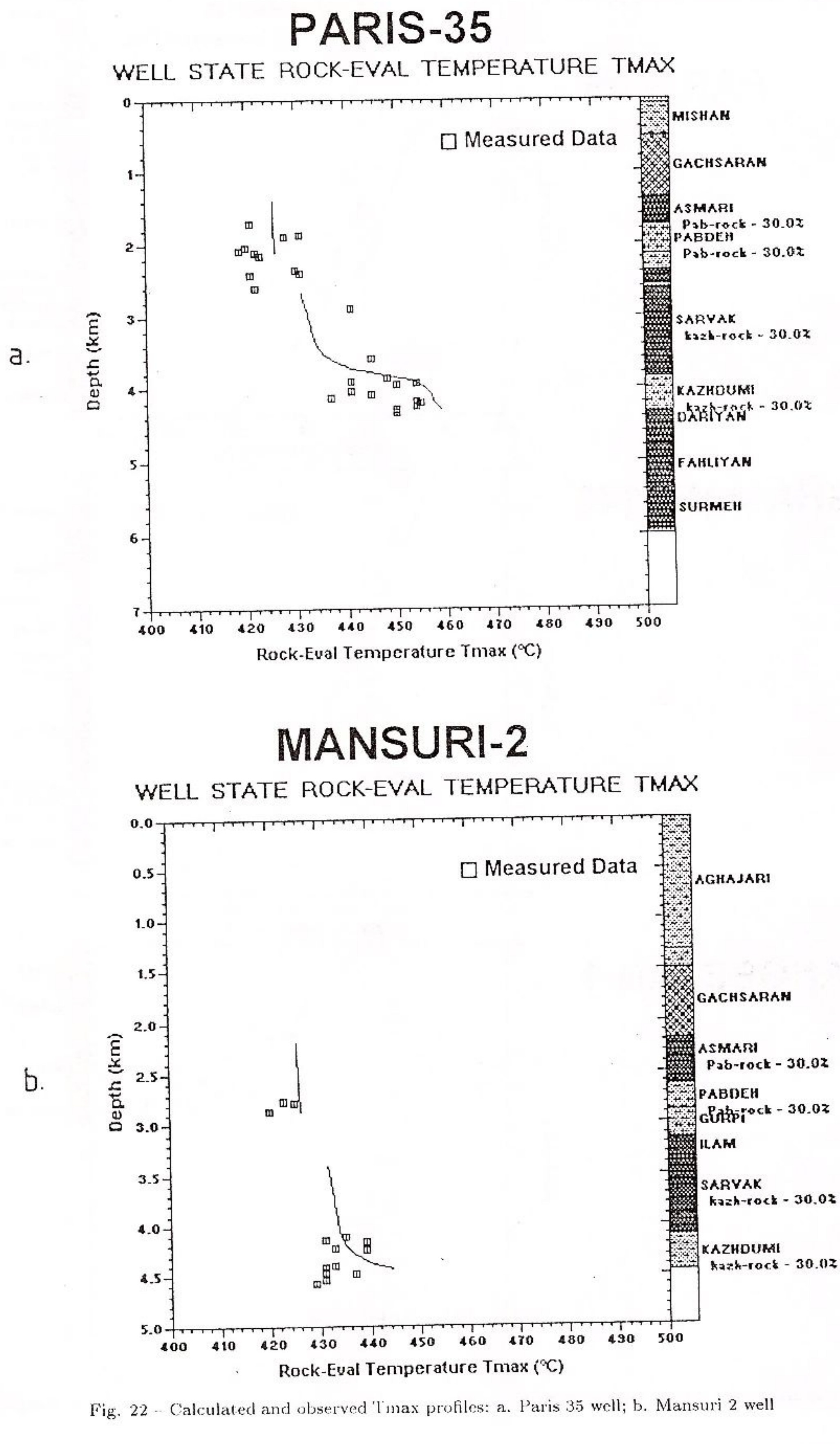
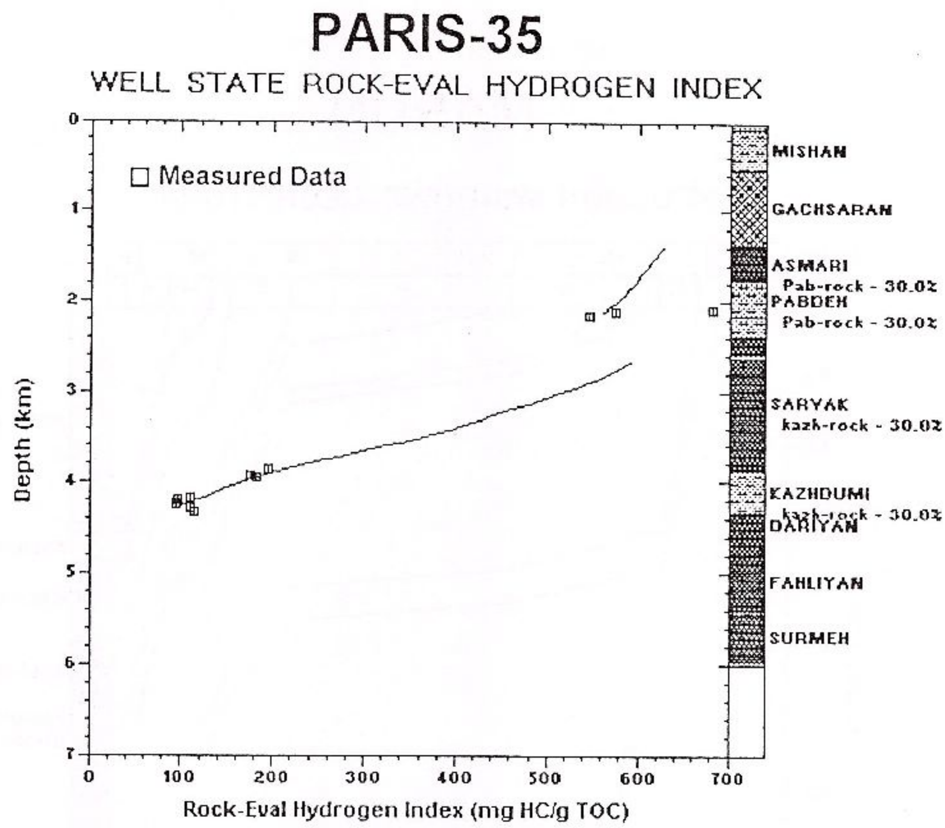


Fig. 22 – Calculated and observed Tmax profiles: a. Paris 35 well; b. Mansuri 2 well

a.



b.

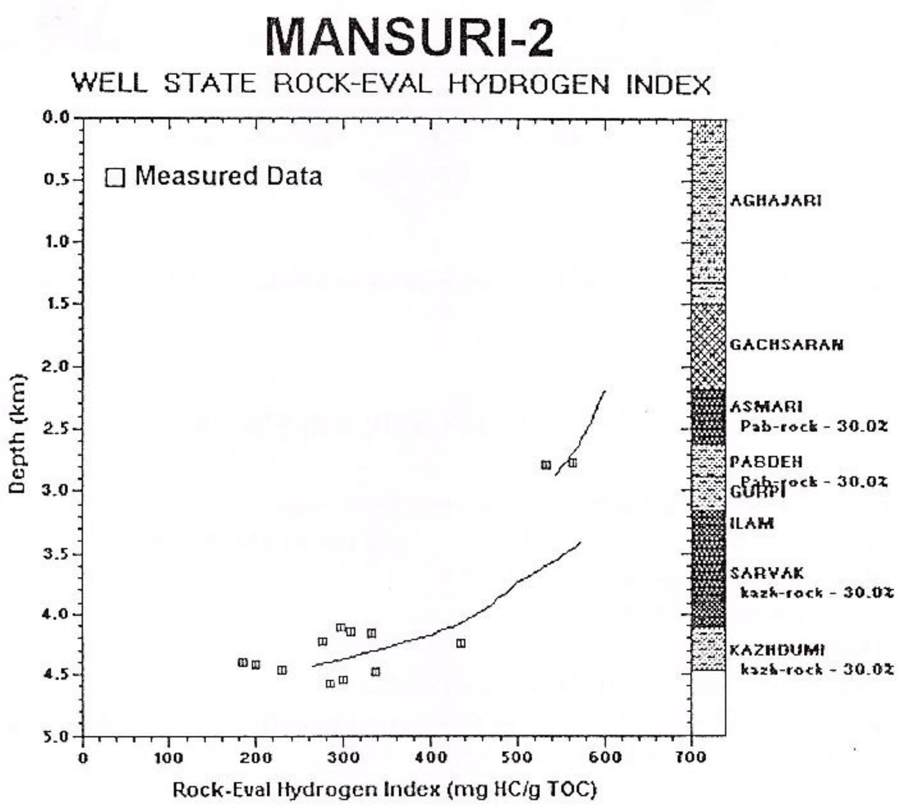


Fig. 23 – Calculated and observed Hydrogen index profiles, Kazhdumi Formation: a. Paris 35 well, b. Mansuri 2 well.

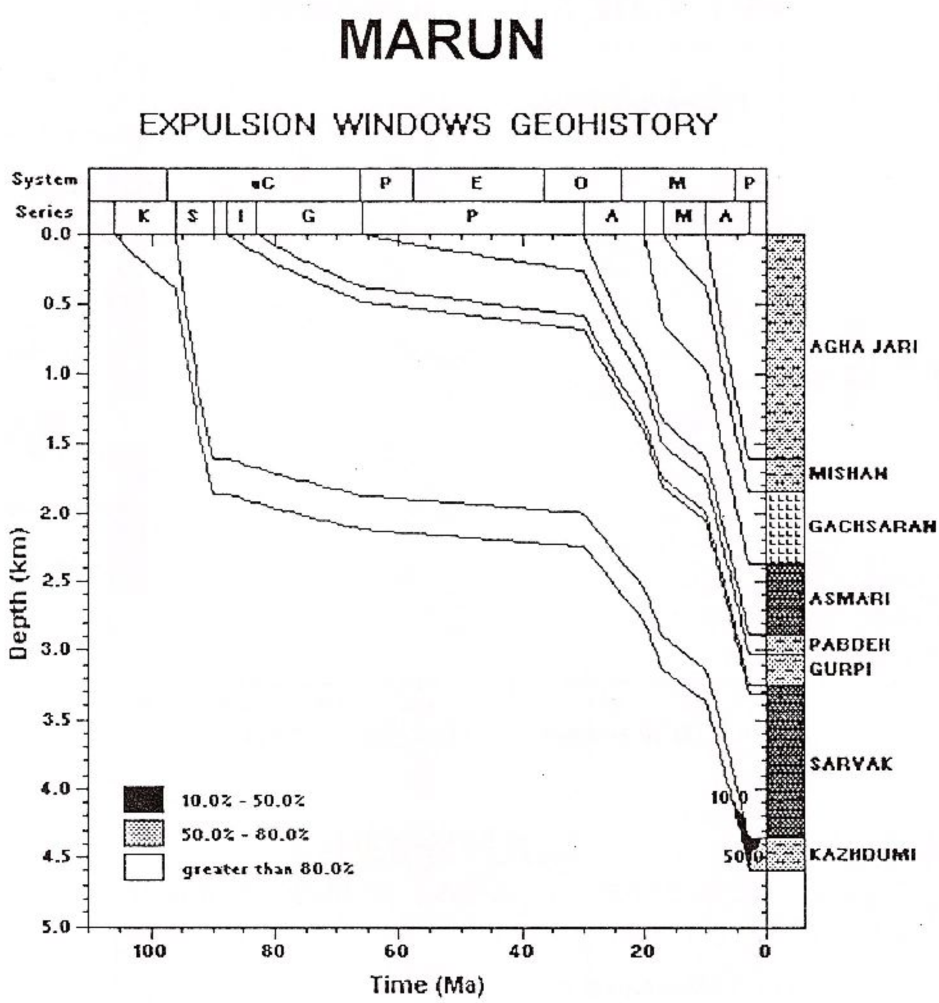


Fig. 24 – Marun burial profile and expulsion windows for a well located close to the top of the anticline.

$$Q_c = h\ (TOC \times HI)\ (\rho R_i \sum S_i\ TR_i)$$

where h is the source rock thickness in the area of drainage

TOC total organic carbon of the source rock per weight of rock

HI its Hydrogen Index

S_i the surface of the slice i

(R the rock specific gravity

TR_i TR calculated by the model for the slice i

If Q_c is expressed in barrels, h in meters, TOC as a proportion HI in kg HC/ton of C, ($\rho_R = 2.5$ t/m3, S_i in Sq km and TR %, Q_c would be:

$$Q_c = 19.65\ 10^3\ h\ (TOC.HI)\ (\sum S_i\ TR_i)$$

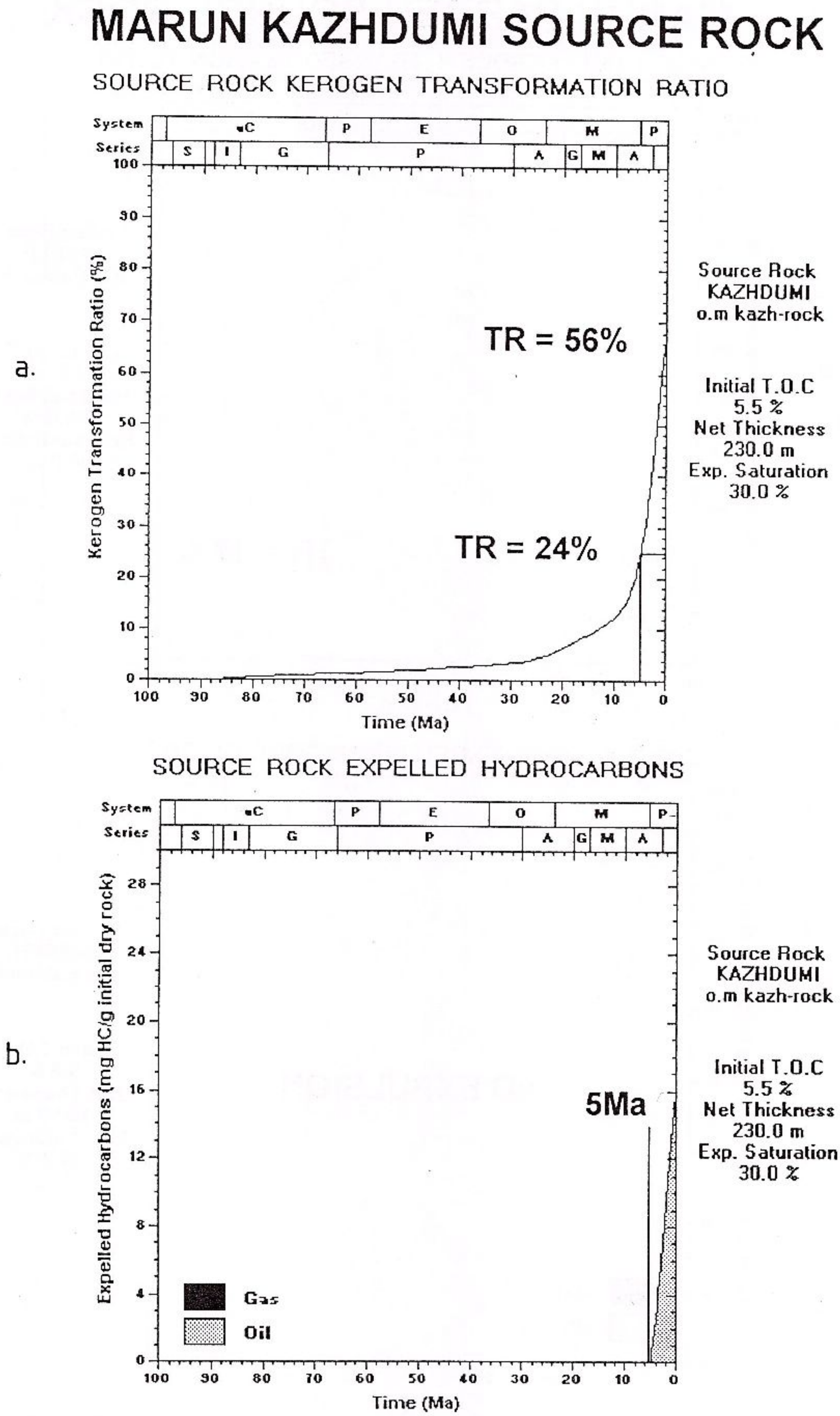


Fig. 25 – Kazhdumi source rock at Marun (top well): a. variation the Transformation Ratio (TR) as a function of time; b. expulsion window

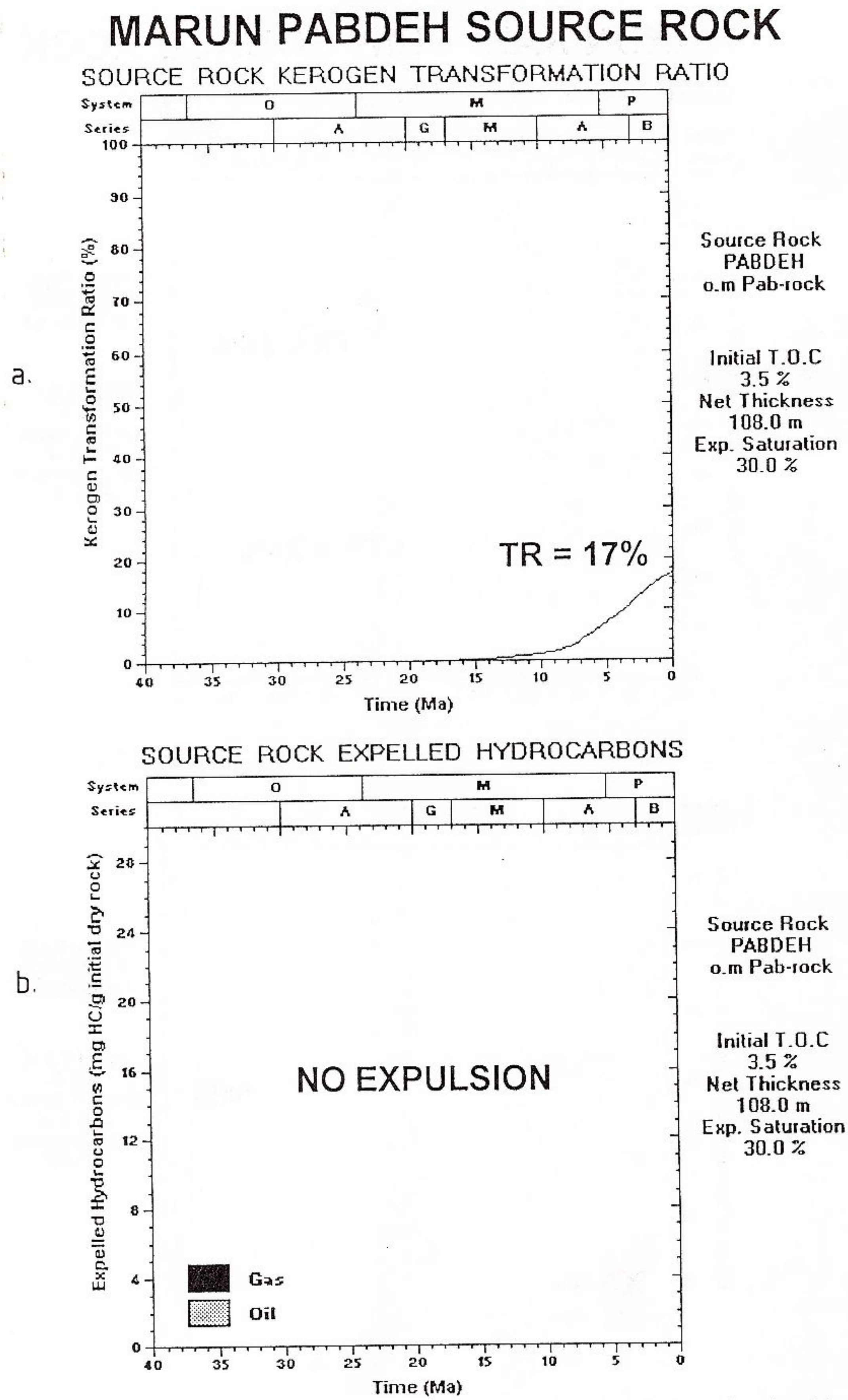


Fig. 26 – Pabdeh source rock at Marun (top well): a. variation of the Transformation Ratio as a function of time
b. expulsion window

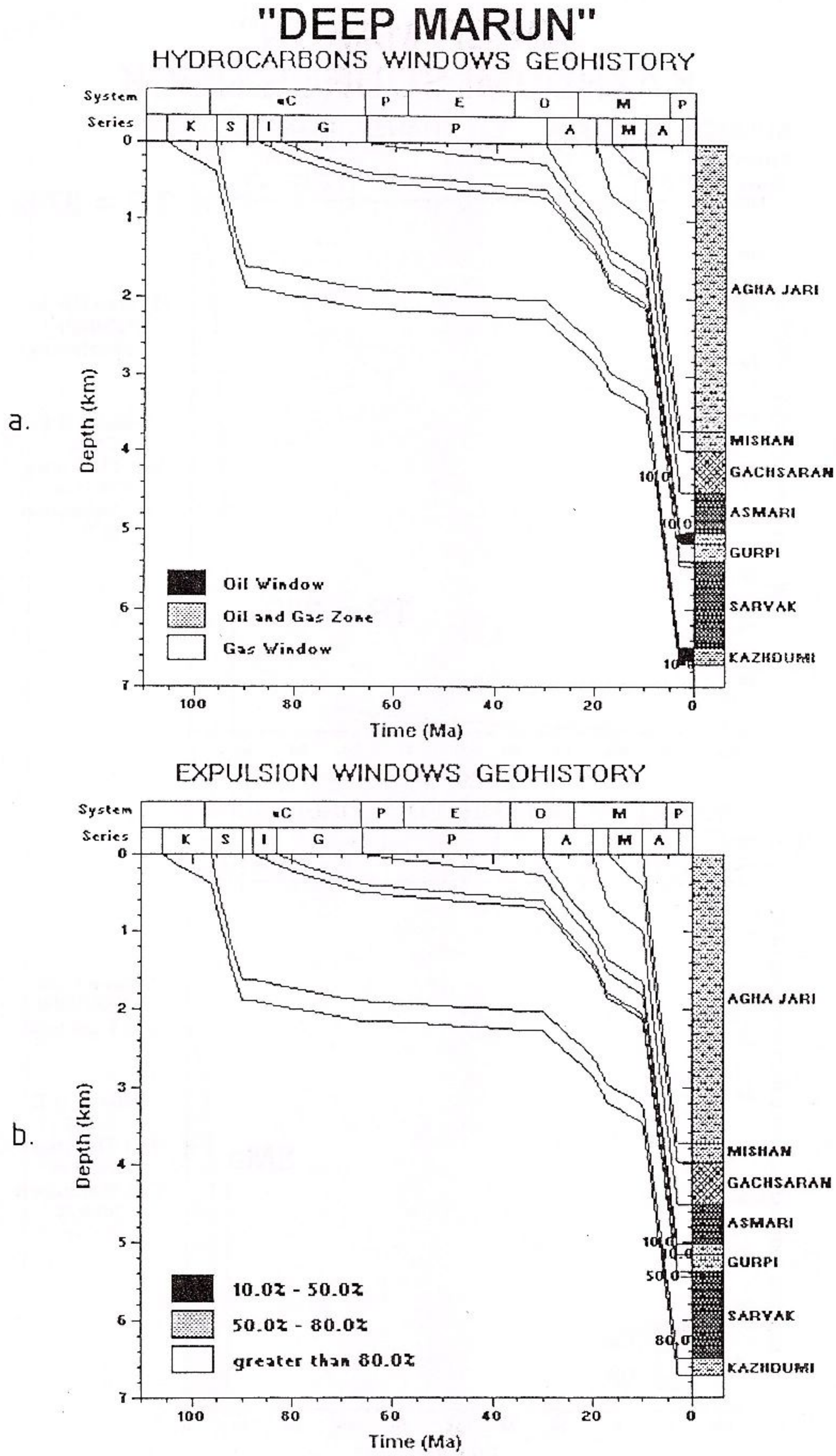


Fig. 27 – Deeper part of the Marun syncline burial profile: a. hydrocarbon window; b. expulsion window

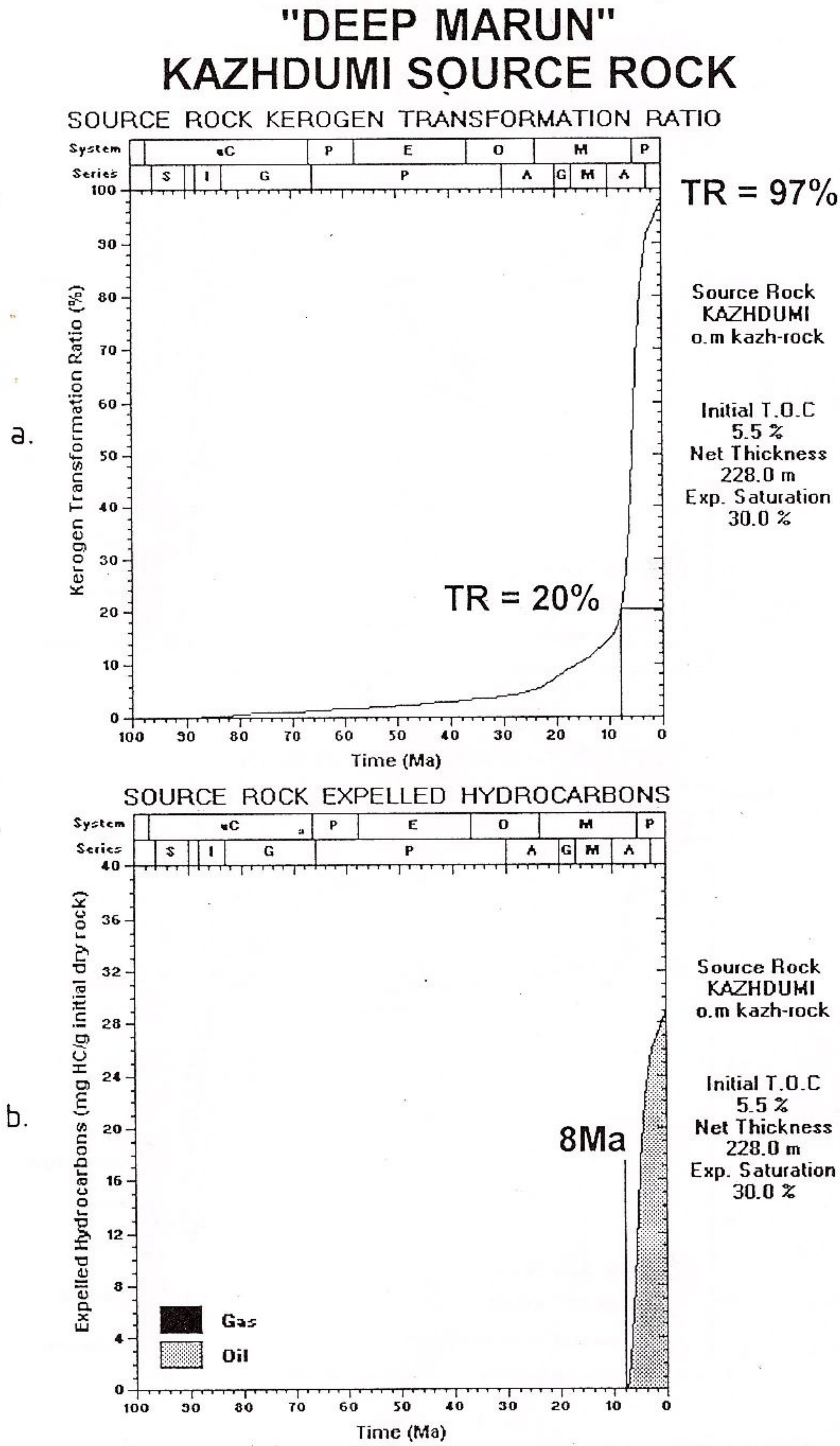


Fig. 28 – Kazhdumi source rock in the deeper part of the Marun syncline: a. variation of TR as a function of time b. expulsion window

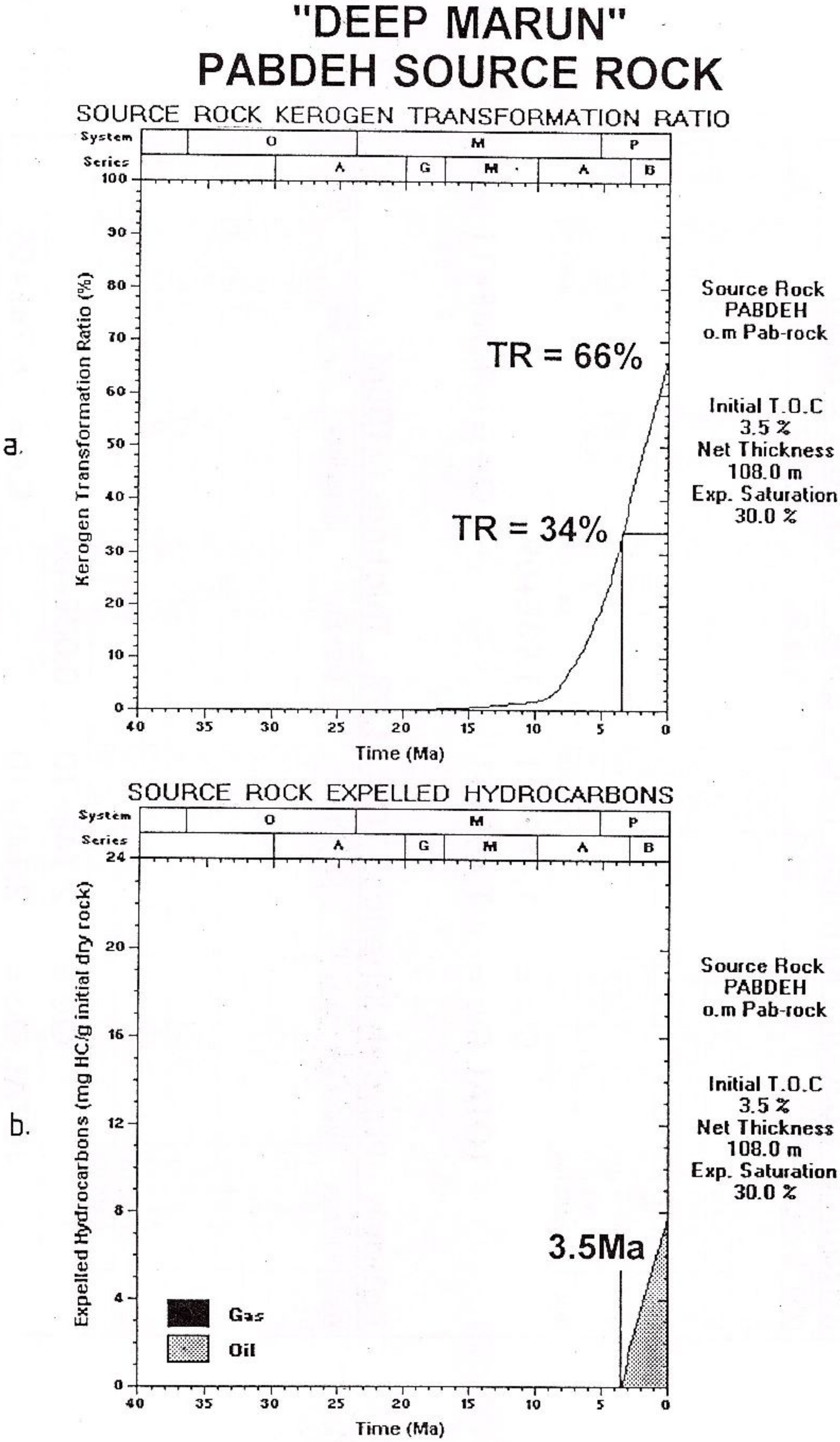


Fig. 29 – Pabdeh source rock in the deeper part of the Marun syncline: a. variation of TR as a function of time b. expulsion window

MARUN-123 (MN)

Kazdumi - Kazdumi Kinetics TOC = 5.5%, Thickness = 228m

Well	Surface	Total-Oil	Gas	Oil Qc	Gas Qc	Expelled	Qe	TR
S0	4.30E+07	13.8		5.93E+08	0.00E+00	9.17	3.94E+08	68.1
S1	7.86E+07	16.18	0.02	1.27E+09	1.57E+06	11.82	9.30E+08	79.5
S2	6.41E+07	17.33	0.04	1.11E+09	2.56E+06	13.27	8.50E+08	85.8
S3	8.65E+07	18.25	0.07	1.58E+09	6.05E+06	14.38	1.24E+09	90.6
S4	1.15E+08	18.91	0.14	2.18E+09	1.62E+07	15.29	1.76E+09	94.4
S5	1.70E+08	19.11	0.24	3.26E+09	4.09E+07	15.87	2.71E+09	96.4
S6	2.97E+08	18.97	0.39	5.63E+09	1.16E+08	16.25	4.82E+09	97.2
S7	1.06E+08	18.68	0.59	1.97E+09	6.22E+07	16.58	1.75E+09	97.8
SUM	**9.60E+08**			**1.76E+10**	**2.45E+08**		**1.45E+10**	

Qc = 1.38E+11 1.93E+09

TOTAL Qc = 1.40E+11 Qe = 1.14E+11

Pabdeh - Pabdeh Kinetics TOC=3.5%, Thickness=108m

Well	Surface	Total Oil	Gas	Oil Qc	Gas Qc	Expelled	Qe	TR
S0	4.30E+07	1.01		4.34E+07	0.00E+00	0	0.00E+00	17.1
S1	7.86E+07	1.26		9.91E+07	0.00E+00	0	0.00E+00	21.1
S2	6.41E+07	1.6		1.02E+08	0.00E+00	0	0.00E+00	26.7
S3	8.65E+07	2		1.73E+08	0.00E+00	0	0.00E+00	33.2
S4	1.15E+08	2.47		2.85E+08	0.00E+00	0.18	2.08E+07	41
S5	1.70E+08	3.06		5.22E+08	0.00E+00	0.89	1.52E+08	50.6
S6	2.97E+08	3.61		1.07E+09	0.00E+00	1.56	4.63E+08	59.4
S7	1.06E+08	4.05		4.27E+08	0.00E+00	2.1	2.22E+08	66.5
	9.60E+08			2.72E+09	0.00E+00		8.57E+08	

Qc = 2.14E+10 0.00E+00

TOTAL Qc = 2.14E+10 Qe = 6.74E+09

Fig. 30 – Field A - Calculation of the oil generated and expelled from the Kazhdumi and Pabdeh source rocks.

The amount expelled would be, when expressed in barrels:

$$Q_E = 19.65\ 10^3\ h\ (TOC.HI)\ (\sum S_i\ Exp_i)$$

Exp_i is the expulsion coefficient or ratio between the amount of oil expelled to the oil generated.

As an example, amounts of oil generated and expelled were calculated for a large size field (designated as Field A) of the Dezful Embayment, from the geometry of its drainage area and the source rock characteristics in the same area.

In the Marun drainage area A, the calculated oil in place generated by the Kazhdumi source rock amounted to 138 billion barrels of oil in-place, from which only 114 billion barrels were expelled. The Pabdeh would have generated 21.4 billion barrels of which only 6.7 billion barrels were expelled. The respective share of Kazhdumi and Pabdeh, for the oil **generated** in area A, would be 85 and 15 %. However, the Kazhdumi **expelled** 95 % of the total amount of oil and the Pabdeh only 5 % (Fig. 30).

The initial amount of oil (and equivalent gas) in-place trapped, mainly in the Asmari reservoir, connected with the Bangestan, corresponds roughly to 50 % of the calculated amount of oil (and equivalent gas) expelled from the Kazhdumi and Pabdeh source rocks in the area of drainage A, as defined by seismic data. The same calculation was done for all fields of the Dezful Embayment. **As an average, slightly less than 20 % of the oil calculated to have been expelled from the source rocks in the Dezful Embayment were found as oil-in-place in the discovered fields.** However, the discussion of the results are out of the scope of this paper.

Conclusion

In the Dezful Embayment, which part of the Iranian Zagros Foothills, excellent source rocks, Kazhdumi (Albian) and Pabdeh (Middle Eocene to Oligocene) coexisted with outstanding reservoirs, Asmari (Early Miocene) and Bangestan (mostly Cenomanian), the Asmari being capped by the efficient evaporite cap rock of the Gachsaran Formation (Early Miocene). Moreover huge whale-back anticlines provide large size traps. **However, the direct charge of these traps was possible only because the oil expulsion from source rocks occurred when Zagros folds were already formed.**

The Zagros folds were formed as the result of the collision of the Arabian Platform with the Central Iran Block which caused the closing of the Neo-Tethys. The chronology of the Zagros orogeny was studied both on the NE edge of the Arabian Platform, close to the plate limit and on the Platform from seismic data. Folds began to grow around 10 Ma ago.

The timing of the oil and gas generation and expulsion was calculated by modelling using burial profiles and kinetic parameters calculated for each of the two main source rocks, from pyrolysis experiments and application of the OPTKIN program. Heat flow assumptions which gave a reasonable fit between observed values of the actual temperature profiles, vitrinite reflectance profiles, source rock parameters, such as Tmax and Hydrogen Indices, and values calculated by the model, were used in this exercise. Oil was found to be expelled from the source rocks at 8 Ma at the earliest in the deeper synclinal areas.

Because of these result, the concept of **drainage area** can be used for modelling, i.e. an area of drainage can be defined, from seismic information and **the oil-in-place to be found in the anticline.** Moreover, the fracturing of the interval Bangestan to Asmari, resulting from the folding facilitated vertical migration towards the Asmari capped by the Gachsaran evaporites. **The same method can be used as a previsional tool for undrilled anticlines, provided the geometry of their area of drainage is known from seismic data and that source rock characteristics can be extrapolated.**

As a conclusion, the unusual gathering of giant oil fields of the Dezful Embayment is in part due to Neogene events, folding as well as oil expulsion, which both occurred during the deposition of the thick molasse-type Agha Jari Formation, from Late Miocene to Pliocene.

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STRATIGRAFIA DEPOZITELOR NEOGENE DIN ZONA HUDEŞTI – PLATFORMA MOLDOVENEASCĂ –

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Key words: Moldavian Platform. Neogene. Lithostratigraphy. Biostratigraphy.

Abstract: *Stratigraphy of the Neogene Deposits in the Hudeşti Area - Moldavian Platform.* The study of the cores from seven drillings, localized in the Hudeşti area, allows the detailed litho and biostratigraphic characterization of the Neogene sedimentary succession, extended on a surface of about 5 q.km. It has been noticed that the Neogene deposits, unconformably overlying the Cretaceous formations, consist only of Upper Kossovian and Lower Volhynian sediments, separated by a marly - limestone marker level. The Upper Kossovian succession is made up of several lithologic sequences, as followes, from bottom to top: calcareous, sandy, gypsiferous and marly sequences. The fossiliferous content of these deposits is represented by foraminifera belonging to the Velapertina Zone (Popescu, 1979) and calcareous nannofossils typical of the Discoaster exilis - NN6 Zone (Syracolithus dalmaticus - NN6c and Calcidiscus leptoporus - NN6d subzones, Mărunţeanu et al., in press). Both fossiliferous communities define the Late Kossovian. The Lower Volhynian formations are also constituted of the following lithologic sequences, from bottom to top: marly - clayey, clayey - siltic, clayey and sandy with pebbles. The three types of mollusc assemblages with *Obsoletiforma ruthenica* + *Mohrensternia*, *Mactra (Sarmatimactra) eichwaldi* + *Ervilia* and *Duplicata* + *Pirenella*, typical of the Early Volhynian, have been identified in these deposits. The same age has also been confirmed by the foraminifera belonging to the Varidentella reussi and Elphidium reginum zones (Popescu, 1995) and by the nannoplankton communities, that characterise the Braarudosphaera bigelowii and Calcidiscus leptoporus centrovalis zones (Mărunţeanu, 1999).

Depozitele neogene ale cuverturii Platformei Moldoveneşti au constituit subiectul unui mare număr de lucrări ştiinţifice, aparţinând în special cercetătorilor şcolii ieşene de geologie. Raritatea deschiderilor naturale, mai ales în partea nordică a platformei, au îngreunat mult descifrarea constituţiei litologice şi corelarea depozitelor miocene.

Pornind de la premiza că datele furnizate de foraje pot întregi imaginea geologică locală sau regională, prezentăm în cele ce urmează un studiu lito şi biostratigrafic detaliat al depozitelor Miocenului mediu, din arealul localităţii Hudeşti. Acest studiu a fost realizat prin investigarea carotelor unui număr de 7 foraje, existente în Litoteca Institutului Geologic al României.

Menţionez faptul că asociaţiile de moluşte, foraminifere şi nannoplancton calcaros au fost determinate de colegii Dr. I. Papaianopol, Dr. Gh. Popescu şi respectiv Dr. M. Mărunţeanu, cărora le mulţumesc pe această cale.

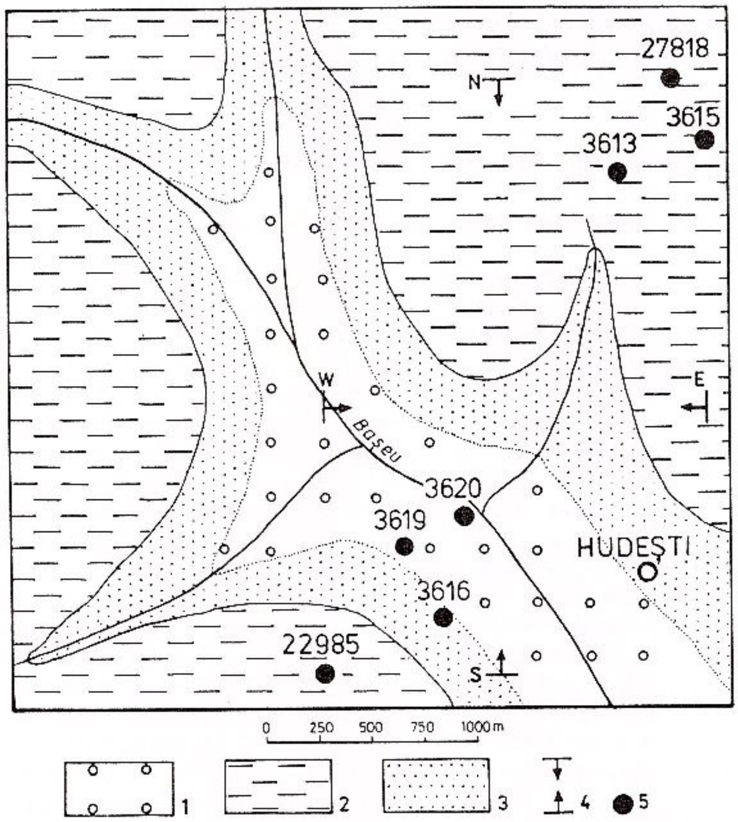
1. Încadrarea geologică a zonei investigate

Zona investigată este situată în partea nordică a Platformei Moldoveneşti, între râurile Siret şi Prut, pe teritoriul localităţii Hudeşti, la izvoarele văii Başeu (fig. 1).

Platforma Moldovenească, parte integrantă a Platformei Europei Orientale (Săndulescu, 1984), este alcătuită dintr-un soclu epicarelian şi o cuvertură ce debutează cu formaţiuni vendiene (Patrulius, Iordan, 1974), urmate de depozite ce aparţin următoarelor cicluri de sedimentare: Cambrian, Ordovician, Devonian, Jurasic superior - Eocretacic, Cretacic superior şi Neogen.

Ciclul neogen cuprinde, în succesiune stratigrafică normală, sedimente badenian - superioare (kossoviene), sarmaţiene şi meoţiene (Mutihac, Ionesi, 1975).

În partea nordică a Platformei Moldoveneşti, între Siret şi Prut, unde se află amplasat şi arealul Hudeşti, ciclul neogen este alcătuit în exclusivitate din depozite badenian - superioare (kossoviene) şi sarmaţian - inferioare (volhyniene). Cele badeniene au fost împărţite de Paghida - Trelea (1969) în trei orizonturi şi



1, Cuaternar; 2, Volhynian inferior – partea inferioară (Buglovian); 3, Volhynian inferior – partea superioară; 4, poziţia secţiunilor geologice; 5, poziţia forajelor studiate

Fig. 1 – Amplasarea forajelor studiate în aria Hudeşti

(Schiţă geologică după Harta Geologică a României 1:200000, foaia I. Darabani)

anume: orizontul conglomeratelor cu silex şi al nisipurilor cuarţoase, ce trec lateral la gipsuri sau anhidrite, orizontul calcarelor şi marnelor cu *Lithothamnium* şi orizontul marnos - argilos cu rare intercalaţii de tufuri sau bentonite.

Succesiunea volhyniană, transgresivă peste cea badeniană (Ionesi, Ionesi, 1983), debutează cu sedimente organogene (Atanasiu, Macarovici, 1950), considerate a aparţine Buglovianului (Ionesi, Ionesi, 1982). În aria localităţii Hudeşti, faciesul organogen este înlocuit de unul terigen, predominant marno - argilos, cu o intercalaţie de tuf (Macarovici, Jeanrenaud, 1958), denumită de Simionescu (1902) Tuful de Hudeşti.

Stiva volhyniană se continuă cu depozite nisipo - conglomeratice (Macarovici, 1958) cu rare intercalaţii argiloase, cunoscute sub numele de "orizontul inferior al Volhynianului" (Macarovici, Jeanrenaud, 1958). La sud de Hudeşti, urmează în continuitate de sedimentare "orizontul superior al Volhynianului" (Macarovici, Jeanrenaud, 1958), separat de cel inferior prin Calcarele de Burdujeni (Ionesi, 1968).

Conţinuturile fosilifere, cantonate în depozitele Miocenului mediu din Platforma Moldovenească, au fost studiate în decursul anilor de numeroşi cercetători (Cobălcescu, 1883; Huică, Lubenescu, 1982; Ionesi, 1968, 1980; Ionesi et al., 1995; etc.), ale căror interpretări biostratigrafice au permis datarea, corelarea şi evoluţia succesiunii sedimentare în timpul Neogenului.

2. Litostratigrafie

În urma studiului carotelor celor 7 foraje amplasate în perimetrul Hudeşti (fig. 1), s-a observat că stiva depozitelor badenian superioare - sarmaţian inferioare admite a fi împărţită în mai multe secvenţe litologice, după tipul predominant de rocă (fig. 2 - 8).

Astfel, formaţiunile badeniene încep cu o secvenţă calcaroasă, alcătuită din calcare masive cu *Lithothamnium*, decimetrice, cu intercalaţii centimetrice de calcare oolitice bogat fosilifere (foraj 27818, fig. 2), gresii (foraj 3615, fig. 3) sau mai rar brecii calcaroase (foraj 3620, fig. 5) şi chiar argile, uneori siltice (forajul

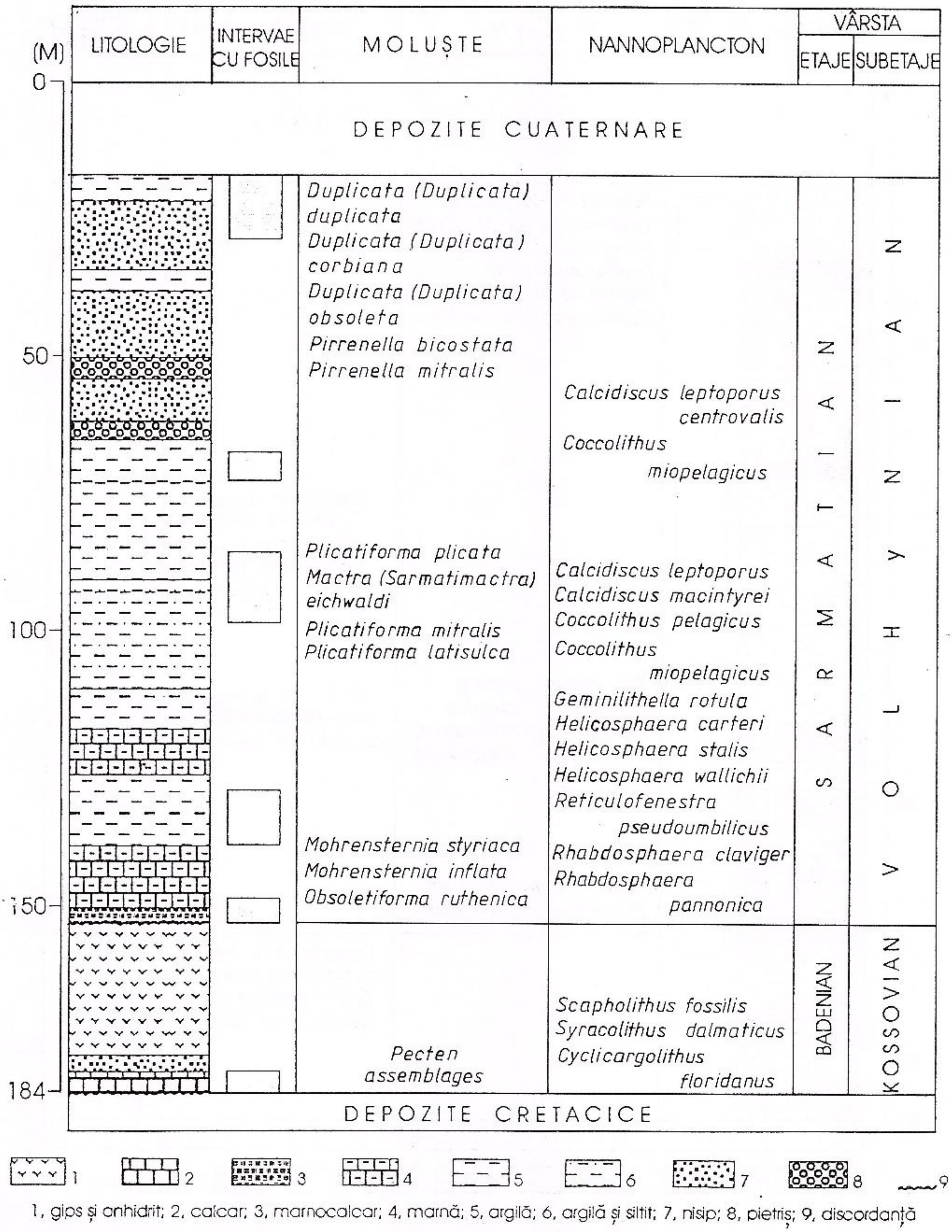


Fig. 2 – Stratigrafia depozitelor neogene. Foraj: 27818 Hudești (230,83 m).

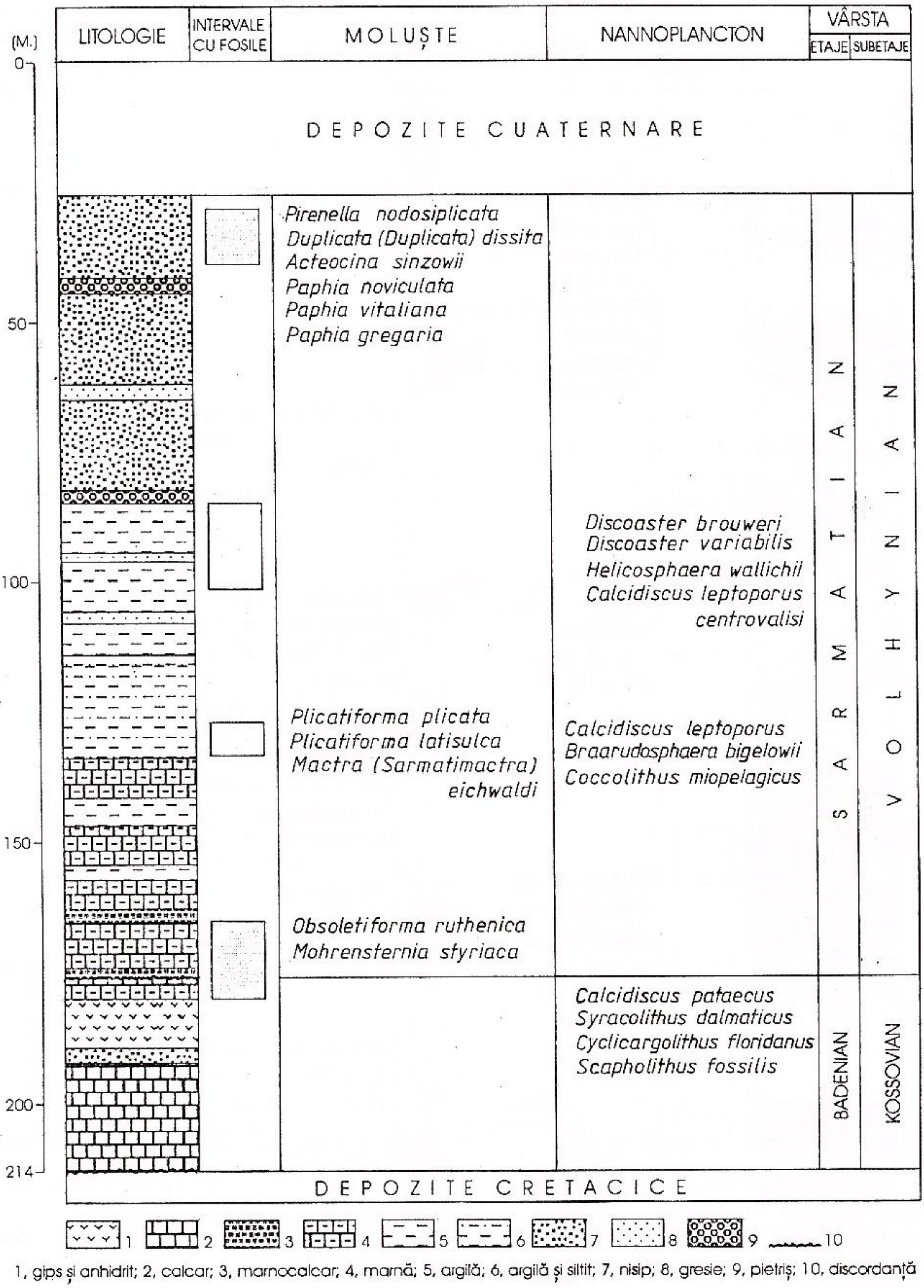


Fig. 3 – Stratigrafia depozitelor neogene. Foraj: 3615 Hudești (253,29 m).

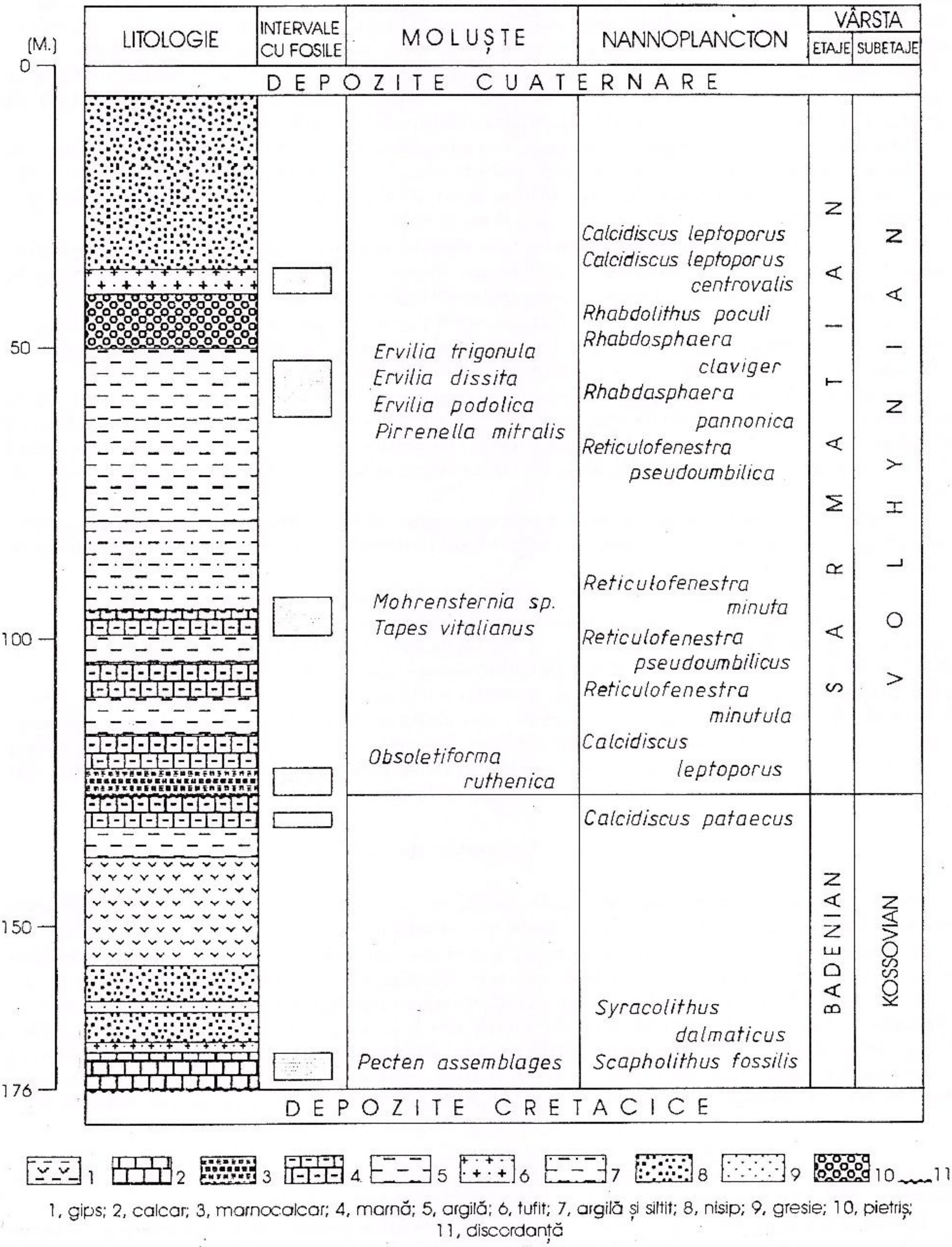


Fig. 4 – Stratigrafia depozitelor neogene. Foraj: 3613 Hudeşti (216,51 m).

3616, fig. 7). Calcarele oolitice, mai frecvente în nordul perimetrului (forajele 27818, 3615 şi 3613, fig. 1), apar sporadic în sudul acestuia. Faciesul calcaros este înlocuit, în extremitatea sudică a zonei investigate, prin marnocalcare cenuşii sau albicioase, compacte (foraj 22985, fig. 8). Într-un singur foraj (3613, fig. 4), în partea superioară a acestei secvenţe, a fost observată o intercalaţie de tufit, de 0,75 metri grosime.

În continuitate de sedimentare urmează secvenţa nisipoasă, alcătuită din nisipuri cenuşii, medio la fin granulare, cu numeroase elemente de cuarţ şi cuarţite. Nisipurile admit intercalaţii de gresii subţiri, de grosimi şi frecvenţe variabile, în care pot fi identificate resturi de Pectinidae (22985, fig. 8).

Urmează, în succesiune stratigrafică normală, secvenţa gipsiferă, ce constă din alternanţe, mai mult sau mai puţin regulate, de gipsuri albe zaharoide şi anhidrite cenuşii - negricioase cu zone brecioase (foraj 3613), între care se pot intercala gresii calcaroase oolitice (foraj 3613) sau marnocalcare cu spărtură concoidală (foraj 3615), sub forma unor strate de maximum 15 cm grosime.

Ultima secvenţă a stivei badeniene, marnoasă, este alcătuită în exclusivitate din marne albicioase masive, frecvent fosilifere, uneori conţinând resturi cărbunoase, eflorescenţe de sulf sau chiar cristale de pirită (în forajele 3613 şi 3615). Lateral, ea are grosimi variabile, efilându-se în forajul 27818 (fig. 2).

Succesiunea sarmaţiană (volhynian - inferioară) poate fi şi ea subdivizată în mai multe secvenţe litologice, interceptate în totalitate mai ales de forajele din nordul ariei investigate (fig. 2-4). Spre sud, ultimele secvenţe sarmaţiene (fig. 5-8) au fost înlăturate parţial sau total de eroziunea post volhyniană.

Secvenţa argilo-marnoasă, dispusă discordant peste depozitele badeniene, are în bază un nivel reper de marnocalcare cenuşii, străbătute de numeroase diaclaze umplute cu gips secundar fibros. Ea este formată din alternanţe de argile şi marne, siltice pe suprafeţele de stratificaţie, ce admit rare şi subţiri intercalaţii de siltite sau chiar marnocalcare (foraj 3615, fig. 3). De regulă în baza secvenţei apar numeroase resturi de macrofaună.

În continuitate de sedimentare se dezvoltă secvenţa argilo - siltică, alcătuită din alternanţe decimetrice de argile fin stratificate şi siltite, puternic micacee. Uneori (forajele 27818 şi 3615) siltitele conţin cuiburi de pirită şi resturi organice vegetale.

Urmează secvenţa argiloasă, constituită în exclusivitate din argile cenuşii, masive.

Cea mai nouă secvenţă volhynian - inferioară este dată de cea nisipoasă, cu intercalaţii frecvente de pietrişuri în partea inferioară şi argile siltice în cea superioară. Ea a fost interceptată numai de forajele 27818, 3615 şi 3613 (fig. 2 - 4), localizate în nordul perimetrului Hudeşti. Nisipurile, medio - granulare până la grosiere, sunt de regulă slab micacee şi conţin numeroase elemente de cuarţ, cuarţite şi calcare. Pietrişurile, fin la medio - granulare, slab sortate, sunt alcătuite din galeţi de roci metamorfice, frecvente cuarţite, calcare albe sau roşii, gresii cenuşii dure sau albe cuarţitice, de tip Kliwa. Într-un singur foraj (3613, fig. 4), în partea inferioară a acestei secvenţe, apare un pachet de aproximativ 1,5 m de tufite albe cu nannofosile calcaroase, ce aminteşte de nivelul reper al Tufului de Hudeşti (Simionescu, 1902).

3. Biostratigrafie

Întreaga succesiune sedimentară a Miocenului mediu, interceptată de forajele studiate, este foarte bogată în resturi fosile (fig. 2 - 8), în special de moluşte şi nannoplancton.

În secvenţele badeniene, calcaroasă şi nisipoasă, au fost observate diferite specii de Pectinidae, însoţite de nannofosile calcaroase aparţinând Zonei Discoaster exilis (Martini, 1971). Ultimele, sunt reprezentate printr-o comunitate cu *Coccolithus pelagicus* (WALLICH), *Coccolithus miopelagicus* BUKRY, *Calcidiscus leptoporus* (MURRAY & BLACKMAN), *Calcidiscus macintyrei* (BUKRY & BRAMLETTE), *Helicosphaera carteri* (WALLICH), *Helicosphaera stalis* THEODORIDIS, *Helicosphaera wallichii* (LOHMANN), *Helicosphaera walbersdorfensis* MÜLLER, *Cyclicargolithus floridanus* (ROTH & HAY), *Reticulofenestra pseudoumbilicus* (GARTNER), *Syracolithus dalmaticus* (KAMPTNER) şi *Scapholithus fossilis* DEFLANDRE. Întreaga asociaţie defineşte Subzona Syracolithus dalmaticus - NN6c (Mărunţeanu et al., in press), tipică Kossovianului superior.

În partea terminală a secvenţei marnoase badeniene (fig. 3, 4, 7, 8), în ansamblul de nannofosile menţionat mai sus, îşi face apariţia specia *Calcidiscus pataecus* (GARTNER) care anunţă debutul Subzonei Calcidiscus pataecus - NN6d, ce marchează sfârşitul Kossovianului (Mărunţeanu et al., 1999).

Într-un singur foraj, 3620 (fig. 5), a fost identificată o asociaţie de foraminifere aparţinând Zonei Velapertina (Popescu, 1979), cu *Velapertina iorgulescui* POPESCU, *Velapertina indigena* (LUCZ.), *Haplophragmoides fragilis* HOGL., *Spiroplectammina marinae* (d'ORBIGNY), *Textularia pala* CZJZ., *Siphotextularia concava* (KARRER), *Martinotiella communis* d'ORBIGNY, *Sigmoilinita tenuis* (CZJZ.), *Triloculina gibba* d'ORBIGNY,

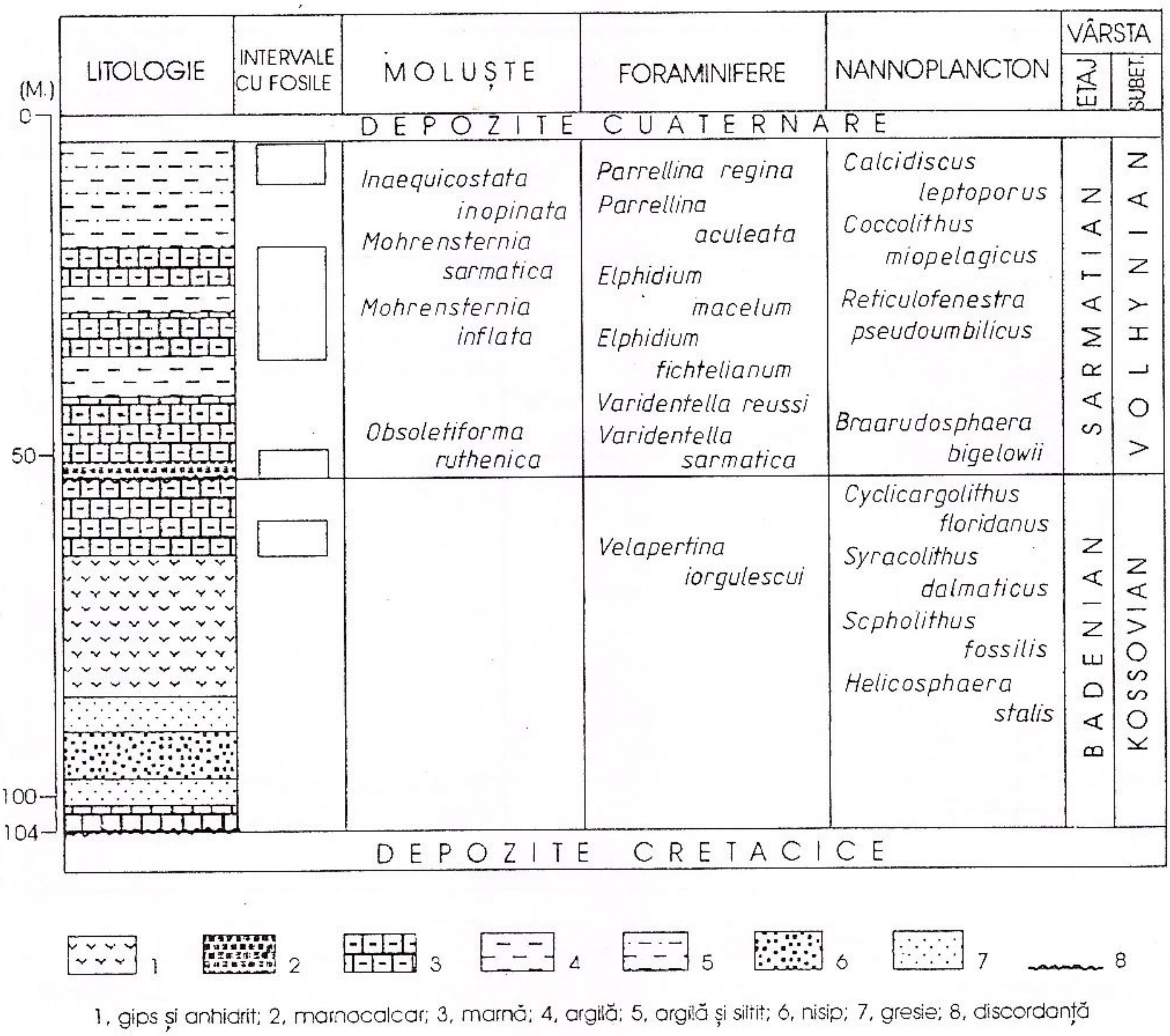


Fig. 5 – Stratigrafia depozitelor neogene. Foraj: 3620 Hudeşti (42,39 m).

Baggatella gutsulica (LIVENTAL), *Cribononion notabilis* (PISHVANOVA).

Depozitele sarmaţiene conţin trei tipuri de comunităţi de moluşte, care în ordine stratigrafică normală se succed astfel: asociaţia cu *Obsoletiforma ruthenica*, cu *Mactra* şi *Ervilia* şi cu *Duplicata* şi *Pirenella*.

Primul tip de asociaţie, caracteristic Buglovianului (Ionesi, Ionesi, 1982) sau debutului Volhynianului (Papaianopol, 1998), a fost identificat în secvenţele marno - argiloasă şi argilo - siltică, fiind alcătuit din *Obsoletiforma ruthenica* (HILBER), *Mohrensternia styriaca* HILBER, *Mohrensternia inflata* (ANDRZEJOWSKI), *Tapes vitalianus* d'ORBIGNY, *Inaequicostata inopinata* GRISCH.

Al doilea tip de asociaţie, tipic pentru Volhynianul inferior (Papaianopol, 1998), se dezvoltă în partea terminală a secvenţei argilo - siltice şi în secvenţa argiloasă. Este definit de ansamblul speciilor *Mactra (Sarmatrimactra) eichwaldi* LASKAREV, *Plicatiforma mitralis* (EICHWALD), *Plicatiforma latisulca* (SIMIONESCU & BARBU), *Plicatiforma plicata* (EICHWALD), *Ervilia trigonula* SOKOLOV, *Ervilia dissita* (EICHWALD), *Ervilia podolica* (EICHWALD).

Ultimul tip de asociaţie, identificat în partea terminală a secvenţei argiloase şi în secvenţa nisipoasă cu pietrişuri, conţine *Pirenella nodosiplicata* (HOERNES), *Duplicata (Duplicata) dissita* (DUBOIS), *Duplicata*

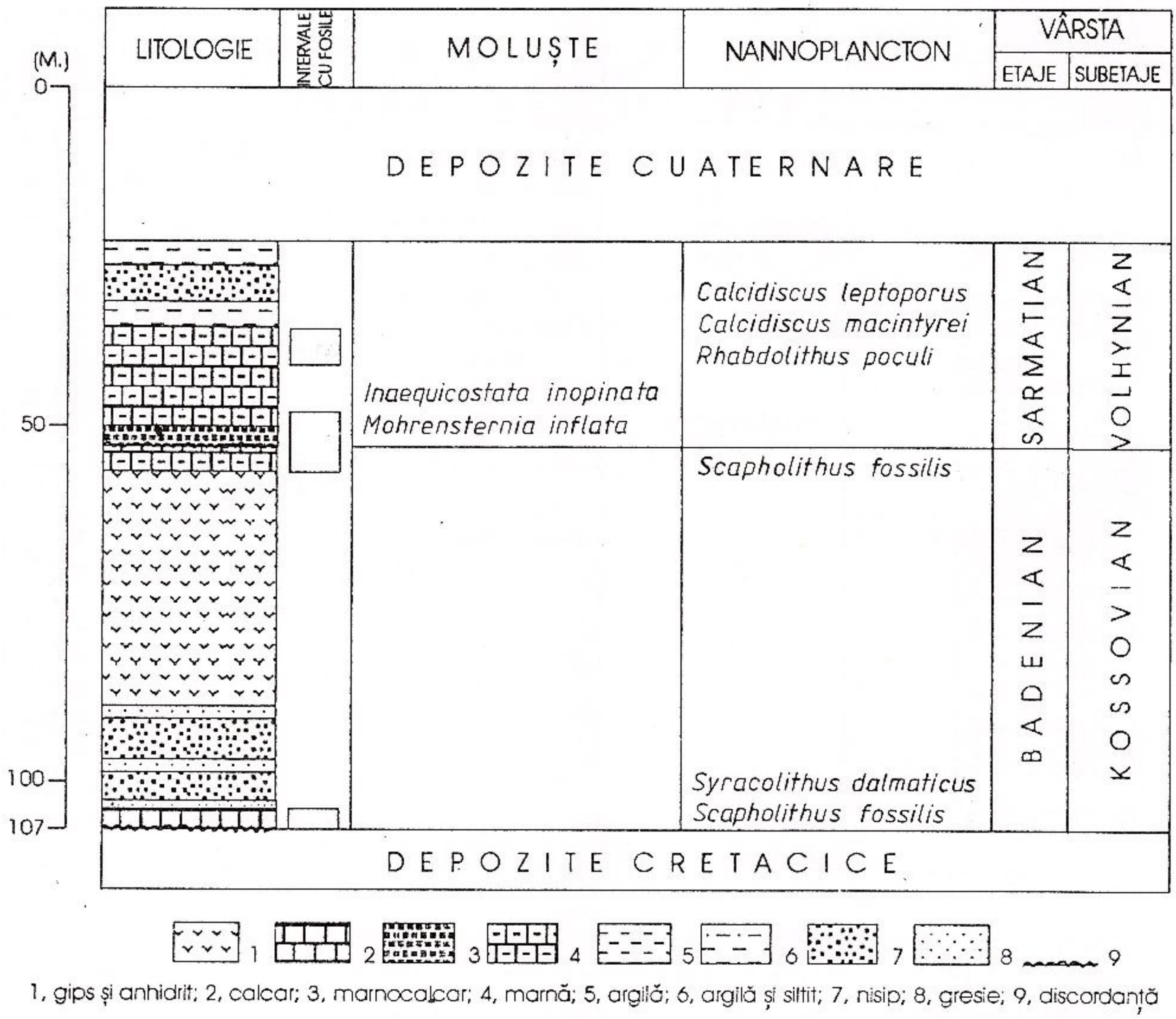


Fig. 6 – Stratigrafia depozitelor neogene. Foraj: 3619 Hudești (142,20 m).

(Duplicata) corbiana (d'ORBIGNY), *Acteocina sinzowii* (KOLESNIKOV), *Paphia naviculata* (R. HOERNES), *Paphia vitaliana* (d'ORBIGNY), *Paphia gregaria* (PARTSCH) și *Plicatiforma mitralis* (EICHWALD). Comunitatea de moluște menționată caracterizează partea superioară a Volhynianului inferior.

Nannofosile calcaroase, care însoțesc cele trei tipuri de asociații de moluște, atestă vârsta volhynian inferioară pentru depozitele cantonatoare. Ele pot fi încadrate în Zonele Braarudosphaera bigelowii și Calcidiscus leptoporus centrovalis (Mărunțeanu, 1998), corelabile cu Zona Discoaster kugleri - NN7 din scara standard de nannoplancton (Martini, 1971).

Primei biozone îi aparțin speciile *Braarudosphaera bigelowii* (GRAN & BRAARUD), *Calcidiscus leptoporus* (MURRAY & BLACKMAN), *Calcidiscus macintyrei* (BUKRY & BRAMLETTE), *Coccolithus pelagicus* (WALLICH), *Coccolithus miopelagicus* BUKRY, *Discoaster brouweri* TAN, *Discoaster variabilis* MARTINI & BRAMLETTE, *Helicosphaera carteri* (WALLICH), *Helicosphaera walbersdorfensis* MÜLLER, *Helicosphaera stalis* THEODORIDIS, *Reticulofenestra pseudoumbilicus* (GARTNER), *Rhabdosphaera pannonica* BALDI - BEKE, *Rhabdosphaera claviger* MURRAY & BLACKMAN și *Sphenolithus abies* DEFLANDRE.

Chiar sub primele nivele de pietrișuri, în asociația de nannoplancton mai sus menționată, a fost identificată prima ocurență a speciei *Calcidiscus leptoporus centrovalis* STRADNER, bioeveniment ce marchează debutul biozonei cu același nume.

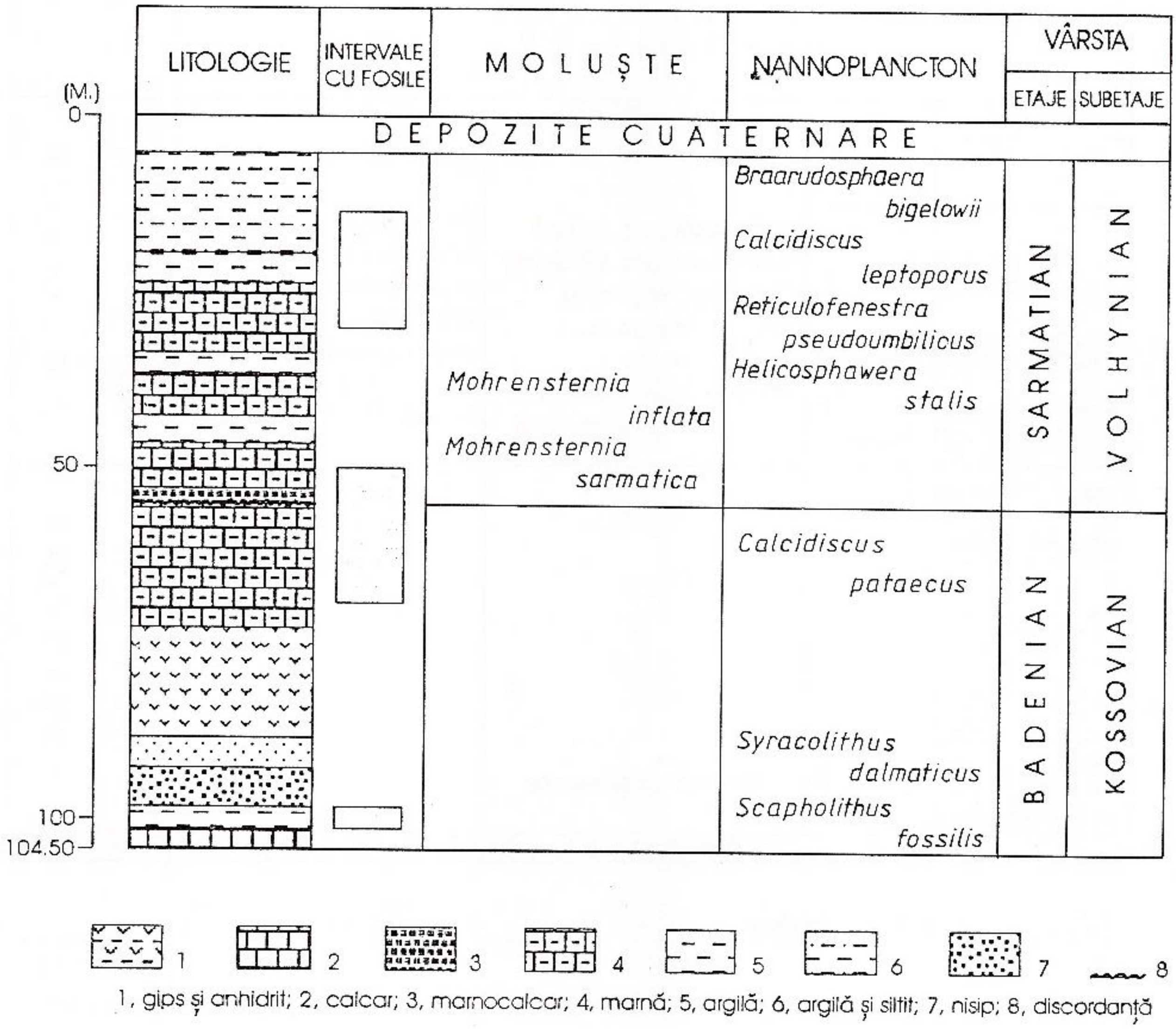


Fig. 7 – Stratigrafia depozitelor neogene. Foraj: 3616 Hudeşti (142,99 m).

În forajul 3620 (fig. 5) secvenţa argilo - marnoasă conţine asociaţii de foraminifere cu *Varidentella reussi* (BOGD.), *Varidentella sarmatica* (KARRER), *Varidentella rotundata* (GERKE), *Varidentella pseudocostata* (VENGL.), *Articulina articulinoides* (GERKE & ISSAREVA), *Bolivina moldava* DIDK., *Porosononion hyalinum* BOGD., aparţinând Zonei Varidentella reussi (Popescu, 1995), ce caracterizează Volhynianul inferior.

Spre topul secvenţei argilo-siltice, interceptată de forajul 3620 (fig. 5), asociaţia menţionată mai sus se completează cu două noi specii, *Elphidium aculeatum* d'ORBIG. şi *Elphidium reginum* (d'ORBIG.), anunţând debutul Zonei Elphidium reginum (Popescu, 1995), volhynian inferioară.

4. Corelarea secvenţelor litologice

Pentru a urmări variaţia laterală a grosimii şi litofaciesului secvenţelor separate în succesiunile de depozite kossovian - superioare şi volhynian - inferioare, au fost proiectate forajele studiate pe două aliniamente, orientate nord - sud şi est - vest, situate la jumătatea distanţei dintre primul şi ultimul foraj (fig. 1).

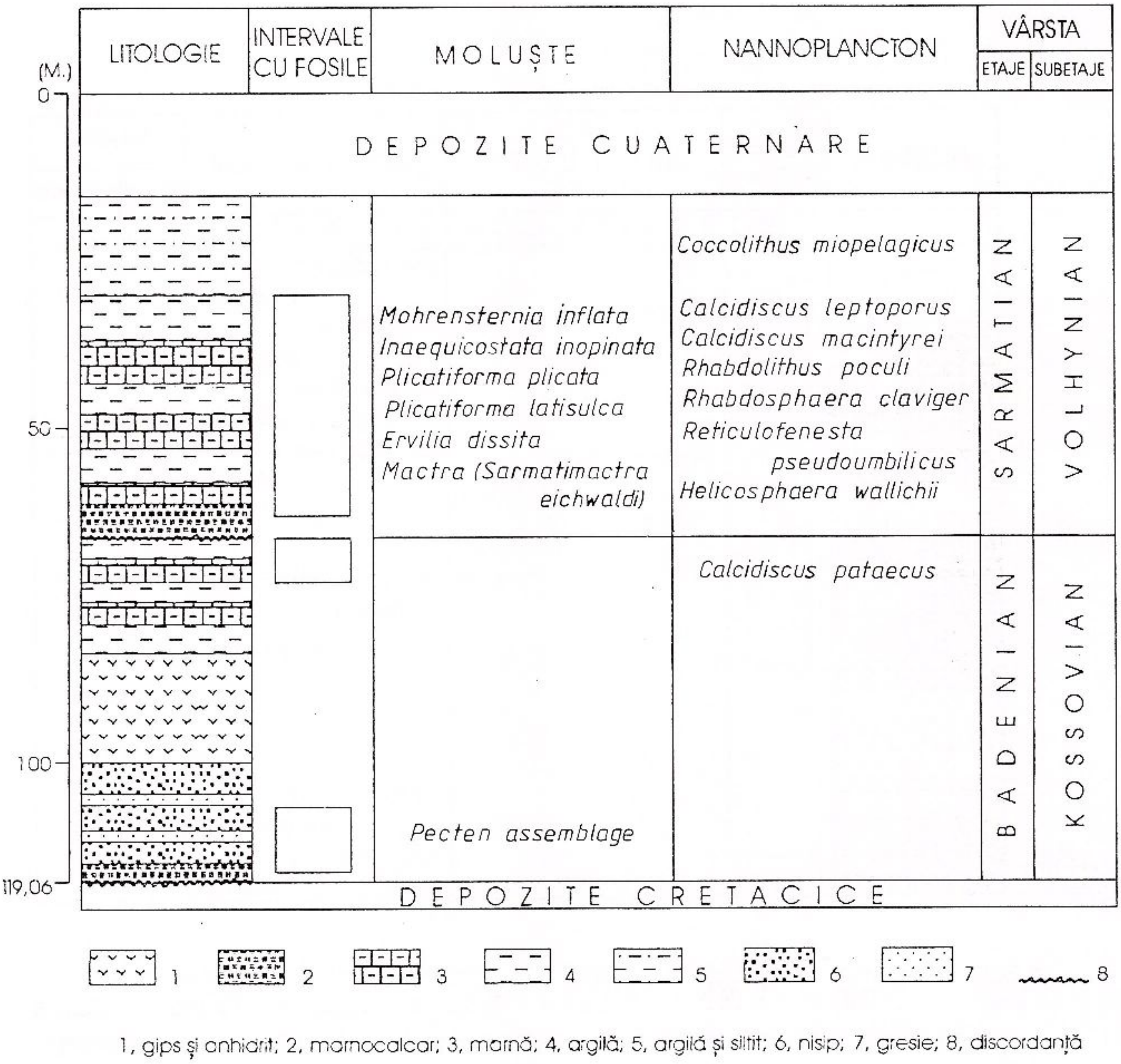
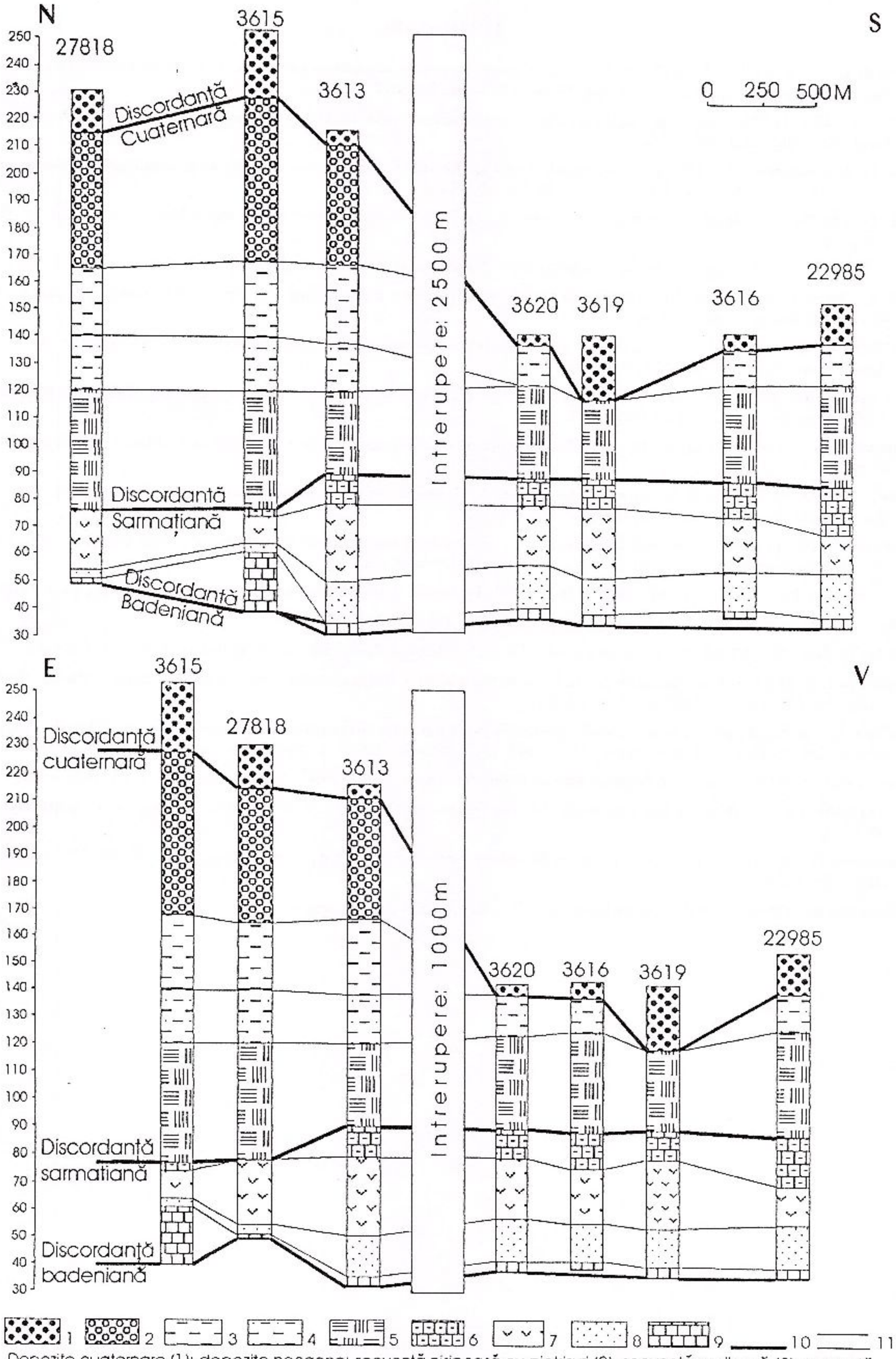


Fig. 8 – Stratigrafia depozitelor neogene. Foraj: 22985 Hudeşti (154,64 m).

Din corelarea secvenţelor litologice dezvoltate in arealul Hudeşti (fig. 9), se pot constata următoarele:

- depozitele kossovian-superioare, discordante, s-au depus peste un paleorelief cretacic, destul de uniform, care prezenta o uşoară ridicare (aproximativ 20 m), la nord de forajul 3615; altitudinea maximă a acestei ridicări se află in apropierea locaţiei forajului 27818;
- aspectul liniei de discordanţă sarmaţiană relevă o coborâre, de aproximativ 15 m, a sedimentelor Volhynianului inferior, la nord şi est de forajul 3613;
- discordanţa sarmaţiană este evidenţiată atât prin dispunerea depozitelor Volhynianului inferior peste diferiţi termeni ai Kossovianului superior cât şi prin lipsa Zonei Lobatula dividens (Popescu, 1995) de foraminifere, identificată de regulă in baza formaţiunilor volhyniene;
- linia extrem de sinuoasă a discordanţei cuaternare şi dispariţia parţială sau totală a ultimelor secvenţe litologice volhyniane la vest şi sud de forajul 3613, pot fi explicate printr-o eroziune diferenţială post Volhynian inferior;
- uşoarele variaţii laterale ale grosimilor secvenţelor litologice separate se datorează probabil frecventelor schimbări laterale de facies, caracteristice depunerilor molasice.



Depozite cuaternare (1); depozite neogene: secvență nisipoasă cu pietrișuri (2), secvență argiloasă (3), secvență argilo-siltică (4), secvență marnoasă cu marnocalcar (5), secvență marnoasă (6), secvență gipsiferă (7), secvență nisipoasă (8), secvență calcaroasă (9); discordanță (10); variația orizontală a grosimii secvențelor (11)

Fig. 9 - Corelarea secvențelor litologice

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THE TEMPORAL CONTENT OF THE NEOGENE COAL-GENERATING CYCLES IN ROMANIA

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Key words: Coal. Climatic cycles. Pliocene. Miocene. Valach cycle. Glaciary phases. Orbital eccentricity. Precessional cycles.

Abstract: The sequential study of the Dacian-Romanian (Pliocene - 5.3-1.8 Ma) coal bearing deposits in the western side of the Dacic Basin (Oltenia) suggests the existence of a maximum development of the coal facies in the lower part of the Romanian (3.2 Ma). This maximum has been considered the opposite of the climatic minimum that corresponds to the Quaternary glaciation (Riss Alpine phase) and leads to the conclusion of the existence of a climatic cycle (named by us **"the Valach cycle"**) with a period of about 4.1 Ma. The cooling period preceeding the Quaternary ones is near the Dacian/Pontian limit. The period of the climatic maximum before the Pliocene one is to be found inside the Pontian (Portaferrian) in which there is (the Lugoj Basin) a coal facies similar to the Pliocene one. Considering the period of these climatic cycles, it is easy to clearly determine at least the moments of maximum heating during the Neogene. So, at the end of the Oligocene the next maximum heating periods can be distinguished, that are marked by the existence of coal facies: Chattian-Lower Aquitanian, Burdigalian, Badenian (Upper Langhian-Kosovian), Sarmatian, Portaferrian and Pliocene. Natural transgressions correspond to these climatic maximum moments and regressions correspond to the opposite moments (the cold ones). The same sequential study has emphasized that the effective generation of the coal beds, inside the Valach cycle is determined by other climatic cycles with a period of about 100,000 years (the short cycles of the orbital eccentricity) that can easily explain the glacial and the interglacial Quaternary phases. Evidently associated transgressions and regressions accompany these cycles, too.

1. Introduction

Studies regarding the Pliocene and Miocene coal facies of the Romanian coal generating basins have led to the problems related to the distinguished cycles temporal content. Such observations, linked with the Pliocene coal complex, have been extended to the other Miocene and Pliocene coal complexes developed in Romania or other neighbouring countries (Serbia, Bulgaria), with coal generating basins; this pointed out interesting results, which can be supported by studies on the Western Europe Neogene coal facies but also on Neogene coal deposits from other continents (Asia, South America, etc.).

It is to note that the first evaluations of the laws regularities, which controlled the occurrence and the development of the coal facies, have allowed to distinguish both a cycle (named "Valach cycle") a relative period of 4.1 Ma, which seems to be a global climatic cycle, and the essential role played by the 100,000 years period cycles and the occurrence of the Neogene coal layers.

In addition, the explorations finding for the Quaternary glaciations appearance and evolution, opened the perspective of middle term global climatic prognosis (one million years).

Obviously, these preliminary results require a thorough study of this kind of research in all imposed directions.

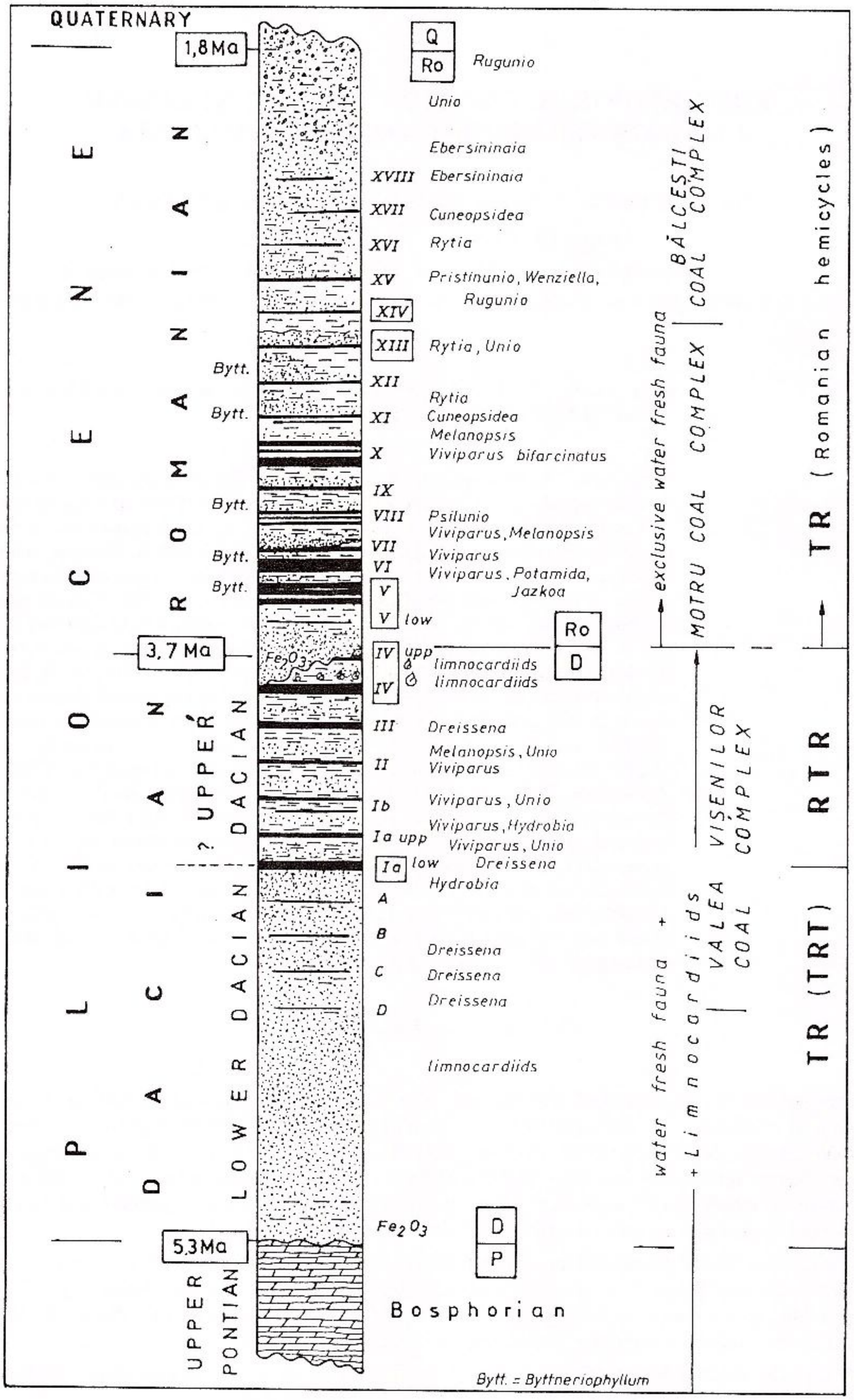

Fig. 1 – Pliocene (Dacian-Romanian) coal complex in the western side of the Dacic Basin (Oltenia).

2. The sequential study of Pliocene coal complex (Dacian-Romanian) in Oltenia. The Valach cycle (4.1 Ma)

The Pliocene coal facies, very occurring in the Dacic Basin (up to 27-28 lignite seams), has been researched, especially in the last years, in the western side of this basin in Oltenia, where it has a spectacular development.

The lignite beds occur since the end of the Lower Dacian, then they are represented in the Upper Dacian and the Lower Romanian and after that they become thinner in the Middle and Upper Romanian.

The various coal complexes distinguished in this stratigraphic level represent in fact the subcomplex of one unitary genetical coal complex of Pliocene age (Dacian-Romanian), the last term of one series that covers the entire Neogene.

The sequential study of these Pliocene coal deposits in Oltenia has mainly emphasized two important realities:

a) The existence of a **maximum development** for the coal deposits - at the beginning of Lower Romanian - given by the development of some thicker and larger seams (Fig. 1).

b) The possibility that the main lignite beds may result from the astronomic and climatic cyclic succession with a periodicity of 100,000 years (the short cycles of the orbital eccentricity).

Obviously, the mentioned maximum supposes the existence of one proper climatic maximum characterized by a warm and wet climate able to provide the vegetation growth for generating the coal beds. We can also suppose the existence in the past of some similar periodical maximums, but also of some climatic minimums, realities that can easily be researched.

We can easily observe that, at least for the Neogene period, the coal facies occurrence depends on certain stratigraphic levels, separated either by other barren intervals or by very thin coal lenses intercalations.

These facts allowed us to presume the existence of one climatic cycle, with its period of about 4.1 Ma, which we called **"Valach cycle"**.

We also observed the following situation: the Pliocene climatic maximum follows the climatic minimum of the Quaternary glacial phase, corresponding to the Alpine Riss phase (Scandinavian Saale phase) (Fig. 2-2').

The former minimum before the Quaternary one is shown by the pregnant cooling period near the Dacian/Pontian boundary, which led to the Mediterranean Sea isolation, being correlated to the "Messinian Salinity Crisis". On the other hand, the climatic Pliocene phase is followed by the climatic maximum of the Middle Pontian.

The last four important moments of this cycle are, at least, clear enough and we can use them for "restricted extrapolations" in the geological history and in the future. Before the Portaferrian period it is possible to mention the development of Miocene coal facies, at least for the maximum moments of the "Valach cycle", which can be followed throughout the Neogene area.

3. The Neogene (Mio-Pliocene) coal complexes in Romania

From a stratigraphical point of view, the coal complexes from the coal bearing deposits of Romania have relatively clear characteristics, which allow us to consider them as belonging to the following stages and substages: Chattian-Lower Aquitanian (Chattian-Egerian), Burdigalian (Eggenburgian-Ottnangian), Middle and Upper Badenian, Middle and Upper Sarmatian (Bessarabian-Kersonian), Middle Pontian (Portaferrian) and Dacian-Romanian (Fig. 3).

3.1. The Chattian-Aquitanian (Chattian-Egerian) coal complex

Stratigraphically, this complex corresponds to the Upper Oligocene and to the Lower Aquitanian and it can include the entire Egerian stage. The Chattian-Aquitanian coal complex, most representative in the **Petroşani Basin** (Fig. 4), is the "horizon 2 (lower producing horizon)" of Chattian-Lower Egerian age including 22 seams (numbered from 0 to 21), in which the most important (considering the thickness and area extent) are 3, 5, 13 and 18. In the next horizon ("horizon 3", Upper Egerian) there are very rare coaly shale or thin coal beds. In this basin there are bituminous coal and cocking bituminous coal.

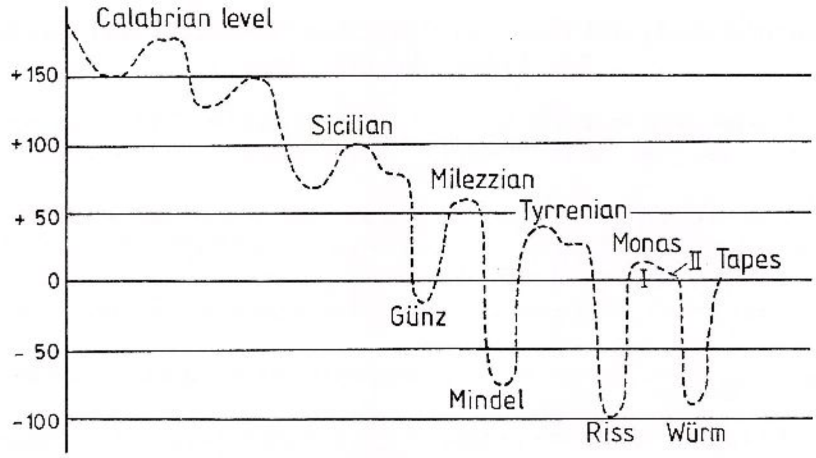

Fig. 2 – The see level variation during the Quaternary (after P. Woldsted, in Saulea, 1967).

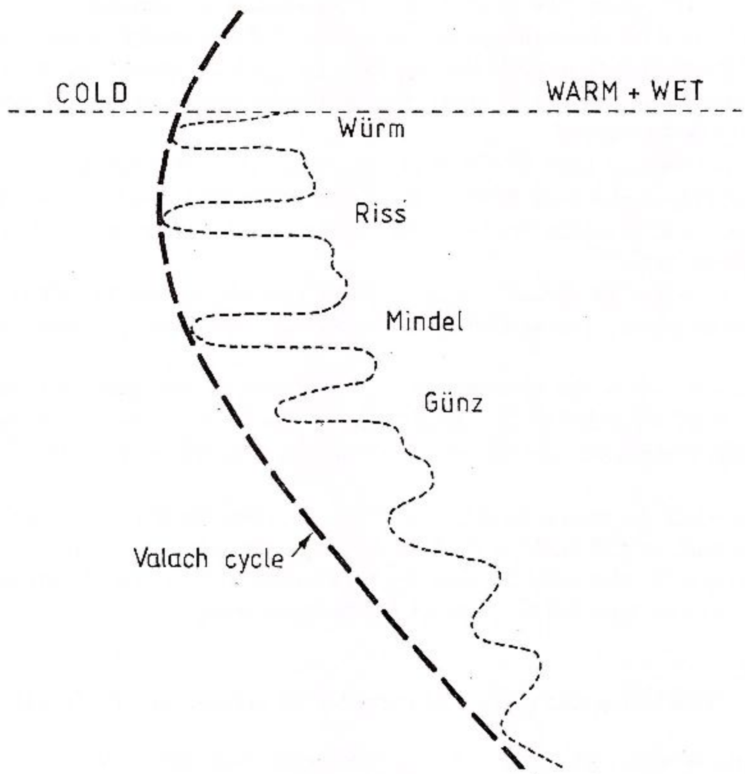

Fig. 2' - Quaternary climatic minimum.

North of the Petroşani Basin the same coal complex (Chattian-Aquitanian) is to be found in the **Haţeg Basin**, where it is represented only by coal intercalation in the lower and middle complex, and north of the Haţeg Basin, in the **Almaş-Agrij Basin**, by the upper (gritty coaly) horizon of the Sinmihai Beds (Upper Egerian). In this last basin the quality of the coal is weaker (brown coal).

3.2. The Burdigalian (Upper Eggenburgian-Ottnangian) coal complex

It is the most non-obvious Mio-Pliocene coal complex in this country. Inside this stratigraphical level (Eggenburgian-Ottnangian) there are coaly clays in the **Vad-Borod Basin** and an "upper producing horizon" (horizon 4) of Upper Egerian-Eggenburgian age in the **Petroşani Basin**, but this can be considered as a younger coal complex. In this "upper producing horizon" of the Petroşani Basin there are about 7-9 thin coal intercalations.

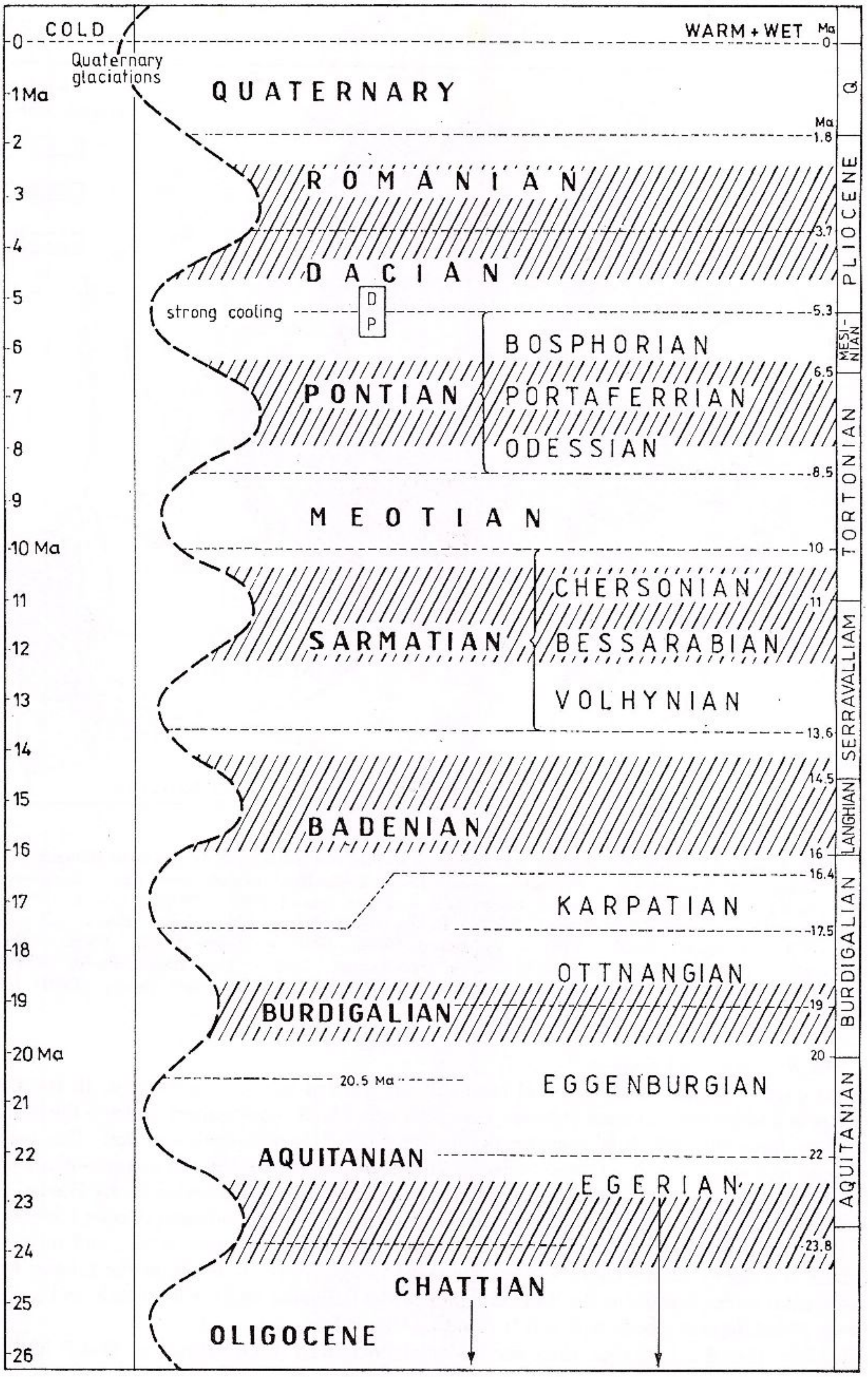

Fig. 3 – The stratigraphic position of Mio-Pliocene coal complexes in Romania.

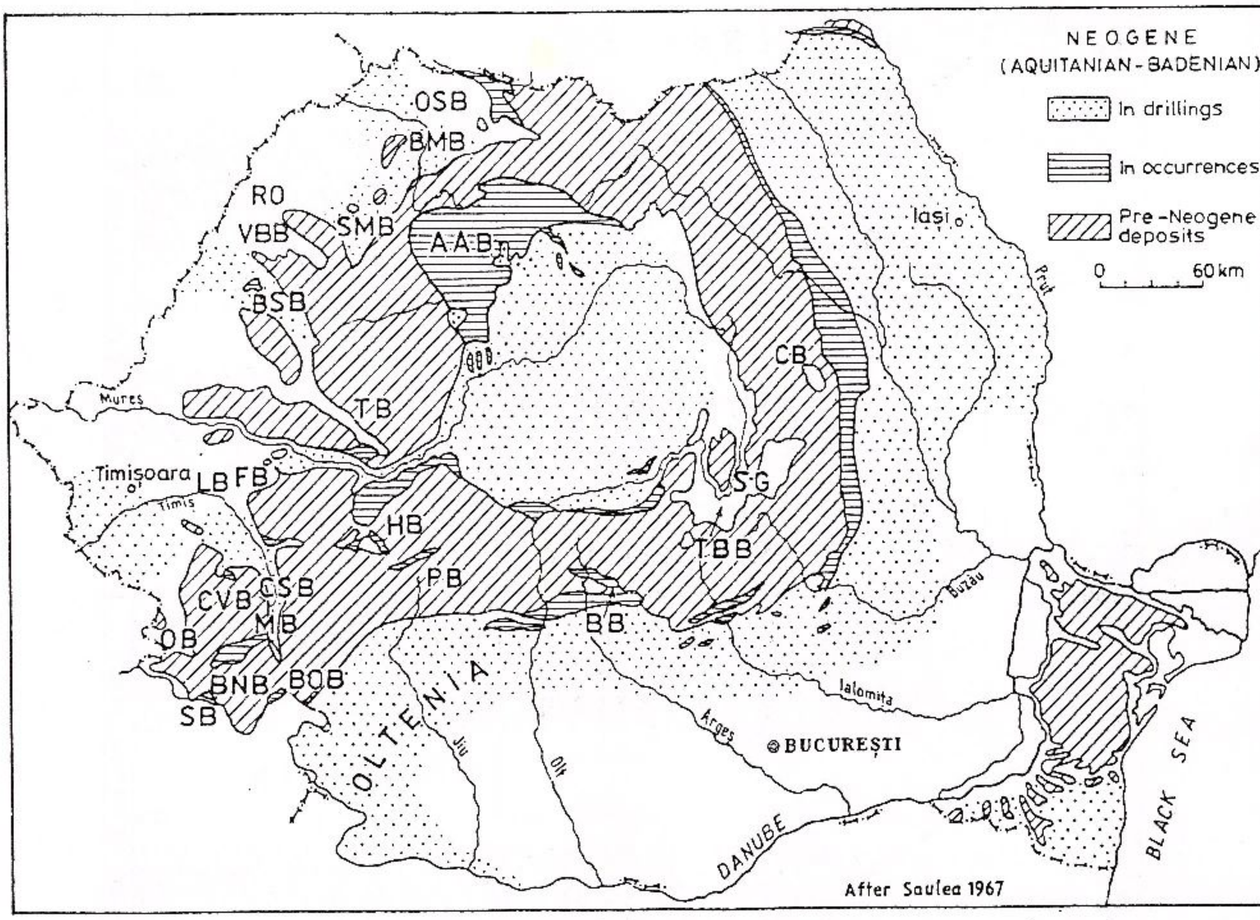

Fig. 4 – The Lower and Middle Miocene deposits and the Mio-Pliocene coal-generating basins in Romania. AAB - Almaş-Agrij Basin; PB - Petroşani Basin; TB - Ţebea-Brad exploitation; CB - Comăneşti Basin; CSB - Caransebeş Basin; LB - Lugoj Basin; FB - Făget Basin; OB - Oraviţa Basin; MB - Mehadia Basin; BNB - Bozovici-Nera Basin; BOB - Bahna-Orşova Basin; HB - Haţeg Basin; SB - Sicheviţa Basin; CVB - Caraşova Basin; VBB - Vad-Borod Basin; SMB - Şimleu Basin; BMB - Baia Mare Basin; OSB - Oaş Basin; RO - Roşiori-Oradea exploitation; TBB - Ţara Bârsei Basin; BSB - Beiuş Basin; SG - Sfântu Gheorghe exploitation; BB - Boteni Basin. (DB) Dacic Basin, (PNB) Pannonian Basin.

3.3. The Badenian coal complex

This complex is present in several coal basins in the western part of the country. In the **Caransebeş Basin** there is a lower coal complex of Lower Badenian age, which is not important from the economic point of view. The upper one, which is economically important, is Upper Badenian in age. This basin contains earthy lignite and woody lignite. Close to this, in the **Mehadia Basin**, there is a marly-sandy horizon with coal, Lower Kossovian in age. There are coals of the same age in the **Bozovici-Nera Basin**, too. To the north, in the **Lugoj Basin**, the deposits contain clayey coals and Lower Badenian (Upper Langhian) volcanic tuffite. In the **Bahna-Orşova Basin** there are Lower Badenian coal deposits too, and in the **Sicheviţa Basin** there is a clayey-tuffitic coal complex, Badenian in age. North of these, in the **Haţeg Basin**, there are coal (lignite) intercalations in the lower complex of the Badenian and further north, in the lower part of the **Ţebea-Brad** deposits, beds A, I and II could be Upper Badenian in age.

South of the Danube, in Serbia, there are Badenian coal facies in **Golubac** and **Donji Milanovac**.

3.4. The Sarmatian (Middle and Upper) coal complex

This complex is obvious in the **Comăneşti Basin** in the East Carpathians, where the coal bearing formations, Middle and Upper Sarmatian in age (Bessarabian-Kersonian), includes up to 10 brown coal seams. In the **Haţeg Basin** there are up to 20 coal levels (xylitic lignite) in the Sarmatian coal deposits. In the **Vad-Borod Basin** the deposits that include exploitable coal beds or only coaly clays are Lower Sarmatian in age. In the **Făget** and **Oraviţa Basins**, there are coal seams or lenses of Lower Bessarabian age. The formation with brown coal in the **Ţebea-Brad** deposits is considered Sarmatian in age (beginning with layer III).

In Serbia, Sarmatian coal-bearing deposits occur in **Kupusiste** and **Mihajlovac**. The coal from **Poljana** and **Bozevac** (considered Pannonian) could probably be Upper Bessarabian-Kersonian in age.

3.5. The Pontian (Middle) coal complex

It is obvious in the **Lugoj Basin** in Romania. In this basin there are 27 seams, grouped in 3 coal complexes (subcomplexes in fact). The most important layers are 3, 9 and 24. 27 coal seams also occur in the **Sărmăşag** deposits (the **Şimleu Basin**), where the 16th layer is the most important and the last 3 layers are found in the lower part of the Upper Pontian. In the rest of the Şimleu Basin there are beds I-X of Lower Pontian age. Pontian lignite beds are to be found in the **Oaş Basin** and the Baia Mare area, too. To the west, the coal bearing deposits of **Roşiori-Oradea** are Upper Pontian in age. In the **Beiuş Basin** corresponding to the Pontian there are 7 lignite beds. In the **Ţara Bârsei Basin**, in the Pontian producing complex, there are coal beds I-IV.

In the southwestern part of the country, in Banat, in the **Făget Basin** corresponding to Pontian, there are thin coal lenses and in **Caraşova Basin** there are coaly shales and a lignite bed.

The same coal complex is present in Serbia too, in **Smederevo-Kovin-Novi-Kostolac** area.

During the Pontian, in the Dacic Basin in Oltenia, between the rivers Olt and Jiu, there are coaly clays and 3 thin lignite layers. Between the rivers Olt and Dâmboviţa, corresponding to the Middle Pontian, there are coal beds in the **Boteni Basin**, which are exploitable in the Schitu-Goleşti deposit, where there are four lignite beds (0-III) of which bed I is the most important. East of these Pontian coal beds and coaly clays occur between the rivers Buzău and Trotuş.

This coal complex began its development at the end of the Lower Pontian (Odessian), it developed very clearly in the Middle Pontian (Portaferrian) and ended at the beginning of the Upper Pontian (Bosphorian).

3.6. The Dacian-Romanian (Pliocene) coal complex

This coal complex is well represented in the Dacic Basin, mainly to the west of the river Olt. Thin coal layers were developed since the Lower Dacian. The seams are characterized by large extension and great thickness during the Upper Dacian and Lower Romanian and by smaller and smaller areas beginning with the Middle Romanian. This complex is also present in the southwestern part of the Dacic Basin, in Bulgaria (**Lom Basin**).

To the east, it is present in the Subcarpathian Hills too, mainly developed during the Upper Dacian and Lower Romanian; it appears even north of the Buzău Valley. In many areas of Oltenia and the Subcarpathian Hills exploiting units, both mines and quarries were developed for the lignite seams. Between the rivers Proviţa and Prahova, up to 28 lignite seams (between 0.1 and 6.1 m thick) were explored of which the first five ones are considered Dacian and the rest Romanian in age.

This coal complex can also be found in the depression areas in the East Carpathians. In the Sf. Gheorghe deposit in the **Ţara Bârsei Basin** beds VI-VII and VIII-X are considered Dacian-Romanian in age. In the eastern part of the Pannonian Basin, inside the Roşiori-Oradea deposits, a coal facies is developed which is Dacian-Romanian in age, too.

4. Neogene coal facies of Europe and of Earth

The Neogene (Mio-Pliocene) coal bearing formations are known both in Europe and in the Parathetys area as well as in different regions of the Earth. This fact entitles us to admit that the climatic cycle (named "Valach cycle") could be recognized on a global scale, and in this case the causes which determine its existence should also have a global character. So, it is necessary to carry out a survey in the occurrence zones of such formations.

4.1. The Parathetys Neogene coal facies

In this area, coal bearing deposits occur even in the Bohem Massif, south of the Metaliferi Mountains on the Carpathian edge, in the East Slovakia brackish facies of clays with coals. In the middle Rhine region (southern) of the sea domain, coast peat deposits are developed (in Miocene), which determined the appearance of the 100 m thick coal bed known in this zone. On the southern edge of the Pannonian Basin, along the Dinaric Mountains, Middle Miocene lacustrine limnic deposits, often with coals, occur.

In the Pontian s.s. a sandy limnic deposit with coals is also known, and in the Hessa-Hanover branch of the Rhine graben, in the Miocene, lacustrine coal facies (the Wetterau Basin) accumulate, which continue even in the Pliocene.

Obviously, we may talk about the Neogene coal complexes of Romania, which belong either to the Pannonian basin, or to the Dacic Basin, or to some intra-Carpathians basins (chapter 3).

4.2. The Neogene coal facies in Western Europe

The lacustrine deposits, sometimes with Miocene coals, occur in some basins situated in the northern extension of the Albanian-Tessalian depression and in Eubeea (Kumi Basin), while the lacustrine coal formations should belong to the Upper Miocene. In Toscana region, marls with lignites in a Tortonian lacustrine facies (probably Upper Miocene) are known, possibly of the same age as the End Miocene marls with lignites in the north-Bethic depression. Between the Elbe and the Vistula, the estuary formation with sands and coals (Braunkohlensande) should be of a Middle-Upper Miocene age.

In this European zone Pliocene coal deposits are also known, such as the lower gray clays with lignites series in the intermontane Apennine Basin (the lacustrine lignite deposits at the edge of the Rhone Depression-Bresse Basin, north of Lyons).

4.3. The Neogene coal facies from Earth

On the Asian continent, coal continental accumulations are known in the Neogene in some small basins. That is the case of the Baikal Lake or of the Sakhalin Island, where the detrital series with coal intercalations should belong to the first part of the Neogene (Miocene).

In the geosyncline Anadir-Kamchatka-Japan-Philippines zone, sea lagoon and continental detrital coal complexes with lava develop in the Neogene. A some of them could be Pliocene in age because the upper part of the Sakhalin Neogene series (probably Pliocene) continues in the Kamchatka and Anadir.

Also, in Asia, in the Sumatra and Java, the Upper Miocene-Pliocene Palembang beds contain lignite seams.

The Neogene coal bearing deposits are also known in South America where, along the Pacific coast of Chile, a Miocene with coal is developed.

5. The main Neogene coal beds developed as an effect of the climatic cycles within a 100,000-year period

The sequential study of the Neogene coal bearing deposits in the western side of the Dacic Basin led us to the conclusion that the main lignite seams of the Dacian-Romanian coal complex follow one another after equal time intervals that can be associated with the astronomical cycles with a period of 100,000 years (the short cycles of the orbital eccentricity).

If we consider the stratigraphical position of each coal subcomplex, the absolute time values corresponding to the limits of the stages and substages and the total number of the main seams of a certain complex, it is easy to deduce that the climatic variations that have determined the generation of these seams cannot be connected to other known cycles besides the above mentioned ones. This hypothesis led us to the conclusion that the generation of about 24 main lignite seams lasted 2.4 Ma, the time difference between two consecutive seams being of 0.1 Ma.

The brief analysis of the other Neogene coal complexes shows that to time intervals between 2 and 3 Ma about 22-23 up to 27-28 distinct coal seams correspond, which underlines the influence of the above mentioned climatic cycles of the coal accumulations. It is remarkable that the amplitude of the climatic cycle generated by the short cycles of orbital eccentricity is greater than that generated by the superior cycle (the Valach cycle), so that the development of the coal seams is very clear at least near the climatic optimum moments of the Valach cycle.

Each cycle with 0.1 Ma period includes two different climatic moments: one warm and wet, and the other cold. The first moment, accompanied by a transgression, is favorable to the development of vegetation that can lead to the generation of a coal bed, and the second moment of the cycle, the cold one, is not favorable

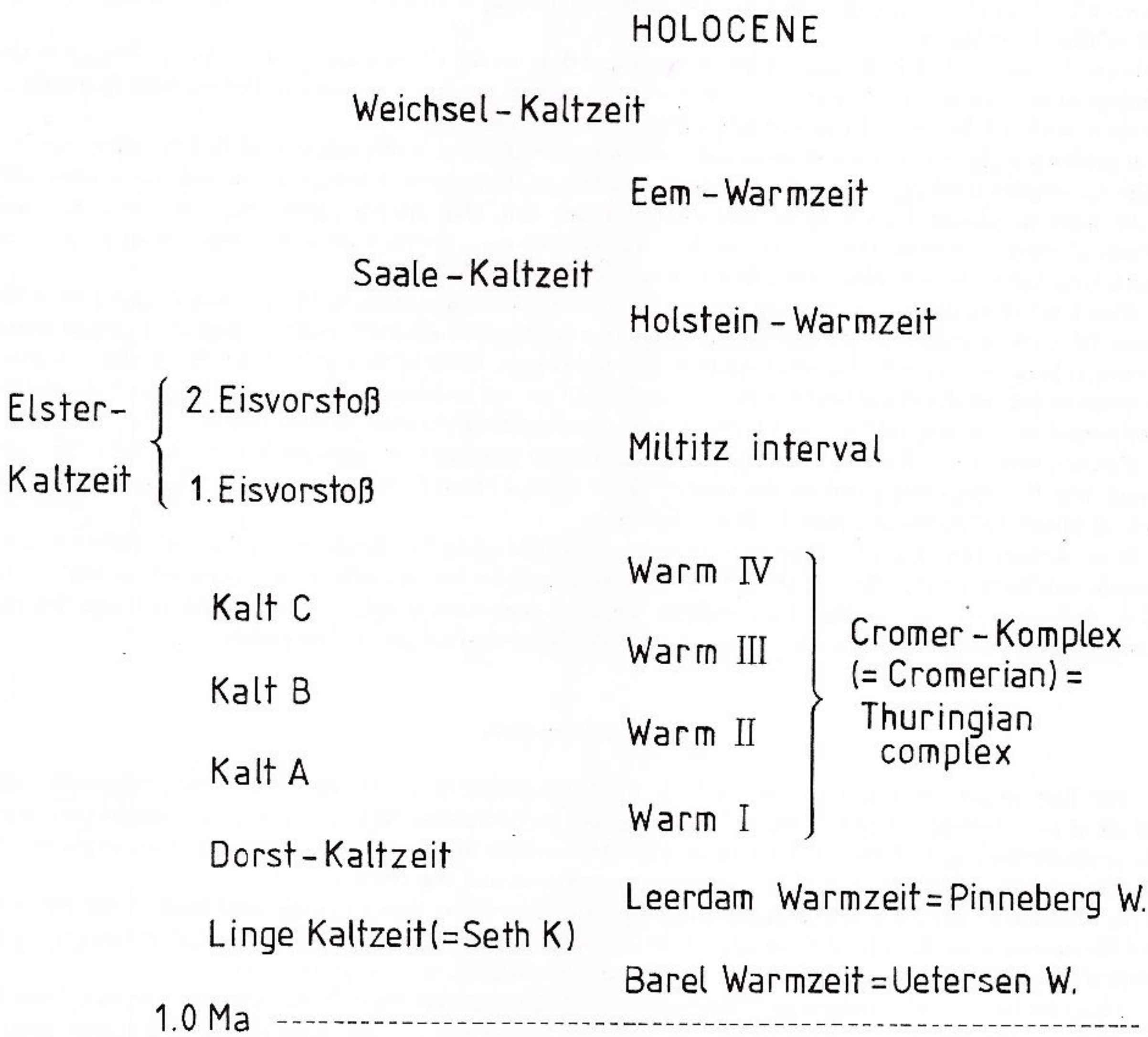

Figure 5

to the coal generation. This second moment is accompanied by regression, and admits the accumulation of coarse material (sands and gravel), because of the intensification of erosional process, which was caused by the regression of the shores. This succession is present in the cold moments of the Valach cycle too, at least in the Quaternary, when the cold moments of the 0.1 Ma cycle correspond to the glacial stages, and the warm ones to the interglacial stages. For the last million of years 10 warm moments and 9 cold moments detectable in the Quaternary of the German Plain can be distinguished (according to the data of Eissmann et al., 1994) (Fig. 5).

The Dutch researchers (1996) assumed that the sedimentation cycles, which led to the generation of the coal seams in the coal-bearing lacustrine deposits from Ptolemais (northwest Greece) could be of the precessional origin so that they could be caused by cycles with a period of about four times shorter than the ones caused by the orbital eccentricity (about 26,000 years).

The lacustrine succession from Ptolemais, dated by Argon 40 and Argon 39 method about 4.19 Ma, corresponds to the lower part of the Pliocene coal complex, probably at the Lower/Upper Dacian boundary in the Dacic Basin.

The effect of the precessional cyclicity is only acceptable for the seams that accompany the main ones from the Pliocene coal complex usually considered as lower or upper, and rarely as independent coal beds or benches of some main layer, which are rarely represented by a single coal level (beds VIII and X in the west of the Dacic Basin).

It can be assumed that the time intervals corresponding to the presessional year are more obvious in the lithological successions that are rich in coal levels. These coal seams correspond to the climatic optimum of the cycle with 4.1 Ma period (the climatic optimum of the Valach cycle).

It is also possible to distinguish some much shorter cycle of about 1000 years period that are characteristic of the "precessional miniglaciations" by a fine observation of the interval structure of the coal layers, especially of the most interbedded ones, those that contain many thin clay intercalations. So, one can distinguish periods of about 500 years that are favorable to the vegetal mass accumulation in alternance with the same length time intervals favorable to sterile (it means clay) accumulation.

This kind of analyses can point to another method for the estimation of the necessary time period for lignite layers accumulation. We can also mention the attempts of estimation of the period of lignite seams forming taking into account the accumulation rate and the thickness of the lignite beds. So, for layer V from the western part of the Dacic Basin, with a thickness of 7 m and an initial thickness of 21 m, an accumulation time period for the vegetal mass of 21,000 years was determined (Pauliuc, Barus, 1982).

Furthermore, it is likely to identify some longer time periods that separate the most important coal seams, which might correspond to the long cycles of orbital eccentricity, any of which being equal to four cycle of about 100,000 years, namely 413,000 years.

It is obvious that the interference of the above-mentioned cycle, characterized by very different time periods, can increase the effect of the warm and wet climate favorable to the coal generation, as well the effect of the cold climate in other time periods. This fact makes more difficult the investigations carried out in order to establish the temporal content of the older Neogene coal-generating cycles.

6. Conclusions

The first important result of this study is the highlighting of a climatic cycle with a relatively large period of time (about 4.1 Ma), named "Valach cycle" (=Wallachian climatic cycle), that might determine the periodic presence of the coal facies at least throughout the Neogene. This cycle also explains the Quaternary glaciations and the Mio-Pliocene transgressions and regressions.

In addition to this, it seems possible to present the hypothesis that the main coal beds of the Neogene coal facies were generated by the climatic cycle, the result of the short cycles of the orbital eccentricity with a 100,000 year period, which also caused the repeated transgressions and regressions.

These cycles have also determined the succession of the glacial phases. It was also possible to determine the effects of the precessional cycles with a 26,000 year period and of some other cycles with known period.

All the above-mentioned conclusions allow the reconsideration of the known points of view regarding the geological evolution of the continental and marine sedimentary basin during the Neogene but also before the Neogene, highlighting the important role played by the climatic factor. Of course, we are searching now to an explanation for the Valach cycle, which we consider for now as an effect of some possible pulsation of the Solar system.

All these observations led us to a better correlation of the Neogene and pre-Neogene deposits with or without coal and also to the initiation of a theory based on the cyclicity of the geological phenomena.

It is also necessary for us to mention that we intend to research thoroughly the Neogene and pre-Neogene coal facies in Europe and all over the world and especially the effect of the climatic cycles on them.

It is interesting to remark here the possibility of the prediction of the geological evolution of the terrestrial globe in the next millions of years considering the cyclicity of the researched phenomena.

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