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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

CORRÉLATION DES UNITÉS LITHOSTRATIGRAPHIQUES ET TECTONIQUES LONGEANT LE RUISSEAU D'ARIEŞ, ENTRE LA VALLÉE DE IARA ET LE MONT GĂINA (MONTS APUSENI)[1]

PAR

ION BALINTONI[2]

Lithostratigraphic and tectonic units. Crystalline series. Someş Series. Codru Series. Biharia Series. Arieşeni Series. Baia de Arieş Series. Lithostratigraphy; Apuseni Mountains — Codru Unit — Gilău Massif.

Abstract

Correlation of the Lithostratigraphic and Tectonic Units along the Arieş, between the Iara Valley and the Găina Mount (Apuseni Mountains). Several tectonic units with metamorphites are found along the Arieş : the Bihor Unit, consisting of the Someş Series and a Mesozoic cover in Bihor facies ; the Finiş Nappe, including the Codru Series and Permian-Mesozoic nonmetamorphosed formations in Codru facies ; the Arieşeni Nappe, constituted of the Arieşeni Series and Permian formations in Codru facies ; the Biharia Nappe consisting of the metamorphites of the Biharia and Belioara series and Permian conglomerates near Băişoara ; the Baia de Arieş Nappe, including the Baia de Arieş Series and the Sohodol marble formation. The Finiş and Arieşeni nappes have been emplaced during the Mediterranean paroxismal tectogenesis and belong to the Codru Nappes System.

Introduction

L'étape actuelle de connaissance du cristallin des monts Apuseni est le résultat de la recherche de plusieurs générations de géologues ; un aperçu historique exhaustif de cette recherche est exposé dans les

[1] Reçue le 28 avril 1982, acceptée le 5 mai 1982, présentée à la séance du 11 mai 1982.

[2] Institutul de Geologie şi Geofizică, str. Caransebeş nr. 1, 78344 Bucureşti, 32.

ouvrages de Dimitrescu (1958, 1966) et dans l'ouvrage de synthèse de Ianovici et al. (1976). C'est pour cette raison-là que nous n'y consignons pas l'évolution des notions et des concepts qui font l'objet du présent article, mais nous nous rapporterons aux récents contributions portant sur la lithostratigraphie et la tectonique de la région en question. Rappelons que les moments principaux de cette évolution peuvent être trouvés dans les ouvrages de : Rozlozsnik (1936), Giușcă (1937), Pálfy, Rozlozsnik (1939), Kräutner (1944), Dimitrescu (1958), Borcoș, Borcoș (1962), Bleahu, Dimitrescu (1963), Dimitrescu (1964), Trif, Stoicovici (1964), Dimitrescu (1966), Dimitrescu, Bleahu (1967), Lupu et al. (1967), Bleahu et al. (1968), Giușcă et al. (1968), Mârza (1969), Visarion, Dimitrescu (1971), Dimitrescu et al. (1974), Ianovici et al. (1976), Dimitrescu et al. (1977), Bleahu et al. (1981), Hârtopanu et al. (1981), Hârtopanu et al. (1982). La conclusion qui en résulte c'est que le cristallin des monts Apuseni est formé des entités lithostratigraphiques qui ont subi un, deux ou trois métamorphismes et qui se trouvent fréquemment en relations tectoniques. Pour tracer les limites géologiques sur l'esquisse tectonique ci-jointe nous avons utilisé le suivant matériel cartographique : Mârza (1969), Dimitrescu et al. (1974), Dimitrescu et al. (1977), Hârtopanu et al. (1982). Les limites envisagées ainsi que les unités lithostratigraphiques et tectoniques ont été étudiées par nous le long des coupes transversales plus importantes du ruisseau d'Arieș et de ses affluents, à partir de la vallée de Iara jusqu'à la vallée de Vidrișoara.

Unités lithostratigraphiques

Sur la légende de l'esquisse tectonique sont représentées les suivantes unités lithostratigraphiques cristallines en superposition, de bas en haut : série de Someș, série de Codru, série d'Arieșeni, série de Biharia, série de Sohodol.

Série de Someș. Suivant Dimitrescu (1966), Giușcă et al. (1968), Ianovici et al. (1976), Hârtopanu et al. (1981), Hârtopanu et al. (1982), la série de Someș représente une pile de roches polymétamorphes d'origine principalement terrigène (paragneiss, micaschistes, divers types de schistes quartzitiques, de rares amphiboles et calcaires cristallins) ; les roches de la série de Someș conservent les traces d'au moins trois métamorphismes préalpins : le premier caractérisé par la présence de la staurotide et du disthène, le deuxième est marqué par l'apparition de la sillimanite en quelques endroits et le troisième fortement régressif, déterminant la chloritisation parfois généralisée des minéraux ferromagnésiens préexistants. La série de Someș a été probablement traversée pendant le deuxième métamorphisme, par les granitoïdes de Muntele Mare à caractère nettement intrusif ; celles-ci sont en général homogènes, ont des limites vives vers les roches environnantes et ont généré une auréole de contact thermique. Nous n'avons pas séparé la série d'Arada de celle de Someș, du fait que nous considérons qu'au moins pour l'aire représentée sur l'esquisse tectonique, l'impression de métamorphisme initial dans la zone de la chlorite laissée par les roches de la série d'Arada est redevable à un rétromorphisme hercynien intense, caractéristique qui a déterminé Dimitrescu (1966) de l'individualiser. Mentionnons en ce sens que Dimitrescu (1966) cite dans la partie supérieure de la série d'Arada un niveau de schistes à grenat chloritisé ; dans tous les cas si on fait des lames minces dans les échantillons contenant de la biotite et de la chlorite peuvent être observés divers stades de choritisation de la biotite.

Série de Codru. Suivant nos observations et les données de Dimitrescu (1958, 1966), Trif (1961), Trif, Stoicovici (1964) et Ianovici et al. (1976), Hârtopanu et al. (1982) nous pouvons conclure que la série de Codru est polymétamorphe, semblable comme évolution à la série de Someş. Nous définissons la série de Codru en tant qu'une séquence de roches d'origine magmatogène basique et terrigène, métamorphisées initialement jusqu'à la zone de la staurotide pénétrées par des granites et intensément migmatisées sur de grandes surfaces au cours d'un métamorphisme ultérieur, affectées ensuite par le rétromorphisme hercynien. La série de Codru ne se parallélise pas avec celle de Biharia et ne constitue pas une partie de celle-ci. Les granitoïdes de Codru peuvent être considérés comme autochtones, ce qui les différencient des granitoïdes de Muntele Mare. Les granitoïdes de Codru se sont intimement mêlés aux métamorphites préexistantes (sur l'aire représentée sur l'esquisse tectonique, surtout des amphibolites) dans une ambiance synmétamorphe à degré élevé, fait qui a engendré des processus d'assimilation et métasomatose. La formation de la sillimanite sur biotite (Trif, 1961) ne met pas en évidence un phénomène de contact entre deux systèmes pétrographiques en contraste thermique, tout comme chez les cornéennes situées à la limite des granites de Muntele Mare, mais des conditions thermodynamiques existantes pour les métamorphites et le matériel granitique envahissant durant leur interaction.

Série d'Arieşeni. Selon Ianovici et al. (1976) la série d'Arieşeni d'âge carbonifère inférieur représente partiellement une séquence volcanogène-sédimentaire basique légèrement métamorphisée pendant l'orogenèse hercynienne.

Série de Biharia. Cette série tout comme la série d'Arieşeni est constituée d'un complexe volcanogène-sédimentaire basique métamorphisé, ayant des affinités de structure avec les séquences ophiolitiques actuelles (Mârza, 1969 ; Ianovici et al., 1976). Dans la série de Biharia prédominent divers types de schistes à porphyroblastes d'albite, avec lesquels s'associent parfois les amphibolites. Sont aussi présentes des lentilles et bandes dolomitiques, à côté des roches graphiteuses et/ou quartzitiques. La série de Biharia est polymétamorphique et transpercée par des granitoïdes. Le métamorphisme initial de la série de Biharia, probablement calédonien, a été de degré inférieur à celui des séries de Someş et de Codru. La série de Biharia de toute l'aire figurée sur l'esquisse tectonique dénote des caractères épizonaux, la biotite étant visible assez rarement, pareillement à un grenat menu, peut-être spessartinique. Les processus rétromorphiques, liés à l'orogenèse hercynienne sont difficiles à déceler, quoique la série de Biharia supporte transgressivement la série de Belioara, métamorphisée initialement au cours de l'orogenèse hercynienne. Les granites pénétrant la série de Biharia peuvent être considérés comme paraautochtones, en se différenciant des granites des Muntele Mare et des granitoïdes de Codru. Bien qu'ils aient une texture gneissique et souvent ils apparaissent sous la forme des pseudocouches en alternant avec les roches vertes, les granitoïdes de la série de Biharia sont dépourvus d'auréole de contact thermique et n'engendrent pas des migmatites. Ils ont réussi toutefois à assimiler du matériel basique et par conséquent ils sont certainement préhercyniens, en sorte que la série de Belioara les transgresse directement à l'ouest du ruisseau Pociovaliştea. Il s'agit donc des granitoïdes syncinématiques mis en place pendant le métamorphisme initial de la série de Biharia et refroidis lentement dans un régime de système thermodynamique fermé, ce qui a favorisé le maintien de l'eau dans le système durant un laps de temps, simultanément à l'évolution des minéraux vers des équilibres de basse température. Ce processus pourrait être décrit en tant qu'un autométamorphisme régressif régional. Par ses caractères et sa grande surface d'affleurement, la série de Biharia représente un excellent repère lithostratigraphique et tectonique. D'après la description ci-dessus, la série de Biharia renferme également sans discontinuités tectoniques ou de tout autre nature, la série de Muncel de la nappe de Codru (Dimitrescu et al., 1974) et le premier horizon de la série de Muncel de la nappe de Muncel-Lupşa (Dimitrescu et al., 1974) ou bien de la nappe de Muncel (Dimitrescu et al., 1977). Les roches respectives sont identiques, du point de vue de la composition minéralogique et de l'histoire du métamorphisme, aux roches de la série de Biharia, avec lesquelles elles font corps commun mais en se différenciant de celles de la série de Baia de Arieş (Muncel), comme on verra plus loin.

Série de Belioara. D'une manière transgressive se dispose sur la série de Biharia (Ianovici et al., 1976) une pile épaisse de roches principalement carbonatées, qui comportent dans leur base des métaconglomérats quartzitiques. Dans la vallée de Poşaga, un niveau de métaconglomérats et métagrès quartzitiques apparaît aussi au-dessus des roches carbonatées. Les schistes quartzitiques sériciteux intercalés dans la masse des roches carbonatées tout comme les métaconglomérats relèvent un faible métamorphisme dans la zone de la chlorite. Il s'agit d'une séquence de roches carbonifères inférieures métamorphisées pendant l'orogenèse hercynienne.

Série de Baia de Arieş. Cette série ressemble de maints points de vue à la série de Someş. Tout comme la série de Someş la série de Baia de Arieş comporte dans sa constitution des roches terrigènes. Elle diffère de celle-ci par l'abondance des roches carbonatées dans la partie supérieure et l'abondance de certains ortogneiss à aspect textural des gneiss oeillés dans la partie inférieure. Les gneiss oeillés ont été décrits par Dimitrescu (1973) dans le proche voisinage du confluent d'Arieşul Mic et Arieşul Mare sous le nom de gneiss de Mihoeşti. La série de Baia de Arieş dénote un métamorphisme initial dans le domaine de stabilité de la staurotide et du disthène, suivi d'un métamorphisme qui a généré en certains endroits de la sillimanite (par exemple sur la vallée de Poşaga) et a été aussi affecté mais d'une manière extensive par le métamorphisme régressif hercynien. La parallélisation de la série de Baia de Arieş avec celle de Someş a été déjà faite par Giuşcă et al. (1968). Dans le présent article, la série de Muncel de la nappe de Muncel-Lupşa (Dimitrescu et al., 1974 ; Dimitrescu et al., 1977), à partir de l'horizon des schistes quartzitiques sériciteux vers le haut, est considérée en tant que base de la série de Baia de Arieş ; les schistes sériciteux chloriteux albitiques, envisagés par les auteurs ci-dessus mentionnés comme horizon basal de la série de Muncel (transgressive sur la série de Biharia) dans la nappe de Muncel-Lupşa, appartiennent en fait à la partie supérieure de la série de Biharia. Nos arguments en faveur de ces affirmations sont : sur la vallée de Pociovaliştea, au-dessus de la série de Belioara ainsi que dans le sommet de Colţu Şesului, aux sources du ruisseau de Sartăş, en surmontant les roches vertes de la série de Biharia, la succession des roches de la base de la série de Baia de Arieş reconnue par Ianovici et al. (1976) est identique aux successions qui peuvent être observées à Mihoeşti et sur la vallée de Vidrişoara au-dessous du mont Găina, dans la série dite de Muncel (Dimitrescu et al., 1974 ; Dimitrescu et al., 1977), au-dessus de l'horizon des schistes sériciteux chloriteux à albite ; à savoir, dans tous les quatre endroits mentionnés apparaissent premièrement des gneiss œillés de type Mihoeşti, surmontés par des micaschistes et paragneiss renfermant des intercalations d'amphibolites, suivis d'une bande de quartzites noirs associés aux roches graphiteuses. Si les gneiss de type Mihoeşti sont présents ou non, suivant que le plan de charriage de la nappe de Baia de Arieş passe par dessous ou bien par dessus de ceux-ci, les micaschistes et surtout les quartzites noirs se continuent sans interruption à partir de la vallée d'Ocolişul à Runc jusqu'au-dessous du mont Găina. Ils ne sont que décalés par des failles comme celle sur la gauche de la vallée Lupşa, ou bien cachés parfois sous la couverture post-tectonique. Un autre argument important pour la parallélisation de la plus grande partie de cette série de Muncel avec la base de la série de Baia de Arieş est que la série de Muncel dans la séquence décrite plus haut ne représente pas un cristallin épizonal, mais une association de roches mésozonales intensément rétromorphisées. A une recherche plus détaillée portant sur toutes les aires figurées par Dimitrescu et al. (1974) et Dimitrescu et al. (1977) comme appartenant à la zone à chlorite de la série de Muncel (excepté les schistes sériciteux chloriteux à albite) on peut rencontrer des grenats reliques de dimensions de plus d'un centimètre et de la biotite à caractère mésozonal. Le processus de destruction du grenat et de la biotite, concomitant à la génération d'un ensemble extrêmement pénétratif de plans S, sur lesquels reposent la chlorite et la muscovite de néoformation, peut être décelé aisément dans les sections minces.

Formations des marbes de Sohodol. Le caractère discordant des marbes de Sohodol reposant sur la série de Baia de Arieş (Muncel) a été reconnu par Ianovici et al. (1976) et figuré par Dimitrescu et al. (1977) sur la Carte géologique de la Roumanie au 1 : 50 000, feuille

Avram Iancu. Du même avis que les auteurs susmentionnés, je crois qu'il s'agit d'une séquence carbonatée paléozoïque moyenne, transgressive sur la série de Baia de Arieş.

En guise de conclusion, nous voulons rappeler que nous avons corrélé les unités lithostratigraphiques métamorphisées affleurant le long de l'Arieş, en se basant sur les suivants caractères corroborés : contenu pétrographique et succession spatiale des entités lithologiques, histoire métamorphique, conditions géotectoniques dans lesquelles se sont accumulées les piles de roches, reflétées dans les caractères du matériel prémétamorphique ; le magmatisme plutonique acide qui a affecté ces unités. Pour ce qui est de la nature des contacts entre les unités lithostratigraphiques, elle sera analysée dans le chapitre suivant. Comme conclusions préliminaires nous pouvons souligner que quelques-unes des séries en contact se distinguent assez bien grâce à ses caractéristiques, de sorte que leur voisinage peut être expliqué seulement du point de vue tectonique.

Unités tectoniques

Sur l'esquisse tectonique sont représentées de bas en haut les suivantes unités tectoniques majeures : unité de Bihor, nappe de Finiş, nappe d'Arieşeni, nappe de Biharia (Lupşa), nappe de Baia de Arieş (Muncel).

Unité de Bihor. Celle-ci est constituée de la série de Someş avec les granitoïdes de Muntele Mare ainsi que du Mésozoïque en faciès de Bihor. L'unité de Bihor est apparemment autochtone (Ianovici et al., 1976) puisqu'on ne connaît pas des unités tectoniques inférieures à cette unité.

Nappe de Finiş (nappe de Codru pour Dimitrescu et al., 1974 ; nappe de Finiş-Ferice-Gîrda pour Ianovici et al., 1976 ; nappe de Gîrda pour Dimitrescu et al., 1977 ; nappe de Finiş-Ferice-Gîrda pour Bleahu et al., 1981). La nappe de Finiş fait partie du système des nappes de Codru, mises en place pendant la tectogenèse paroxismale méditerranéenne. La nappe de Finiş comporte dans sa constitution la série de Codru et du sédimentaire d'âge permien-mésozoïque en faciès de Codru. Le contact anormal avec l'unité de Bihor est mis en évidence par les observations suivantes : a — à partir de Valea Caselor de Cîmpeni vers l'ouest, la série de Codru chevauche le sédimentaire de l'unité de Bihor ; b — à l'est du point susmentionné, la série de Codru pratiquement non rétromorphisée dans cette région repose sur la partie supérieure de la série de Someş intensément rétromorphisée pendant l'orogenèse hercynienne (série d'Arada pour Dimitrescu, 1966) ; c — à la limite d'entre les séries de Codru et de Someş qui peut être expliquée par la différence d'intensité du rétromorphisme, les granitoïdes de Codru disparaissant brusquement ; d — vers le sud-est de Muntele Mare, les granitoïdes de Muntele Mare viennent en contact directement avec les granitoïdes de Codru, les conditions de genèse et la mise en place de ces deux types de roches étant complètement différentes ; e — la série de Codru à son tour, à partir de Valea Caselor de Cîmpeni vers l'ouest, supporte le sédimentaire permo-mésozoïque en faciès de

Codru ; f — le contact entre les séries de Someş et de Codru est marqué par des milonites ; g — la série de Codru prend contact avec des termes différents de la série de Someş. Une bonne partie de cette argumentation peut être retrouvée dans l'ouvrage de Dimitrescu de 1966.

Nappe d'Arieşeni. Selon Dumitrescu et al. (1977) la nappe affleure rien que dans la partie ouest du périmètre figuré par nous sur l'esquisse tectonique. Elle est constituée de roches de la série d'Arieşeni et du Permien en faciès de Codru, étant une nappe du système de Codru.

Nappe de Biharia. Elle comporte les séries de Biharia, de Belioara transgressive sur celle de Biharia et les conglomérats permiens de Băişoara. D'après notre avis, la nappe de Biharia renferme le long de l'Arieş une partie de la nappe de Muncel-Lupşa de Dimitrescu et al. (1974) et Dimitrescu et al. (1977). Pour comprendre la nature du contact entre la série de Biharia (qui constitue la partie inférieure de la nappe de Biharia) et la série de Codru de la nappe de Finiş, nous présenterons les arguments qui repoussent l'hypothèse proposée par Dimitrescu et al. (1974) et Ianovici et al. (1976), à savoir que la série de Codru pourrait représenter la partie inférieure de la série de Biharia, migmatisée par les granitoïdes de Codru. Ainsi : a — il y a entre les deux séries une forte discordance de degré de métamorphisme initial (roches à staurotide en contact avec des roches dépourvues même de la biotite, sans pouvoir démontrer leur caractère rétromorphe) ; b — dans la série peuvent être décelées les traces de trois métamorphismes, tandis que dans celle de Biharia les traces de deux métamorphismes ; c — les granitoïdes de la série de Codru s'associant à une zone de migmatisation intense s'achèvent brusquement le long du contact avec la série de Biharia; d — ces migmatites peuvent contenir de la sillimanite voire au contact avec les roches de la série de Biharia où comme nous avons déjà mentionné, il est impossible de mettre en évidence la biotite ; e — les granitoïdes de la série de Codru et ceux composant la série de Biharia appartiennent à des ambiances génétiques et évolutives différentes ; f — la série de Biharia prend contact avec divers termes lithostratigraphiques de la série de Codru. Cette situation a été expliquée par le caractère transgressif du matériel prémétamorphique de la série de Biharia sur les roches déjà migmatisées de la série de Codru. Du fait que la migmatisation de la série de Codru ne semble pas être précalédonienne et le métamorphisme initial de la série de Biharia ne peut pas être postcalédonienne, cette supposition devient improbable. En admettant toutefois que la migmatisation de la série de Codru serait précalédonienne, la transgressivité de la série de Bihaira sur les migmatites de la série de Codru ne peut pas être acceptée puisque : a — le contact entre les deux séries est jalonné des milonites arrivant à 1000 m d'épaisseur ; b — à partir de Valea Caselor de Cîmpeni vers l'ouest, entre les deux séries s'interposent premièrement des formations permiennes et ensuite la nappe d'Arieşeni, fait qui relève indubitablement un contact anormal ; c — les milonites développées sur le contact entre les séries de Codru et de Biharia sont situées dans la prolongation directe du Permien qui affleure à Cîmpeni ; d — les propriétés de la série de Codru et celles de la série de Biharia, d'un côté et d'autre du Permien et de la nappe d'Arieşeni, sont exactement celles longeant d'un côté et d'autre la zone milonitique ; e — des termes lithostratigraphiques différents de la base de la série de Biharia viennent en contact avec la série de Codru. Ce fait observé par Dimitrescu et al. (1977) a été expliqué par l'attribution de certains secteurs de la série de Biharia à la série de Muncel, considérée également transgressive sur les séries de Biharia et de Codru. Cette hypothèse est dépassée actuellement dans la perspective du présent ouvrage. Donc, le contact entre la série de Biharia et celle de Codru a les caractéristiques d'un plan de charriage.

Nappe de Baia de Arieş (Muncel). Cette nappe est formée de la série de Baia de Arieş et de la formation des marbres de Sohodol. Selon notre opinion la nappe de Baia de Arieş comprend outre la nappe de Baia de Arieş dans l'acception de Ianovici et al. (1976) la partie supérieure de la nappe de Muncel-Lupşa (Dimitrescu et al., 1974 ; Dimitrescu et al., 1977). Dans le bassin de Vidrişoara, le plan de charriage d'entre les nappes de Biharia et de Baia de Arieş coïncide à celui tracé par Giuşcă (1937). Tel qu'on peut observer sur l'esquisse tectonique, la nappe de Baia de Arieş se continue sans discontinuité de Baia de Arieş jusqu'à la crête de Biharia, nappe séparée depuis 1936 par Rozlozsnik sans lui donner un nom. Un seul argument est suffisant pour démontrer la position en nappe de charriage de la série de Baia de Arieş à savoir : cette série, à métamorphisme initial de degré plus élevé et conservant les traces de trois métamorphismes, surmonte les séries de Belioara et de Biharia, à degré de métamorphisme initial plus réduit et affectées rien que d'un ou de deux métamorphismes. Il s'agit par conséquent d'une relation anormale. Cet argument est complété et renforcé par les observations suivantes : aux environs de Băişoara et sur la vallée de Vidrişoara, les roches de la série de Baia de Arieş exemptes de granites, prennent contact directement avec les granitoïdes de la série de Biharia ; des roches terrigènes ou migmatiques de type gneiss de Mihoeşti de la série de Baia de Arieş, reposent sans transitions sur les roches vertes de la série de Biharia. Il est question également d'une relation anormale, ces roches ayant des ambiances géotectoniques d'accumulation différentes et éloignées les unes des autres. Ensuite à Băişoara, entre la série de Baia de Arieş et celle de Biharia s'interposent des conglomérats permiens ; la partie supérieure de la série de Biharia et celle inférieure de la série de Baia de Arieş présentent des termes lithostratigraphiques et pétrographiques différents le long du contact de ces deux séries ; tout comme en d'autres cas, ce contact est jalonné des milonites ; la série de Baia de Arieş peut débuter par les mêmes termes, autant sur la série de Belioara, où le contact tectonique est incontestable, que sur la série de Biharia, là où le contact tectonique n'a pas été reconnu jusqu'à présent.

Admettons que toutes les nappes en question sont des nappes de cisaillement. C'est le cas pour les nappes de socle. L'image représentée sur l'esquisse tectonique met en évidence la grande distribution aréale et la continuité de ces nappes de charriage sur des surfaces considérables, étant envisagées de cette manière d'après l'ordre de grandeur à côté des nappes de socle connues dans les Carpathes Orientales et Méridionales. En raison de plusieurs points de vue, les nappes de Biharia et de Baia de Arieş ressemblent aux nappes de charriage préalpines de la zone cristallino-mésozoïque des Carpathes Orientales. On pourrait mentionner là-dessus l'absence de leur couverture sédimentaire mésozoïque — nous avons surtout en vue la nappe de Biharia — et la participation à leur constitution d'une seule série cristalline préhercynienne. Sans être convaincantes, ces ressemblances pourraient être considérées dans le cas de l'attestation d'un âge paléozoïque supérieur pour la tectogenèse paroxismale qui a engendré ces nappes, hypothèse soutenue par Bleahu et al. (1981). Pour ce qui est des nappes de Finiş et d'Arieşeni, elles font partie du système de Codru et sont d'âge méditérranéen. La couverture post-tectonique pour ces nappes débute par le Sénonien en faciès de Gosau.

Conclusions

En guise de conclusion notons que : l'individualisation de la série de Codru à titre de pile de métamorphites comparable à la série de Someş en ce qui concerne son histoire géologique ; le passage des séries de Biharia et de Muncel de la nappe de Codru (Dimitrescu et al., 1974) à la base de la série de Biharia de la nappe de Biharia ; jonction de l'horizon inférieur de la série de Muncel de la nappe de Muncel-Lupşa (Dimitrescu et al., 1974 ; Dimitrescu et al., 1977) à la partie supérieure de la série de Biharia de la nappe de Biharia, et la parallélisation du reste de la série de Muncel de la même nappe avec la série de Baia de Arieş ; l'argumentation de la continuité entre la nappe de Muncel du Bihor central et la nappe de Baia de Arieş est marquée par l'atribution de cette unité à la partie supérieure de la nappe de Muncel-Lupşa d'une part et par la continuité entre la nappe de Biharia et la partie inférieure de la nappe de Muncel-Lupşa, d'autre part.

Je tient à exprimer mes remerciements à I. Hârtopanu pour l'amabilité de nous mettre à la disposition les échantillons recueillis des coupes transversales à travers les unités lithostratigraphiques présentes dans la zone, bordant les affluents de gauche de l'Arieş, entre Valea Caselor de Cîmpeni et la vallée de Lupşa.

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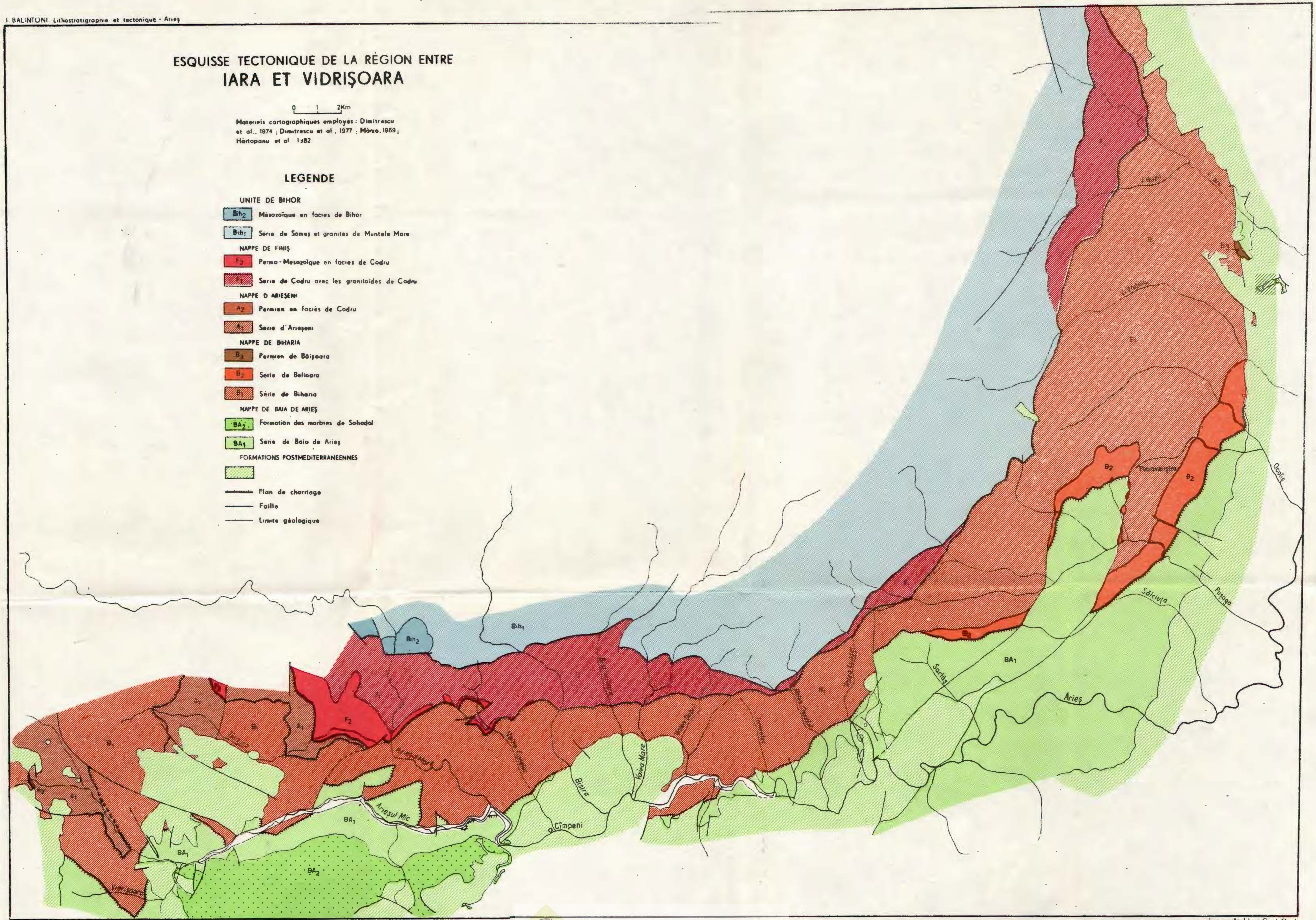

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

NOTE SUR LA PRÉSENCE DE QUELQUES LAMBEAUX DE RECOUVREMENT DE LA NAPPE DE FINIŞ, DANS LE BASSIN DE LA VALLÉE DU DRĂGANU (MONTS APUSENI)[1]

PAR

ION BALINTONI[2]

Finiş Nappe. Someş Series. Tectonic unit. Tectogenesis. Alpine orogenesis; Apuseni Mountains — Bihor Unit — Vlădeasa Massif.

Abstract

On the Presence of Some Outliers of the Finiş Nappe in the Drăganu Valley Basin (Apuseni Mountains). In the Drăganu Valley, the Someş Series, overlain by the Mesozoic in Bihor facies, underlies tectonically the Mezeş Series, overlapped by the Permian in Codru facies. The characteristics of this tectonic unit lead to the conclusion that it is the Finiş Nappe of the Codru system, emplaced during the pre-Gosau, Meso-Cretaceous paroxismal tectogenesis.

Introduction. Dimitrescu (1959) et Cîmpeanu et Cîmpeanu (1968) séparent dans la région du bassin de la vallée du Drăganu deux séries cristallines : mésométamorphique et épimétamorphique. Ces deux séries se trouveraient en relation de discordance lithostratigraphique et de métamorphisme, celle épimétamorphique représentant la couverture transgressive métamorphisée ultérieurement à celle mésométamorphique, Sur la Carte géologique de la Roumanie au 1 : 50 000 (feuille Remeţi — 1973 — et feuille Ciucea — 1974), la série mésométamorphique est dénommée série de Someş et celle épimétamorphique, série de Mezeş. Entre les deux séries apparaît une limite de transgression.

Discussion de la structure. L'auteur considère que la série de Mezeş est entraînée dans une nappe de charriage reposant sur la série de Someş, parce que ces séries cristallines en superposition supportent des formations non métamorphiques complètement différentes comme âge

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[2] Institutul de Geologie şi Geofizică, str. Caransebeş nr. 1, 78344, Bucureşti, 32.

et faciès. Ainsi, aux sources du ruisseau Onăcăi, la série de Someș supporte des sédiments mésozoïques à compter du Trias inférieur, alors que sur le cours inférieur du même ruisseau et dans la vallée du Drăganu, la série de Mezeș est surmontée par des conglomérats polygènes permiens d'une puissance de quelques centaines de mètres, suivis de tufs rhyolitiques soudés de même âge, le Mésozoïque faisant défaut. La séquence mésozoïque reposant sur la série de Someș est caractéristique pour l'unité de Bihor, tandis que le Permien de type affleurant aux sources du ruisseau Onăcăi peut être parallélisé à la couverture de la nappe de Finiș du système des nappes de Codru, tel qu'il a été décrit par Ianovici et al. (1976) et Bleahu et al. (1981). Il s'ensuit donc, que dans la région en question, l'unité de Bihor préserve des lambeaux de recouvrement de la nappe de Finiș (Finiș-Ferice-Gîrda, in Bleahu et al., 1981), bien connue le long de la vallée de l'Arieș (nappe de Gîrda). Un autre argument pour cette parallélisation est la présence du cristallin dans la constitution des lambeaux de recouvrement de la vallée du Drăganu, si on tient compte que du système des nappes de Codru rien que la nappe d'Arieșeni, située structuralement bien plus haut, comporte du cristallin dans sa constitution. En outre le cristallin de la nappe d'Arieșeni diffère totalement de celui de la série de Mezeș.

Cette argumentation structurale, quoique suffisante pour soutenir mon point de vue est aussi confirmée par les observations suivantes :

1. Le Permien manque sur la série de Someș ;

2. Nulle part la série de Mezeș ne supporte des sédiments mésozoïques ;

3. Il n'y a pas d'endroit où le contact entre les séries de Someș et de Mezeș soit recouvert par des formations permiennes ou bien mésozoïques antérieures au Crétacé supérieur ;

4. Sur le versant gauche de la vallée du Dașoru, la série de Mezeș chevauche évidemment le Mésozoïque de type Bihor ;

5. Les quartzites werféniens des sources du ruisseau Onăcăi apparaissent entre la série de Someș et celle de Mezeș. Cela est observable sur les versants du ruisseau, la stratification des quartzites werféniens étant parallèle à la base de la série de Mezeș qui les recouvre.

6. Les relations structurales des sources du ruisseau Onăcăi ne sauraient expliquer par une faille de type de celle tracée sur la carte de Cîmpeanu et Cîmpeanu (1968) et sur la feuille Ciucea (1974). Ils ont figuré une faille verticale qui soulève le compartiment est, formé seulement des roches de la série de Mezeș, tandis que le compartiment ouest n'est constitué, au-dessous du Mésozoïque, que des roches de la série de Someș. Etant donné que les roches de la série de Mezeș ont des pendages de 60° environ vers l'est, elles devraient apparaître plus vers l'ouest dans le compartiment ouest, au-dessous du Mésozoïque, ce qui n'est pas vrai. Le plan de charriage a été peut-être confondu avec une faille à cause de sa grande inclinaison. Ce fait met en évidence un intense plissement postcharriage, à orientation NNE-SSW à peu près, observable également en d'autres parties du nord de la Transylvanie.

Une conséquence importante de la présence des lambeaux de recouvrement de la nappe de Finiş dans la vallée du Drăganu est que le massif de Gilău apparaît dans une grande fenêtre tectonique, étant dépassé entièrement, au moins par la nappe de Finiş, au cours de la tectogenèse paroxysmale pré-Gosau. L'ouverture de cette fenêtre tectonique, logée au-dessus du massif granitique Muntele Mare, a été favorisée de l'érosion différentielle due au ploiement diapir postérieur aux charriages de ce massif granitique.

Constitution de la nappe de Finiş. Comme on a mentionné ci-dessus, la nappe de Finiş comporte une série cristalline apparemment épimétamorphique, série de Mezeş, des conglomérats polygènes rouge-violacés et tufs rhyolitiques soudés permiens. Longeant la vallée de l'Arieş, le cristallin de la nappe de Finiş est représenté par la série de Codru (Hârtopanu et al., 1982), qui a dans sa constitution des amphibolites et des roches terrigènes, traversées et migmatisées par les granitoïdes de Codru pendant un métamorphisme ultérieur. Le degré initial de métamorphisme de la série de Codru a atteint la zone de la staurotide. La série de Mezeş est surtout terrigène, dépourvue de granites et migmatites et à degré de métamorphisme initial évident dans la zone de la chlorite, au moins dans la vallée du Drăganu (Ştefan et al., 1974). Cependant Cîmpeanu et Cîmpeanu (1968) citent de la biotite en plusieurs endroits. Quelques lames minces taillées dans des roches à biotite recueillies du ruisseau Onăcăi ont mené à la conclusion qu'il s'agit d'une biotite de mésozone, chloritisée dans la plupart au cours d'un événement ultérieur : elle a été déformée et fracturée durant le processus de foliation où se stratifient la chlorite et la muscovite de néoformation. Autrement dit, il est question que dans ce cas se développe une séquence polymétamorphique, le dernier métamorphisme étant intensément rétromorphique. Par conséquent, il est possible d'admettre que la série de Mezeş représente une partie non migmatisée et non granitisée de la série de Codru.

La position charriée de la série de Mezeş est bien argumentée par Horvath (in Marinescu et al., 1980), voire dans le massif de Mezeş où elle repose en position horizontale sur le Mésozoïque du faciès de Bihor, transgressif sur la série de Someş. Les lambeaux de recouvrement de la nappe de Finiş du bassin de la vallée du Drăganu se continuent autant vers le sud que vers le nord de l'aire représentée sur la carte jointe à cette note.

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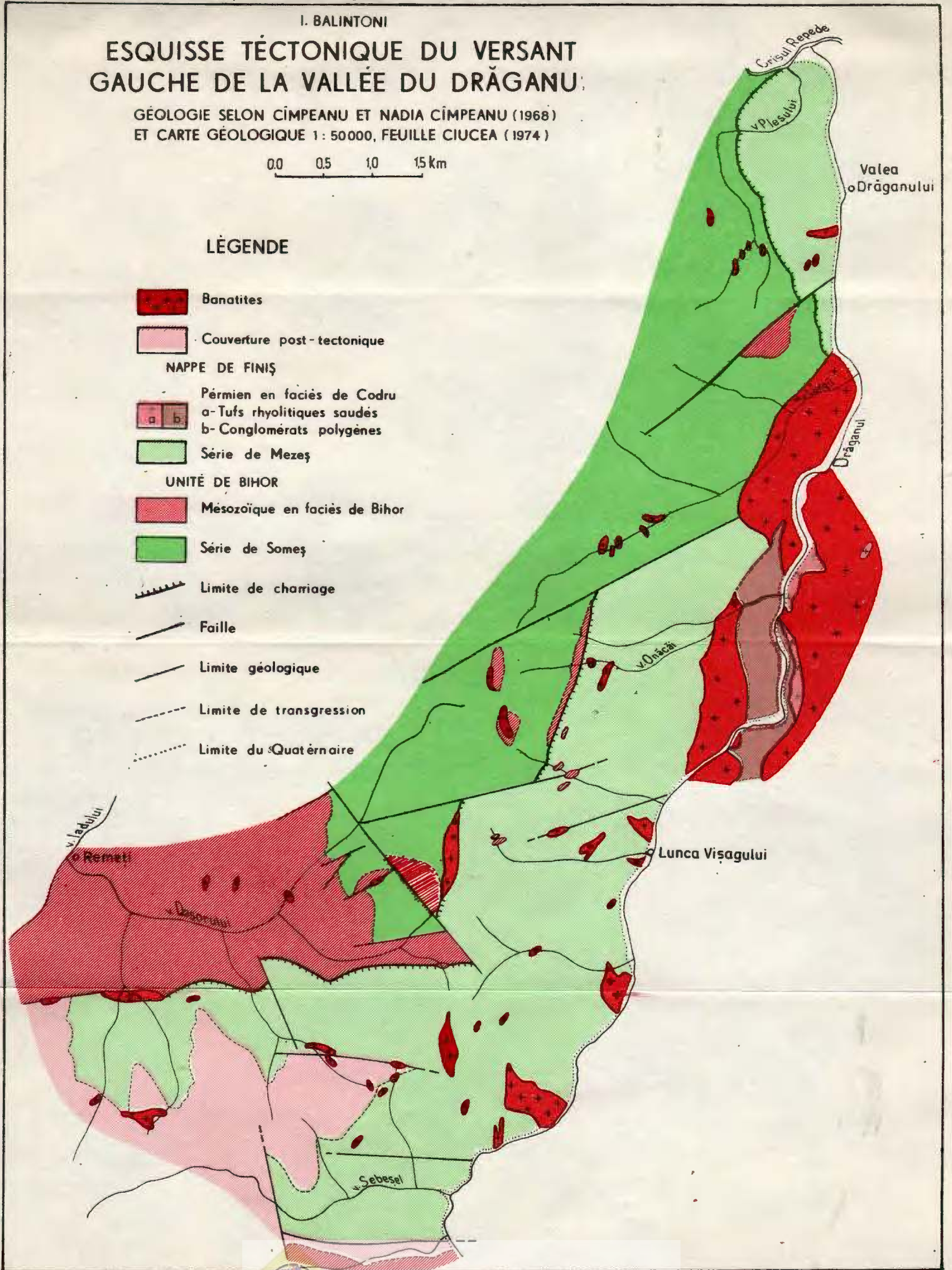

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METAMORPHISM AND DEFORMATION — FURTHER INDICATORS IN ESTABLISHING THE LITHOSTRATIGRAPHIC SUCCESSION OF SOME POLYCYCLIC FORMATIONS [1]

BY

VIORICA IANCU [2]

Stratigraphy of crystalline formations. Lithostratigraphy. Tectonics and magmatism; South Carpathians — Crystalline Getic Domain; Crystalline Danubian Domain.

Sommaire

Métamorphisme et déformation — indicateurs supplémentaires pour établir la succession lithostratigraphique de certaines formations polycycliques. L'étude détaillée des formations métamorphiques polycycliques des Carpathes Méridionales a permis de séparer des séquences à évolution spécifique des événements métamorphiques et de déformation. La détermination de l'âge des intervalles de sédimentation aussi bien que des principales orogenèses qui ont affecté ces formations s'appuie sur des études comparatives prenant en considération les relations réciproques entre des formations à évolution différente, vu les données d'âge connues.

The main data, rendered here as a synthetic table, have been used as arguments for the separation of lithostratigraphic units in a paper presented at the 12th Congress of the Carpatho-Balkan Geological Association (Iancu, 1983). The table has not been annexed to the mentioned paper for lack of space.

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The data refer to several lithostratigraphic units (formations and groups) thoroughly studied on certain areas in the western part of the South Carpathians. Those formations are to be found in the pre-Alpine basement of the main Alpine tectonic units : Danubian, Getic and Lower Supragetic (Pl. I). The present configuration of the Supragetic nappes in the Dognecea zone will constitute the subject of another paper (Iancu, 1985).

The above-mentioned Table presents the main metamorphic events (M), characterized by the specific parageneses and minerals (column 3) and deformational ones (e.g. sets of planes S, column 5, fold generations, column 6). Further characterizations are brought by the description of the most typical features (mineral lineations and their relationships with the fold axes, transposition types, etc. ; column 7) as well as of the associated migmatites and magmatites (column 8). Recent papers, with a general or symetric character referring to the South Carpathians, are due to Savu et al. (1978 a, b), Bercia (1975), Berza (1978). A more recent lithostratigraphic correlation of the Carpathian metamorphites was carried out by Kräutner (1980), following further details given in the Guidebook to Excursion A1 (Kräutner et al., 1981) as a result of the researches accomplished in the South Carpathians on the occasion of the 12th Congress of the Carpatho-Balkan Association.

The present paper deals with the probable ages of the sedimentary (premetamorphic) sequences (Tab., col. 1) and of the main metamorphic events associated with pre-Alpine orogenic cycles (col. 2).

Mention should be made that when estimating the approximate ages all the age determinations (radiogenic, paleontologic and palynologic) known in the relevant literature referring to the study formations have been taken into account.

Our conclusions are rough if one considers the possible space and time variations of the T and P parameters (variation of the geothermal gradients, variation of the superposition degree of the thermal maximums with the deformational ones, diachronous zonations associated with the same metamorphism or zonations superposed to areas with a different previous evolution, etc.).

The paleontological and palynological age determinations are due to : Stănoiu (1972, 1973), Năstăseanu (1975), Visarion (unpublished data), Maier, Visarion (1976). All this refers to the metamorphosed Paleozoic formations. Their present relations with the Precambrian basement are sometimes direct, other times they may be inferred from comparative studies related to deformational and metamorphic elements.

For the Precambrian metamorphic formations and the associated granitoids the radiogene ages known from the papers of Soroiu et al. (1970), Bagdasarian (1972), Mînzatu et al. (1975) have been taken into account. The latest and more interesting radiogene data are found in the paper of Pavelescu et al. (1979), which also includes determinations on zircon of different generations using the U/Pb method. The paper also deals with the interpretation of the results obtained on minerals with a different (pre-metamorphic relics, metamorphic minerals) or very complicated (restructured or regenerated minerals) history.

A succession in time of the main metamorphic events associated with active orogenic periods on the partly rigid zones, either on orogenic reactivated zones or on the new sedimentary areas, has been described on the basis of the synchronism, succession and superposition relationships of events of metamorphism, deformation, migmatisation and magmatism.

Mention should be made of the nonhomogeneous character, on a regional scale, within the same tectonic units, of entities regarded as stratigraphic markers in respect of the number, type and succession of the main metamorphic and deformational events.

One may distinguish areas in which the oldest metamorphites (pre-Caledonian) are represented as well as areas of pre-Devonian (Caledonian) regional remetamorphism. The latter are sometimes reworked in the Variscan orogenesis like the formations with Caledonian progressive metamorphism.

Thus, in the Lower Danubian Unit (according to Berza et al., 1983), Precambrian formations (Drăgşan and Lainici-Păiuş series) occur, up to now considered with a unique regional metamorphism in the almandine amphibolite facies (Iancu, 1974 ; Berza, Seghedi, 1975 a) and the cordierite amphibolite facies, respectively (Savu, 1970 ; Berza, 1978), affected by Variscan retrometamorphism.

The evidence obtained by mineralogical and microstructural studies attests the effects of high-grade, superposed metamorphism, under different thermodynamic conditions determined by the change of the tectonic regime, during the pre-Caledonian orogeneses (Tab.). The first metamorphism of the basic and intermediary rocks from the dyke system separated by Berza and Seghedi (1975 b) within the Lainici-Păiuş Series is assigned to the Caledonian orogenesis.

Likewise, a first phase of regional retromorphism and the dynamic metamorphism at the chlorite zone level (blastomylonites), which affected the Drăgşan and the Lainici-Păiuş series, are assigned to the Caledonian orogenesis.

The Getic Nappe — Bahna and Porţile de Fier outliers — provides direct information on the regional remetamorphism of the Sebeş-Lotru Series (Group), under the almandine amphibolite facies conditions. This metamorphism may be correlated with a significant structural reorganization proved by the relict structural elements (Hârtopanu, 1978 ; Iancu, Hârtopanu, 1979, 1982).

In the Lower Supragetic Units there occur formations of the Sebeş-Lotru Group — Russo-Săndulescu et al., 1978 — (Tilva Drenii augen gneisses formations) in which the effects of the pre-Caledonian metamorphisms are maintained almost unchanged. Nevertheless, there are also zones in which the Precambrian formations (Bocşiţa-Drimoxa Formation = micaceous gneisses complex, Codarcea, 1931 ; Bocşiţa-Drimoxa Series, Constantinof, 1980 — and Valeapai Formation) are strongly affected by remetamorphism in the greenschist facies with advanced transposition and metamorphic differentiation, considered to be associated with the Caledonian orogenesis. The superposed effects of the Variscan orogenesis are displayed by a new generation of folds and microfolds with plane-axial cleavages, partly penetrative, accompanied by mineral neoformation.

TABLE WITH METAMORPHIC A

	Probable age		Metamorphic events (M), specific mineral paragenesis	Corresponding planes (S)
	Sedimentation	Metamorphism		
				SUPRAGET
MONIOM GROUP Cârșie Formation Valea Satului Fm	Lower Carb. Devonian (D)	Alpine Variscan polyphasic (Sudetic and Erzgebirge)	Contact metamorphism Dynamic metamorphism M_1b: Blastomylonitic recrystalization M_1a: Low green schists facies, selective: cl, ab, ep, ac, ser, ca.	S_3: blostamylonitic foliation S_2: axial plane cleavages, lo recryst. (calcite) S_3 S_1 ‖ S_2 S_1: Oriented M_1 min., part penetrative, differentiated S_i: well-preserved bedding
VODNIC GROUP Tîlva Mare Fm Dognecea (terrigenous) Fm Rafnic (volcanogenous) Fm	Ord-Silur? (O—S) Lower Cambrian (Cb 1)	Alpine Variscan Caledonian (Taconian Ardenian?) Early Caledonian (Sardinian?)	Dynamic metam., contact metamorphism M_2 Low-grade remetam., local, mainly dynamic M_1b Tarde (inter) kinematic ab and subsequent dynamic metamorphism M_1a progressive green schists facies cl, ep, ac stilpn. (oriented dynamic phases) The same in matrix and in ab porphyroblasts	S_2: partial reorientation, par penetrative S_1: oriented M_1 min.; Si II in ab porphyroblasts S_i: modified by common foldi with basement (mesometan
(BÎRZAVA GROUP?) Bocșița-Drimoxa Fm	Upper Proterozoic (?) (Pr 3) (?)	Alpine Variscan Early Caledonian Assyntic Pre-Assyntic (Grenvillian)	Contact metam. Dynamic metam. M_3b Green schists facies, static: cl, turm q M_3a Green schists facies, regional remetam: ab, cl, mu, cp, ac- M_2 (interkinematic): and, ab-olig, q M_1(MP) almandine-amfibolitic facies, relict: gr, bi_1, hb_1, mu_1, plag.	S_4: shear planes S_3: crenulation cleavages, part reoriented min. S_2: regional transposition a crystallization of M_3 min, wi metamorphic differentiatio S_1: oriented, relict, M_1 min e closed in interkinemat plag.
SEBEȘ-LOTRU(?) GROUP: — Valeapai Fm.	(Middle?) Upper Proterozoic Pr (2?)3	Alpine Variscan Early Caledonian Assyntic (?) Grenvillian	M_5 (banatitic contact) px, oliv, bi, co, kfsp. M_4 retrogressive, local. M_3(LP), retrogressive, regional cl, ac, ab, ep, ca M_2(LP), synkinematic: and, bi_2 mu_2hb_2 M_1/M_2? (MP) relict: bi_1, gr, plag, hb_1	S: dynamic, local S_4: partly penetrative S_3: regional remetam., transpos tion, q segregation, orient M_3 min. S_2: synkinematic recryst, ne min., deformed M_1 mi S_1: relict, destroyed
— Tîlva Drenii (augen gneisses) Fm.	(Middle?) Upper Proterozoic Pr (2?)3	Alpine Grenvillian Pre-Grenvillian	M? (banatitic contact) co, and, bi, mu M_2 (MP), isofacial regional reorganisation: plag, bi_2, q, mu M_1 (MP), relict: gr, plag, kfsp. (deformed augen), hb_1, ky, bi_1	S?: local laminations S_2: oriented M_2 min. S_1: destroyed, deformed, reli metam. porphyroclasts
				GETI
Jidoștița (Ivanu) Fm.	Upper Proterozoic (Pr 3)	Early Caled? (Variscan) Assyntic Pre-Assyntic	M_3: retrograde, cl. zone, local M_2: (LP), static: cord, sill (fibrolite) M_1: Prograde bi ≠ gr zone (MP?)	S_3: oriented M_3 min., loc S_2: axial plane cleavage, parti reorientation, local S_1: oriented bi ≠ S_0 Common for basement
SEBEȘ-LOTRU GROUP (Bahna, Porțile de Fier outliers)	(Middle?) Upper Proterozoic Pr(2?)3	Early Caled? (Variscan?) Assyntic Grenvillian Pre-Grenvillian	M_4: Local remetam. bi → cl zones M_3 (LP) static, areal overprint; an, co, sill + kfsp (neo-zoning) M_2 (MP): ky lt, gr, bi, mu. Regional, isofacial remetamorphism M_1 (MP): ky_1 $sill_1$, gr, bi_1, hb_1, px. Relict paragenesis and pyroxenitic rocks (eclogitoides, quartzites)	S_θ: dynamic metam. S_3: local, partly penetrativ reoriented relics and ne crystalliz. ab, bi, q S_2: regional reorganisation a transposition (NE/SW) S_1: local preserved on mesos pic scale.

DEFORMATIONAL FEATURES

Deformational elements generations of folds (B)	Important features	Igneous intrusion Migmatisation	Known ages
U N I T S (sensu stricto)			
B_1 Mullion structures around the congl. pebbles and boulders B_2: syn — S, asymmetric folds B_1: Folding of S_0, locally preserved.	Incipient and differentiated, mecanic transposition. Well-preserved bedding and sedimentary, magmatic structures $Lh_2 \parallel LiS_0S_2$ $Lh_1 \neq Lh_2$	Banatitic veins Pre-metam, magmatic rocks (subsequent low-grade metam) a. basic — intermediate dyke system b. allochthonous, ultrabasic rocks.	Palynologic ages: Carboniferous Devonian (Visarion, Iancu, 1985)
B_2: Symmetric folds, post S_1 and ab porphyroblasts B_1: synchronous with M_1	Folded and obliterated contact with basement Common folding and metam. with basement. Variscan overprint Polyphasic prograde metam. (green schists facies, stilpnomelane and biotite zones) S_0 modified by metamorphic differentiation and transposition	Sodic metasomatism (?) (ab porphyroblasts) Deformed magmatic rocks (granodiorite-diorite small dykes)	
B_3: post S_2, inclined or recumbent folds B_2: synmetam. (M_2) transposition folds (similar, intra — S_2) B_1: "dead" folds, refolded and sheared	Composed, nonplanar structures Incomplete transposition on S_2 Deformed q segregations with regional extent Polystadial minerals: relict metam. porphyroclasts of olig. and albite porphyroblasts (post — S_2) Repeated metamorphic differentiation	Premetamorphic magmatic intrusion (small scale) Arteritic migmatisation (,,effet du socle")	
B_4: post — S_3, concentric B_3: syn — S_3, intrafolial. similar to concentric, refolded B_2: syn — S_2, microscopic, folds B_1: relict, refolded folds	Deformed andalusite replaced by kfsp Regional remetamorphism, quartz segregations Repeated transposition B_3 axis $\parallel Lm_3 \parallel Li(S_2/S_3)$ Distorted, older lineations	Banatitic intrusions with contact metamorphism (bi_3, and, cord, kfsp) and metasomatic hydrothermal metamorphism Relict metatectic migmatisation	
Deformed, augen kfsp. (with S shape) Mullion aggregates	Relict metam. porphyroclasts and recrystallised metam. matrix Polycrystalline (deformed kfsp and recrystallized plagioclases) augen feldspars	Relict metam. migm. (metamorphosed migmatites)	
U N I T S			
B_2: concentric, symmetric folds (refolding) B_1: Tight folds associated with penetrative foliation in. basement	Local retrogressive metam. Static overprint Prograde regional metamorphism producing overprint of basement (at bi zone, level)	Arteritic synmetamorphic migmatisation (,,effet du socle")	
$B_4 \perp B_3 \parallel B_2 \neq B_1$ B_4: regional, E—W, open. B_3: post S_2, inclined to recumbent folds B_2: transposition, syn — S_2 folds with NE—SW trending, concentric to similar type B_1: relict, pre — S_2, refolded folds with low symmetry	Deformed migmatites Superposed zoning Local transposition on S_3 Advanced transposition on S_3 Lm_2, NE—SW trending, $\parallel (B_2)$ axis Relict Lm_1E—W orientation, rotated by B_2 folds; Lm_1 — Lm_2 Repeated remetam. of eclogitoids Composed, nonplanar, inhomogeneous structures.	Polystadial pegmatites III Re — migm. in M_3 high grade zones II Syn — M_2 metatectic migm. local intrusion of granitoid rocks I Metatectic migmatisation (syn — M_1)	Rb: Sr: 850 m.y. (Bagdasarian) A: 1000—1100 m.y.-metamorphic magmatic event or B: 1600 m.y.; detrital, premetamorphic zircon 340 m.y.-Hercynian retromorphism (Pavelescu et al.)

1	2	3	4	5
				DANUBIAN
PALEOZOIC (TULIŞA) GROUP Poiana Mică Fm	Lower Carb. —Devonian (C_1—D)	Variscan	Uncertain data with respect to: metamorphic events M_1: Low green schists facies : cl , ser, q, ac, turm, ru.	S_2: p.a. cleavages, penetrative in basement S_1: oriented M_1 min. S_0: deformed bedding
Valea Izvorului Fm	Ord-Silur. (O—S)	Caledonian (uncertain)		
DANUBIAN COMPLEX (Peri-Caledonian belt) LAINICI-PĂIUŞ GROUP and (Granitoid batholits) (pre-Silur, dyke system)	Upper Proterozoic (Pr 3)	Alpine Variscan Early Caledonian Assyntic (Danubian) Grenvillian	M_5 Dynamic metam. M_4 Local, retrogressive M_3 Retrogressive overprint with areal extent: cl, ab, ac, stilp. ep M_2 (LP) synkinematic: bi, mu Fe oxides, graphite, hb_2, an, co, sill M_1 (MP), relict: gr, bi_1, hb_1 + + plag, clpx.	S_5 common foliation with Lower Paleozoic S_4 blastomylonitic zones; S_3 crystallization of M_3 min. S_2 syndeformational crystallization of M_2 min. S_1 relict M_1 min. disrupted, reoriented
DRĂGŞAN GROUP (Caledonian belt)	(Middle?) Upper Proterozoic	Variscan Early-Caledonian Assyntic Grenvillian Pre-Grenvillian	M Green schists facies, local remetam. M_4 Green schists facies overprint in metam. rocks; adaptation of magmatic rocks M_3 (LP) contact metam: sill, mu, Q M_2 (MP): isofacial reorganisation and metastable persistence of st, gr, bi, hb + plag. M_1 (MP) relict: ky, st, bi_1, gr_1, orto + clpx, hb_1 + plag.	S_4 shear planes — blastomylonitic foliation S_3 oriented cl, ab, ac, ep. S_2 regional foliation and layering S_1 microscopic orientation of disrupted mineral relics, locally preserved on mesoscopic scale.

Generally, the retromorphism areas are those within which there are still serious problems related to the existence of Lower Paleozoic series or formations unconformably overlying the Precambrian basement.

In the Lower Supragetic Units there is a contact between a pre-Devonian sequence (Vodnic Group) and the Proterozoic basement, representing an initial unconformity, affected by folding and metamorphism (Iancu, 1983). This Lower Paleozoic sequence, partly representing the green phyllites series (Radu-Mercus, 1962) or the Valea Caraşului Series (Constantinof, 1980), has not been separated up to now by the Precambrian basement occurring in the pre-Alpine Vodnic anticline axis. It has been included either in the Lescoviţa Series (Devonian-Carboniferous, according to Maier, 1979) or in the Valea Caraşului Series (Ordovician-Silurian, cf. Constantinof, 1980) by comparison with formations dated in the Locva massif.

Similar questions occur in connection with the recognition of sequences of different ages within the Precambrian series, e.g. the recognition of unconformities modified by deformations or subsequent transpositions or obliterated by recrystallizations.

6	7	8	9
NITS			
Deformed macrofossils Penetrative shear or cleavage planes in basement	Sedimentary relict structures Sediments of continental type Important stratigraphic and metam discordance in relation with Precambrian basement	Pegmatitic veins	Paleontologic ages: after Stănoiu (1972) — Devonian—Lower Carb. — Ordovician-Silurian
B_3: post — S_2 microfolds B_3: intrafolial, tight, synmetam folds B_1: regional, open.	Evident effects of Caledonian orogenesis Synkinematic LP metam Repeated intrusion of granitoid rocks S_2 regional foliation, common for syn — S_2 granitoids.	III Emplacement of basic dyke system (Early Caledonian) II Post-metam intrusion and arteritic migmatisation metasomatic feldspatisation I Syn — M_2 intrusion of granitoids (Tismana 610 m.y.), Assyntic or Danubian event, after Pavelescu et al.	K-Ar ages for: muscovite, biotite, amphibole — 522 554 m.y. Tismana-Novaci granitoids — 610 m.y. (U/Pb — for zircon) — after Pavelescu et al. (unpublished data with author's assignment)
B_3: regional, post S_2, synchronous with Cerna granitoids emplacement B_3: synmetam with S_2 axial plane foliation B_1: relict folds and min. lineation very scarce on mesoscopic scale.	Polystadial mineral adaptation of granitoid rocks Isofacial, regional remetamorphism Very advanced transposition and reorganisation on S_2. Two sets of mineral lineations in amphibolitic rocks (graben structutes)	IV Pegmatitic and rhyodacitic dykes III Arteritic migmatisation II Syn B_3-granitoid intrusion I Pre-granitoid migmatisation	

It is considered that the individuality and the evolution in time of a lithostratigraphic unit may be inferred also from the comparative study of the metamorphism, deformation and migmatization elements. A first attempt was made for formations included in the Sebeş-Lotru Series (Iancu, Hârtopanu, 1979), individualized as Jidoştiţa Series (Conovici, Conovici, 1978). A similar formation (Ivanu Formation) is to be found in the Godeanu outlier, as well.

Inferences on the synchronism, succession or superposition relationships of major metamorphism and deformation events can be regarded as further arguments when separating lithostratigraphic units of different ages, which might be considered supergroups (Iancu, 1983).

The evolution in time of such units can be different for various areas or rock-volumes. Such arguments must be correlated with information on age, sedimentology (facies, thicknesses, etc. in case of the weakly metamorphosed formations), petrochemistry, etc.

In areas affected by high-grade metamorphic processes, often reiterated, the stratigraphic markers are hardly, even impossible, followed on a mesoscopic scale. In those cases the petrographic markers (generally considered as bedding) as well as the structural ones allow a sequential reconstitution of the geological events, that is of the premetamorphic lithostratigraphic units.

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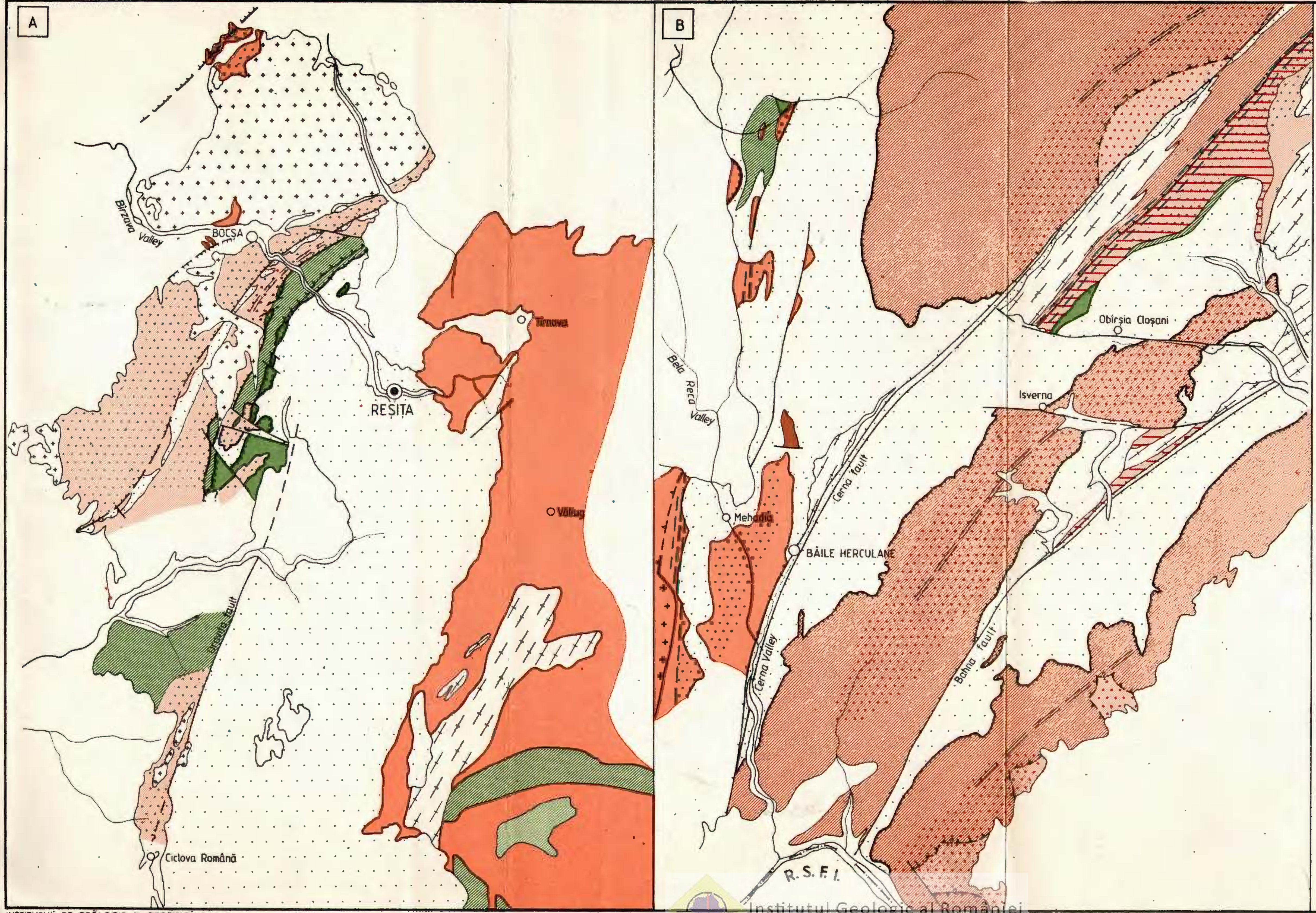

STRUCTURAL SKETCHES OF THE WESTERN AREAS OF THE SOUTH CARPATHIANS

(after the Map of the S. R. Romania, scale 1 : 200 000, modified according to personal data ; further information after Bercia, Hârtopanu, 1980)

LEGEND

- Cenozoic sedimentary formations
- Upper Paleozoic - Mesozoic sedimentary formations
- Middle Paleozoic (Devonian - Lower Carboniferous) formations deformed and metamorphosed during the Variscan orogenesis
- Pre - Devonian Paleozoic formations with Caledonian deformations and metamorphism, reactivated during the Variscan orogenesis
- Precambrian formations (Middle - Upper Proterozoic), with pre-Caledonian polymetamorphism: nondifferentiated (a), formations with two deformation phases associated with Barrovian-type metamorphism (b), formations with one deformation phase associated with initial metamorphism of Barrovian type (c)

AREAS AND ZONES OF REACTIVATION

- Pre - Caledonian low-pressure remetamorphism: a. static ; b. synkinematic
- Reactivated zones in the Caledonian orogenesis: a. folding associated with metamorphic reorganization ; b. shear zones accompanied by recrystallization (blastomylonites)
- Upper Cretaceous - Palaeogene (banatitic) magmatites
- Paleozoic granites
- Pre - Caledonian granitoids
- Alpine nappes (nondifferentiated in time)
- Pre - Alpine tectonic lines (nondifferentiated)
- Reversed faults
- Intra - Miocene longitudinal faults
- Faults (in general)
- Alignment of pre - Alpine recumbent folds (frontal zone)

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

LOWER SUPRAGETIC NAPPES OF THE BANAT, MONIOM-DOGNECEA ZONE [1]

BY

VIORICA IANCU [2]

Getic Nappes. Bocşa Nappe. Moniom Nappe. Supragetic units. South Carpathians. Crystalline Getic Domain. Locva Mountains and crystalline islands of the West Banat; Sedimentary Getic Domain — Locva-Semenic Region.

Sommaire

Nappes supragétiques inférieures du Banat, zone de Moniom-Dognecea. L'esquisse tectonique de la zone de Moniom-Dognecea comporte la configuration et la distribution des nappes de Bocşa et de Moniom, nappes de socle appartenant au groupe de nappes supragétiques inférieures. Ces nappes se trouvent en position externe et inférieure envers l'unité de Timiş. Dans le bassin de Bîrzava, les rapports de superposition sont, de l'ouest vers l'est : nappe de Bocşa, nappe de Moniom, nappe de Reşiţa. Au sud de la vallée de Bîrzava, la nappe de Bocşa surmonte celle de Moniom, se disposant directement sur les unités de couverture : nappe de Reşiţa, unité de Lupac-Dealul Vremii, appartenant au système gétique.

The tectonic sketch, which constitutes the object of this paper (Pl. I), synthesizes the structural relationships between the Supragetic Units and those of the Getic Nappe in the Moniom-Dognecea zone. With a view to pointing out the contribution in the clearing out of the Alpine structure of the study area an evolution of the ideas and of the previous cartographic materials will be presented further on.

The existence of a significant nappe with crystalline, in a position geometrically superior to the Getic Nappe (Murgoci, 1905), was first mentioned by Popescu-Voiteşti (1929) and later specified by Cantuniari (1930). The nappe was rendered evident — in the Banat — by the pointing out of the outliers of the green crystalline overlying the Upper Carboniferous deposits, probably emplaced during the Mesocretaceous time.

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In the Ezeriş-Dognecea zone, Codarcea (1931) distinguished two tectonic units overthrusting each other: the "micaceous gneisses complex" in the west and the "green rocks complex" in the east. The crystalline schists slightly overlap the Carboniferous deposits along a subvertical plane.

In 1934 Streckeisen correlated the Alpine units of the South Carpathians and pointed out : I. autochthonous massifs ; II. Getic Nappe ; III. upper nappes. On his map the upper nappes occur in the Banat, Poiana Ruscă, in the north of the Sebeş-Cibin massif, the Făgăraş Mts and the Cozia and Leaota massifs.

In 1967 Codarcea, Lupu, Dessila-Codarcea and Lupu took over the idea of the underthrust of the Getic crystaline by an uper unit, called the "Supragetic Nappe". The plane in the base of the Supragetic Nappe follows the Sasca line (pointed out by Schréter in 1910) and continues, with a route similar to that mentioned by Streckeisen, up to the Olt Valley. East of the Olt Valley, the route continues north of the Leaota massif, assigned to the Getic Nappe as it displays a similar Mesozoic cover.

Săndulescu (1975) reported a cover nappe (Sasca-Gornjak) between the Supragetic and Getic nappes, mostly consisting of Permian and Triassic formations. The author correlated the Supragetic Nappe, regarded as a large Alpine nappe, with the Morava Nappe in Eastern Serbia (Grubič, 1974).

Năstăseanu (1978) pointed out three east-trending units in the inner part of the Getic Realm : Reşiţa Nappe and Dealul Vremii and Lupac digitations. They consist of Upper Paleozoic-Mesozoic formations overlapped — from the west — by the Dognecea Nappe ("green crystalline" nappe, sensu Cantuniari, 1930).

The cartographic image of the Banat Supragetic Units is practically unchanged as compared to the one initially given by Cantuniari, Codarcea, Streckeisen (Maier, 1974, 1979 ; Constantinof, 1972). The denominations used lately for the two units (Kräutner et al., 1978, 1981 ; Iancu, 1983, 1984) are : Bocşa Unit for the western nappe and Locva Unit for the eastern one.

Iancu (1983) brought arguments in favour of the existence of an internal structural unit (nappe), west of the Locva and Bocşa units, in a superior position as against the latter, called "Timiş Unit". The more complicated inner structure of the Alpine nappes was revealed pointing out a thrust fault (Ezeriş-Colţan) through which the Bocşiţa-Drimoxa Subunit overlaps the Mesozoic deposits of the Buchin Unit. Two pre-Alpine tectonic units were also mentioned : Valea Satului-Locai blastomylonite overthrust (Variscan) inside the Moniom Nappe and the overthrust which gets the Tîlva Drenii gneisses into tectonic contact with the Bocşiţa-Drimoxa Formation.

Năstăseanu, Maksimović (1983) correlated the tectonic units in the inner part of the Getic Realm north and south of the Danube and distinguished several nappes (from west to east) : Bocşa (= Bocşa Unit)-Morava (= western part of the Morava Nappe) Nappe, Dognecea (= Locva Unit)-Luznica (= eastern part of the Morava Nappe) Nappe, Reşiţa Nappe, and Sasca-Gornjak Nappe.

The present configuration of the Moniom-Dognecea zone is rendered on the Plate I. It presents the extension and the relationships of the two Supragetic Units of the Banat, with the cover nappes of the Getic Nappes System. The sedimentary cover of the Getic Nappe — "Reşiţa-Moldova Nouă" zone — consists of Mesozoic and Paleozoic (Upper Carboniferous-Permian) formations, folded together and involved into thrust structures during the Alpine orogenesis.

All this are the results of the mappings carried out within the Bocşa sheet (Iancu, in Russo-Săndulescu et al., 1982) and the Lupac sheet (Năstăseanu, Iancu, Russo-Săndulescu, unpublished data).

The petrographic and lithostratigraphic content of the above-mentioned units had been rendered in several previous papers : Iancu, 1983, 1984 — metamorphic formations — and Năstăseanu, 1978, 1981 — sedimentary formations.

The units regarded as nappes, found in superposition relations are (from west to east), as follows : Bocşa Nappe and Moniom Nappe, belonging to the Lower Supragetic Nappes System. They overlap the Reşiţa Nappe and the Lupac-Dealul Vremii Unit, assigned to the Getic System.

1. *Bocşa Nappe*. The eastern boundary of the Bocşa Nappe was known as following a linear course, connecting the Colţan tunnel (Bîrzava Valley) with the Dognecea Valley, and from here it turned much to the west. The present paper brings significant modifications as regards the configuration and extension of the Bocşa Nappe. The eastern prolongation with about 5—6 km of the nappe is marked by several outliers, partly described by Cantuniari, assigned to the "green crystalline" (= Monion). This unit also includes the crystalline overlying the Vodnic-Bichinecea Half-window assigned in the previous papers together with the Moniom Nappe to the Locva Unit.

Recent data (Iancu, 1984, 1985) point out the complex petrographic and lithostratigraphic content of the Bocşa Nappe. This unit is constituted of Proterozoic formations with pre-Caledonian polymetamorphism (Tîlva Drenii Formation, Bocşiţa-Drimoxa Formation) and Lower Paleozoic epimetamorphic formations (Vodnic Group : Tîlva Mare quartzitic formation, Vodnic albite porphyroblasts schists formation and Dognecea terrigene formation).

2. *Moniom Nappe*. This nappe has been up to now structurally correlated with the Locva crystalline and called "Locva Unit". It is well individualized in the Bîrzava Valley basin (Moniom Brook, Valea Satului, Locai Brook), Ferendia Valley basin and the upper basin of the Dognecea Valley. In the Dognecea-Lupac sector the Moniom Nappe is overpassed by the Bocşa Nappe directly overlying the cover units of the Getic Nappe (regarded as a large basement nappe).

In order to specify the individuality and southward extension of this structural unit, provided that the correlation with the Locva Unit becomes difficult, we propose the denomination "Moniom Nappe".

The Moniom Nappe consists of Devonian formations (Valea Satului Formation) and Lower Carboniferous formations (Cîrşie metaconglomerates formation, assigned by Maier to the Devonian, 1979), dated on the basis of palyno-protistologic content by Visarion (Iancu, Visarion, 1985). These rocks were affected by low-grade metamorphism — greenschist facies — and deformed during the Variscan orogenesis (Pl. II). The contact between the two formations is marked by a zone of "tectonic mixture" affected by dynamic metamorphism (blastomylonites) and accompanied by small allochthonous bodies of serpentinites. Within the Valea Satului volcano-sedimentary formation mention should be made of a system of magmatite bodies (dykes and veins) — gabbros, diorite-granodiorites deformed and recrystallized under the greenschists facies conditions. The two mentioned types of formations are supposed to have accumulated under different geotectonic conditions and brought into direct contact during the Variscan (pre-Westphalian) orogenesis. The Devonian sequence belongs, in our opinion, to a Variscan paleosuture zone. Our statement is based on specific deformational elements (Iancu, 1984, 1985) and must be supported by further arguments based on petrographical, geochemical and sedimentological studies.

Evidence on the moment or moments of emplacement of the Alpine nappes in the study zone is supplied by the youngest formations affected by the thrust planes and their relationships with banatites.

The youngest formations belonging to the Mesozoic cover of the Supragetic nappes in the Banat are of Lower Cretaceous age. They occur on the Bocşa Nappe, in the Buchin Subunit, and are overlapped by the Bocşiţa-Drimoxa Subunit along the Ezeriş-Colţan-Ocna de Fier Fault (previous to the banatites emplacement). Formations of the same age and of Upper Cretaceous are overthrust by the crystalline of the Timiş Unit in the north-westernmost part of the Bocşa massif. As this thrust line, which affects the Upper Cretaceous, is overlain by Badenian formations the thrusting of the Timiş Unit over the Bocşa Nappe should be linked to the Laramian paroxismal phase.

The recrystallized sandstones and limestones from the Harca Ravine (Ferendia Valley), affected by the laminations which mark out the basal overthrust of the Bocşa Nappe, constitute the only known occurrence (Codarcea, 1931 ; Iancu, 1983) of Paleo-Mesozoic cover (?) on the Moniom Nappe. The mentioned overthrust is straightened out in the westernmost contact zone with the Moniom Nappe, and the mylonites within this zone are penetrated by banatites and affected by the thermal metamorphism generated by them, giving rise to recrystallized mylonites with biotite.

The mylonites in the base of the Moniom Nappe and the Paleozoic formations of the cover units are also affected by the thermal metamorphism generated by banatites.

It is to be mentioned that the Bocşa and Moniom nappes are previous to the banatites emplacement, as first pointed out by Codarcea.

The latest evidence concerning the banatites in the study zone (Russo-Săndulescu et al., 1982) allow the structural correlation of the Dognecea-Ocna de Fier main body with the Bocşa 3 type, on which radiogene ages of 65—56 m.y. (Paleocene) have been obtained using the K/Ar method.

All this as well as the data proving the emplacement of the cover nappes in the Upper Cretaceous (post-Albian, according to Năstăseanu,

1978) situate the thrusts of the units involved between the Albian and the Paleogene (Austrian, Mediterranean or Laramian phases).

At present, considering the data from the "Reşiţa-Moldova Nouă Zone", no specifications can be made on the synchronism or the succession of the emplacement of the two Supragetic nappes regarded either separately or in relation with the Getic cover nappes.

The regional structural correlations (with the Rusca Montană Basin or the southern zone) require a special attention as to the correct spatio-temporal "connection" of the system of overthrust and faults and implicitly of the units generated in a certain phase. The resuming of the whole structure in the intra-Miocene phase (well represented south of the Danube — Năstăseanu, Maksimović, 1983) makes more difficult the clearing up of the regional structural relationships.

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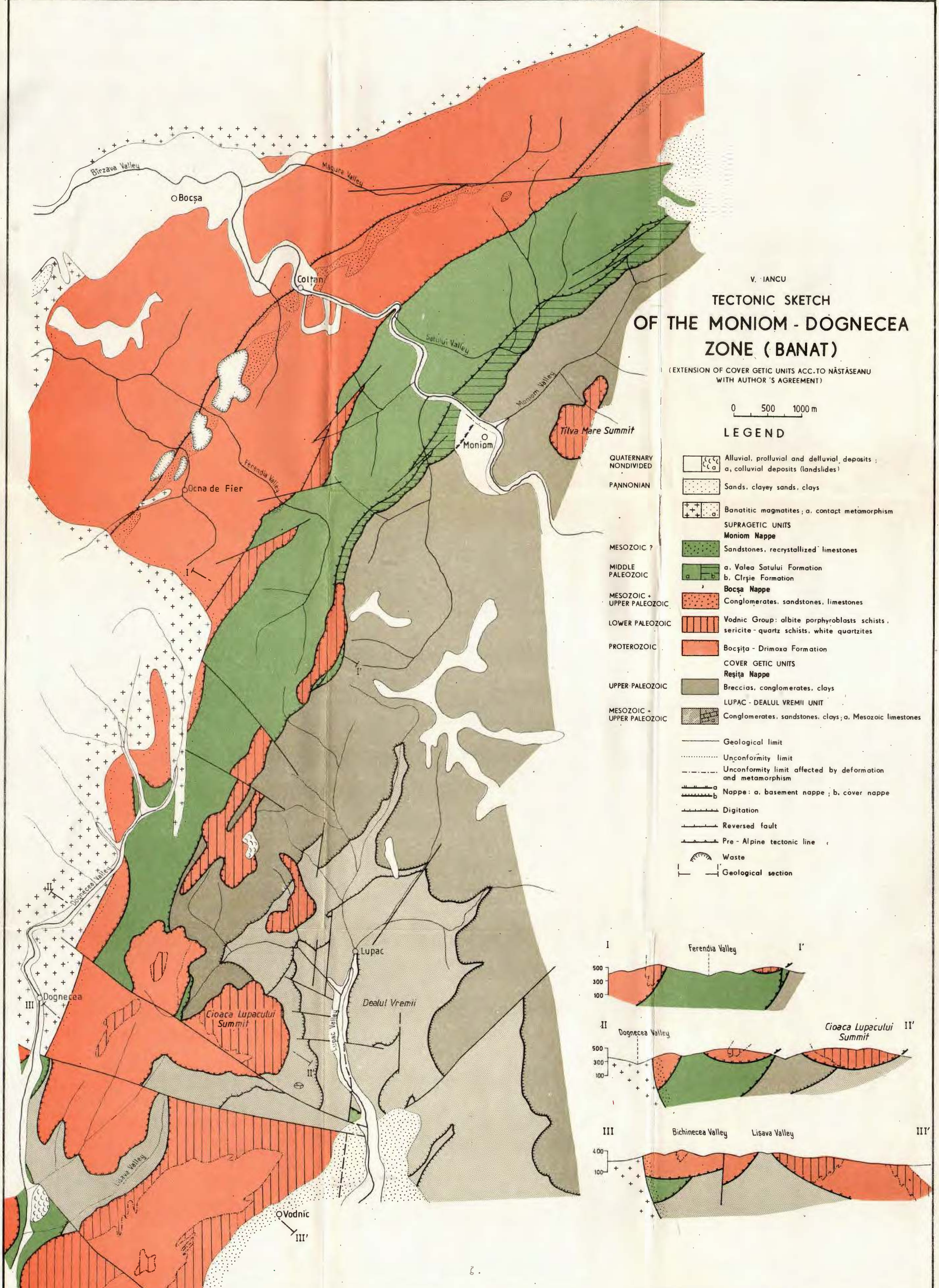
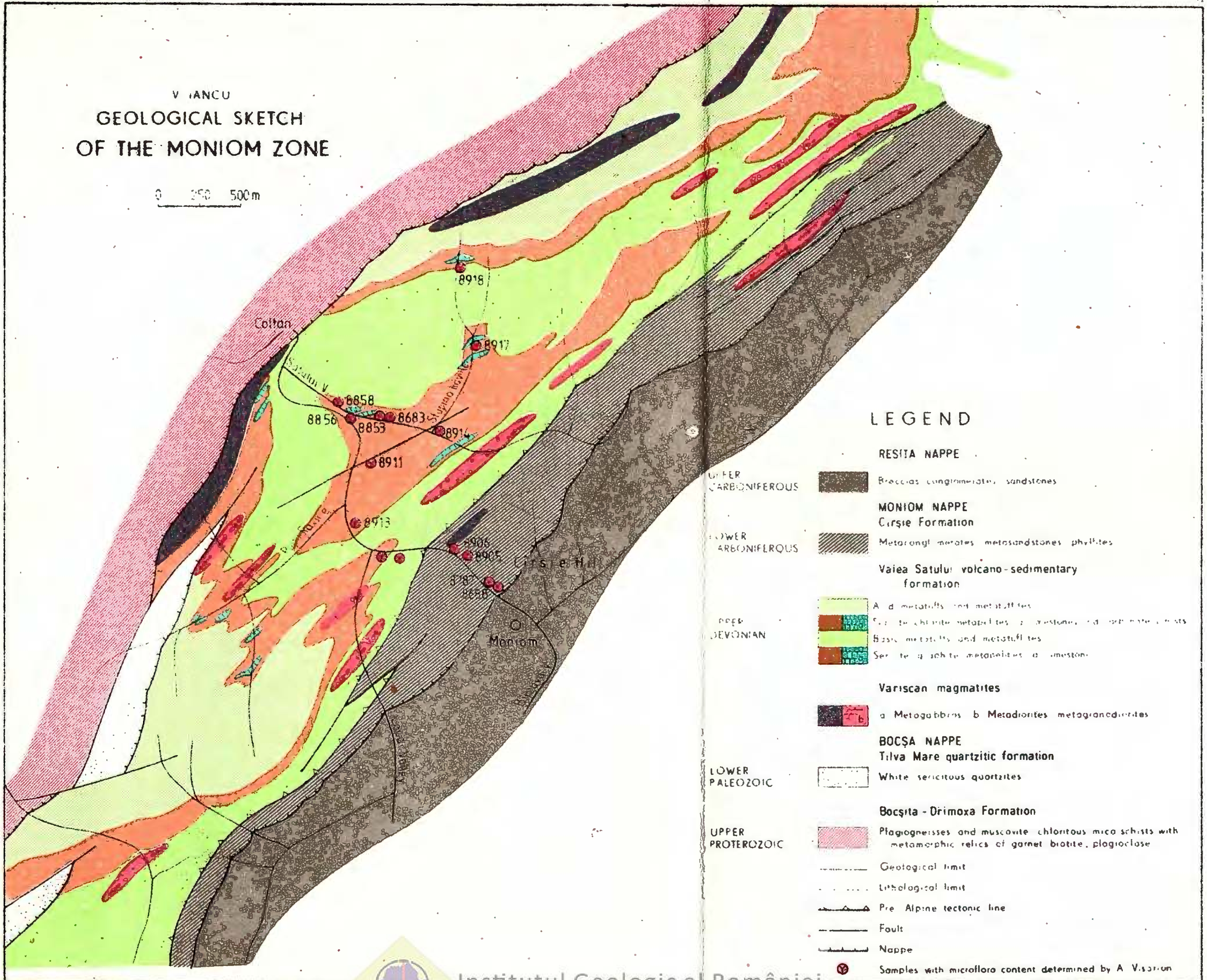

V. IANCU
TECTONIC SKETCH
OF THE MONIOM - DOGNECEA
ZONE (BANAT)
(EXTENSION OF COVER GETIC UNITS ACC. TO NĂSTĂSEANU
WITH AUTHOR'S AGREEMENT)
0 500 1000 m
LEGEND
QUATERNARY NONDIVIDED
Alluvial, prolluvial and delluvial deposits ; a, colluvial deposits (landslides)
PANNONIAN
Sands, clayey sands, clays
Banatitic magmatites ; a, contact metamorphism
SUPRAGETIC UNITS
Moniom Nappe
MESOZOIC ?
Sandstones, recrystallized limestones
MIDDLE PALEOZOIC
a, Valea Satului Formation
b, Cîrşie Formation
Bocşa Nappe
MESOZOIC + UPPER PALEOZOIC
Conglomerates, sandstones, limestones
LOWER PALEOZOIC
Vodnic Group: albite porphyroblasts schists, sericite - quartz schists, white quartzites
PROTEROZOIC
Bocşiţa - Drimoxa Formation
COVER GETIC UNITS
Reşiţa Nappe
UPPER PALEOZOIC
Breccias, conglomerates, clays
LUPAC - DEALUL VREMII UNIT
MESOZOIC + UPPER PALEOZOIC
Conglomerates, sandstones, clays ; a, Mesozoic limestones
Geological limit
Unconformity limit
Unconformity limit affected by deformation and metamorphism
Nappe : a, basement nappe ; b, cover nappe
Digitation
Reversed fault
Pre - Alpine tectonic line
Waste
Geological section
Bîrzava Valley
Măgura Valley
Bocşa
Colţan
Satului Valley
Moniom Valley
Moniom
Tilva Mare Summit
Ferendia Valley
Ocna de Fier
Dognecea Valley
Dognecea
Lupac
Dealul Vremii
Cioaca Lupacului Summit
Lupac Valley
Lişava Valley
Vodnic
I
I'
II
II'
III
III'
Ferendia Valley
Dognecea Valley
Cioaca Lupacului Summit
Bichinecea Valley
Lişava Valley
500
300
100
400



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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

L'OLISTOLITE DE CALCAIRE NÉOTRIASIQUE DE BEJAN (AU NORD DE DEVA — MONTS APUSENI DU SUD)[1]

PAR

MARCEL LUPU, DENISA LUPU[2]

Olistostrome. Olistolith. Barremian. Aptian. Halobia-bearing limestone ; Apuseni Mountains — Mureș Mountains — Zarand Mountains.

Abstract

The Upper Triassic Limestone Olistolith of Bejan (North of Deva — South Apuseni Mountains). In the Bejan olistostrome formation, of Barremian-Lower Aptian age, north of Deva, an olistolith of Upper Carnian-Lower Norian age has been discovered. It consists of grey pelagic limestones partly of *Halobia*-bearing coquinas. The facies of the limestone provides differences with the Upper Triassic sequences appearing in the Moma Nappe (North Apuseni Mountains) or in the Transylvanian Nappe System in the East Carpathians.

A l'occasion des levés géologiques éxecutés pour la carte au 1 : 50.000, feuille Deva, sur l'aire comprise entre la vallée de Boholt à l'E et jusqu'à peu près la vallée de Căbești à l'O, le méridien de la localité de Fornădia au N et celui de la localité de Bejan au S, a été séparée une unité attachée initialement à celle de Căbești, unité dénommée ultérieurement, par l'un des auteurs de la présente note (Lupu, 1983 ; Lupu, in Bleahu et al., 1981) unité de Bejan. Le principal constituant lithologique de cette unité est la formation de Bejan, de type wildflysch, où on a identifié l'olistolite de calcaire néotriasique. C'est celui-ci qui fait l'objet de cette note.

La formation de Bejan est constituée d'une matrice principalement argileuse, écailleuse en partie, de couleur gris foncé-noirâtre, difficile à déceler du fait du degré avancé de recouvrement du terrain. Subordonnément, apparaissent des grès quartzeux à ciment calcaire, lenticulaire, représentant des boudines sédimentaires.

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[2] Institutul de Geologie și Geofizică, str. Caransebeș nr. 1, 78344, București 32.

La frappante ressemblance de la matrice de la formation de Bejan aux principaux termes de la formation de Căbeşti, avec laquelle elle vient en contact vers le N, nous a mené à lui accorder l'âge barrémien-aptien inférieur.

Les olistolites de la formation de Bejan appartiennent notamment aux calcaires micritiques, gris, parfois violacés, de type couches à *Aptychus,* tout comme les basaltes et les calcaires massifs néojurassiques. Peu nombreuses, les olistolites ont en revanche des dimensions relativement grandes, de quelques mètres à des dizaines de mètres cubes.

La formation et respectivement l'unité de Bejan chevauchent au N l'unité de Căbeşti, le long d'une dislocation à caractère de faille inverse abrupte.

L'olistolite de calcaire néotriasique apparaît dans le versant sud de la colline de Dumbrava, à 500 m N du confluent de la vallée de Fornădia avec la vallée de Caianu.

La surface d'affleurement de l'olistolite est de 8 m^2 environ. Le calcaire est gris clair, blanchâtre, à aspect parfois vitreux.

Il contient une association de *Halobia* dont les exponants appartiennent à l'intervalle Carnien supérieur-Norien inférieur : *Halobia* ex. gr. *superba* Majsisovics, *H. rugosa* Gümbel, *H.* ex. gr. *radiata hyatti* Gruber, *H.* (?) cf. *clari* Gruber.

Par endroits, le calcaire a un aspect de type coquina, représentant des agglomérations de coquilles, dans la majorité fragmentées, ce qui a mené à des estimations vu les caractères morphologiques d'un groupement spécifique.

L'analyse en section mince a mis en évidence un calcaire micritique, parfois à radiolaires, dont le milieu de dépôt peut être interprété en tant que pélagique.

Les implications structurales-faciales de la présence du calcaire néotriasique sont encore difficiles de préciser. Initialement (Săndulescu, 1980), on a fait une corrélation entre la formation de Căbeşti — considérée à cette date-là comme porteuse d'olistolites — et le wildflysch de la nappe bucovinienne.

En faisant une analyse comparative entre le calcaire néotriasique de Bejan et celui du Trias supérieur des territoires environnants, on constate des ressemblances tout comme des différences. Ainsi, Patrulius et al. (1980) étudiant le Trias des séries transylvaines distinguent la série de Zimbru et y signalent des calcaires finement granulaires à *Halobia* porteux, comme toute la série susmentionnée, d'accidents siliceux mais qui n'apparaissent pas dans l'olistolite de Bejan.

Dans les monts Apuseni du Nord, notamment dans la nappe de Moma on a signalé (Patrulius et al., 1976) des niveaux décimétriques de calcaires micritiques de teinte rose dans le calcaire de type Tisovec. Le calcaire de Bejan est gris clair et entièrement pélagique.

En fait, l'impédimenta majeur dans la corrélation faciale est l'olistolite de calcaires qui représente un intervalle stratigraphique réduit.

C'est pour cette raison qu'il ne faut exclure ni la possibilité de l'existence d'un faciès pélagique du Néotrias, dissemblable par conséquent des faciès connus jusqu'à présent dans les territoires environnants.

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EXPLICATION DES PLANCHES

PLANCHE I

Fig. 1 — *Halobia* ex gr. *superba* Mojs., 3 ×, Bejan-Deva (monts Apuseni du Sud), Carnien supérieur-Norien.

Fig. 2 — *Halobia rugosa* Gümbel, 3 ×, Bejan-Deva (monts Apuseni du Sud), Carnien supérieur-Norien.

Fig. 3, 4 — *Halobia* ex gr. *radiata hyatti*, 3 ×, Bejan-Deva (monts Apuseni du Sud), Carnien supérieur-Norien.

PLANCHE II

Fig. 2 — *Halobia rugosa* Gümbel, 3 ×, Bejan-Deva (monts Apuseni de du Sud), Carnien supérieur-Norien.

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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

Project 195 : Ophiolites and Lithosphere of Marginal Seas

STRUCTURAL, PETROGRAPHIC AND GEOTECTONIC STUDY OF THE SHEETED DYKE COMPLEX IN THE MUREȘ ZONE, DUMBRĂVIȚA-BAIA-BĂTUȚA-JULIȚA REGION (APUSENI MOUNTAINS)[1]

BY

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Ophiolites. Dyke. Ocean-floor basalts. Gabbros. Dolerites. Petrological study ; Apuseni Mountains — Mureș Mountains — Drocea Mountains.

Sommaire

Etude structurale, pétrographique et géotectonique du complexe de sheeted dykes de la zone de Mureș, région de Dumbrăvița-Baia-Bătuța-Julița (monts Apuseni). Le complexe de sheeted dykes se développe dans la partie ouest de la zone de Mureș. Sa présence dans le cadre de cette zone est une preuve péremptoire que ses ophiolites se sont formées dans une zone de fond océanique et dans des conditions de spreading. Les sheeted dykes sont généralement orientés sur la direction N 78°—86° O et ont des pendages variant entre 70° S et 80° N. L'épaisseur des dykes oscille entre 0,5 et 2 m et plus rarement entre 5 et 10 m Ils présentens des structures différentes (de hyalopilitique à hipidiomorphegrenue). Le complexe est formé de sheeted dykes de basaltes, dolérites, gabbros, spilites, felsites albitiques, quartz-diorites pegmatoïdes, granophyres, hyalobasaltes filoniens et plagio-granites albitiques. Au complexe de sheeted dykes s'associent les corps de gabbros de Julița et Bătuța.

Introduction

The ophiolitic rocks in the Mureș Zone drew the geologists' attention since the last century because, besides their scientific significance, pyrite and chalcopyrite mineralizations are associated, in places, with them. In 1952 the discovery that some gabbro bodies in the Mureș Zone contain vanadiferous titanomagnetite stimulated the study of the ophiolites. The researches carried out in the period 1981—1982 had as

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an objective the sheeted dyke complex of the ophiolitic series in the Dumbrăvița-Baia-Bătuța-Julița region (Pl. I), a formation first established as a separate unit in Romania by Savu in 1981 (Savu, 1983) and described in detail in the report on 1982 (Savu et al., 1982 c). The present paper is based on the first part of the mentioned report.

History of Researches

Disparate data on the "melaphyric" or "diabasic" rocks occurring in the region are known before the World War One. It seems that till the World War Two the region was not properly investigated. A map, scale 1 : 100 000, published by Socolescu (1941), exceeds a little westwards the meridian of the locality of Julița, where the mentioned author plotted Triassic diabases. In 1953, Papiu described the Mesozoic sedimentary rocks and the "basic effusive series" constituted of submarine lava flows intruded by gabbros and diorites. The map and a summary of this paper were published only in 1959. In the same period, Lațiu (1957) described the processes of albitization and epidotization of the basic rocks.

In 1953 Savu began a systematic study of the ophiolitic rocks in this part of the Drocea Mts which he continued later on (Savu, 1957, unpublished data ; Savu et al., 1979). On these occasions he established the vein or dyke character of some rocks occurring in this region (Savu, 1953, unpublished data), pointing out the presence of parallel dykes (Savu et al., 1979) or sheeted-dykes (Savu et al., 1982 b).

Geological Setting of the Region

The region lies in the west-northwestern part of the Mureș Zone, covering the axial, ophiolitic part of this zone. Two rock series can be distinguished : the ophiolitic rock series, formed on the floor of the Mureș Ocean during the Jurassic till the end of the Callovian and the flysch deposit series, formed during the Upper Jurassic and Lower Cretaceous in the northern marginal trough of the ocean zone, as a result of the bilateral subduction processes which began concomitantly with the Late Kimmerian movements (Savu, 1962 b ; Savu, Udrescu, 1973 ; Savu, 1983).

1. The flysch deposits series is represented here by its upper part, of Lower Cretaceous age, constituted of alternations of beds made up of marls, clayey marls with calcite veins (Papiu, 1953 ; Savu et al., 1979).

2. The rocks of the ophiolitic series form two magmatic complexes: (a) the ocean-floor basalt complex and the (b) sheeted dyke complex (Savu, 1983).

As shown on the annexed map (Pl. I) the Jurassic ophiolite rocks thrust the Lower Cretaceous flysch formations northwards, as pointed out by Papiu (1953) and established by Savu (1957 ; 1983 ; Savu et al., 1979). On the thrust plane are involved jaspers and red argillites, which formed during the Upper Jurassic, constituting a red complex of characteristic rocks situated in the base of the flysch which at present occurs under the thrust ophiolites, somewhere in depth, from where jaspers and argillites have been torn off. The red complex reappears to the southwest (Laleşinţ) and northeast (Lupeşti-Buceava) of the region of study. This thrust began to form probably during the Austrian movements and was finished off during the Laramian movements.

There are dip faults involved both in the thrust plane of the ophiolites over the Lower Cretaceous flysch and in the ophiolitic series

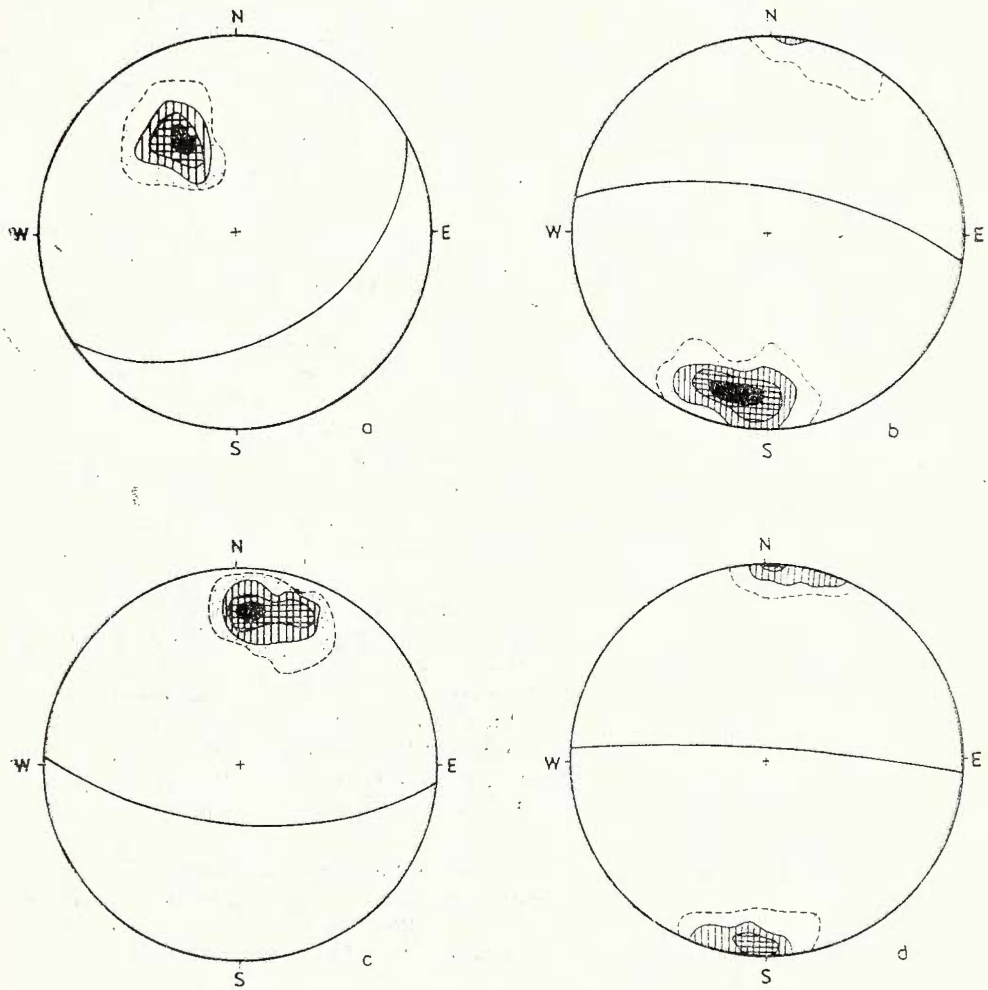

Fig. 1 — Structural diagrams.
a, position of the jasper and red clay intercalations in the Valea Calului Brook basalts (17 planes) — isolines : 0.5-2-4-6%; b, position of the dykes on the Valea Mare Brook (N. Bălcescu) and Vălcuţa Valley (70 planes) — isolines : 0.5-4-10-20%; c, position of the dykes on the Frunţii Brook and Bătuţa Quarry (20 planes) — isolines : 0.5-2-4-5%; d, position of the dykes on the Baba Valley and Juliţa Quarry (23 planes) — isolines : 0.5-2-8%.

(Pl. I). The fractures strike NW-SE and belong to a fault system typical of this SW part of the Apuseni Mountains.

Ocean Floor Basalt Complex

In normal position, the ocean-floor basalt complex is situated at the upper part of the ophiolitic series and represents the upper complex (0_1) (Savu, 1983). In the region under discussion this complex is tectonically included between the sheeted dyke complex and the overlapped Lower Cretaceous flysch. For this reason the rocks of this complex form a basaltic strip along the thrust plane between Baia and Mureș, extending NE and SW.

This complex is mainly constituted of basaltic rocks often in pillow-lava facies. On the summit between the Huleiul and Calul valleys, narrow bands of red argillites and jaspers with veinlets of silica and zeolites are intercalated between the basaltic flows. They trend N 62° E/45° S, approximately conformably with the thrust plane (Pl. I), as shown on the diagram in Figure 1 a. As the basaltic rocks of this complex underwent strong tectonic efforts, being situated in the base of the nappe and thrusting over the "autochthon", they are usually tectonized, mylonitized and brecciated. Veinlets of silica, calcite and zeolite deposited on the fissures formed between them so that at present the pillow-lava structures are hardly recognized.

The basaltic complex (O_1) consists of several varieties of basalts : hyalobasalts, variolites, spilites, and basalts.

1. The crust of the pillow-lava separations, about 2-3 cm thick and affected by radial fissures as against the pillow shape, is represented by hyalobasalt. It consists of a groundmass of devitrified glass, pigmented with a fine, opaque magnetite powder, including amygdaloid-like separations made up also of limpid devitrified glass and rare microphenocrysts of plagioclase and very rarely of augite. In other cases the vitreous groundmass contains few plagioclase microlites and microphenocrysts, the latter displaying acicular terminations. The microphenocrysts can be albitic, the rocks corresponding to spilites as regards their origin. There are hyalobasalts in which the plagioclase laths are grouped, constituting a primary form of variolitic or intersertal texture.

2. The interior of the pillow-lava separations and some lava flows are constituted of intersertal basalts consisting in a network of albitized or sericitized plagioclase laths, in the meshes of which there is devitrified or chloritized glass impregnated with a fine powder of iron oxides and small augite crystals.

Sheeted Dyke Complex

Most of the region is constituted of a complex of parallel dykes formed on the Mureș Ocean floor (Savu, 1983). This complex is represented by a succession of dykes, whose position and structure can be observed clearly in the quarries of Bătuța, Julița and on Valea Mare Brook (N. Bălcescu) where a forest road have recently been built, as well as in the rare outcrops on the Baba and Vălcuța valleys and Fruntea Brook (Pls. I and II, Figs. 1, 2).

The sheeted dykes generally strike N 78° to 86°W and their dippings vary between 70°S and 80°N (Fig. 1 b, c, d). The WNW-ESE strike of the sheeted dyke complex, slightly differ from the present trending of the Mureș Zone (in this region approximately ENE-WSW or E-W, like that of the Julița gabbroic body) might be a characteristic only of the tectonic block located between the Căpruța-Bacău Fault (Savu, 1983) and another fault situated eastwards. This unconformity is also pointed out by the position of the maximums on the diagrams in Figure 1, out of which the first one indicates the trending of the

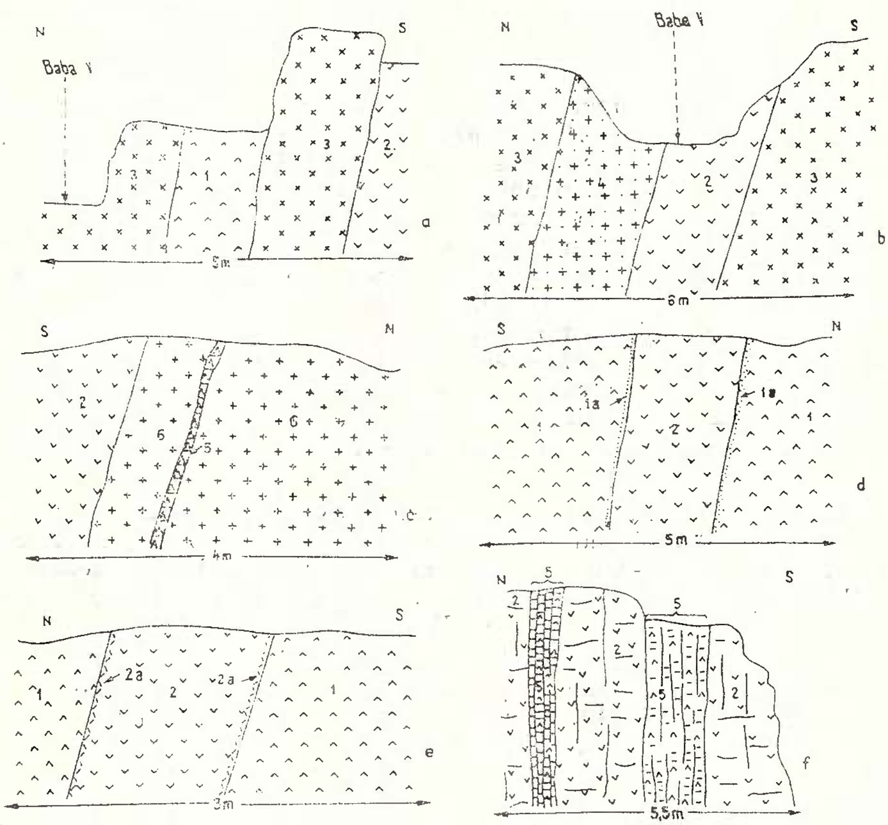

Fig. 2 — Structure of the sheeted dyke complex.

a, b, dykes on the Baba Valley ; c, hyalobasalt (5) intruding a gabbro dyke (6) on the Valea Mare Brook (N. Bălcescu) ; d, chilled zone (1 a) of two basalt dykes (1) at the contact with a dolerite dyke (2) on the Valea Mare Brook (N. Bălcescu); e, marginal brecciation zone (2 a) of a dolerite dyke (2) at the contact with two basalt dykes (1) on the Valea Mare Brook (N. Bălcescu) ; f, two hyalobasalt dykes (5) alternating with dolerite dykes (2) displaying several fissure systems. 1, basalt ; 2, dolerite ; 3, granophyre ; 4, gabbroporphyrite ; 5, hyalobasalt ; 6, gabbro.

northwestern margin of the ophiolitic plate thrusting over the flysch formation and the other three — the trending of the sheeted dyke complex in this region.

The dykes intrude, on the same direction, the Julița gabbroic body (Savu et al., 1982 b) which should have come also from a more important dyke, like the Dumbrăvița gabbroic bodies. The magma intruded on the separation surface between the ocean-floor basalt complex and the sheeted dyke complex and then it became differentiated in situ giving rise to the present stratified, almost tabular body (Savu et al., 1982 b).

The thickness of the sheeted dykes varies from 0.5 to 2 m, more rarely 5 to 10 m, as in the case of the gabbro and granophyre dykes. They consist of varied rocks, among which the intergranular basalts and dolerites are predominant (Fig. 2 a, b and Pl. II, Figs. 1, 2).

Gabbro dykes, among which gabbros with vanadiferous titanomagnetite (Savu, 1953 ; 1972 ; 1973) predominate in the basin of the Calul Valley and its tributaries — Grunețul Mare, Grunețul Mic and Pîrîul lui Coti — and granophyre dykes intruded by parallel dykes of dolerites, basalts, gabbroporphyrites and hyalobasalts are characteristic in the Baba Valley basin. The same rock types occur also in the parallel dykes in the upper basin of the Valea Mare Brook (N. Bălcescu) and Vălcuța Valley (Pl. I). In these zones there also occur irregular veins of aplitic albite and plagiogranites and albite quartz-diorites, with a small thickness. They are the last differentiated products as they penetrate the basic rocks on longitudinal protofractures, along which the latter are brecciated. Fragments of these breccias are included in the vein mass and altered into basic hornfelses.

Granophyres and other more basic rocks are penetrated by dykes of devitrified basalts and hyalobasalts, the latter showing thicknesses ranging from 1 to 4-5 cm. These veins or small dykes (dykelets) are parallel to the sheeted dyke complex, as one can see on Valea Mare Brook (N. Bălcescu), Vălcuța and Baba valleys (Fig. 2 c).

As regards the structure of the parallel dykes of basic rocks it is to be mentioned that they display a narrow chilled zone on one margin (Moors, Vine, 1971) or on both margins, and in the interior they consist of basaltic and doleritic rocks with a common structure (Fig. 2 d). In the chilled zone the rock might appear as an aphanitic basalt, tending to hyalobasalts, with phenocrysts of albitized plagioclase included in a hyalopilitic groundmass formed of chloritized glass and microlites of albitic plagioclase, disposed divergently or even parallel to the dyke walls. It is to be mentioned that unlike the thicker basaltic dykes which have one or two chilled zones, represented by glassy basalt, only on margins, the hyalobasalts dykelets — being thin (0.5 to 5 cm thick) — consolidated entirely as glassy rock whose groundmass includes only rare plagioclase microcrystals or microphenocrysts. However, there are cases when the margins of some basaltic dykes are slightly brecciated due to the friction on the spreading plane before the injection of a new dyke or before the intrusion of the latter when the magma was already viscous, almost consolidated (Fig. 2 e).

On the margins of the hyalobasaltic thin dykes one can observe flow planar structures (Balk, 1937), within which the fine grains of iron oxides are concentrated in very thin bands which under the microscope appear less transparent than other bands with which they alternate and which are mostly made up of devitrified glass. Inside these dykes the plagioclase microlites and microphenocrysts are almost parallel to their walls. The contacts between the hyalobasaltic dykes and those of basalts or dolerites are usually plane, but there are cases when the two rocks seem to interpenetrate, indicating that hyalobasalt dykes intruded when the preceeding dyke was not entirely consolidated.

On margins, dykes can be albitized due to the juvenile solutions rich in H_2O and Na_2O or to the sea water percolating in depth on the spreading plane which was being opened until a new lava (dyke) filled the space, heated and brackish water which became vapours and attacked the rocks from the furrow walls in the spreading zone or even the dyke mass which was being consolidated.

There are also spilitic dykes coming from magmas formed during the differentiation of the tholeiitic magma, finally giving rise to albitic granophyres and plagiogranites. Parallel, thin small veins or dykes, consisting of albitic felsites are very rarely found.

All dykes, especially the doleritic and basaltic ones, display a system of fissures parallel to their walls and two systems perpendicular to them and also to one another. The first system of fissures is always obvious in case of the hyalobasaltic veinlets or dykelets (Fig. 2 f).

The radiogene age of the sheeted dyke complex is of 180 m.y. (Herz et al., 1974).

Petrography of the Sheeted Dyke Complex

The sheeted dyke complex consists of basalts, dolerites, gabbros, spilites, albitic felsites, pegmatoid quartzdiorites, granophyres, vein hyalobasalts and aplitic albite plagiogranites. It is worth mentioning that the rocks of this complex usually underwent phenomena of ocean floor metamorphism. The plagioclase is frequently saussuritized or albitized, and the pyroxene underwent phenomena of uralitization, epidotization — sometimes actinolite being formed — and chloritization, when grains of iron and titanium oxides are separated.

1. Basalts are usually represented by rocks with a granulation less than 1 mm and an intergranular texture. They consist of plagioclase laths (An_{55}) (47-48%), twinned after albite law. Augite crystals, isometric and often idiomorphous, partly altered into clinochlore or pennine are to be found in the spaces of the plagioclase network. A small percentage (6%) of the rock mass is represented by very fine grains of magnetite and pyrite.

Sometimes basalts contain globular separations, formed of glass and plagioclase microlites described previously by the author of this paper at Lalesinţ (Savu, Manea, 1982) and Visca (Savu et al., 1985, in press) ; they have for basalts the same significance as chondrites for meteorites.

2. Dolerites are rocks with ophitic or ophitoporphyritic texture and consist of plagioclase laths (An_{55}), usually more than 1 mm long, which form a network, over which xenomorphous or poikilitic crystals of clinopyroxene (augite) or magnetite develop. Porphyritic dolerites including plagioclase phenocrysts are rarely found. Quartz dolerites are characterized by small amounts of quartz.

3. Gabbros and gabbrodolerites form parallel dykes and small bodies well represented in the Valea Calului Brook basin (Dumbrăvița), also found in the whole region. At Julița and Bătuța they constitute two bodies of greater significance, within which gabbros occur in association with olivine gabbros and hyperites. They are black coloured and display a hipidiomorphous-grainy (gabbros) or ophitic (gabbrodolerites) texture. They are characterized by the predominance of plagioclase as against the melanocrate minerals (Tab.) as well as by its idiomorphous character, as it is always the first one which crystallizes from the tholeiitic magma.

3.1. Gabbrodiorites consist of a network of plagioclase crystals (An_{50}), more then 3 mm long, idiomorphous and divergently arranged, in places affected by saussuritization phenomena. The melanocrate mineral is a clinopyroxene, on margins altered into uralite and chlorite. By their granulation gabbrodolerites make the transition between gabbros and dolerites.

3.2 Gabbros are phaneritic rocks with a hipidiomorphous-grainy texture, consisting of plagioclase, clinopyroxene, magnetite and secondary minerals (Tab.). Plagioclase (An_{58}) occurs as tabular crystals, polysynthetically twinned and in places saussuritized. In certain rocks plagioclase exceeds the amount of melanocrate minerals, sometimes reaching 67% of the volume and for this reason the rocks resemble

TABLE

Mineralogical composition of gabbros and gabbrodolerites

Minerals/rocks	Gabbro-dolerite	Gabbro-dolerite	Gabbro	Magnetite gabbro	Gabbro	Gabbro
Plagioclase	50.00	47.50	53.90	43.00	63.00	48.00
Diopside	22.93	20.00	5.50	10.00	2.00	9.95
Uralite	21.00	21.00	39.00	24.00	29.00	42.00
Chlorite	—	—	1.00	—	0.50	—
Quartz	—	5.00	—	—	—	—
Magnetite	0.67	6.50	0.60	23.00	5.50	0.05
Total	100.00	100.00	100.00	100.00	100.00	100.00

the gabbroanorthosites. Clinopyroxene ($c \wedge Ng = 44°$), displaying hematite separation after (100) face, is xenomorphous as it always crystallizes after plagioclase. Its secondary alterations lead to the successive formation of the following minerals : green hornblende-uralite-acicular actinolite-chlorite iron oxides and leucoxene being separated. Secondary hornblende has the following pleochroism : Ng=green-bluish ; Nm=green-brown ; Np=yellowish-slightly greenish. Sometimes in the clinopyroxene from gabbros one can observe small separations of brown hornblende. Magnetite is also xenomorphous, being the last magmatic mineral crystallized.

3.3. Gabbros with vanadiferous titanomagnetite (ferrogabbros) are typical especially of the Grunețul Mare Valley ; however, they also occur in other dykes as well as in the Julița gabbroic body (Savu et al., 1982 b). The rocks are formed of tabular crystals of plagioclase (An_{55}), corroded on margins during the crystallization of the melanocrate minerals, and show a more acid, narrow marginal zone. Diopside and magnetite are xenomorphous, being formed after the plagioclase crystals. The gabbros with vanadiferous titanomagnetite (ferrogabbros) include the following succession of Fe, Ti and V. minerals (Savu, 1973) : vanadiferous titanomagnetite ($\pm$ ulvöspinel), ilmenite, rutile, hematite and sphene. In some gabbros magnetite represents 23% (Tab.) or even 25% of the volume (Savu, 1972). Sometimes gabbros also comprise a small amount of quartz (Tab.).

As previously mentioned (Savu et al., 1982 b) the rocks of the lower horizon of the Julița gabbroic body often display cummulate structures, e.g. the poikilitic development of the clinopyroxene and vanadiferous titanomagnetite crystals, minerals which crystallized later from the intercumulus liquid (Wager, Brown, 1968) and included plagioclase crystals previously "sedimented" on the bottom of the intrusion.

4. In some dykes gabbros resemble pegmatoid rocks and for this reason their granulation is coarser, the mineralogical components often exceeding 1 cm in length. Sometimes a primary green hornblende can occur beside clinopyroxene, quartz and saussuritized plagioclase.

5. Spilites form dykes which alternate with those of basalts and dolerites, rocks similar as regards the texture, but different because the plagioclase laths consist of albite (An_{8-10}) and the melanocrate minerals are replaced by chlorite. They are usually richer in plagioclase than the basalt-dolerite rocks and sometimes they contain xenomorphous quartz.

6. Albite felsites are greenish aphanitic rocks, holocrystalline, which consist of elongated albite crystals arranged as garlands, more seldom as rosettes and include very elongated and thin titanaugite crystals ($c \wedge Ng = 47°$) ; some of the augite crystals are divergently arranged and others as garlands, like the albite laths (Pl. III, Fig. 3). The larger magnetite crystals, some of them skeleton-like, are very seldom found. In some cases the clinopyroxene laths have been replaced by chlorite lamellae and consequently the whole rock is impregnated by sticks and fine grains of iron and titanium oxides.

7. Pegmatoid quartz diorites are rocks with a divergent texture, with the following composition : plagioclase (43.20%), clinopyroxene (2.80%), secondary hornblende (40%), quartz (6%), magnetite (8%). Plagioclase occurs as idiomorphous, elongated crystals, displaying twins after albite and albite-Karlsbad law. In some rocks they consist of labradorite (An_{52}) and in others of andesine (An_{32}) or are completely albitized. Crystals of clinopyroxene and of secondary hornblende and

grains of xenomorphous quartz are found among the plagioclase crystals. These rocks make the transition towards granophyres.

8. Granophyres form dykes of rocks with a divergent texture (Pl. III, Fig. 1), rich in albite plagioclase (An_5), including albite (45%), fine-twinned albite (5%), quartz (13%), green hornblende (24%), magnetite (4%), accessory minerals (epidote, apatite, ilmenite, magnetite) (9%). Plagioclase crystals are strongly elongated (1.5/0.5 cm) and display twins after albite law. Their outlines are irregular as albite is intergrown on margins with quartz forming myrmekitic or micrographic textures developed in the interstices of the albite crystals. Sometimes, these interstices contain a fine-twinned albite of low-temperature (schachbrettalbite) which might indicate the pre-existence of a potash feldspar. The melanocrate mineral is a green amphibole, occurring as very elongated and thin crystals, often replaced by actinote, epidote and chlorite. This amphibole displays the following optic properties: Ng=green-yellowish; Nm=greenish-brown; Np=yellowish; $c \wedge Ng = 15\text{-}17°$. Apatite crystals, thin and much elongated, with hexagonal section, occur as well. They have previously been described as micropegmatites (Savu, 1962 a).

9. Vein hyalobasalts (hyaloferrobasalt) are black aphanitic rocks, formed of a devitrified glass mass, containing numerous fine magnetite grains and rare microcrystals or microphenocrysts of plagioclase (An_{45-50}). Some of the latter are thin and elongated (0.5-1.5 mm) and twinned after the albite or albite-Karlsbad law. Clinopyroxene microphenocrysts occur very seldom. When the veins are thin (1 cm) or on the margins of the thickest ones, the plagioclase microcrystals or microphenocrysts tend to form a fluidal texture (Pl. III, Fig. 2).

10. Aplite-albite plagiogranites represent leucocrate rocks formed of quartz (30%), albite (46-50%), amphibole (5%), fine-twinned albite (6%), accessory and secondary minerals (12%). Quartz grains have a slightly irregular extinction. Plagioclase (An_{5-8}) forms dim hipidiomorphous crystals, twinned after albite, albite-Karlsbad and seldom pericline law. Some albite crystals form myrmekitic intergrowths with quartz as in granophyres. The interstices of the plagioclase crystals include, in places, fine-twinned albite, of low temperature, which seems to have replaced a potash feldspar. Melanocrate mineral is an amphibole, frequently replaced by pistacite and chlorite with the following optic properties: Ng=green-brownish; Nm= brown-greenish; Np= =yellowish; $c \wedge Ng = 22°$. Magnetite constitutes fine grains.

On the Vălcuța Valley these rocks pass to a more melanocrate facies enriched in amphibole in the quartz detriment. They correspond to albite-bearing quartz diorite.

Both granites and diorites include xenoliths of basic rocks (Pl. IV, Figs. 1, 2) metamorphosed under conditions of amphibole — or pyroxene hornfelses (Turner-Verhoogen, 1960). Two mineral parageneses can be distinguished within them:

1. Plagioclase (An_8) — green hornblende — magnetite
2. Plagioclase — diopside — magnetite.

The first paragenesis occurs more frequently, the second one is found only in xenoliths of the diorite rocks whose amphibole hornfels mass contains small portions constituted of pyroxene hornfelses. In hornfelses, plagioclase is usually saussuritized and packed with zoisite and kaolinite. Amphibole is found in allotriomorphous or hipidio-

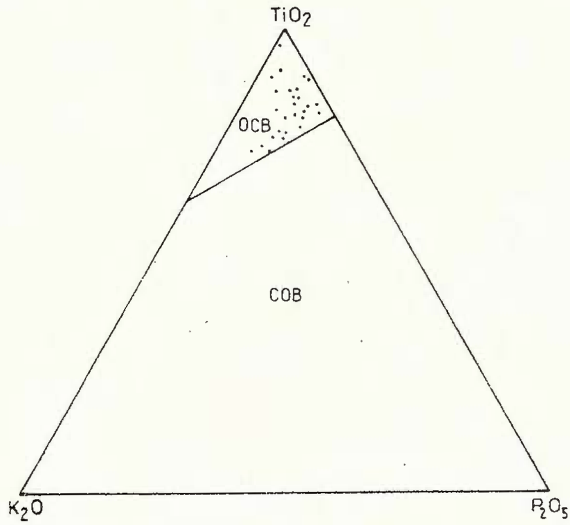

Fig. 3 — TiO_2-P_2O_5-K_2O diagram (acc. to Pearce et al., 1975)

morphous crystals displaying the following optic properties: Ng=green-brownish ; Nm=brown-greenish; Np=yellowish; c $\wedge$ Ng=14°. Magnetite forms very fine and panallotriomorphous grains.

Geotectonic Conditions of the Sheeted Dyke Complex Formation

In the Mureş Zone the sheeted dyke complex (O_2) formed during the spreading of the Mureş Ocean (Savu, 1983). Its presence in this zone proves that the ophiolites of the southern Apuseni Mountains formed under ocean-floor conditions. This conclusion is clearly inferred also from the diagram in Figure 3 (Savu et al., 1982 a) on which the rocks of the mentioned complex plot.

The relationships among the dykes which follow one after another clearly point out their formation under conditions of the spreading process as also indicated by the distribution of the values of the rock differentiation index on the map of the zone northwest of Bata (Savu, Manea, 1982 a).

In a previous paper (Savu, 1983) was pointed out that this complex might be spread along the whole Mureş Zone at least as much as the overlying ocean-floor basaltic complex (O_1), its lava flows being supplies by dykes whose magma rises from depth. This inferrence is based on the fact that within the basaltic complex (O_1) sills and dykes of dolerites and spilites, as well as bodies or dykes of gabbros also supplied by such dykes are frequently found. The gabbroic bodies trend NE-SW, like the dykes, which represents a further argument in favour of the presence of the sheeted dyke complex (O_2) under the basaltic complex (O_1).

One can observe at the surface that nowadays the Mureş Zone is the result of the unrest geological history of this zone. For this reason two elements of great significance related to the primary structure of the Mureş Zone are still obscure.

1. The first question refers to the possible existence of a third and a fourth complex — the gabbro (O_3) and peridotitic (O_4) complexes — overlain by the sheeted dyke complex, known in the classical ophiolitic zones in the world. The existence of numerous gabbroic bodies and of some peridotite bodies encompassed in the two known complexes makes us suppose the existence of a gabbro and a peridotite complex at the depth of 1 to 2 km.

2. The second question refers to the actual place of the axial zone (median ridge) of the Mureş Ocean, which ceased to function at the beginning of the closing phase of this ocean.

The excentric position of the sheeted dyke complex as against the actual axial part of the Mureş Zone (Savu, 1983) might rather indicate that the ophiolites had been affected by folding phenomena like the sedimentary formations in the marginal basins. In the present case one may suppose that this complex rose from depth along the northwestern margin of the Mureş Zone during the thrusting of the ophiolites over the Upper Jurassic-Lower Cretaceous flysch formations.

As already mentioned, in the region under discussion, the sheeted dykes trend WNW-ESE, a direction which makes an acute angle with the direction of the elongation of the Mureş Zone. It was supposed that this orientation might be characteristic only of the block lying between two dip faults, striking NW-SE, one located on the Mureş Valley between Căpruţa and Bacău and another one eastwards, indicating that this block also underwent a short clockwise rotation.

It is possible that the unconformity between the trending of the sheeted dyke complex and the actual northwestern margin of the Mureş Zone might have been determined by the asymmetric break of the oceanic crust on the two margins of the zone, according to the model previously elaborated (Savu, 1983). The external segments of the oceanic crust being subducted under the axial oceanic plate, beside a segment of the sialic part of the plates, the latter are at present in contact with different terms of the axial oceanic crust. Therefore, if the actual outcropping area of the sheeted dyke complex, lying between Troaş and Bata (35/10 km), represented the "median" ridge of the Mureş Ocean, one of the following two possibilities should be accepted : either that the Mureş Ocean spreading took place asymmetrically, SSE respectively, or that the break of the oceanic crust was entirely asymmetrical, its northwestern side being mostly involved in the subduction process and consumed. In the second alternative it is possible that the breccious aspects of the ophiolitic rocks on the northwestern border of the Mureş Zone and especially of those belonging to the basaltic complex (O_1) might be determined not only by the tectonic efforts during the thrusting of the ophiolites over the flysch formation but also by those which led to the breaking of the oceanic crust along

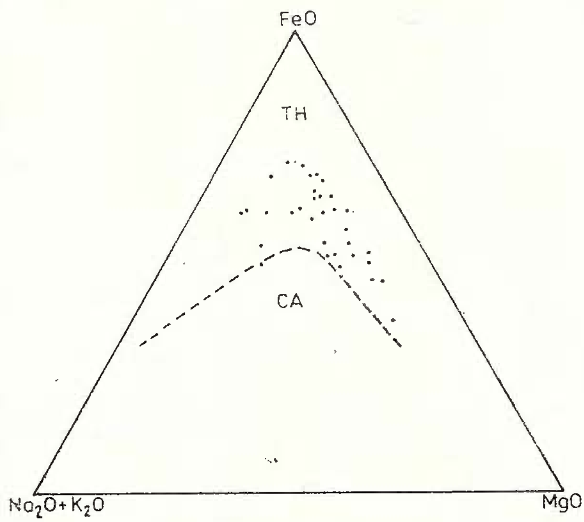

Fig. 4 — FeO-MgO-Na_2O+K_2O diagram (acc. to Irvine and Baragar, 1971).

this margin. The researches carried out northeastwards in the Lupeşti-Troaş region, and southwestwards in the Bata region might clear up this problem.

The sheeted dyke complex originated in a tholeiitic magma (Fig. 4, Savu et al., 1982 a) formed in the upper mantle. This magma underwent intense differentiation processes, from the primary magma towards magmas enriched in iron which gave rise to the vanadiferous titanomagnetite gabbros (ferrogabbros) and magmas rich in silica, natrium and volatiles, which gave rise to albitic granophyres and plagiogranites. This trend of differentiation of the tholeiitic magmas which gave rise to sheeted dyke complexes is the normal one, being mentioned by other researchers, as well (Thayer, 1978 ; Coleman, 1977 a, b).

Conclusions

Several general conclusions can be inferred from this paper :

The occurrence of the sheeted dyke complex in the Mureş Zone represents a peremptory proof that its ophiolites originate in an ocean-floor zone.

In this area the sheeted dykes trend generally N78° to 86°W and dip 70°S to 80°N.

The thickness of the dykes varies between 0.5 and 2 m more seldom 5 to 10 m ; the rock textures range from hyalopilitic to hipidiomorphous-grainy.

The complex consists of parallel dykes of basalt, dolerite, gabbros, spilites, albitic felsites, pegmatoid quartz-diorites, granophyres, vein (dykelet) hyalobasalts, albitic plagiogranites.

With the sheeted dyke complex are related two more important gabbroic bodies, e.g. those at Julița and Bătuța.

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EXPLANATION OF PLATES

Plate II

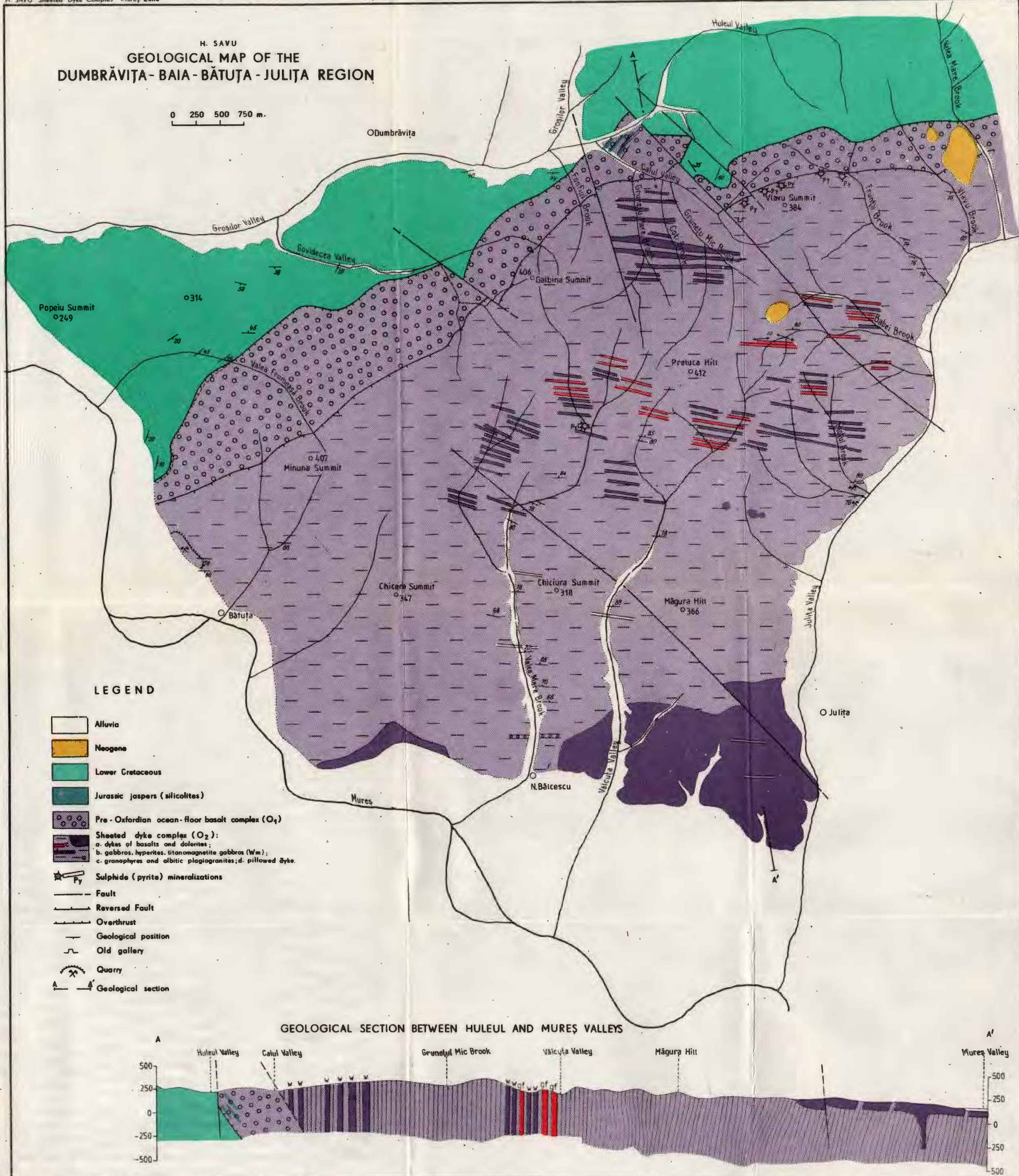
Fig. 1 — Sheeted dykes in the Bătuţa Quarry (eastern part).
Fig. 2 — The same sheeted dykes.

Plate III

Fig. 1 — Divergent structure of a granophyre on the Baba Valley, at the contact with a hyalobasalt dykelet (2.5 cm thick).
Fig. 2 — Vein hyalobasalt (hyaloferrobasalt) on the Vlavu Valley. × 35 ; N ||.
Fig 3 — Albite felsite on Dealul Rîpii Hill. × 35, N ||.

Plate IV

Fig. 1 — Albite-quartzdiorite with xenoliths of thermally metamorphosed basic rocks on the Vălcuţa Valley.
Fig. 2 — Albite plagiogranite with xenoliths of thermally metamorphosed basic rocks on Valea Mare Brook (N. Bălcescu).



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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

IUGS UNESCO IGCP

Project 195 : Ophiolites and Lithosphere of Marginal Seas

TECTONIC POSITION AND ORIGIN OF ALPINE OPHIOLITES IN THE MEHEDINȚI PLATEAU (SOUTH CARPATHIANS) WITH SPECIAL REGARD TO THOSE IN THE PODENI-ISVERNA-NADANOVA REGION [1]

BY

HARALAMBIE SAVU [2]

Ultramafic rocks. Ophiolites. Mesozoic. Plate tectonics. Alpine tectonics. Ocean floor spreading. Subduction ; South Carpathians — Crystalline Getic Domain — Mehedinți Plateau.

Sommaire

Position tectonique et origine des ophiolites alpines du plateau de Mehedinți (Carpathes Méridionales) avec considération spéciale sur celles de la région de Podeni-Isverna-Nadanova. L'évolution géologique alpine du plateau de Mehedinți, respectivement de la chaîne des Carpathes Méridionales, s'est déroulée en deux étapes principales : (1) étape jurassique-crétacé inférieure et (2) étape crétacé supérieure. Pendant la première étape, prend naissance une zone océanique entre la microplaque transylvaine et la microplaque moesienne ; c'est ici qu'au cours de l'expansion du fond océanique se forme une croûte océanique typique (J_1-J_2). Pendant le Jurassique moyen et supérieur a lieu la fermeture de la zone océanique par un processus du subduction ; par la suite s'est formée une olistostrome jurassique-néocomienne (J_2-Cr_1) à olistolites de roches basiques et ultrabasiques, à éléments de schistes cristallins, gabbros, basaltes etc. Ces olistolites représentent les restes d'obduction de la croûte océanique jurassique démembrée, qui a été soumise au processus de subduction au-dessous des Carpathes, où elle a été consommée pour la plupart. L'olistostrome, y compris les olistolites, est affectée d'un processus d'anchimétamorphisme probablement néokimmérien. Durant la deuxième étape se produisent de nouveaux processus de subduction et des mouvements orogéniques. C'est maintenant que s'est formé

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le wildflysch crétacé supérieur à olistolites des formations mésozoïques antérieures et du soubassement précambrien de l'autochtone danubien. Sont mises en place les nappes de Cerna et Severin, probablement comme nappes de décollement (olistonappe), suivies par le charriage de la nappe gétique. Il s'ensuit que la croûte océanique prend naissance pendant le stade d'expansion de l'océan (Océan carpathique), tandis que le mélange ophiolitique se forme durant le stade de fermeture de celui-ci. Le mélange ophiolitique se forme dans la zone de friction d'entre la plaque continentale et la croûte océanique qui était entraînée en subduction, zone dont sont arrachées et transportées par obduction des blocs de mélange ophiolitique, roches basiques et ultrabasiques, schistes cristallins et d'autres, qui constituent des olistolites dans l'olistostrome jurassique-néocomienne (J_2-Cr_1).

Introduction

This paper presents the results of the researches carried out on the ophiolites of the Mehedinţi Plateau, represented by ultramafic and basic rocks. These results were first presented in the geological report of Savu et al. (1982) and this study is based on the first part of the mentioned report.

The first general researches on the geology of the Mehedinţi Plateau are due to Drăghiceanu (1885), Ştefănescu (1888), Mrazec (1895), and Murgoci (1905). They were successful as in that period was established the nappe structure of the South Carpathians (Murgoci, 1905). The studies carried out in the period between the two World Wars by Streckeisen (1934) and Codarcea (1940) brought new data for the outlining of the South Carpathians tectonics. Codarcea (1940) gave a clear image on the geology and tectonics of the Mehedinţi Plateau. He pointed out that between the Getic Nappe and the sedimentary formations of the Danubian Autochthon there is a parautochthon, represented by the Severin Nappe, consisting of Upper Jurassic and Lower Cretaceous sedimentary deposits intruded by serpentinite-ophiolite bodies.

After the World War Two mention should be made of the researches accomplished by Drăghici and Drăghici (1956), Mercus (1959), Trifulescu and Mureşan (1962), Andrei et al. (1971) and Matsch et al. (1972) in the Mehedinţi Plateau. In 1962, Drăghici brought new data on the stratigraphy of the region and pointed out the important fractures which cross the Bahna Syncline.

Mărunţiu et al. (1978) stated that the serpentinite bodies which would form an ophiolitic mélange are, however, bodies intruded at the tectonic level of the sedimentary rocks through a process of diapirism (Mărunţiu, 1983).

Recently, Stănoiu (1982) has considered that the sedimentary formations encompassing the ultramafic rocks of the Mehedinţi Plateau, Isverna-Brebina zone, would not belong to the Upper Jurassic and Lower Cretaceous as indicated by Codarcea (1940) and Codarcea et al. (1961) on the published maps, but to the Upper Cretaceous.

Nevertheless, Stănoiu (1973) reported a specimen of *Calpionella* from the above-mentioned rocks.

The ophiolitic rocks of the Mehedinţi Plateau were also presented in the synthesis papers of Savu (1967, 1968, 1980) as well as in two guidebooks on ophiolitic rocks (Savu et al., 1978 ; Cioflica et al., 1981) elaborated on the occasion of international meetings held in Romania.

Remarks on the Geology of the Region

The general geological maps of the Mehedinţi Plateau and its surroundings (Baia de Aramă Map, scale 1 : 200 000 ; Obîrşia Cloşani, Tismana and Nadanova maps, scale 1 : 50 000) indicate that the sedimentary formations of the Danubian Autochthon, overlain by the Severin Parautochthon, transgressively overlap the basement of Precambrian crystalline schists, which would correspond to the Lainici-Păiuş Series, possibly also the Drăgşan Series, as well as the associated granitoids.

Sedimentary Deposits of the Danubian Autochthon. The Precambrian and Paleozoic basement is overlain by Lower Jurassic deposits made up of conglomerates, feldspathic and quartzitic sandstones with local intercalations of siltitic argillites, in places pyrophyllitic. There follow Upper and Middle Jurassic, Neocomian and Barremian-Aptian limestones, which gave rise to karstic phenomena in numerous places in this region (Isverna, Balta). These formations are successively overlain by the Nadanova Formation of Cenomanian-Turonian age (Pop. 1973) with a preflysch aspect. There follow the Upper Cretaceous deposits in wildflysch facies (Drăghici, 1962), with an olistostrome character (Stănoiu, 1982) which we shall denominate Upper Cretaceous olistostrome. It includes olistoliths of limestones, sandstones and argillites from the Jurassic-Neocomian basement, as in the northern part of the region (Pl. I). However, olistoliths of crystalline schists and small olistoliths of rocks torn off from the Severin Nappe (Stănoiu, 1982), as well as basic rocks (Savu, 1980) — which cannot always be mapped because of their, small sizes — are observed.

Formations of the Severin Nappe. The Severin Nappe, separated as such by Codarcea (1940), was contested in this region by Stănoiu (1982), its formations being entirely assigned to the Upper Cretaceous wildflysch, within which three horizons are distinguished : the lower horizon, represented by the base of the wildflysch mentioned above, overlain by the second horizon, consisting of argillites with olistoliths of basic and ultramafic rocks, which would correspond to the Upper Jurassic Azuga Beds occurring in the base of the Severin Nappe sensu Codarcea (1940), and finally the third horizon, developed west of the Bahna Outlier, between Obîrşia Cloşani and Podeni, which would correspond to the Lower Cretaceous formations of the Sinaia and, possibly, Comarnic beds on Codarcea's (1940) scheme.

Our field and laboratory researches pointed out that the so-called middle horizon (Stănoiu, 1982), the Azuga Beds, respectively, in the base of the Severin Nappe (Codarcea, 1940), consists of an anchimetamorphic pelitic series, represented by phyllitic rocks, as emphasized by Codarcea, too, since 1940. The anchimetamorphic character of the Azuga Beds is obviously in contrast with that of sedimentary rocks of the Upper Cretaceous wildflysch formations tectonically overlain by them. Moreover, along the Coşuştea Valley and especially west of Isverna, a strong shearing plane can be observed clearly in the base of the Azuga Beds, also pointed out by Drăghici in 1962. Considering all this, as well as the fossil remnants reported by Codarcea (1940) and Stănoiu (1973) in formations assigned to the Azuga Beds, we maintain the Severin Nappe within which according to us the Azuga Beds form a part of the Jurassic-Neocomian olistostrome (J_2-Cr_1) including the olistoliths of classical ophiolitic mélange (i.e. mélange in mélange) and basic and ultramafic rocks.

Bahna Outlier. It consists of crystalline schists of the Sebeş-Lotru Series and for this reason this outlier is considered to represent a rest of the eroded Getic Nappe.

After the nappe formation, the region was affected by transversal faults. One of them is the Coşuştea Fault in the north of the region (Pl. I). It has a E-W trend and shifts the southern block with more than 3.5 km to the west as against the northern one. Another fault, less significant, occurs west of Podeni, where it affects the Bahna crystalline outlier so that the overlain serpentinites are exposed (Pl. I).

Geology and Structure of the Ultramafic Rock Bodies

The bodies of serpentinitic rocks are the result of the anchimetamorphism of the ultramafic rock olistoliths, exclusively encompassed within the Azuga Beds complex in the base of the Severin Nappe, also affected by the same weak metamorphism.

The Azuga Beds form the lower complex with an olistostrome character of the Jurassic-Neocomian olistostrome. The binding material (matrix) of this olistostrome displays a peculiar phyllitic-satinated aspect and consists of anchimetamorphic schists, greenish or grey, more seldom reddish, within which small bands with a more sandy-carbonatic character are intercalated. Small intercalations of grey limestones are to be found in the Coşuştea spring zone. Although recrystallized, as a result of the anchimetamorphism, they still preserve remnants of *Calpionella*, difficult to be determined. In the Obîrşia Cloşani-Mărăşeşti-Ponoare zone the anchimetamorphic schists of the Azuga Beds include, in their lower part, an olistoplate of basalts in pillow lava facies (Pl. III, Fig. 1, 2), also anchimetamorphosed. For this reason most of the ultramafic rocks olistoliths are to be found in the Azuga Beds complex at a stratigraphic level superior to the basalts, as pointed out by Stănoiu, too (1982).

The phyllitous-satinated schists display an obvious schistosity and, in places, are microfolded during the movements which led to their folding and anchimetamorphism, and which might have been the Late Kimmerian movements. Conformable lenses and unconformable veins of exudated quartz are observed within them. The mentioned schists consist of anchimetamorphic argillites, metasiltites, metasandstones, and metacarbonatic rocks; in places the rocks display characters of "ardesian schists". They include phylosilicate minerals of pyrophyllitic and sericitic type forming small scales disposed with the (001) face in the plane of the schist foliation. Quartz veinlets intrude the rock, some of them being undulated like the ptygmatic veins (Pl. IV, Fig. 1), which points to a folding with concentric sliding which affected the schists during the anchimetamorphism and the veinlets formation.

Ultramafic Rock Bodies. They represent olistoliths of different sizes — from small fragments and blocks of several metres in diameter up to large bodies of 2—3 km long — constituted of ultramafic rocks or an ophiolitic mélange with serpentinitic matrix insedimented in the Jurassic olistostrome complex. They occur in zones in which the formations of the Jurassic-Neocomian olistostrome crop out, both west and east of the Bahna Outlier; it makes us presume that such olistoliths are also to be found under this outlier. A proof of this way of emplacement of the ophiolitic bodies is the fact that they do not metamorphose the anchimetamorphic schists at the contact. Thus, the origin of the so-called "cold intrusions of ultramafic rocks" (Hess, 1955) in the sedimentary or metamorphic formations of the Earth's crust is cleared up at least partly. Due to the anchimetamorphism phenomenon, the bodies

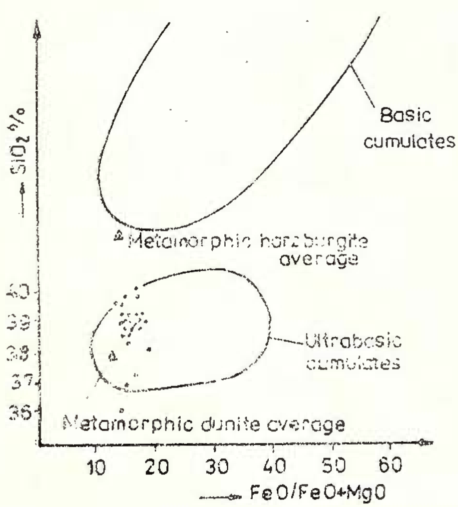

Fig. 1 — SiO_2-FeO/FeO+MgO diagram (after Coleman, 1977 b).

of ultramafic rocks and of ophiolitic mélange are almost entirely serpentinized (Pl. IV, Fig. 3), the water necessary for serpentinization coming from sediments. That is why on the diagrams the serpentinized ultramafic rocks are situated nearby the metamorphosed dunites (Fig. 1), the metamorphosed peridotite field, respectively (Fig. 2). The ophiolitic mélange includs exotic blocks of terrigene crystalline schists, amphibolites, hornblendites and pegmatites or of gabbros, basalts and even blocks of the Azuga Beds as well as jaspers (Pl. IV, Fig. 2) intimately included during the formation of the ophiolitic mélange (Savu et al.,

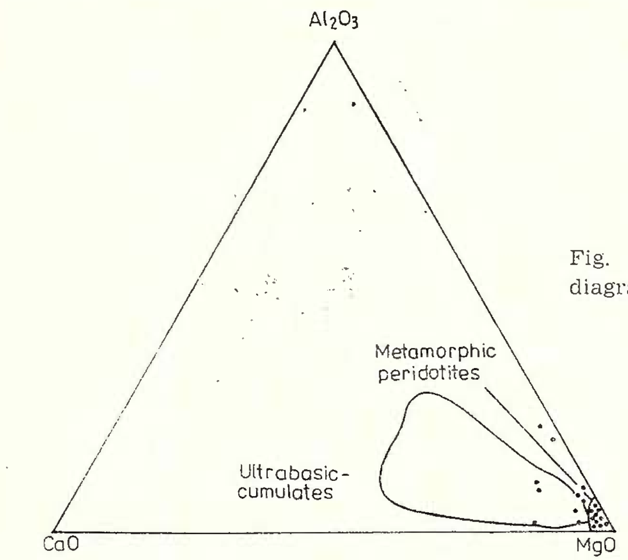

Fig. 2 — Al_2O_3-MgO-CaO diagram (after Coleman, 1977 b).

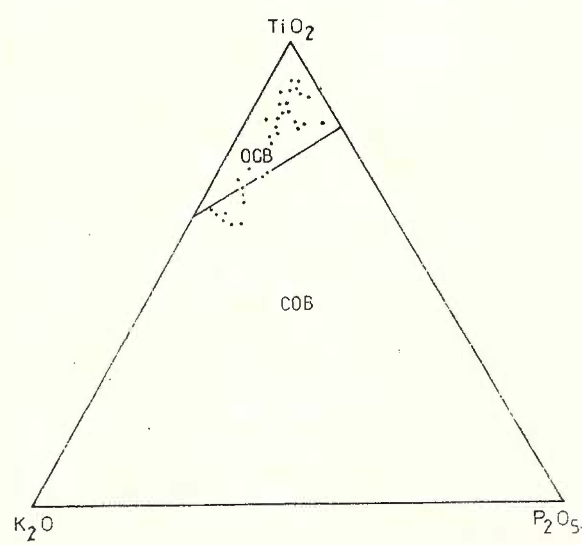

Fig. 3 — TiO_2-P_2O_5-K_2O diagram (after Pearce et al., 1975).

1978) in the obduction zone of the Carpathian Ocean (Pl. II). Olistoliths from the same rocks also occur isolated in the Jurassic olistostrome.

The bodies of serpentinized ultramafic rocks are flattened and elongated like any other pebble from a sedimentary formation more or less metamorphosed (Cloos, 1946). Some bodies were phacoidized, others lenthened and divided into several smaller, boudine-like lenses like those on the right-hand side of the Coşuştea Valley, southwest of Isverna (Pl. I), where the ultramafic rocks do not form a continuous bed as indicated on older maps (Codarcea, 1940). The bodies of ultramafic rocks are conformable with the foliation, that is with the layering of the anchimetamorphic terrigene rocks of the Aguza Beds. On margins, they display phenomena of more intense shearing determined by the mechanical reaction between the two rock types with different competencies — the terrigene schists and the ultramafic rocks olistoliths, more or less plastic, deformed and serpentinized during the folding and anchimetamorphism of the Jurassic-Neocomian olistostrome.

The serpentinite bodies display two categories of structural elements : (a) primary, (b) secondary.

a. The primary structural elements were determined by processes underwent by the rocks from the initial deposit on the Carpathian ocean-floor at their formation. They represent planary elements (Balk, 1937) and have been determined by phenomena of differentiation and rhythmical layering similar to the structures described by Wager and Brown (1968). As mentioned by Mărunţiu et al. (1978) the rhythmical layering occurs in places and it can be observed only in zones in which the rocks have been affected by weathering, provided that they are less serpentinized. In this case on the rock surface thin bands, richer in olivine (dunitic), alternating with bands richer in pyroxenes, come out in bold relief. Small agglomerations of fine pyroxene crystals, like schlieren, can be observed in the homogeneous rocks under the microscope.

b. The secondary structural elements formed during the Jurassic anchimetamorphism, concomitantly with the rock serpentinization in the secondary deposit within the latter occurred as olistoliths in the Jurassic-Neocomian olistostrome situated on the edge of the oceanic plate or in a secondary basin, as well as during the development of the Upper Cretaceous tectonic processes which led to the nappe emplacement. Thus, two categories of structural elements can be distinguished. The first category includes an obvious foliation formed during the plastic stage of the rocks. It becomes more prominent along certain planes and gives rise to shearing zones of ultramafic rocks. Both the foliation and the parallel shearing planes are conformable with the foliation of the surrounding anchimetamorphic terrigene schists. The foliation of the ultramafic rocks was obviously determined by processes of folding and anchimetamorphism which affected the Jurassic-Neocomian olistostrome. The tectonic processes leading to the emplacement of the nappes determine the formation, in the ultramafic and terrigene rocks, of some cleavage planes as well as of mylonitization, brecciation and faulting phenomena (Pl. IV, Fig. 4).

Origin of Ophiolitic Rocks

As already mentioned, the ultramafic rock bodies occur as ex-olistoliths included into the anchimetamorphic Jurassic-Neocomian olistostrome. There are several hypotheses on the origin and emplacement of these ultramafic bodies in the mentioned formation. Thus, they were supposed to represent intrusive bodies (Codarcea, 1940) or bodies emplaced by diapirism (Mărunţiu, 1983), therefore autochthonous rocks which do not come from an oceanic crust (Stănoiu, 1982). They were also regarded as olistoliths of ultramafic rocks in an Upper Cretaceous olistostrome. It implies their origin in a subduction zone of the same age (Stănoiu, 1982) a fact out of question as there is no Upper Cretaceous oceanic zone.

Neither opinion explains satisfactorily the origin and the tectonic position of the ophiolitic zone in the Mehedinţi Plateau as at the contact of the bodies do not occur any phenomena of thermic metamorphism on the binder of the anchimetamorphic olistostrome or geometric relations of magmatic or diapiric intrusion, the bodies being concordant with the terrigene rocks. Likewise, the anchimetamorphic character of the Jurassic olistostrome, in a striking contrast with the aspect of sedimentary rocks of the subjacent Upper Cretaceous deposits, excludes the possibility that the ultramafic rocks might represent olistoliths in an Upper Cretaceous olistostrome.

The bodies of ultramafic rocks are undoubtedly olistoliths, as we shall see further on, the oceanic zone in which they originate as well as the olistostrome including them formed and developed in another space and time. In order to understand these phenomena one has to admit that the Alpine geological evolution of the South Carpathians has been achieved during two stages or cycles : (1) Jurassic-Lower Cretaceous stage and (2) Upper Cretaceous stage.

Jurassic-Lower Cretaceous Stage. The evolution of the Mesozoic magmatism in this area of the Carpathians as well as its geotectonics were gradually outlined by us in 1980, when we pointed out that the ophiolites of the Mehedinţi Plateau represent ocean-floor rocks, of Jurassic-pre-Oxfordian age and the magmatites of the Arjana and Ţarcu zones are submarine island-arc volcanics, of Upper Jurassic age (Savu, 1980). The ocean-floor character of the magmatites was also admitted by Cioflica et al. (1981).

In the sixties it was pointed out that the initial magmatites of the South Carpathians formed in the "Carpathian Geosyncline" lying between the Moesian Platform and the Getic Domain crystalline (Savu, 1967, 1968). Later on, this "geosyncline" was considered to be the result of the evolution of an Alpine oceanic zone (Rădulescu, Săndulescu, 1973 ; Herz, Savu, 1974 ; Bleahu, 1974). Recently, this oceanic zone has been considered to be located between the Moesian Microplate and the Transylvanian Microplate (Savu, 1980), the latter including both the Getic Domain and the Danubian one.

This oceanic zone probably formed at the end of the Triassic or the beginning of the Jurassic, so that in the first part of the latter period took place the spreading of the ocean floor and the formation of a classical oceanic crust (Abbate et al., 1972 ; Coleman, 1977 a, b ; Thayer, 1977 ; Griffin, Varne, 1980) as shown in Plate IIa. It probably consisted of an ocean-floor basaltic complex (O_1), made up of basaltic lavas (pillow lavas), with small intercalations of pyroclastics, jaspers

(Pl. IV, Fig. 2) and pelitic or aleuropelitic rocks, probably a sheeted dyke complex (O_2), a gabbro (O_3), and a peridotitic lower complex (O_4) in association with dunites. The character of ocean floor tholeiites of the rocks belonging to the basaltic complex is clearly rendered on the diagram in Figure 4 ; it was also mentioned in previous papers by Savu (1980) and Cioflica et al. (1981). This oceanic zone was probably symmetrical with the Zvornic oceanic zone in Yugoslavia (Dimitrievič, Dimitrievič, 1973) as compared with the Carpatho-Balkan or Serbo-Macedonian microcontinent (element).

The closing of the oceanic zone probably started in the Middle Jurassic, by a subduction process (Pl. II b). The subduction of the oceanic floor under the Carpathian mobile zone took place under specific conditions due to the closeness of the Moesian Microplate to the Transylvanian Microplate. Concomitantly took place the obduction (Christensen, Salisbury, 1975) of the ophiolitic rocks torn off from the oceanic crust, giving rise to a classical ophiolitic mélange, with serpentinitic matrix, including the blocks of crystalline schists pulled out from the eastern margin of the Transylvanian Microplate, and possibly of Jurassic pelitic deposits occurring in the depression of the subduction zone (Severin Trough). The blocks and plates of basaltic rocks of the upper ophiolitic complex of the oceanic crust and those of ultramafic rocks and of ophiolitic mélange are insedimented as olistoliths in these pelitic sedimentary deposits, which continue their formation in the Jurassic sea which was invading the Transylvanian Microplate too, the whole eastern zone of the Danubian Autochthon, respectively. It is to be mentioned that olistoliths of serpentinites included in Mesozoic sedimentary deposits weakly metamorphosed are also known on the Valea Craiului, Muntele Mic and Arjana zone. The Jurassic-Neocomian olistostrome (J_2-Cr_1) is thus formed, including the Azuga and Sinaia beds with olistoliths from the Alpine ophiolites and the ophiolitic mélange of the South Carpathians.

There is evidence that the olistoliths of basic and ultramafic rocks and especially those of ophiolitic mélange originate in an oceanic zone situated to the east : a. olistoliths of such rocks abound in the eastern part of the Severin Nappe and they become larger and larger ; b. these olistoliths do not occur, for example, in the Jurassic deposits in the western part of the Danubian Autochthon but they appear in its eastern part even in the Jurassic deposits which do not pertain to the Severin Nappe (Muntele Mic) ; c. the presence of the olistoliths of ultramafic rocks and of ophiolitic mélange in the anchimetamorphic schists of the Jurassic olistostrome can be explained only by obduction from the east. If, on the contrary, we considered that the olistoliths of ophiolitic rocks were obducted from the west, we should admit their formation during the Upper Cretaceous (Stănoiu, 1982), when the Getic Nappe thrust to the east, or during the Upper Jurassic, which is not true (Săndulescu, 1980). Provided that the ophiolitic rocks of the Severin Nappe had been emplaced by processes of intrusion or diapirism, which is not true, their transport from the west concomitantly with the Severin Nappe should be admitted.

As the olistoliths of basic and ultramafic rocks as well as of ophiolitic mélange are insedimented in the Jurassic olistostrome one can infer that in that oceanic zone subduction processes took place at the end of the Middle Jurassic or the beginning of the Upper Jurassic, which were not connected with the emplacement of the Getic Nappe formed much later. Similarly we explained the present position of some Alpine ophiolites in the East Carpathians, obducted from the Siret Ocean in the east and deposited on the crystalline schists of the Transylvanian Microplate — subsequently torn out in several nappes — or insedimented in the flysch deposits (Savu, 1980 ; Savu et al., 1984 b).

It is odd that, besides the basalt olistoplates in the Obîrşia Cloşani and Mărăşeşti-Ponoare zones and the olistoliths of ultramafic rocks and of ophiolitic mélange, no olistoliths are known to come from the sheeted dyke complex. It is possible that this complex might be missing in the zones where the olistoliths come from. The appearance of the basaltic olistoplates in the lower part of the Jurassic-Neocomian olistostrome and of the olistoliths of ultramafic rocks and of ophiolitic mélange at an upper level (Stănoiu, 1982) indicate, in our opinion, the order of their tearing out from the two complexes of the oceanic crust, during the obduction.

The Jurassic oceanic crust is now completely subducted (Savu et al. 1985 a, Fig. 6) and mostly consumed, so that between the Transylvanian Microplate and the Moesian Microplate there are very few ophiolitic rocks to be detected by geophysical devices. As a matter of fact the two plates are in a contact of "continent/continent collision", concealled by the sedimentary deposits of the Precarpathian Depression.

In the western part of the Middle and Upper Jurassic sea appears a submarine (Savu, 1980) withinplate or island-arc volcanism, whose products are at present intercalated in the Mesozoic sedimentary deposits in the Arjana and Ţarcu zones, assigned by Năstăseanu (1967, 1980) to the Toarcian-Oxfordian interval. This volcanism develops inside the Transylvanian Microplate (Pl. II b). Its products are represented by agglomerates, tuffs and more seldom lava flows intercalated in sedimentary deposits, sometimes veins, too. They are constituted of porphyritic basalts, often amygdaloidal, spilites, bostonites, trachytes and keratophyres, rocks found in the Cozia Mount (Cornereva) (Savu et al., 1978) and in the Muntele Mic Mount.

In the eastern trough in which the Jurassic-Neocomian olistostrome formed — also the origin zone of the Severin Nappe — an anchimetamorphic process took place which affected the olistostrome deposits and determined the serpentinization of the Jurassic ultramafic rocks and of ophiolitic mélange of the same age. This weak metamorphism arose probably during the Late Kimmerian movements or towards the end of the Upper Jurassic (?). Lower Cretaceous formations (Codarcea, 1940), referred to the Sinaia and Comarnic beds, are conformably developed. They are not affected by the metamorphic processes or at least not equally to the Jurassic ones, the Azuga Beds, respectively.

In the Vardar zone, in Yugoslavia, Dimitrievič and Dimitrievič (1973) studied a similar ophiolitic mélange, of Upper Jurassic age, supposed to have formed in the Middle Jurassic simultaneously with the beginning of the closing process of the Zvornic ocean zone. More southwards, in Macedonia, Mercier and Vergely (1972) also pointed out an ophiolitic mélange, of Mesozoic zone. According to them it underwent an Upper Jurassic-Eocretaceous deformation, during which it was metamorphosed ; one can infer that in these zones a classic ophiolitic mélange formed and was metamorphosed concomitantly with that from the Carpathians. The Zvornic oceanic zone, which generated the ophiolitic series at the expense of which the mélange in the Vardar zone formed, developed also in the first part of the Jurassic.

2. The second stage of evolution of this Carpathian zone started with the formation of the Nadanova Beds, transgressively overlying the Lower Cretaceous carbonatic formations, and continued with the formation of the wildflysch and the nappe formation. The chaotic structure of the wildflysch indicates that during the Upper Cretaceous the sedimentation occurred under conditions of strong tectonic mobility, a fact proved also by the abundance of blocks of crystalline schists and older sedimentary rocks. They also indicate that in the sedimentary basin several geanticlines or shoals of the floor of this basin rose (Pl. II c), from which olistoplates or formations of olistonappe (Schwab, 1979) type appeared, e.g. that of the Cerna Nappe (Stănoiu, 1982), insedimented in the Upper Creaceous deposits (Pl. I). During the wildflysch deposition, under conditions of intensification of the subduction processes of the Moesian Microplate under the South Carpathians an island-arc volcanism develops, probably an equivalent of that occurring in the Rusca Montană. Its products have recently been described by Drăghici (1982) in the Brebina Valley.

There is evidence of the development of such subduction processes in the Middle and Upper Jurassic and the Upper Cretaceous, as follows : a. the existence of the Precarpathian Depression, formed along the subduction zone and whose crystalline basement is, in the Slatina zone, at a depth of 2000 m and sinks towards the Carpathian Chain, where it exceeds 8000 m in depth, north of Tg. Jiu (Map of the Metamorphites in the Carpatho-Balkan-Dinaric Realm, scale 1 : 1 000 000, Budapest, 1976) ; b. vergency of the structures rendered evident by recent papers (Savu, Hann 1980 ; Savu et al., 1984 a ; Kräutner et al., 1981) pointed out that, like in the East Carpathians, the Danubian Autochthon displays a nappe and a scale structure, striking towards the platform, in this case the Moesian Microplate, subducted (underthrust) under the Carpathian Chain concomitantly with the East-European continent, whose integral part was since the end of the Early Kimmerian movements. Some of these nappes, overlapping the Schela Formation (Schela Nappe, Arșeni-Drăgoești Thrust), occur on the very external margin of the Danubian Autochthon ; therefore one should admit that the Getic and Danubian realms display nappe or scale structures, among which the Getic Nappe is the most significant one but not the only one ; c. the presence of a volcanism in the eastern part of the Danubian Autochthon in the Middle and Upper Jurassic and of another one in the Upper Cretaceous, which could be explained only if a subduction from east to west, under the Transylvanian Microplate, is admitted.

As indicated on the sections in Plate I, in the succession of the geological structure the Upper Cretaceous wildflysch is overlain by the Severin Nappe. According to some researchers (Codarcea, 1940 ; Stănoiu, 1982) it came from the western part of the sedimentary basin and would thrust over the Upper Cretaceous wildflysch. In our opinion it might have been formed in the eastern part of the Jurassic-Neocomian sedimentary basin (J_2-Cr_1) and from here torn off during the Subhercynian or Laramian foldings and pushed over or in the Upper Cretaceous deposits. Up to now, in the study area no Upper Cretaceous deposits overlying the Severin Nappe have been mentioned except Stănoiu's statement that the whole unit is of Upper Cretaceous age. Our opinion is that the presence of the olistoliths of rocks belonging to the Jurassic-Neocomian olistostrome as well as to other rocks of the Severin Nappe in the Upper Cretaceous wildflysch (Stănoiu, 1982) indicate that the Severin Nappe was moving during the sedimentation of this wildflysch and consequently it might represent an olistonappe, too (Pl. II c).

From all this one can infer that the Severin Nappe, including numerous and huge olistoliths and olistoplates of ophiolitic rocks, could come only from a trough east of the Danubian Autochthon as a nappe or olistonappe, because more westwards the Jurassic deposits do not contain olistoliths of ophiolitic rocks.

All the Mesozoic structures are finally thrust by the Getic Nappe which includes the Bahna Outlier too and overlaps the whole edifice. The formation and motion of the nappe were determined by a subduction (underthrusting) process of the Moesian Microplate under the Carpathian zone (Savu et al., 1985 a) and its final emplacement is due to Laramian movements (Codarcea, 1940).

Inferences

Several ideas can be inferred from this paper, as follows :

The Alpine geological evolution of the Mehedinţi Plateau, the South Carpathian Chain, respectively, was achieved during two main stages : (1) Jurassic-Lower Cretaceous stage and (2) Upper Cretaceous stage.

In the first stage (J_1-J_2) an ocean zone appears between the Transylvanian Microplate and the Moesian Microplate, within which a classical ocean crust (ophiolitic series) formed during the spreading process.

In the Middle and Upper Jurassic the oceanic zone is closed by a subduction process, giving rise to a Jurassic-Neocomian olistostrome (J_2-Cr_1) with olistoliths of basic, ultramafic rocks and of ophiolitic mélange with elements of crystalline schists, gabbros, basalts, etc. coming from the dismembered ophiolitic plate.

One can certainly state that the olistoliths of basic and ultramafic rocks and of ophiolitic mélange represent the obducted fossil remnants of the Jurassic ocean crust which was mostly subducted under the Carpathians and consumed.

The Jurassic olistostrome (the olistoliths inclusive) is affected by a process of anchimetamorphism, probably Neokimmerian.

In the second stage (Cr_2) subduction (underthrusting) processes and orogenic movements develop ; appears the Upper Cretaceous wildflysch with olistoliths from the previous Mesozoic formations and the Precambrian basement of the Danubian Autochthon.

There are emplaced the Cerna insedimented nappe (olistonappe), probably of sliding and the Severin Nappe, followed by the thrusting of the Getic Nappe over all the previous units.

Finally, there results that the ophiolitic series or the ocean crust is formed during the ocean spreading (Carpathian Ocean), and the ophiolitic mélange during the closing stage ; the latter appears in the friction zone between the continental plate (Transylvanian Microplate) and the ocean crust, which was being subducted, resulting in the obduction of blocks of ophiolitic mélange and of basic and ultramafic rocks, crystalline schists, etc., insedimented as olistoliths in the Jurassic-Neocomian olistostrome (Severin Nappe).

[3] The Bahna Outlier could belong to one of the geanticlines or shoals rising in the Upper Cretaceous basin (Pl. IIc), from where it was pushed more eastwards over the Upper Cretaceous wildflysch formations and the Severin Nappe, respectively. The olistoliths of crystalline schists of the Danubian Autochthon series in the wildflysch and the foldings which continued after the emplacement of these nappes, at present occurring together in a syncline (Pl. I) mentioned by Murgoci since 1910, are in favour of the above-mentioned situation. However, it is only a working hypothesis and if its truthfulness is proved we will consider that the root zone of this outlier would be : (a) the present outcropping zone of the Neamţu Series, and in this case the nappe would represent the crystalline schists basement of the Cerna Olistonappe ; (b) the existing strip of crystalline schists lying between the spring area of the Cerna and Topleţ valleys, which might represent a scale enrooted westwards, not an outlier of the Getic Nappe as represented on the maps.

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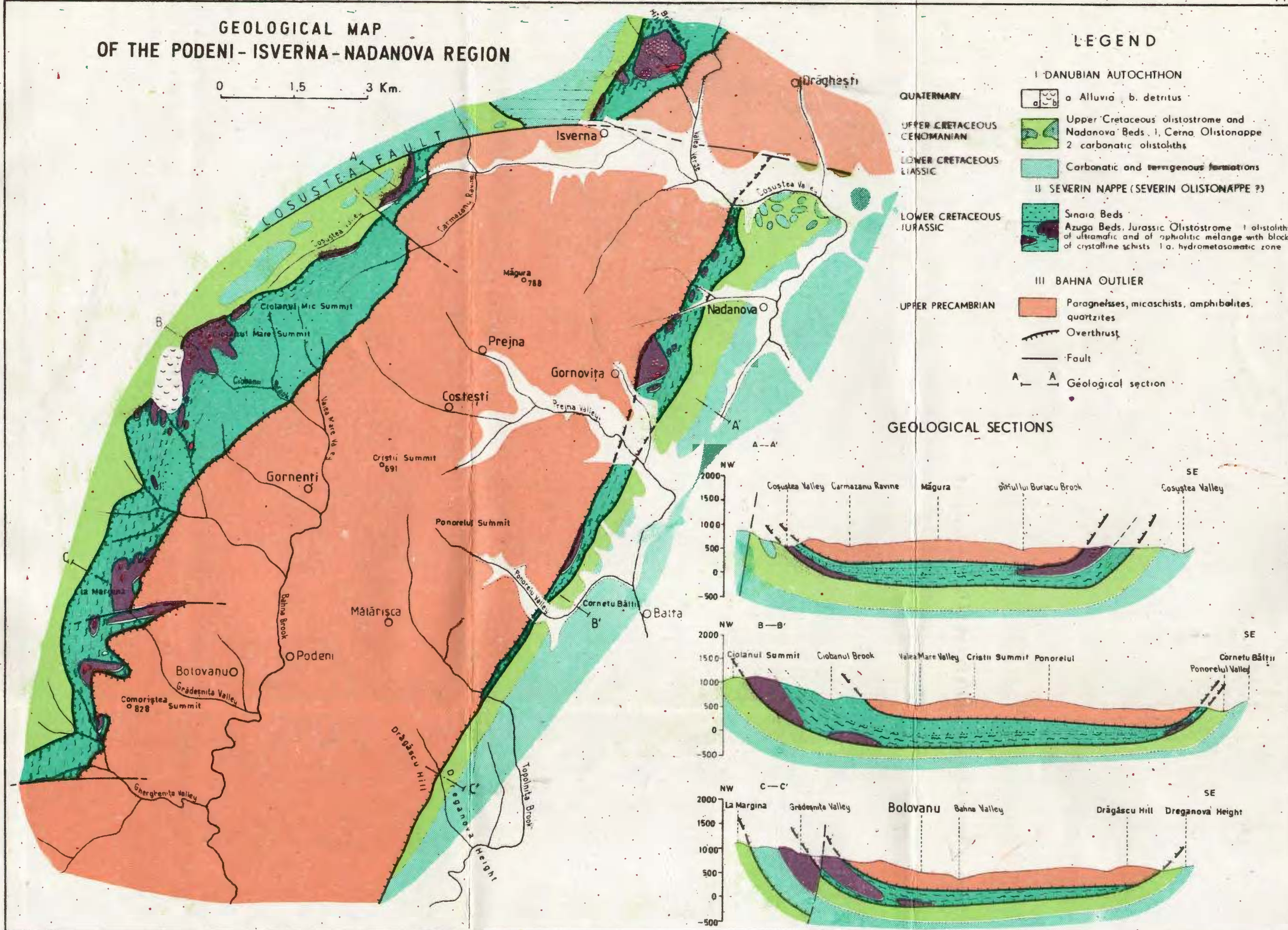
EXPLANATION OF PLATES

Plate III

Fig. 1 — Pillow-lava on the Băroaia Valley.
Fig. 2 — Pillow-lava on the Turcu Valley.

Plate IV

Fig. 1 — Anchimetamorphic schist with ptygmatic veinlets of exuded quartz. Valea lui Dragu. N ||, × 30.
Fig. 2 — Jasper with radiolarian remnants flattened and recrystallized. Gornova. N ||; × 35.
Fig. 3 — Deformed serpentinite with a bastitized orthopyroxene crystal. Gornova. N ||, × 30.
Fig. 4 — Serpentinite displaying two foliations (S_1 and S_2). Spring area of the Hoanca Valley. N ||, × 30.



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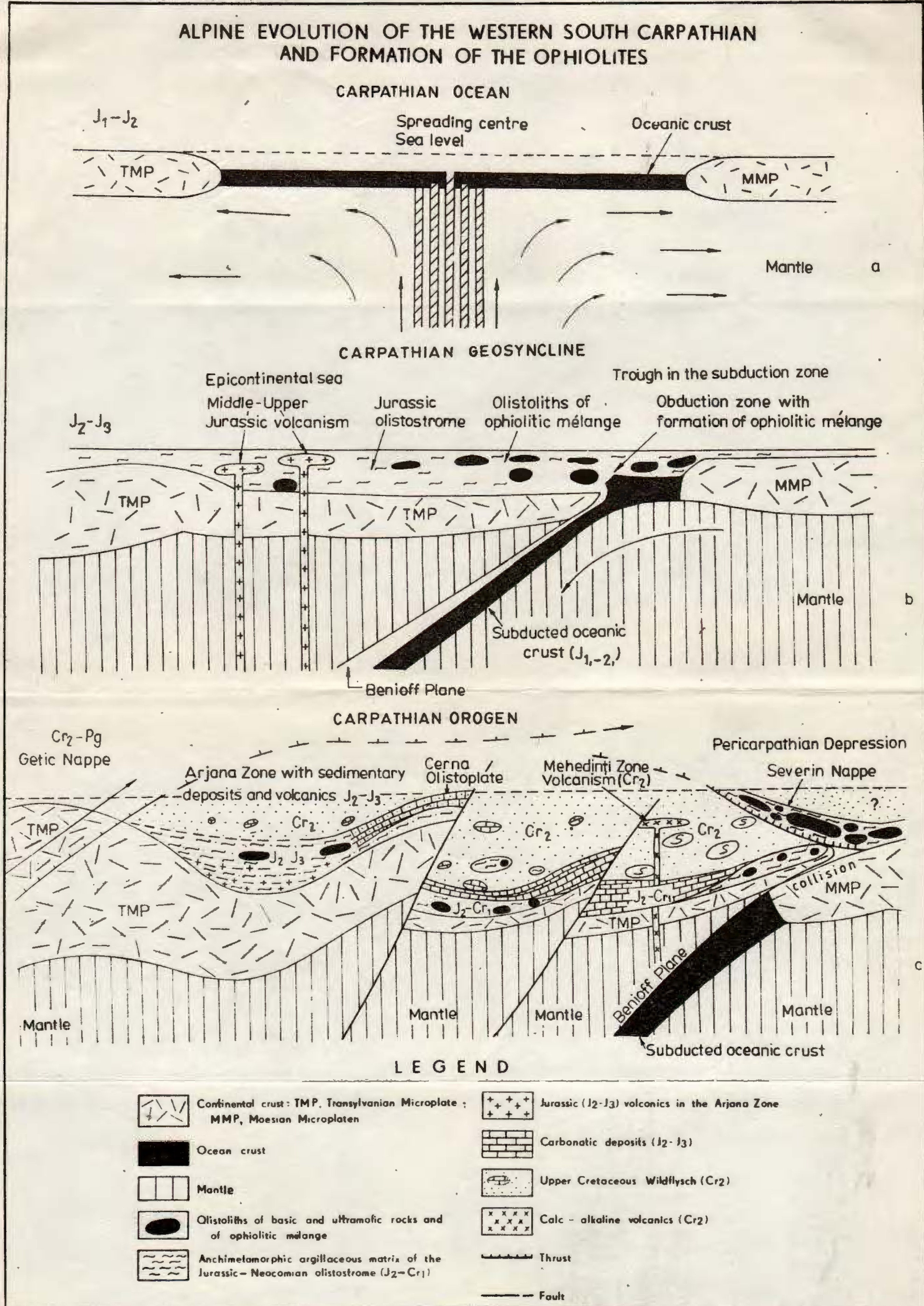

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

CONTRIBUTIONS TO THE STUDY OF PRE-TRIASSIC MAGMATITES IN THE MOESIAN PLATFORM [1]

BY

HARALAMBIE SAVU [2], DUMITRU PARASCHIV [3]

Precambrian. Granitoid plate tectonics. Island arc volcanism. Calc-alkali volcanism. Petrochemistry. Zirconium ; Getic Plateau ; Romanian Plain ; Dobrogea — Central Dobrogea ; South Dobrogea.

Sommaire

Contributions à l'étude des magmatites prétriasiques de la plate-forme moesienne. La plate-forme moesienne (microplaque moesienne) est formée d'un soubassement de schistes cristallins précambriens — où sont cantonnés des corps de roches granitoïdes et une couverture sédimentaire. On trouve au niveau du Permien les produits d'un volcanisme d'arc insulaire. Cette série volcanique se caractérise par des contenus de Zr plus grands qu'en d'autres séries similaires. Le magma qui a généré les volcanites s'est formé à une grande profondeur sur un plan Benioff.

The synthesis papers elaborated up to the present (Paraschiv, 1979, 1982) pointed out that the Moesian Platform corresponds to a very active sedimentary area, within which a succession of deposits, more or less complete, accumulated beginning with the Cambrian and continuing up to the Quaternary. These deposits have been grouped into several lithostratigraphic units (Paraschiv, 1982), found also in the central-northern sector of the platform (Pl.).

A sequence of mostly terrigenous deposits, with a thickness of about 2000-2200 m, which constitute the Ialomiţa Group, is indivi-

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dualized in the base of the sedimentary cover. The Ialomiţa Group consists of three units, regarded as formation : Mangalia Quartzites — Cambrian-Ordovician, Ţăndărei Argillites — deposited in the graptolite facies in the Lower Ordovician-Eodevonian — and Smirna Sandy Quartzites — Eifelian, locally Eodevonian or even Pridolian, which resemble the Old Red Sandstone.

In normal succession there follows the Călăraşi Formation, practically entirely carbonatic, which can reach thicknesses of 2800 m. It is of Givetian-Viséan age, locally also Lower Namurian.

The sedimentary cover of the platform continues with another strong sequence of mostly detrital deposits which constitutes the Vedea-Jiu Group. This group — with thicknesses which can reach 3700 m — consists of two lithostratigraphic units : the Vlaşin Formation and the Roşiori Formation. The former (100-1000 m thick) consists of mostly paralic deposits belonging to the Viséan, Namurian, Westphalian and probably Stephanian. The Roşiori Formation is identical with what has previously been called (Grigoraş et al., 1963) the "Lower Red Series". It is made up of quasicontinental deposits, locally lagoonal or lacustrine, which can reach up to 2700 m in thickness. The basement — belonging to the Permian, possibly the Upper Carboniferous — includes, besides argillites, sandstones, conglomerates, anhydrites, locally limestones and salt, some levels of effusive magmatites, a fact which determined its defining as "Ciureşti Sedimentary-Volcanogene Complex". The present paper will refer mostly to this lithostratigraphic term. Over the magmatite sequence two horizons were delimited within the Roşiori Formation : Brădeşti Sandy Horizon (50-500 m) and Viişoara Pelitic Horizon (20-30 m). Almost all the limits of the mentioned terms are heterochronous.

The stratigraphic succession continues with the Alexandria Formation, mostly carbonatic, of Mesotriassic age, locally Lower Triassic and possibly Permian age, and the Segarcea Formation (Upper Triassic) overlain transgressively and unconformably by Jurassic, Cretaceous, Paleogene, Neogene, and Quaternary deposits.

The structure of the Moesian Platform is obviously marked by the disjunctive tectonics. Two tendencies can be observed within the fault network which divides the platform into numerous compartments: an eastwestern tendency and a northwest-southeast one. The block structure is associated with plicative elements, in places with overthrust tendencies, clay diapirs, morphostructures and deformations generated by the differential compactness of the rocks. The whole structural mosaic is involved in several major deformations, uplifts and sunk zones, whose origin is still unsatisfactorily explained.

Besides the sedimentary cover, the boreholes drilled for the prospection and exploration of hydrocarbon deposits also reached, in some places, the crystalline schists basement of the Carpathian Foreland, as well as numerous intrusive magmatite bodies. Considering the less importance of the plutonic rocks as against oil and gas, the researches carried out up to now generally pointed out their presence and macroscopic analysis without any complex and systematic studies.

However, the relevant literature gives some information on the "Moesian eruptive".

A first synthesis on the intrusive massifs of the Moesian Platform was effectuated by Barbu and Dăneţ (1970). According to them the intrusive magmatites which, beside the crystalline schists, form the platform basement, are grouped into two sectors separated by the Olt Fault. The western sector — Balş-Dioşti — consists mostly of basic differentiated rocks of the gabbro and meladiorite type associated with granodiorites and quartz diorites. The eastern sector — Slatina-Mogoşeşti — is represented by granitic-like rocks with numerous structural and mineralogical variations. The Balş eruptive massif was emplaced in the Paleozoic (undefinite), and the Slatina-Mogoşeşti one in the Lower Paleozoic.

According to Paraschiv (1974) the intrusive magmatites were emplaced in several stages, the Breton phase inclusively. Mutihac (1975), considering Barbu and Dăneţ's accounts, appreciates that the emplacement of the Balş-Optaşi-Slatina magmatic body probably took place during the Baikalian orogenesis and they seem to represent the late orogenic plutonic magmatism of the Baikalian cycle.

More recently, Pătruţ et al. (1980) consider that the intrusive magmatites beside metamorphites form the Precambrian basement of the platform. All the occurrences of intrusive rocks constitute a unitary body — a huge batholite — with meladiorites in its nucleus, surrounded by diorites s. str. quartz-diorites-tonalites and granodiorites to the east. Granites (Oporelu, Priseaca, Mogoşeşti), with a microgranite aureole in the base and bordered by meso- and epizonal metamorphic schists to the north, are found to the nord-east.

Our observations indicate that the Precambrian basement consists of a series of crystalline schists, initially highly metamorphosed, which nowadays are intensely retromorphosed in places even phylonitized. This series consists of alternations of biotite paragneisses (Străjeşti, borehole 3745, m 3758—3761) and amphibolites (Balş, borehole 100, m 2831—2832 ; Fig. 1). Due to the retromorphism process, the feldspars from these rocks are sericitized, and melanocrate minerals such as biotite and amphibole, are replaced by chlorite, iron oxides and leucoxene being separated. Micas from micaschists recrystallized into smaller muscovite lamellae.

The bodies of granitoid rocks (granites and granodiorites) encompassed in the crystalline schists and the Lower Paleozoic formations were also affected by the later metamorphic processes. These rocks were reached by drillings in several places, among which only those from the boreholes (Fig. 1) Pădureţi (borehole 3215, m 5505—5508), Mogoşeşti (borehole 921, m 4500—4502), Străjeşti (borehole 3004, m 3446—3447) and Balş (borehole 100, m 2665—2665.5) were analysed. The later regeneration process, also affecting the crystalline schists, determined the complete sericitization of the feldspars from the rocks and the total chloritization of biotite, as well as the partial recrystallization of the quartz grains.

Recently, Paraschiv et al. (1982) have published the first results of the isotopic age determinations, according to which the metamorphic basement is characterized by values of 543 ± 17 — 566 ± 11 m.y. Taking into account that the study rocks are diaphthorized, the above-mentioned authors consider the Moesian Crystalline of Precambrian

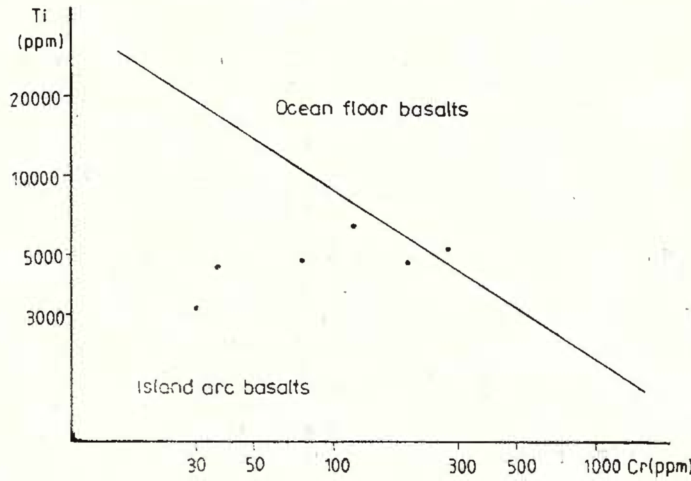

Fig. 1 — Ti-Cr diagram, acc. to Pearce (1975).

age. As regards the intrusive rocks the analyses indicated 350 — 371 ± 11 m.y. for granites and meladiorites and 281 — 289 ± 10 m.y. for granodiorites, diorites and amphibolites. As the granitoids had been influenced by events subsequent to their emplacement, although the values obtained are included in the Hercynian tectono-magmatic cycle, some of them might be older.

Pre-Triassic Volcanism

As regards the effusive magmatites it is worth mentioning the synthesis paper published by Paraschiv (1978), in which four successive levels of volcanics are pointed out : quartz and feldspar porphyries (in the base), followed by eruptive rocks, mostly basic (basalts, melaphyres), again porphyries (in one borehole), and finally basalts beside melaphyres and pyroclastics. The first three magmatic sequences — which we are dealing with in this paper — are assigned to the Permian and the last one belongs to the Middle Triassic. A sample (borehole 5 Strîmbeni) referred to the second phase of Permian eruptions (basic rocks) indicated 243 ± 12 m.y. (Paraschiv et al., 1982).

With a view to completing the knowledge on the pre-Triassic magmatites in the Carpathian Foreland, the eruptive rocks of the Moesian Platform have been studied thoroughly and systematically, the present paper being such an exemple. The pre-Triassic volcanism starts as mentioned above, with an episode of acid volcanics and continues with basic, intermediary and spilitic rocks, bored by drilling on numerous structures among which we have studied only those coming from the boreholes quoted in Table.

a. The eruptive rocks of the first episode are represented by flows of rhyolites and vein rocks with a hypabyssal character, quartz-diorite porphyries and granodiorites porphyries.

The rhyolites or quartz porphyries found at Leu (borehole 4207, m 2514—2519), of a pinkish colour, display a porphyric texture and a massive structure. They consist of a microcrystalline groundmass formed of quartz and feldspars, with small chlorite nests formed at the expense of biotite and rare magnetite grains. Quartz phenocrysts, strongly corroded by magma, are to be found in this groundmass.

The granodiorite porphyry (Ciureşti, borehole 2, m 2556-2557) displays a porphyric texture and a massive structure, formed of holocrystalline mass with isometric crystals of quartz and sericitized feldspars and rare magnetite grains. Phenocrysts of dim, decalcified plagioclase displaying repeated twinning float in the groundmass. Crystals of potash feldspar, also altered, are seldom found.

TABLE

Trace elements (ppm) in volcanic rocks

No.	1	2	3	4	5	6
Rock type	Spilites	Basalt	Basalt	Spilites	Basaltic andesite	Albitophyre
Locality	Morunglav	Morunglav	Recea	Caracal	Pădureţi	Gliganu
Borehole No. Depth (m).	81 2692—2693.5	81 2742—2744	2402 2864—2867	20 3095—3096—	5215 5252—5253	4242 4100—4102
Ti	4800	5300	4900	6500	4600	3200
Ni	95	115	32	68	26	20
Co	22	32	14	13	14	8
Cr	190	275	75	120	36	30
V	135	120	105	110	80	26
Sc	23	23	14	13.5	12.5	10
Zr	210	2220	225	240	400	210
La	30	44	56	50	48	86
Y	18	24	19	27	25	16
Yb	2.0	1.6	1.5	1.9	2.0	1.0
Ba	700	320	750	550	850	850
Sr	1000	650	1000	240	550	320

The quartzdiorite porphyry (Dobrein, borehole 35, m 2206—2207) also displays a porphyritic texture, being made up of a groundmass and plagioclase feldspar phenocrysts. The groundmass consists of fine crystals of repeated-twinned plagioclase, quartz as grains with irregular contour and magnetite. The plagioclase phenocrysts (An_{38-40}) are also repeated-twinned and have a characteristic zonal structure composed of several zones. Some phenocrysts are partly resorbed due to the reaction with the magma and others have their central part altered and replaced by chlorite. One can very rarely observe microphenocrysts of augite and biotite as large lamellae, resorbed on margins, numerous magnetite grains being formed. Biotite pleochroism is as follows: Ng = Nm = dark brown; Np = yellowish. It includes magnetite crystals. Certain chlorite and magnetite pseudomorphoses indicate after their appearance that the rock initially contained also an amphibole, subsequently replaced.

b. The volcanics of the second episode are represented by basalts, spilites, basaltic andesites and albitophyres, for which the radiogene age analyses indicate 240—280 m.y.

Basalts are porphyritic rocks, made up of a pilotaxitic groundmass with a fluidal texture, in which elongated plagioclase phenocrysts (An_{50}), twinned after the albite law, float. The groundmass consists of elongated microcrystals of plagioclase, augite and magnetite. Basaltic andesites also display a porphyritic texture. They are constituted of a hyalopilitic mass with fluidal texture made up of elongated crystals of plagioclase, augite, magnetite and glass. Microphenocrysts of plagioclase (An_{45}), twinned after the albite law and very rarely the pericline law, as well as of idiomorphous augite, displaying repeated twinning after phase (100), are to be found within the groundmass.

Spilites or albitized rocks have an amygdaloid structure and a fluidal texture. They are often invaded by calcite masses filling especially the amygdals. They consist of microlites of albitic plagioclase, especially with fluidal orientation, devitrified glass and magnetite. The plagioclase microlites are turbid, altered, sericitized and decalcified.

The Gliganu albitophyre (Tab.) is a felsitic rock with a porphyric texture; it consists of a groundmass with fluidal texture, made up of elongate albite microcrystals and small magnetite grains with irregular contour. It includes albite phenocrysts (An_8), elongated, dim and sericitized, displaying repeated twinning as well as fine magnetite grains.

Although the petrographic evidence points out that the pre-Triassic volcanics constitute a calc-alkali volcanic series, the rocks belonging to the second episode of volcanic activity (Tab.) have been analysed by C. Udrescu from the Institute of Geology and Geophysics in order to specify the conditions under which they occurred as well as to characterize in geochemical respect the rocks analysed.

Table shows that Ti, Ni, Co, Cr, V and Sc, as well as the lantanides display contents similar to those of the island arc calc-alkali volcanic series. Except for La, which is a bit higher like in the within-plate volcanic series, all the other elements are lower in the albitophyre from no. 6. The island arc volcanic character of these rocks results from the diagram in Figure 1 on which volcanics plot in the field of this type of rocks. The Ba contents, high enough, which characterize the island arc volcanics according to Miyashiro (1975), lead to the same conclusion.

A particularity of this rock series, as against other series of island arc volcanics, is the higher Zr content, which varies from 210 to 400 ppm like in the withinplate volcanic series. The origin of such an enrichment is still obscure. It might be the result of the contamination of magmas during their penetration of the sialic cover of the Moesian plate. Because of this high Zr content, on the diagram in Figure 2 the rocks plot outside the field C of the island arc volcanics, towards the zircon corner. A similar behaviour is observed in case of the rocks on the diagram in Figure 3. Here, in spite of the low contents in titanium, similar to those from other series of island arc volcanics, the rocks occur outside the field of the island arc basalts and andesites, the basaltic andesite (no. 5) being situated even outside the diagram.

The series under discussion displays higher contents, as compared with other series of calc-alkali volcanics, in case of Sr, too, an element ranging between 240 and 1000 ppm. The mean values of the contents is of 626 ppm. If this value is plotted on the diagram established by Hart et al. (1970), there results that the magma from which the pre-Triassic volcanic series come formed at a depth of 700 km, at a level approximately similar to that at which formed the magmas which gave rise to some volcanic series in New Zealand and Japan.

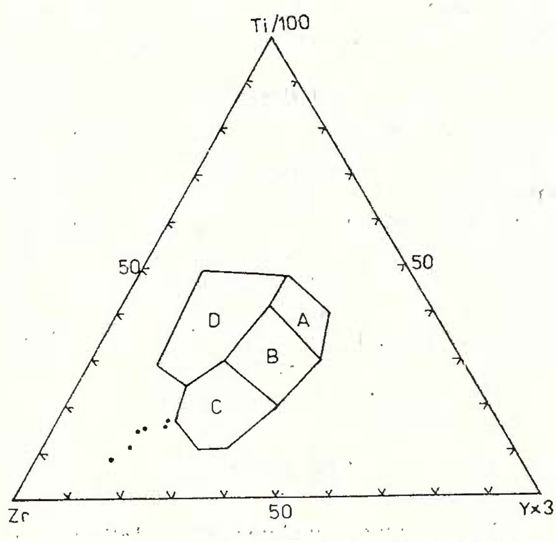

Fig. 2 — Ti-Y-Zr diagram, acc. to Pearce and Cann (1973).

It is very difficult to fancy a model which might explain the formation of the magma in a hot spot zone or on a Benioff plane nowadays when the configuration of the European continent has changed as compared with its image at the end of the Paleozoic. Considering the great depth at which the magmatic chamber formed, one may state that the subduction zone should have been very remote,

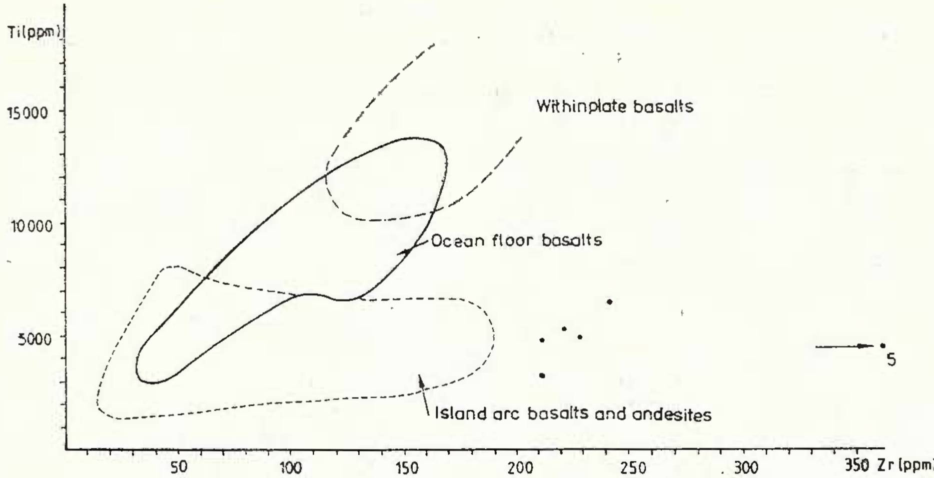

Fig. 3 — Ti-Zr diagram, acc. to Pearce and Gale (1977).

so that according to the dipping of the Benioff plane the magmas formed on it erupted within the Moesian Platform. This subduction zone could be found somewhere south of the Balkans or on the south-western border of the East-European continent.

Conclusions

The basement of the Moesian Platform (Moesian microplate) consists of Precambrian crystalline series, within which bodies of granitoid rocks are encompassed.

Products of an withinplate or island arc volcanism occur in the post-Namurian — pre-Triassic sedimentary cover.

The calc-alkali volcanic series is characterized by Zr contents higher than in other similar series.

The magma from which the volcanics come formed at a great depth.

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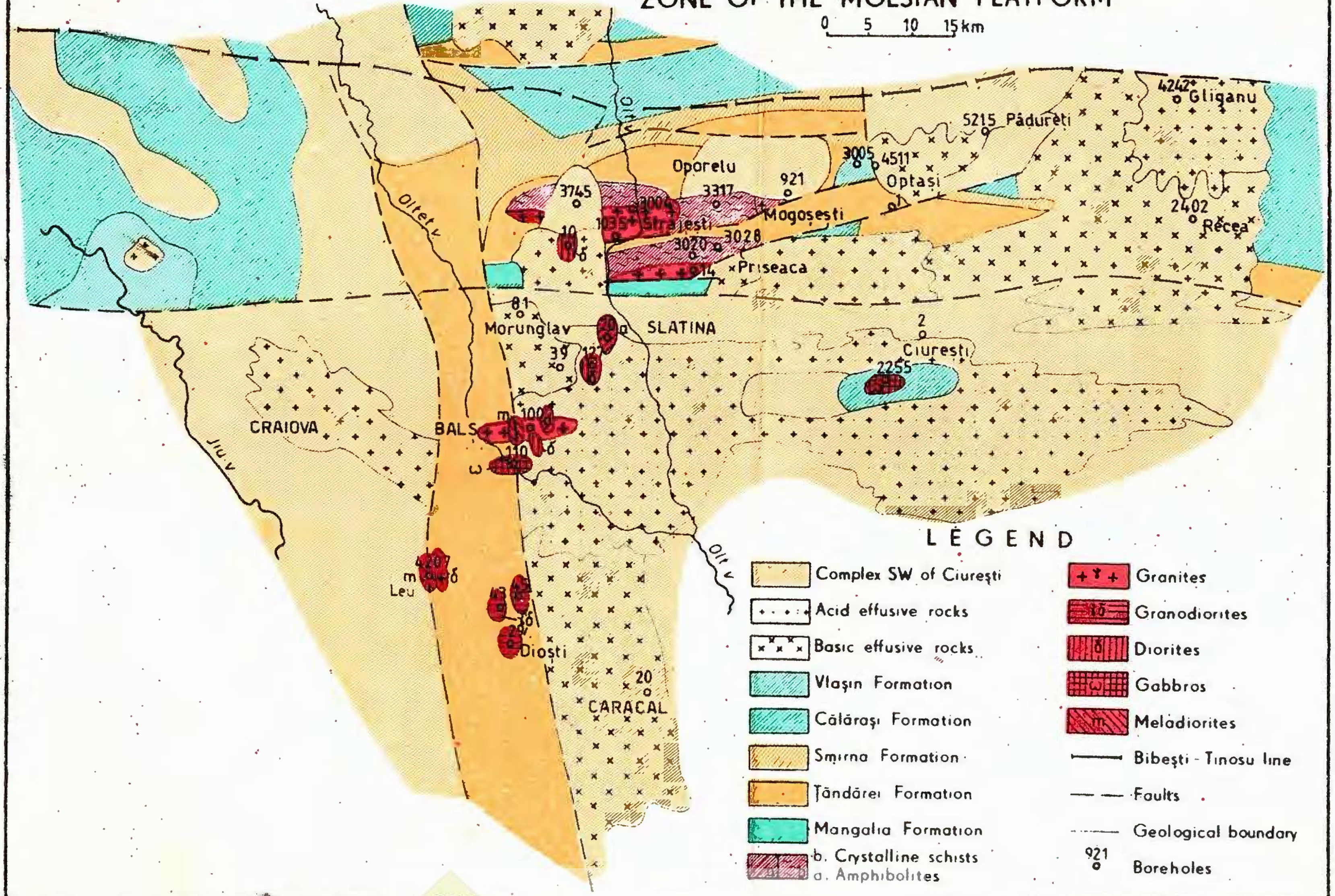
H. SAVU, D. PARASCHIV

DISTRIBUTION OF PRE-TRIASSIC FORMATIONS IN THE CENTRAL-NORTHERN ZONE OF THE MOESIAN PLATFORM



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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

CONTRIBUTIONS À LA CONNAISSANCE DES NAPPES CRÉTACÉES DES MONTS DU MARAMUREŞ (CARPATHES ORIENTALES)[1]

PA

MIRCEA SĂNDULESCU[2]

Nappe. Cretaceous. Dacides. Tectonic unit. Flysch. Scale; sedimentary klippe. Tectogenesis. East Carpathians — Crystalline-Mesozoic Zone — Maramureş Mountains; Inner Flysch Zone — Maramureş Mountains.

Abstract

Contributions to the Knowledge of the Cretaceous Nappes in the Maramureş Mountains (East Carpathians). The Cretaceous nappes in the Maramureş Mts belong, on the one hand, to the Median Dacides (Crystalline-Mesozoic Zone) and, on the other hand, to the External Dacides (Inner Flysch Zone). The former are basement nappes, thrust during the Mesocretaceous tectogenesis, and the latter are obduction nappes, which resemble cover nappes, thrust during the Mesocretaceous (Black Flysch Nappe) and the Laramian tectogenesis (Ceahlău Nappe). Each nappe is described in detail as well as the mutual relationships between them.

Introduction

Les premières données concernant les monts du Maramureş remontent au siècle dernier (Zapalowicz, 1886). Les recherches modernes sont reprises par Bleahu (1955, 1962) dans la partie centrale et occidentale et par Ciornei (1970) dans la partie orientale (vallée du Vaser). Bleahu distingue plusieurs unités tectoniques (unité cristallino-mésozoïque, unité de l'éruptif basique, unités de Rahov et de Corbu); Ciornei reconnaît les fenêtres tectoniques de Făina, de Şuligu et par-

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tiellement de Ştevioara. Plus tard, l'unité de l'eruptif basique est dénommée unité du Flysch Noir (Bleahu et al., 1968). Les premières contributions sur la stratigraphie des formations mésozoïques des monts du Maramureş sont dues à Bleahu (1955, 1962) autant pour la zone cristallino-mésozoïque que pour la zone du flysch. Iliescu et al. (1968), Săndulescu (1975) et Mitrea et al. (1979) apportent de nouvelles données sur la stratigraphie de ces formations. Les séries métamorphiques sont étudiées par Ciornei (1970), Pitulea (1967), Zincenco (1971, 1982) et Zincenco, Balintoni (1978).

Les nappes crétacées des monts du Maramureş ont été également figurées dans les synthèses portant sur la zone cristallino-mésozoïque des Carpathes Orientales (Băncilă, 1958 ; Săndulescu, 1972, 1975 ; Dumitrescu, Săndulescu, 1970 ; Bercia et al., 1976 ; Săndulescu et al., 1981).

Les données concernant les roches basiques reviennent à Bleahu (1955, 1962) et Săndulescu et al. (1979, 1980).

Le schéma tectonique que nous adoptons dans cette note est celui déjà proposé (Săndulescu, 1972, 1975, 1980), suivant lequel on distingue dans la zone cristallino-mésozoïque un système de nappes de socle — le système des nappes centrales est-carpathiques — qui comprend (de haut en bas) : nappe bucovinienne, nappe sub-bucovinienne et nappes infrabucoviniennes. Au-dessus de la nappe bucovinienne reposent les nappes transylvaines (nappes d'obduction) connues seulement en dehors des monts du Maramureş (Rarău, Hăghimaş, Perşani), où elles ont été épargnées par l'érosion. A l'extérieur des nappes centrales est-carpathiques (qui représentent les Dacides médianes) se développent les nappes daciques externes (nappe du Flysch Noir et nappe de Ceahlău). Une couverture post-tectonique (postnappe) qui débute avec le Cénomanien (ou bien le Vraconien) recouvre en discordance les nappes centrales est-carpathiques aussi bien que le nappes transylvaines.

Nappes centrales est-carpathiques

La plus profonde nappe infrabucovinienne connue dans les monts du Maramureş (et en général dans les Carpathes Orientales centrales) est cele affleurant dans la fenêtre de Vaser ; elle correspond à l'unité de Belopotok des monts de Rahov (Ukraine) (Hain et al., 1968 ; Bîzova et al., 1971). Les dépôts mésozoïques de cette unité reposent en discordance sur les formations métamorphiques et appartiennent, probablement, au Lias (Zincenko y a trouvé une bélemnite — données inédites). C'est une séquence à prédominance schisteuse (silts et pélites noirâtres à intercalations de grès minces) surmontant des grès quartzeux noirâtres (ruisseau de Botizu). Dans la vallée de la Tisa cette séquence schisteuse est surmontée en discordance par la série de Dovgorun (cf. Bizova), jurassique moyen-supérieur, métamorphisée. Suivant Zincenco (1982) le Mésozoïque de la fenêtre de Vaser est également métamorphisée (contenant du chloritoïde), ce qui met en évidence la corrélation des deux unités (Vaser et Belopotok). Un autre élément distinctif est l'absence du Trias de l'unité de Belopotok-Vaser.

Une autre unité infrabucovinienne profonde affleure dans la fenêtre de Pentaia. C'est ici que la série schisteuse noirâtre liasique représente le sommet de la séquence mésozoïque à laquelle s'ajoutent, vers la base, des dépôts triasiques inférieurs (?) (grès quartzitiques) et moyens (dolomies et calcaires bitumineux), ces dernières étant traversés par des sills d'andésites et de basaltes aphanitiques (Săndulescu et al., 1979). Les formations triasiques reposent en discordance sur des formations métamorphiques qui à leur tour chevauchent, dans les parties profondes de la fenêtre, une autre unité à schistes cristallins surmontés par des calcschistes limonitiques, peut-être paléozoïques. Cette unité inférieure semble être un équivalent de celle de Vaser-Belopotok (où des calcschistes sont également connus), en marquant ainsi la position paléogéographique plus interne de l'unité de Pentaia.

Aux nappes infrabucoviniennes appartiennent également les formations mésozoïques des unités de Petriceaua et de Poleanca. Une séquence permi-triasique est connue dans l'unité de Poleanca ; c'est d'ailleurs l'unité dont le Permien est le mieux développé de tous les unités centrales est-carpathiques.

Le Permien de l'unité de Poleanca montre deux séquences : celle inférieure à grès, silts et argiles sombres, celle supérieure généralement de couleur rouge (grès et silts) à intercalations de conglomérats et brèches polymictiques claires ; des basaltes et des rhyolites sont interstratifiés dans la séquence supérieure. Un Permien ayant une constitution semblable est connu (Hain et al., 1968 ; Bîzova et al., 1971) dans les écailles situées entre le front de la nappe de Delovetzk (= nappe sub-bucoviniene) et celle de Belopotok, dans le massif de Rahov.

Le Trias de l'unité de Poleanca décrit aussi par Bleahu (1955), débute par des grès quartzitiques clairs (Trias inférieur ?) suivis de dolomies bitumineuses sombres, litées et de dolomies massives (coupe du ruisseau de Chiroaia) (Trias moyen). Il est difficile de décider si les quartzites de la base du Trias sont concordants ou discordants par rapport aux dépôts permiens.

L'unité de Petriceaua est caractérisée par une séquence relativement bien développée du Trias moyen (et supérieur ?). La majeure partie est constituée de calcaires massifs à zones ou rognons dolomitiques (remarqués par Bleahu, 1955, 1962). Ils surmontent des calcaires lités gris sombre, des calcaires et des dolomies bitumineuses (par endroits à rognons dolomitiques) et des calcaires rouges. A la base de la succession carbonatée se développent des dolomies massives qui surmontent des schistes bariolés et des grès quartzitiques clairs (du Trias inférieur ?). Dans la coupe du ruisseau de Repedea, au-dessous des grès quartzitiques et reposant en discordance sur les roches métamorphiques de la série de Bretila, affleurent des grès rouges, polymictiques qui peuvent être éventuellement attribués au Permien. Sur cette coupe il est également difficile de trancher la discontinuité ou bien la continuité entre le Permien et le Trias. De toute manière le Permien de l'unité de

Petriceaua est bien plus mince (une dizaine de mètres) que celui de l'unité de Poleanca (plusieurs centaines de mètres).

Un petit lambeau de recouvrement affleurant à mi-pente sur le versant gauche de la vallée de Polanski semble également appartenir à l'unité de Petriceaua. Là, au-dessus des métamorphites fortement écrasées suivent des grès quartzitiques clairs et des calcaires et dolomies litées, grises ou gris sombre.

La succession triasique de l'unité de Petriceaua ressemble à celle de type Marghitul (Bîzova et al., 1971) du massif de Rahov où des calcaires blancs sont bien développés. Pourtant, le Permien de la suite de Marghitul se rapproche de celui connu dans l'unité de Poleanca, montrant le caractère intermédiaire de cette série par rapport aux deux unités des monts du Maramureş.

Les successions reconnues dans les unités de Pentaia, Poleanca et Petriceaua aussi bien que celle de Marghitul étaient considérées auparavant (Bleahu, 1955, 1962 ; Mitrea et al., 1979) une série unique caractérisant une seule unité. Elles sont pourtant typiques pour des unités différentes, provenant du domaine infrabucovinien. Les corréler aux successions du synclinal marginal, donc à la nappe bucovinienne (Mitrea et al., 1979) est une erreur évidente.

Dans la zone des sources du ruisseau de Ştevioara (bassin du Vaser) nous avons rencontré une séquence sédimentaire qui présente des affinités infrabucoviniennes. Au-dessus des schistes métamorphiques rétromorphisés (série de Bretila ?) reposent des quartzites clairs (4 m) et des dolomies litées, sombres (60 à 70 m) ; la succession rappelle partiellement celle de Pentaia.

En systématisant nos observations sur les unités infrabucoviniennes on peut considérer que la disposition des unités primaires — avant les charriages — était (de l'extérieur vers l'intérieur) : Vaser-Belopotok, Pentaia, Petriceaua, Poleanca (Marghitul entre les deux dernières).

Le lambeau de recouvrement de Stînişoara (Fig. 1) présente selon nos observations des particularités qui le rapproche de l'unité de Roziss du massif de Rahov. En effet, c'est au-dessous des grès quartzitiques du Trias (inférieur ?) qu'affleurent des phyllades verdâtres et rougeâtres identiques aux schistes de Roziss de la vallée de la Tisa. La succession comprend aussi — surmontant les grès quartzitiques — des dolomies et des calcaires massifs médiotriasiques. Un mince liseret de schistes cristallins fortement tectonisés se trouvent par endroits à la base du lambeau (partie NW de celui-ci). Reposant directement sur la nappe du Flysch Noir et n'ayant des rapports directs avec les autres unités centrales est-carpathiques, la position du lambeau de Stînişoara est difficile à préciser dans les monts du Maramureş. En raison de la situation connue dans la vallée de la Tisa (Hain et al., 1968 ; Bîzova et al., 1971) on peut toutefois admettre que le lambeau de Stînişoara est le reste d'une unité située au-dessous de la nappe sub-bucovinienne (Delovetzk), appartenant donc aux unités infrabucoviniennes. Nous avons reconnu — dans la vallée de Botizu — des phyllades semblables à celles du lambeau de Stînişoara coincées entre la nappe sub-bucovinienne et les formations jurassiques de la fenètre de Vaser.

La nappe sub-bucovinienne affleure largement dans les monts du Maramureş y étant représentée surtout par ses roches métamorphiques préalpines („complexe" de Vaser avec ses quartzites de Gliganu, série de Tibău — Zincenco et al., 1976 ; Săndulescu, Zincenco, in Vâjdea et al., 1981). La séquence sédimentaire sub-bucovinienne est restreinte à

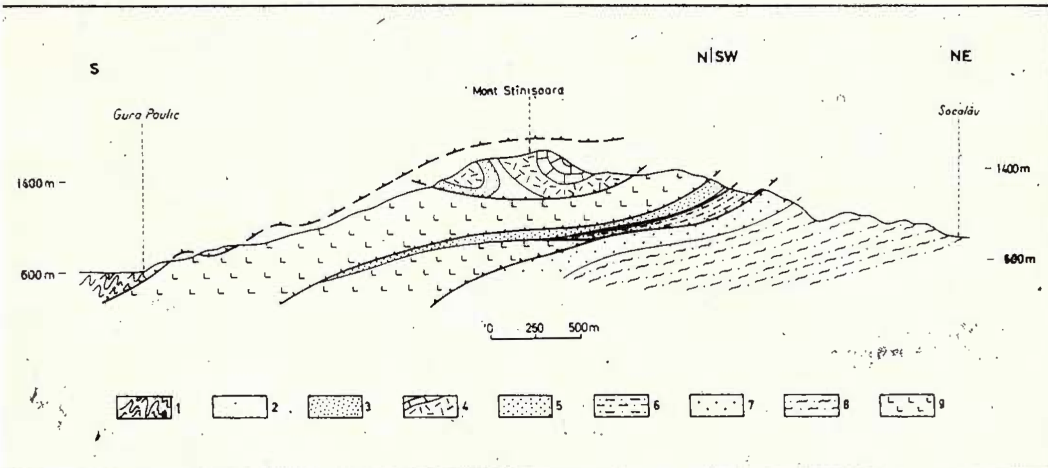

Fig. 1 — Coupe géologique de Gura Paulic-Stînişoara-Socalău.
1, schistes cristallins ; 2, série de Roziss ; 3, Werfénien ; 4, Trias moyen ; 5, flysch gréseux calcarénitique (type de Vinderelu) ; 6, flysch schisteux (Flysch Noir) ; 7, flysch gréseux ; (type de Bistra) ; 8, flysch calcaire (couches de Sinaia) ; 9, spilites ; basaltes et tufs basiques.

quelques lambeaux épargnés par l'érosion, situés dans le bassin du ruisseau de Vaser.

Sur Plaiul Comanului la succession débute par des brèches angulaires à blocs de schistes cristallins (brèches de Hăghimaş — Carbonifère supérieur), surmontées par des grès rouges polymictiques permiens, des quartzites clairs éotriasiques (?) et des dolomies massives médiotriasiques. Elle repose soit directement sur les formations métamorphiques du „complexe" de Vaser soit sur les roches carbonatées de la série de Tibău. Cette dernière situation est également connue dans le mont de Stînişoara Căţelii situé entre les sources du Vaser et la vallée de Pîrcălab (sur la frontière soviéto-roumaine).

Une séquence qui semble appartenir également à la nappe sub-bucovinienne affleure sur Piciorul Babei au-dessus des schistes métamorphiques du „complexe" de Vaser. Des dolomies massives médiotriasiques y sont surmontées par des schistes phylliteux verts et rouge griotte, à intercalations, à leur partie inférieure, de silicolites vertes. Au-dessus, suivent des calcaires lités noirâtres et rouges. Dans une certaine mesure cette succession rappelle celle de la fenêtre de Tomeşti où les dolomies sub-bucoviniennes sont aussi surmontées par des schistes bariolés à silicolites. Nous les rangeons dans le Trias moyen (Ladinien ?). Il est à remarquer, toutefois, que les schistes phylliteux de

Piciorul Babei ont un degré de métamorphisme (anchimétamorphisme) plus élevé que celui des schistes de Tomeşti. On peut à la rigueur les comparer aux phyllades de Roziss, mais dans ce cas-là la tectonique devient très compliquée.

La nappe bucovinienne a été préservée dans les monts du Maramureş sous la forme de lambeaux de recouvrement plus ou moins étendus. Celui de Burloaia (Zincenco, 1971) et le plus grand. Il y a encore celui de Bardău (Săndulescu, 1972, 1975 ; Bercia et al., 1976) et aussi ceux qui affleurent dans le mont de Lostun et sous la couverture postnappe de la crête de Comanova-Gnăteasa. Seul le dernier montre la séquence sédimentaire bucovinienne (sur les autres celles-ci étant enlevée par l'érosion) constituée de grès quartzitiques clairs et de dolomies massives.

Nappes daciques externes

Dans les monts du Maramureş affleurent deux des nappes daciques externes connues dans les Carpathes Orientales : nappe du Flysch Noir et nappe de Ceahlău.

Nappe du Flysch Noir. Cette unité a été reconnue par Bleahu (1955, 1962) sous le nom de „unité de l'éruptif basique" et plus tard (Bleahu et al., 1968) sous celui de „unité du Flysch Noir". Mitrea et al. (1979) la considèrent une digitation de la nappe bucovinienne. Bleahu (1955, 1962) a séparé dans cette nappe plusieurs écailles constituées du complexe basique, des formations de Mihailec et de Vîrtop, du Flysch Noir et du flysch de Vinderelu. Il est le premier à avoir y identifier des stromatites. Ces derniers temps nous avons remarqué des successions particulières pour certaines écailles de la nappe (Săndulescu, 1975 ; Săndulescu et al., 1979, 1980).

Généralement, la semelle de chaque écaille est constituée par le complexe basique, étudié par Russo-Săndulescu, Medeşan et Bratosin (les résultats sont consignés in Săndulescu et al., 1979 ; 1980, 1981) qui y ont séparé des basaltes, tufs, tuffites, schallsteins et stromatites que des sills et dykes doléritiques. Ces auteurs ont également constaté que les pillow-lavas et généralement les basaltes sont prédominants dans les écailles de Farcău et de Budescu ; dans celle de Mihailec-Paulic et celle de Rica I-Socalău des cinérites basiques sont associées aux basaltes, tandis que dans les écailles de Rica II (la plus externe) et d'Obnuj-Vîrtop (la plus interne) ce sont les cinérites qui prédominent.

Dans l'écaille d'Obnuj-Vîrtop, (Fig. 2) le complexe basique est surmonté par la formation d'Obnuj (Săndulescu, 1975) à prédominance argilo-siltique, noirâtre, riche en graphyte, avec des intercalations de grès quartzitiques clairs ou sombres, grès micacés minces, grès limonitiques, calcaires noirâtres (bitumineux sulfatés par endroits), lydiennes noires et calcaires sidéritiques. Sur les pentes sud du mont de Rugaşu, à la base de la série d'Obnuj, les quartzites clairs sont mieux développés (60 à 70 m). Leur absence ou leur puissance réduite dans d'autres coupes est probablement due à des processus tectoniques. En effet il y a des coupes (pentes sud du mont de Rugaşu, vallées d'Obnuj et de Polanski, fenêtre de Polanski) où la base de la formation d'Obnuj est laminée, marquant un certain décollement tectonique par rapport au complexe basique plus compétent.

La formation d'Obnuj renferme aussi des klippes sédimentaires (ruisseau de Vinderelu, pentes sud du mont de Rugașu) constituées de quartzites, dolomies, calcaires lités ou massifs d'âge triasique (de type des roches connues dans les nappes infrabucoviniennes). Leur présence est également soulignée dans cette formation par des intercalations de type „slumps".

Nous l'avons aussi trouvée à l'est de la vallée de Ruscova. Dans le mont de Lutoasa, elle constitue une petite écaille qui surmonte celle de Budescu. Dans la fenêtre de Făina, la formation d'Obnuj est développée dans l'écaille supérieure de la nappe du Flysch Noir qui y affleure.

Le problème de l'âge de la formation d'Obnuj n'est pas simple. Bleahu (1955, 1962) y cite des exemplaires d'*Entolium*. Năstăseanu a trouvé (lors d'une excursion de l'Institut géologique en 1972) des exemplaires de *Gryphaea* (sur le ruisseau d'Obnuj). Nous avons rencontré dans les calcaires de la partie moyenne de la formation des tintinides (*Crassicolaria massutiniana, Cr. brevis, Cr. parvula, Calpionella alpina*) d'âge tithonique supérieur. Mitrea et al. (1979) y citent des foraminifères et une association palynologique d'âge jurassique (en général). On pourra donc conclure qu'on est en présence d'une série qui débute dans le Jurassique (moyen ou inférieur) et qui monte jusqu'au Tithonique supérieur, voire dans le Néocomien. Mais on peut également envisager des fossiles plus anciennes rédéposées au cours du Tithonique. Retenons donc que la formation d'Obnuj est certainement d'âge tithonique (+ néocomien) et que la présence du Jurassique moyen et inférieur reste encore une question à résoudre.

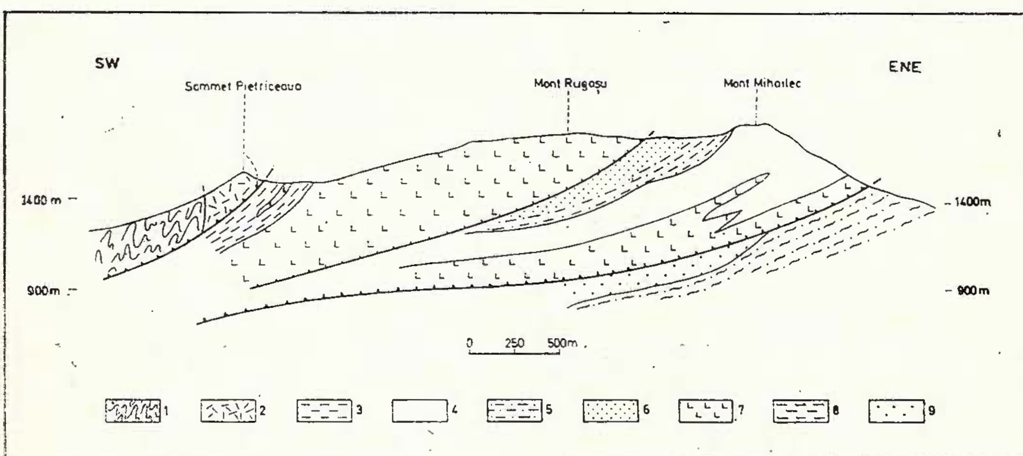

Fig. 2 — Coupe géologique de Pietriceaua-Mihailec.
1, schistes cristallins ; 2, Trias ; 3, formation d'Obnuj ; 4, formation de Mihailec-Vîrtop ; 5, Flysch Noir ; 6, flysch de Vinderelu ; 7, spilites, basaltes, tufs basiques ; 8, couches de Sinaia ; 9, flysch gréseux (type de Bistra).

L'écaille de Farcău a une position externe par rapport à celle d'Obnuj-Vîrtop. Elle s'étend du ruisseau de Chiroaia à travers le mont de Farcău jusqu'au cirque glacière de Vîrtop. Celle-ci groupe les deux types lithologiques que Bleahu (1955, 1962) a décrit sous les noms de formation de Mihailec et formation de Vîrtop. Nous avons préféré un seul nom, vu que les stromatites qui caractérisent la première formation et les calcaires bréchiques de la deuxième formation sont intimement associés ; cette association varie sensiblement d'une coupe à l'autre. Outre les roches mentionnées ci-dessus, dans la formation de Mihailec-Vîrtop s'intercalent des calcaires finement lités, tufs, tuffites et roches basiques. L'épaisseur de chaque type pétrographique est variable d'une coupe à l'autre que sur la même coupe d'une séquence à l'autre. Il est intéressant de souligner que les calcaires bréchiques contiennent des débris de quartz, schistes cristallins épi- et mésozonaux et roches basiques. Ceci montre deux sources du matériel détritique fonctionnant simultanément ou bien une seule source à constitution géologique compliquée comportant des schistes cristallins et des roches basiques.

Nous avons déterminé dans les calcaires bréchiques *Pseudocyclammina, Nautiloculina, Nodosaria* et *Glomospirella ?* (indiquant le Jurassique), mais les restes fossiles sont en général fortement diagénisés. En nous rapportant à des calcaires organogènes d'âge oxfordien connus dans l'unité de Kameny Potok (Hain et al., 1968 ; Bîzova et al., 1971), nous pourrions accorder le même âge au moins à une partie de la formation de Mihailec-Vîrtop.

La formation de Mihailec-Vîrtop est surmontée par le Flysch Noir. C'est un flysch à rythmes binaires constitués de grès calcaires ou argileux, fins, laminés ou largement convolutes et de silts et argiles noirâtres (souvent graphiteuses) ou gris sombre. Le Flysch Noir perd souvent (dans les écailles de Farcău, Mihailec-Paulic et Budescu) son caractère rythmique, se rapprochant de celui de la formation d'Obnuj. A l'ouest du lac de Vinderelu, le Flysch Noir renferme des klippes sédimentaires (calcaires lités, probablement triasiques, mais aussi calcaires bréchiques jurassiques).

Surmontant la formation de Mihalec-Vîrtop et ayant des affinités avec la formation d'Obnuj (au moins une partie de celle-ci) le Flysch Noir peut être considéré d'âge tithonique-néocomien. Remarquons que Bleahu et al. (1968) l'ont traité de jurassique supérieur-néocomien et Mitrea et al. (1979) citent du ruisseau d'Obnujul Farcăului une association palynologique d'âge jurassique supérieur-crétacé inférieur.

La plus jeune séquence de l'écaille de Farcău est le flysch de Vinderelu séparé par Bleahu (1955, 1962). C'est un flysch gréseux à rythmes binaires, de plus en plus grossier et massif vers sa partie supérieure. Il succède, semble-t-il, sans discontinuité au Flysch Noir. Bleahu (1955, 1962) suppose pour ce flysch l'âge barrémien-aptien que nous adoptons ; nous voulons remarquer que Iliescu et al. (1968) ont trouvé des blocs de calcaires à orbitolines dans les arénites du flysch de Vinderelu.

L'écaille de Mihailec-Paulic (Fig. 2) montre une succession semblable à celle de Farcău, à part les différences de constitution du complexe basique (signalées par Russo-Săndulescu in Săndulescu et al., 1979, 1980). Le Flysch Noir y renferme aussi des klippes sédimentaires, constituées de grès quartzitiques (Werfénien ?), schistes calcaires de type schistes de Campil et dolomies médiotriasiques. Quelques klippes sont constituées de grès rouges permiens.

L'écaille de Budescu représente à elle seule la nappe du Flysch Noir entre les vallées de Lutoasa et de Botizu. Au-dessus du complexe basique y suivent la formation de Mihailec-Vîrtop, le Flysch Noir et le flysch de Vinderelu. Sur certaines coupes (versant gauche de la vallée de Budescu), dans la formation de Mihailec-Vîrtop s'intercalent des séquences schisteuses noirâtres qui rappellent le Flysch Noir, révélant des passages latéraux entre les deux formations. Une telle hypothèse semble expliquer le large développement du Flysch Noir dans le bassin du ruisseau de Botizu et dans le mont de Budescu qui renferme aussi des klippes sédimentaires de roches triasiques.

Une succession particulière comporte les écailles de Rica (Rica I-Socalău et Rica II). C'est là que le complexe basique est divisé en deux séquences par une barre calcaire — le calcaire de Rica. Il est nettement développé d'une épaisseur de 30 à 50 m dans l'écaille de Rica II (la plus externe), étant constituée de micrites litées ou schisteuses, à teinte claire ou rouge, à radiolaires et calcisphaeres ; nous y avons rencontré des tintinides aplaties (par le stress). Dans l'écaille de Rica I-Socalău, le calcaire de Rica est plus mince (10 m), mais des intercalations du même type apparaissent dans la séquence supérieure du complexe basique jusqu'au-dessous du Flysch Noir. Ce dernier est d'une puissance réduite dans les écailles de Rica et il est moins graphiteux que celui des écailles plus internes. Il est surmonté par le flysch de Vinderelu (Rica II). L'âge tithonique possible du calcaire de Rica et le développement réduit du Flysch Noir suggèrent que dans les deux écailles de Rica le complexe basique peut monter jusqu'au Tithonique.

Dans la partie septentrionale de la crête de Stînişoara, au front de l'écaille de Mihailec-Paulic, affleure une écaille constituée du Flysch Noir et du flysch de Vinderelu. Plus ou moins à la limite entre les deux flyschs se trouvent un lambeau de roches basiques et une klippe de calcaires lités (Trias ?), qui semblent être des klippes sédimentaires.

Dans la fenêtre de Ştevioara affleurent deux lambeaux que nous considérons appartenir à la nappe du Flysch Noir. Le plus important est constitué de formations sédimentaires (grès et conglomérats quartzitiques, argiles noirâtres, calcaires rouges kimméridgiens à *Saccocoma*), surmonté géométriquement par des pillow-lavas. Il n'y a pas de critères pour décider si la succession est normale ou renversée. Pourtant, la deuxième possibilité n'est pas exclue si l'on compare les conglomérats et les grès quartzitiques avec ceux qui surmontent le complexe basique du mont de Rugaşu. Un autre lambeau est constitué du flysch de Vinderelu à la base duquel affleurent quelques klippes (sédimentaires ? !) constituées de calcaires triasiques bitumineux à chailles. Une corrélation de ces deux lambeaux avec les écailles de la nappe du Flysch Noir est bien difficile. Si l'on admet la position renversée du lambeau à pillow-lavas on pourrait alors le comparer à celui d'Obnuj-Vîrtop.

Nappe de Ceahlău. Bleahu (1955, 1962) a séparé à l'extérieur de la nappe du Flysch Noir deux unités : celle de Rahov et celle de Corbu. La dernière a été comparée par Băncilă (1958) à la nappe du flysch curbicortical (est-interne). Plus tard, Dumitrescu, Săndulescu (1968), Ştefănescu (1967), Săndulescu (1972, 1975) ont considéré que les deux unités apartiennent à la nappe de Ceahlău, représentant deux digitations de celle-là.

Le flysch de Sinaia n'affleure que dans la digitation interne. Il a une constitution monotone (flysch à rythmes binaires de grès calcaires et marnes, avec des intercalations de marnocalcaires ou bien de calcschistes), fait qui ne permet pas une claire division lithostratigraphique. Cependant, dans certains points nous avons reconnu à la partie terminale de ce flysch les brèches polymictiques qui sont caractéristiques pour le flysch de Sinaia supérieur, de Moldavie.

Le flysch de Sinaia est surmonté par un flysch gréseux, massif, pouvant être comparé au flysch de Bistra de la vallée de Bistriţa. Il est bien différent du flysch de Vinderelu qui n'apparaît pas dans la nappe de Ceahlău.

Le flysch de Corbu prend part à la constitution de la digitation externe. C'est un flysch gréso-schisteux alternant à des séquences de flysch gréseux qui deviennent de plus en plus massives vers la partie supérieure (grès de Copilaşu). De ce point de vue il ressemble au flysch de Sînmartin-Bodoc connu dans les monts de Ciuc et Bodoc. Les ammonites trouvées par Marinescu (1979) et par Bucur (fide Marinescu, 1979) *Prodeshyesites* cf. *germanicus*, *P. pseudokiliani*, *Deshyesites* cf. *involutus* et *Acanthoplites* sp.) à différents niveaux attestent l'âge barrémien-aptien de ce flysch, en renforçant ainsi les suppositions antérieures.

Couverture postnappe

Sans entrer dans les détails nous tenons à remarquer que les nappes centrales est-carpathiques sont cachetées par une couverture postnappe qui débute par le Cénomanien, voire le Vraconien. Dans une seul endroit (mont de Lutoasa) cette couverture recouvre le contact tectonique entre les nappes centrales et celles des Dacides externes.

Considérations tectoniques

Nous nous arrêterons ci-dessous à quelques problèmes concernant la structure et l'évolution géotectonique des nappes crétacées des monts du Maramureş.

Domaines géotectoniques. Les nappes crétacées proviennent de deux grands domaines : celui central est-carpathique ou des Dacides médianes et celui des Dacides externes (mégasillon dacique externe — Săndulescu, 1973). Le premier est un domaine à croûte continentale bordé vers l'intérieur du domaine océanique transylvain (Săndulescu, 1975, 1980). De ce dernier proviennent les nappes transylvaines, érodées dans les monts du Maramureş, mais présentes dans les monts de Rarău, de Hăghimaş et de Perşani, au-dessus de la nappe bucovinienne.

Le domaine dacique externe situé à l'extérieur de celui des Dacides médianes s'est fort probablement individualisé seulement au Jurassique (Lias ?) en tant qu'une zone de rift intracontinental (du type des rifts de l'Afrique orientale-mer Rouge-mer Morte) où une croûte amincie (Carpathes Orientales) même océanique par endroits (Carpathes Méridionales) a été générée. Une preuve de cette affirmation sont les caractères géochimiques du complexe basique qui dénotent les traits de basaltes de type „within-plate" (suivant les études de Russo-Săndulescu, Medeşan et Bratosin in Săndulescu et al., 1979, 1980). L'absence des roches ultramafiques vient souligner ce caractère. Il est difficile de préciser le début du rifting puisqu'on ne saurait établir le moment du début de l'évolution du complexe basique. En le considérant au début du Jurassique ou plus tard, dans le Jurassique moyen, nous ne croyons pas nous tromper trop. Cette activité magmatique a continué jusqu'au Tithonique en quelques endroits. Elle doit caractériser non seulement l'aréal de la nappe du Flysch Noir mais aussi celle de la nappe de Ceahlău (les roches basiques du mont de Pietros, U.R.S.S., situées à la base de la nappe de Ceahlău sont une preuve en ce sens).

A la fin de l'Aptien ou durant l'Albien la sédimentation du flysch dans le mégasillon dacique externe s'arrête. C'est le commencement des tectogenèses mésocrétacées.

Origine des klippes sédimentaires et leur mise en place. La fréquence et par endroits les dimensions remarquables des klippes sédimentaires enrobées dans la formation d'Obnuj ou dans le Flysch Noir montre que ce processus tectono-sédimentaire était assez répandu. Ce qui frappe dans le cas de ces klippes sédimentaires est le fait qu'elles ont été englobées dans des formations néojurassiques-néocomiennes bien avant les premiers serrages tectogénétiques. Bien qu'une partie des klippes aient une lithologie „neutre" il y en a assez qui comportent les lithofaciès des formations triasiques infrabucoviniennes. En précisant qu'avant le début du rifting du domaine dacique externe celui-là pourrait héberger une sédimentation triasique (et permienne) semblable au domaine infrabucovinien voisin, nous nous demandons si les klippes pouvaient provenir de l'intérieur du mégasillon dacique, externe lui-même. En fait, si par les processus de fracturation accompagnant le rifting, des blocs plus ou moins importants constituaient à certains moments des escarpements actifs, on pourrait trouver les aires d'origine des klippes sédimentaires. Ce modèle découle du fait que les unités infrabucoviniennes n'avaient pas encore bougées de leurs positions primaires et, de plus, les unités les plus externes sont dépourvues des dépôts triasiques (Vaser-Belopotok). Un autre argument vient appuyer cette hypothèse : la présence dans le Flysch Noir des klippes de calcaires bréchiques dont l'origine dans le mégassillon dacique externe est indubitable. Les escarpements dont nous venons de parler devraient être situés dans la partie interne de la zone de sédimentation de la nappe du Flysch Noir vu que les klippes sédimentaires se trouvent dans les écailles internes de celle-ci (Obnuj-Vîrtop, Farcău, Mihailec-Paulic, Budescu) et font défaut dans celles externes (Rica).

Au sujet de ce problème il faut encore rappeler l'abondance des débris de quartz et de roches métamorphiques dans certains niveaux de la formation d'Obnuj et dans la formation de Mihailec-Vîrtop. Les érosions rapides et fortes des escarpements mentionnés peuvent en donner l'explication.

L'âge des charriages et la succession des processus tectogénétiques. En s'étayant sur des situations connues dans les monts du Maramureş et pour des raisons de corrélation régionale l'âge mésocrétacé des nappes centrales est-carpathiques ne fait aucun doute. Ce sont des charriages de socle, épigliptiques (précédés par une période d'érosion). Le développement et la distribution discontinues de la plupart des nappes infrabucoviniennes nous suggèrent qu'après leur charriage elles ont été soumises à une érosion antérieure à la mise en place de la nappe sub-bucovinienne qui recouvre tout ce „puzzle" d'unités infrabucoviniennes. Une seconde période d'érosion a eu lieu avant le charriage de la nappe bucovinienne sur laquelle étaient déjà charriées les nappes transylvaines. Les charriages se sont donc succédés de l'extérieur vers l'intérieur (pour le domaine central est-carpathique) et de bas vers le haut. Le mécanisme le plus acceptable pour ce modèle est le sous-charriage successif des unités.

Un métamorphisme de haute pression et de basse température (hP/bT) existant dans la nappe du Flysch Noir (Săndulescu et al., 1979, 1980), accompagné d'une schistosité parallèle aux plans de chevauchement des écailles de la nappe, peut être considéré contemporain aux mêmes tectogenèses mésocrétacées, notamment aux premières saccades de celles-ci. En réalité, ce métamorphisme n'affecte ni les unités charriées par dessus la nappe du Flysch Noir, ni la nappe de Ceahlău située au-dessous.

L'écaillage et le métamorphisme des formations constituant la nappe du Flysch Noir peuvent donc être placés au début des tectogenèses mésocrétacées. Son charriage est ultérieur aux écaillages, par suite il a lieu à la fin des serrages mésocrétacés ou bien il est contemporain du charriage de la nappe de Ceahlău, c'est-à-dire à la fin du Crétacé supérieur. Nous n'avons pas des arguments directs pour résoudre ce problème, mais nous sommes plutôt inclinés de considérer le charriage de la nappe du Flysch Noir d'âge mésocrétacé, avec la possibilité que des reprises du plan de charriage aient eu lieu à la fin du Crétacé.

D'ailleurs le fait que la nappe sub-bucovinienne, par exemple, dépasse la nappe du Flysch Noir, étant charriée directement sur celle de Ceahlău (vallée de Lutoasa, fenêtres de Lostun et de Ştevioara) renforce la conclusion mentionnée ci-dessus.

Corrélations régionales. Des unités que nous venons de séparer dans la zone cristallino-mésozoïque des monts du Maramureş, à savoir celles de Vaser, de Stînişoara et celle sub-bucovinienne ont leur équivalent dans les nappes de Belopotok, de Roziss et de Delovetzk du massif de Rahov.

La nappe du Flysch Noir correspond à l'unité de Kameny Potok et/ou à l'unité de Civcin. Dans le mont de Civcin nous supposons l'existence de deux écailles de la nappe du Flysch Noir, celle de Budescu surmontée par une écaille qui pourrait être l'équivalent de l'écaille d'Obnuj-Vîrtop.

La digitation interne de la nappe de Ceahlău est corrélable avec la digitation de Ciuc en Moldavie ou avec l'unité de Rahov en Ukraine. La digitation externe correspond à ce que les auteurs soviétiques séparent sous le nom d'unité de Burkut ou unité de Porkuletz. Au sud de la vallée de Moldova, elle peut être l'équivalent de la digitation de Bodoc ou bien de celle de Durău.

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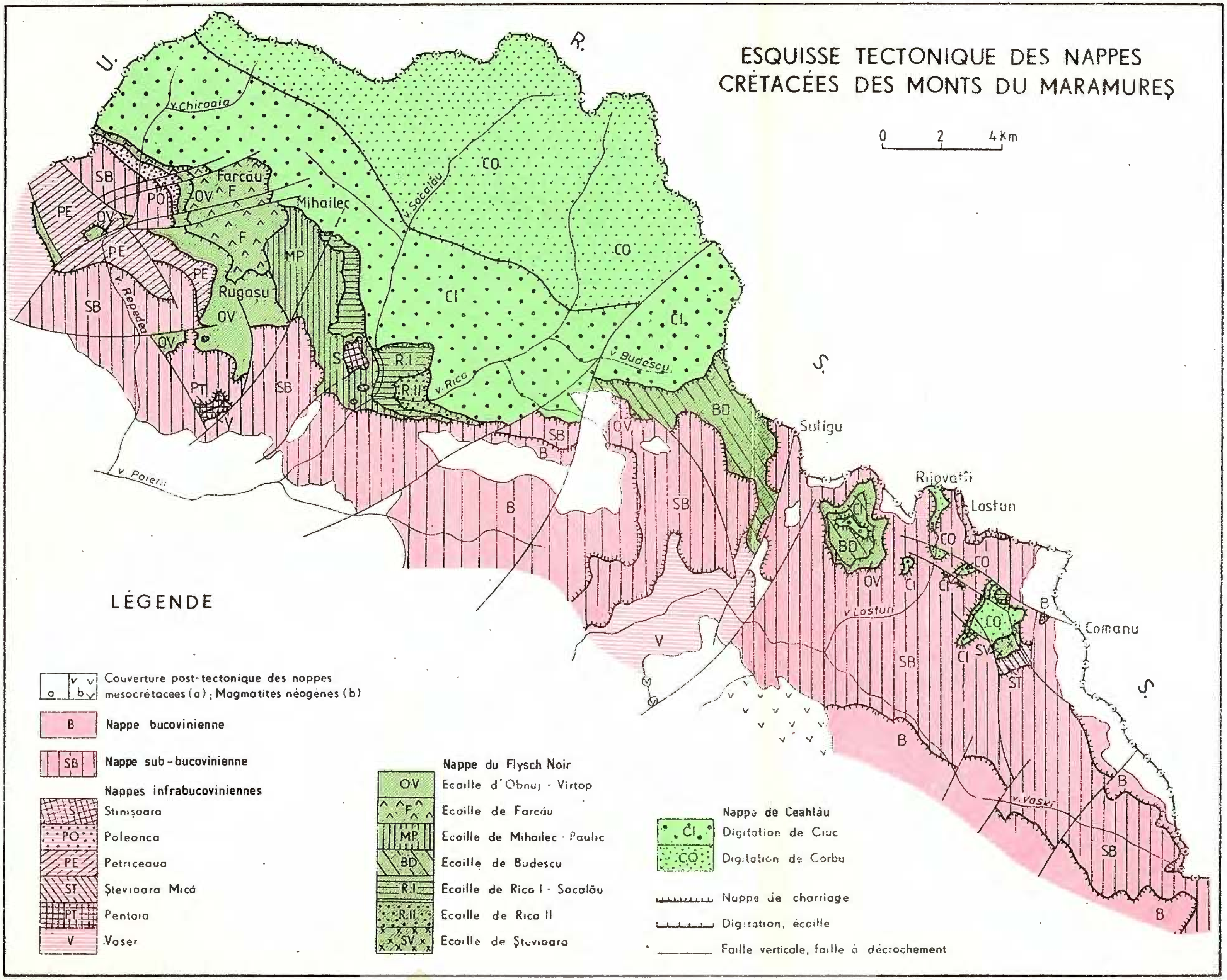

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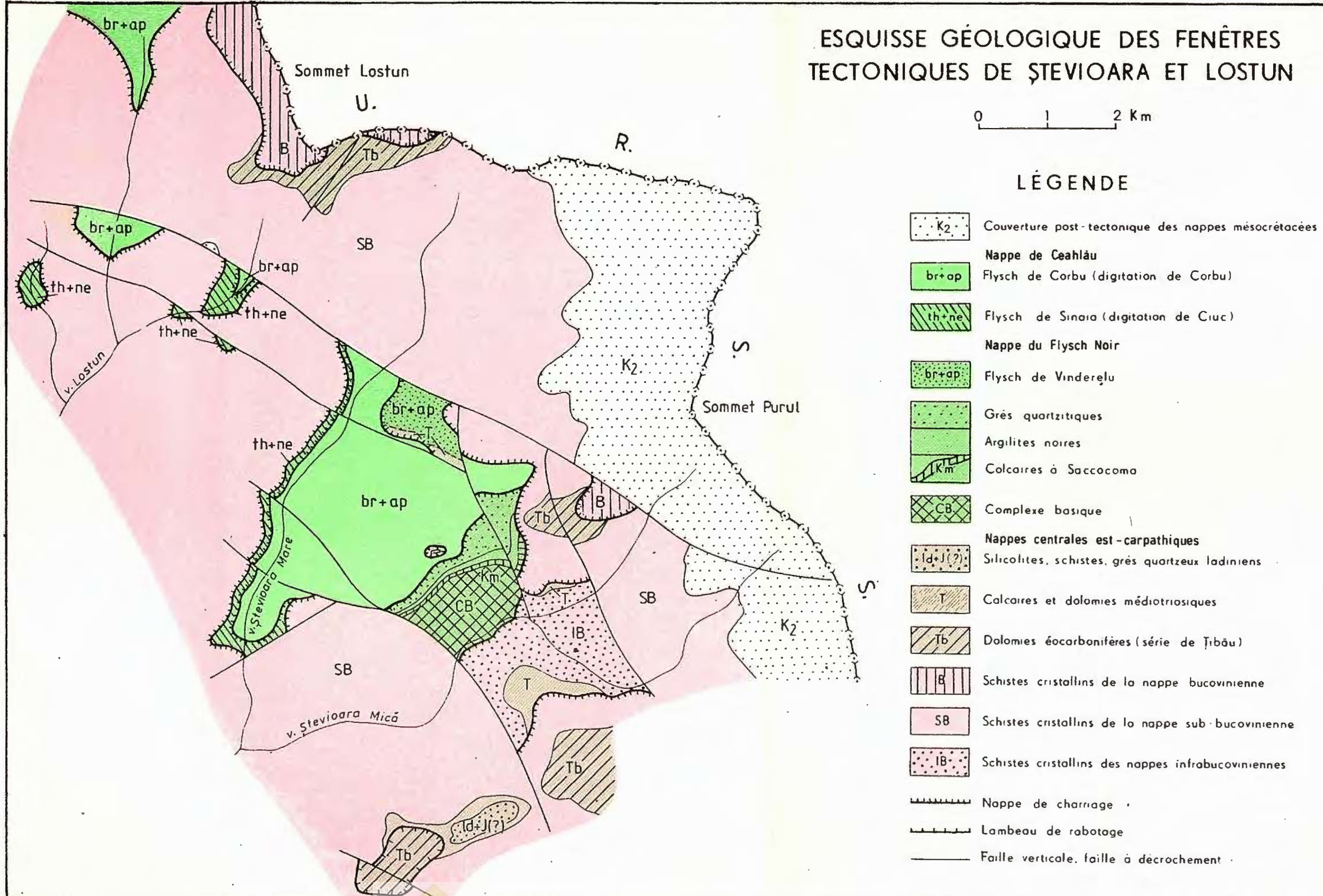
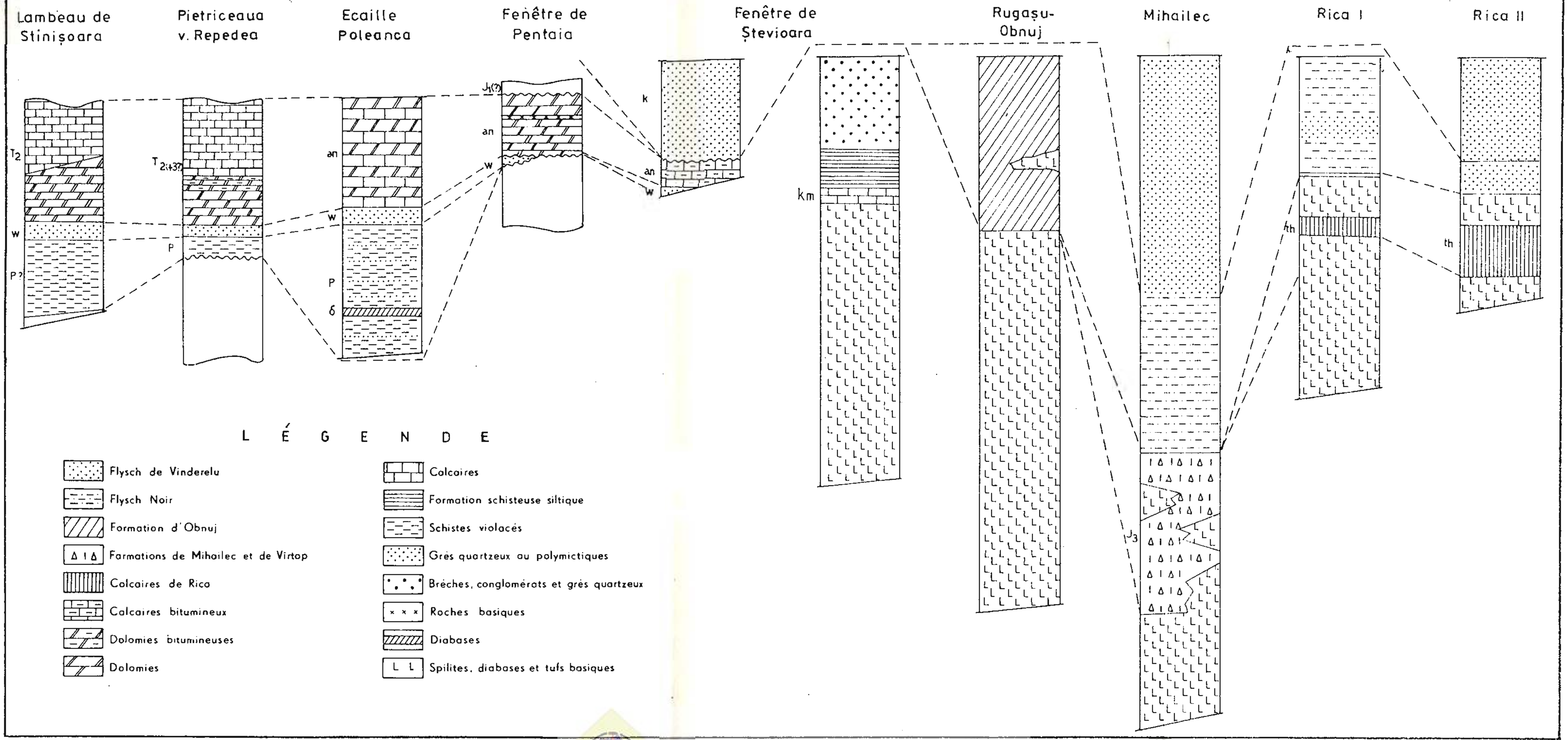
ESQUISSE GÉOLOGIQUE DE LA NAPPE DU FLYSCH NOIR ENTRE LES VALLÉES DE CHIROAIA ET RICA


ESQUISSE GÉOLOGIQUE DES FENÊTRES TECTONIQUES DE ŞTEVIOARA ET LOSTUN
0 1 2 Km
LÉGENDE
K2 Couverture post-tectonique des nappes mésocrétacées
Nappe de Ceahlău
br+ap Flysch de Corbu (digitation de Corbu)
th+ne Flysch de Sinaia (digitation de Ciuc)
Nappe du Flysch Noir
br+ap Flysch de Vinderelu
Grès quartzitiques
Argilites noires
Km Calcaires à Saccocoma
CB Complexe basique
Nappes centrales est-carpathiques
Id+J(?) Silicolites, schistes, grès quartzeux ladiniens
T Calcaires et dolomies médiotriasiques
Tb Dolomies éocarbonifères (série de Ţibău)
B Schistes cristallins de la nappe bucovinienne
SB Schistes cristallins de la nappe sub-bucovinienne
IB Schistes cristallins des nappes infrabucoviniennes
Nappe de charriage
Lambeau de rabotage
Faille verticale, faille à décrochement
Sommet Lostun
U.
R.
S.
S.
Sommet Purul
v. Lostun
v. Ştevioara Mare
v. Ştevioara Mică
br+ap
th+ne
SB
K2
Tb
B
T
CB
Km
IB
Id+J(?)

CORRÉLATION DES FORMATIONS MÉSOZOÏQUES DE LA RÉGION DE REPEDEA - RICA - ŞTEVIOARA



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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

DISCUȚII PRIVIND VÎRSTA CALCARELOR DE PE VALEA TAIȚEI (DOBROGEA DE NORD)[1]

DE

ANTONETA SEGHEDI[2]

Middle Triassic. Limestones. Lithostratigraphy. Tectonic window. Consul Beds. Dorbogea — North Dobrogea — Măcin Zone.

Abstract

The Age of the Limestones on the Taița Valley — A Discussion. On the western bank of the Taița Valley (Măcin Unit of North Dobrogea), between Horia and Balabancea, micritic limestones interbedded with green and red shales occur. There are great lithofacial differences between these rocks and the Paleozoic and Triassic deposits of the Măcin Unit, but they are much alike the Consul Beds (Middle Triassic). The Consul Beds, southwest of Balabancea, are tectonically overlain by metamorphic rocks of the Măcin Unit (Boclugea Group) and represent a tectonic window of the Consul Unit.

În unitatea de Măcin a Dobrogei de nord, pe malul drept al Văii Taitei, între localitățile Horia și Balabancea, aflorează discontinuu formațiuni calcaroase a căror vîrstă a fost foarte divers interpretată. Astfel, Pascu (1904) le consideră triasice, Murgoci (1914) le încadrează la Devonian, Ianovici et al. (1961) și Mirăuță (1962) le atribuie seriei filito-cuarțitice (seria de Boclugea) cambrian ?-ordoviciană, Baclau, Pîrvu (1963) le asimilează cu Silurianul, iar Mirăuță (1966) le descrie la complexul de tranziție între „seria rocilor bazice verzi" — seria de Megina, atribuită Cambrianului — și seria filito-cuarțitică.

Pe malul vestic al lacului Taița, calcarele formează o ivire izolată în masa depozitelor de loess, aflorînd pe cîteva zeci de metri lungime. Sînt calcare micritice, albicioase în suprafața de alterare, slab cenușii

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[2] Institutul de Geologie și Geofizică, str. Caransebeș nr. 1, 78344 București 32.

în spărtură proaspătă şi formează strate de 1—5 cm, avînd lamine milimetrice sau intercalaţii de 1—2 cm de şisturi argiloase roşii şi verzi. Stratificaţia (S_0) este puternic deformată de microcute strînse pînă la izoclinale, astfel încît ele se desfac în plăci după planele de clivaj axial al cutelor şi nu după planele de stratificaţie. Clivajul S_1 este şi el deformat de microcute kink, uneori conjugate, ale căror plane axiale sînt verticale şi au orientarea NE-SV. Planele S_1 sînt orientate NV-SE, avînd căderi mari spre NE sau SV. Aceleaşi calcare apar şi pe drumul spre Balabancea, la baza dealului Coada Muchii, avînd intercalaţii argiloase verzui sau cenuşii, puternic strivite. Microcutele de tip kink ale acestor roci sînt descrise de Murgoci (1914), Mirăuţă (1966) şi Bălaşa, Miheşan (1974). Contactul cu formaţiunile metamorfice din dealul Coada Muchii nu este deschis, dar la 10 m vest de calcare aflorează şisturi muscovitice şi cuarţite ale seriei de Boclugea, cutate pe direcţia EV şi puternic milonitizate.

Microscopic, calcarele sînt constituite dintr-o masă de calcit fin cristalizat, fără a urmări o orientare preferenţială. Cînd sînt prezente intercalaţii argiloase, microcutele sînt vizibile, iar clivajul axial ce le însoţeşte crenulează puternic benzile argiloase din şarniere, confundîndu-se pe flancuri cu stratificaţia. Deformarea nu este însoţită de blasteză metamorfică.

În suitele de roci metamorfice din unitatea de Măcin, o foliaţie metamorfică preexistentă este afectată de deformări ulterioare succe-

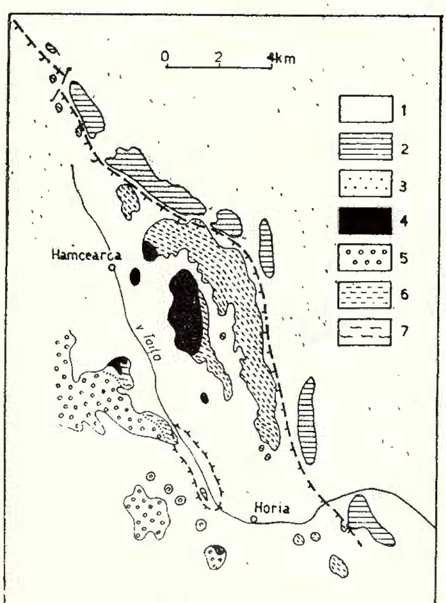

Fig. — Schiţa geologică a zonei Hamcearca-Horia
1, loess ; unitatea de Consul ; 2, strate de Consul ; 3, Werfenian ; unitatea de Măcin ; 4, roci magmatice ; 5, formaţiunea de Carapelit ; 6, seria de Boclugea ; 7, seria de Megina.

Esquisse géologique de la zone de Hamcearca-Horia.
1, loess ; unité de Consul ; 2, couches de Consul ; 3, Verfénien ; unité de Măcin ; 4, roches magmatiques ; 5, formation de Carapelit ; 6, série de Boclugea ; 7, série de Megina.

sive (Seghedi, 1983), deci apartenenţa calcarelor de pe Valea Taiţei la seria de Boclugea sau la seria de Megina este exclusă atît pe considerente litologice, cît şi structural-metamorfice. Rocile în care mai poate fi recunoscută stratificaţia iniţială sînt formaţiunile paleozoice începînd cu Silurianul. Deşi există roci carbonatice în fiecare din aceste formaţiuni, ele diferă de calcarele de pe Valea Taiţei. Astfel, formaţiunea de Cerna (Silurian) conţine calcare negre grafitoase, cu pirită, sugerînd formarea într-un mediu euxinic (Mirăuţă, Mirăuţă, 1962) ; în formaţiunea de Bujoarele (Devonian inferior) apar şisturi calcaroase, gresii calcaroase gălbui sau brune, calcare albe şi cenuşiu-negricioase cu entroce, în Geddinian, şi calcare grezoase gălbui, fosilifere, în Coblenţian (Mirăuţă, 1966) ; formaţiunea de Carapelit în facies de tip Camena cuprinde gresii calcaroase şi calcare oolitice (Mirăuţă, 1967). Depozite roşii şi verzi sînt cunoscute în formaţiunea de Carapelit (Rotman, 1917 ; Cădere, 1924 ; Mirăuţă, Mirăuţă, 1962 ; Mirăuţă, 1967), dar sînt în general detritice grosiere.

Intercalaţiile argiloase roşii şi verzi din calcarele de pe Valea Taiţei sugerează apartenenţa lor la Triasic (S. Rădan, comunicare verbală). Depozitele triasice ale unităţii de Măcin aflorează la Camena — formaţiunea de Uspenin (Grădinaru, 1981) — şi îmbracă un facies predominant organogen — calcare micritice şi calcare argiloase în plăci, biomicrite şi biolitite (Grădinaru, 1981).

Dacă nu pot fi asimilate cu depozite carbonatice ale unităţii de Măcin, calcarele de pe Valea Taiţei sînt foarte asemănătoare cu depozitele calcaroase ale formaţiunii stratelor de Consul ce aflorează în apropierea liniei Luncaviţa-Consul, la NE de Nifon. Stratele de Consul (Patrulius et al., 1974) sînt depozite nefosilifere, pentru care se admite vîrsta triasică medie. În dealul Consul, aceste roci sînt intens deformate de cute tectiforme (Patrulius et al., 1974), iar la NE de Nifon, dealul Pîrlita, ele prezintă acelaşi stil tectonic şi aceeaşi succesiune a proceselor de deformare ca cea descrisă pe malul lacului Taiţa. Stratele de Consul cuprind şi calcare fin granulare cenuşii, cu lamine milimetrice de şisturi calcaroase verzui sau cenuşii şi şisturi argiloase verzui în strate de 1—3 cm (Mirăuţă, 1966 ; Patrulius et al., 1974).

Atribuind calcarele de pe Valea Taiţei formaţiunii stratelor de Consul (formaţiunea calcarelor detritice — E. Mirăuţă et al., 1981), apariţia lor pe Valea Taiţei de sub şisturile cristaline ale seriei de Boclugea nu poate fi interpretată decît ca o fereastră tectonică a unităţii de Consul şi constituie o confirmare a caracterului de pînză al unităţii de Măcin, aşa cum a fost definită de Mirăuţă (în Patrulius et al., 1971).

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DISCUSSIONS SUR L'ÂGE DES CALCAIRES DE LA VALLÉE DE TAIŢA (DOBROGEA DU NORD)

(Résumé)

Les calcaires gris à intercalations de schistes argileux verts et rouges, affleurant sur le versant ouest de la vallée de Taiţa, entre Horia et Balabancea, ne peuvent pas être assimilés aux dépôts paléozoïques ou triasiques de l'unité de Măcin. Ces roches ressemblent bien aux couches de Consul — triasiques moyennes — et leur présence dans l'unité de Măcin, en position géométrique inférieure aux schistes cristallins de la série de Boclugea, est considérée en tant qu'une fenêtre tectonique de l'unité de Consul.

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A NEW TECTONIC IMAGE OF THE MESOZOIC DEPOSITS OF THE CODRU-MOMA MOUNTAINS BETWEEN CRIȘU NEGRU AND RÎPOASA VALLEYS (NORTHERN APUSENI) [1]

BY

MIHAI ȘTEFĂNESCU, ȘTEFANA PANIN, CAMELIA TOMESCU [2]

Mesozoic. Lithostratigraphy. Alpine tectonics. Tectonic units. Nappes. Digitations. Apuseni Mountains — Codru Unit — Codru-Moma Mountains.

Sommaire

Une nouvelle image tectonique des dêpôts mésozoïques des monts Codru-Moma, entre Crișul Negru et Valea Rîpoasa. La présente note essaie de préciser la stratigraphie et les rapports tectoniques des formations mésozoïques qui affleurent entre les vallées de Crișu Negru et de Rîpoasa (monts de Codru). Du point de vue structural, ces formations se rattachent à deux unités du système des nappes de Codru, notamment à la nappe de Finiș (qui comprend les digitations d'Armanu et de Dumbrăvița) et à la nappe de Șeasa. La nappe de Finiș s. str. est caractérisée par des formations détritiques éotriasiques ainsi que par des formations carbonatées et détritiques néotriasiques. La nappe de Șeasa, récemment mise en évidence par les auteurs, chevauche celle de Finiș s. str., étant caractérisée par des dépôts détritiques (prépondérants) à intercalations calcaires triasiques moyennes et supérieures. Les deux nappes ont une couverture sédimentaire commune — couverture de Pontu — qui comprend des dépôts rhétiens supérieurs et liasiques. La puissante formation détritique tithonique-néocomienne — couverture de Valea Mare — achève le cycle sédimentaire de cette zone.

Introduction

The Codru Moma Mountains are situated in the north-western part of the Apuseni Mountains. From the structural point of view, the formations constituting them belong to the Codru Nappes System.

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It is a long time (1911) since thrusting phenomena were demonstrated in the Northern Apuseni Mountains by Szontagh et al. (fide Bleahu, 1976). In the Codru Mountains, the nappe structure was recognized by Rozlozsnik (1936).

During the last years intensive investigations have been carried out in the Codru-Moma Mountains. Among the papers referring to this area we shall quote only the most important ones, belonging to the following authors : Bleahu et al. (1957—1960, unpublished data), Bleahu (1965), Bleahu, Patrulius (1967), Bleahu et al. (1970), Bleahu et al. (1971), Patrulius et al. (1971), Panin, Tomescu (1974), Patrulius, Pajaud 1974), Bleahu (1976), Ianovici et al. (1976), Patrulius et al. (1979 a), Patrulius et al. (1979 b), Bleahu et al. (1979), Bleahu et al. (1981).

According to the conclusion of the mentioned authors, in the Codru-Moma area there are five nappes, as follows : Finiş, Dieva, Moma, Vaşcău, and Coleşti.

Our paper deals only with the deposits formerly included into the Finiş Nappe.

Lithostratigraphical Aspects

The Finiş Nappe was considered as a unitary tectonic unit until 1960, when Bleahu et al. (unpublished data) discovered a singular condensed facies of the Triassic deposits outcropping between the Armanu and Hăigaş valleys. In 1970 Bleahu et al. described and included a peculiar calcareous detrital sequence into the Finiş Nappe, which later will be named „Codru Formation“ (Patrulius, Pajaud, 1974).

The field evidence recently obtained in the area between the Rîpoasa Valley (to the south) and the Crişu Negru Valley (to the north) lead to a different stratigraphical and, consequently, structural interpretations of the Finiş domain, in respect to those already published.

Thus, in the above-mentioned area, the Triassic deposits display two main distinct sequences, as follows :

1. a) North of the Izbuc Valley and west of the Finiş Valley a sequence, made up of several members, develops : pink or yellowish quartzitic sandstones and red shales interbeds (Werfenian) ; black dolomites — Lower Dolomite — (Anisian) ; black, bedded limestones, frequently grading into dolomites, grey or yellow (when weathered) shales — Roşia Formation — (Uppermost Anisian-Carnian) ; white dolomites — Upper Dolomite (Upper Carnian ?-Norian) ; white, grey, red, medium- to thin-bedded limestones and/or dolomites and white reef limestones rock assemblage we shall denominate "Tisa Formation" (Norian) ; from the lithological point of view, it seems to be an equivalent of the "Obîrşia Izbucului Beds“ (Bordea et al., 1975), pro parte ; usually, at the boundary between the "Tisa Formation" and the next member, red breccias and shales are to be found. The last member of the Triassic sequence is represented by red-violaceous, green and grey shales, yellow-greenish or violet siliceous sandstones, sometimes breccias — Carpathian Keuper Formation (Rhaetian).

b) South of the Hăigaş Valley, this lithostratigraphic leitmotif displays some singularities, as follows (Pl. I : 1 b) : each member becomes thinner and, consequently, the whole sequence is more condensed ; here, the dolomitization of the black limestones belonging to the Roşia Formation is almost absent ; within the Carpathian Keuper Formation the presence of yellow dolomite intercalations can be observed. In the Armanu Valley, connected to the Carpathian Keuper Formation, some basic rocks crop out. South of the Santalpar Crest, this condensed sequence displays a thickening tendency.

2. Between the Ursu Valley and the source of the Rîpoasa Valley, a sequence less complete than the above mentioned one crops out, having, at the same time, a different constitution (Pl. I : 2). It starts with grey, grey-whitish, medium- to thin-bedded limestones, which sometimes contain cherts, corresponding to the upper part of the Roşia Formation (Carnian-Early Norian). There follows black or yellow (when weathered) argillaceous shales and marls, grey or black hard calcareous sandstones, among which lenses of grey limestones are developed, an assemblage described as the Codru Formation (Patrulius, Pajaud, 1974), largely outcropping in the Şeasa Valley (in the former geological literature — Şasa Valley). The next part of the sequence is made up of thick-bedded, grey or whitish limestones (Dachsteinlike limestones), red-violaceous limestones and dolomites, red shales and calcareous breccias with red matrix (Norian). Finally, the topmost component of the sequence contains violaceous, red and grey or greenish shales, yellow dolomites, yellow and black sandstones and black calcareous breccias — Carpathian Keuper Formation (Rhaetian).

All the sequences above described have a common and unitary cover — Pontu Cover — made up of dark grey to black limestones with corals and black shales — Kössen Formation [3] — (Upper Rhaetian), black thin- to medium-bedded limestones, black shales and calcareous, dark grey sandstones ; fossiliferous, red limestones, encrinitic limestones, white-yellowish limestones (Liassic). After a gap the sedimentation is resumed during the Upper Jurassic (white, thick-bedded limestones, white thin-bedded limestones with cherts and jaspers). The Lower Cretaceous deposits (grey shales and sandstones, locally fossiliferous calcarenites) are following. Both the Upper Jurassic and the Lower Cretaceous deposits represent the Valea Mare Cover.

Structural Aspects

As we have already said, the Finiş Nappe was a long time considered as a unitary tectonic unit. In the area of interest seven years ago Bleahu et al. (1971) separated three tectonic units — Şoimi Unit, Finiş-Gîrda Unit and Dumbrăviţa Unit — without drawing any tectonic boundary between the Dumbrăviţa and Finiş-Gîrda units because of the lack of field evidences.

Due to our detailed mapping, it is now possible to clearly distinguish, on the studied territory, two different nappes (Pl. II), namely the Finiş Nappe s. str. (lower), corresponding to the first examined sequence (1 a, b), and the Şeasa Nappe (upper), made up of the second sequence (2).

The Finiş Nappe continuously develops between the Crişu Negru Valley and the Brătcoaia carstic zone, and probably southwards, down to the Moneasa Valley. Even within its territory we could, however, recognize an important thrusting plane which separates the non-condensed normal sequence (1 a, beneath) from the condensed one (1 b, above) (Pl. I). Thus, we are in the presence of two digitations — Dumbrăviţa [4] and Armanu — of one and the same nappe, namely the Finiş Nappe. The importance of the Armanu Digitation thrusting on the Dumbrăviţa one is pointed out by the fact that between them a thin tectonic slice is interposed. Due to the fact that the slice is made up of the same condensed deposits as the Armanu Digitation, it has an overthrust slice ("lambeau de poussé") character.

The Armanu Digitation basal plane was traced from the Hăigaş Valley eastwards, up to the right side of the Armanu Valley. East of the Armanu Valley, the tectonic plane does not crop out, being covered by a pile of deposits beginning with Upper Rhaetian (Kössen Formation) and grading up into Liassic rocks. This pile covers both the Dumbrăviţa and the Armanu digitations, without interruptions or facial changes, a fact which proves that the thrusting of the Armanu Digitation over the Dumbrăviţa one has taken place before the Kössen Formation deposition. The uppermost Rhaetian-Liassic deposits are, therefore, representing a post-tectogenetic (post-thrusting) cover (Pontu Cover), common for both digitations.

South of the Ursu Valley, the Finiş Nappe, represented by the Armeanu Digitation, is thrust by the Şeasa Nappe More precisely the Roşia Formation (Carnian) or the Codru Formation (Norian), belonging to the Şeasa Nappe, thrust the Carpathian Keuper Formation (Rhaetian) of the Finiş Nappe. This fact makes clear the tectonic contact between the two nappes. The Codru Formation, earlier considered (Patrulius, Pajaud, 1974 ; Ianovici et al., 1976 ; Patrulius et al., 1979 ; Bleahu et al., 1979 ; Bleahu et al., 1981) as a characteristic formation of the Finiş Nappe (in spite of its absence in the Finiş Valley) belongs, in fact, to a higher nappe, namely to the Şeasa Nappe.

North of the Ursu Valley, the above-mentioned contact is obliqually cut by a vertical fault beyond which it is very difficult to recognize it because only the Carpathian Keuper Formation outcrops there. Taking into account some lithological details of this formation, it seems, however, that the frontal line of the Şeasa Nappe reaches the right side of the Finiş Valley. Here, the Carpathian Keuper Formation is overplated by the same Pontu Cover which develops uninterruptedly on both the Finiş and the Şeasa nappes. In spite of the difficulties in following the limit of these two units on the Carpathian Keuper outcropping surface, we think that the overthrusting of the Şeasa Nappe took place before the uppermost Rhaetian (before Kössen Formation). This means that the uppermost Rhaetian-Liassic deposits are to be considered as a post-tectogenetic (post-thrusting) cover, both for the contact between the two digitations of the Finiş Nappe, and for the contact between the Finiş Nappe and the Şeasa Nappe.

Mention should be made of the fact that the deposits building up the Finiş and Şeasa nappes, together with their common post-overthrusting cover (including the Upper Jurassic-Lower Cretaceous deposits — Valea Mare Cover [5]) were later affected by younger tectonic movements during the Austrian and Mediterranean phases.

Comments

Summarizing the above-discussed problems, one can say that instead of the one unit — the Finiş Nappe (in the former meaning), there are now two units, namely the Finiş Nappe *s. str.* and the Şeasa Nappe, the former being made up of the two digitations — Dumbrăviţa and Armanu, respectively.

The Neogene Beiuş Basin represents an impediment for the attempt in following farther eastwards, in the Pădurea Craiului and the Bihor Mts, the above-mentioned units. The correlation possibilities between the Codru-Moma and the Bihor-Pădurea Craiului Mts have been discussed many times during the last years by Bleahu (1963), Patrulius et al. (1971), Bleahu (1976), Ianovici et al. (1976), Patrulius et al. (1979), Bleahu et al. (1981).

We do not intend to discuss these correlation possibilities but only to try to find eastwards, on the other side of the Beiuş Basin, the possible equivalents of the Finiş and Şeasa nappes, as they have been distinguished by us in the Codru-Moma Mts.

Taking into account the fact that both the Finiş and the Şeasa nappes are characterized by the large development of the Carpathian Keuper Formation, we have first to look for those units which include this formation in the Bihor and Pădurea Craiului Mts. These are : Vălani Nappe (Patrulius, 1971), Ferice Nappe (Rozlozsnik, 1936), Ferice Slice and Următ Nappe (Bordea et al., 1975). Generally speaking we can say that the three above-mentioned units can be correlated with the Finiş-Şeasa Group. However, as already known (Pl. I), the Finiş Nappe differs from the Şeasa one by the presence of the dolomitic-calcareous sequence at the stratigraphical level at which in the last unit a detrital-calcareous formation (Codru Formation) appears. Therefore, we have to look now for those units containing similar deposits in the Bihor and Pădurea Craiului Mts. These are : Roşia-Sohodol Slice (Patrulius et al., 1971) as well as the Ferice Nappe (according to Bordea et al., 1978). These last two units are certainly equivalent with the Şeasa Nappe. Mention should be made of the fact that the Ferice Nappe has more similarities with the southern part of the Şeasa Nappe, developed outside the studied area, in the Moneasa zone.

As concerns the Următ Nappe, it is difficult to say to which unit it belongs, because the deposits older than the Carpathian Keuper are missing.

The Ferice Slice, which displays an upper dolomite-Carnian Keuper, probably represents the Finiş Nappe *s. str.*

The so-called Corbeşti Outlier (Patrulius et al., 1979) has an incomplete sequence, due to the erosional phenomena. For this reason it has been considered as belonging either to the Arieşeni Nappe (most frequently), or to the Dumbrăviţa Unit (Bleahu et al., 1971), or even to the Finiş Nappe (Patrulius et al., 1971). Not far south of its outcropping area, a borehole (Corbeşti 4008) beneath the Neogene deposits revealed the presence of a sequence (Bleahu et al., 1971) similar to that of the Finiş Nappe *s. str.* This drilling proves that the Finiş Nappe *s. str.* extends northwards, beneath the Beiuş Basin, just near the Corbeşti Outlier. Taking into account that the north-east compartment of the Corbeşti Fault had a positive movement one can say that the Corbeşti Outlier also represents the Finiş Nappe, as it has already been said in a former interpretation by Patrulius et al. (1971).

Even if the Vălani Nappe has deposits in Carpathian Keuper facies, its petrographic position, as Patrulius has already said (1971) is intermediary between the Bihor Autochthon and the Finiş Domain.

Thus, following Patrulius' opinions, the Văleni Nappe displays a double character, having Triassic deposits kindered to those of the Codru Domain and Jurassic deposits similar to those of the Bihor Autochthonous Domain. This situation can easily be explained if one accepts that during the pre-Rhaetian tectonic movements, the former area, corresponding to the present Vălani Nappe, comes by the Bihor Autochthon, thrusting it partially and having the possibility to act as a unitary domain later on.

Instead of conclusions we want to emphasize that the most important new data pointed out in the present paper are both the delimitation of two digitations of the Finiş Nappe (Dumbrăviţa and Armanu, respectively) and the discovery of the Şeasa Nappe.

[3] We use this denomination in the lithostratigraphic meaning in which it was employed in the geological literature concerning the Apuseni Mountains.

[4] This denomination was first used by Bleahu et al. (1971). We use the same denomination with a different tectonic meaning.

[5] Valea Mare Cover corresponds to the Valea Mare Formation (Patrulius et al., 1976).

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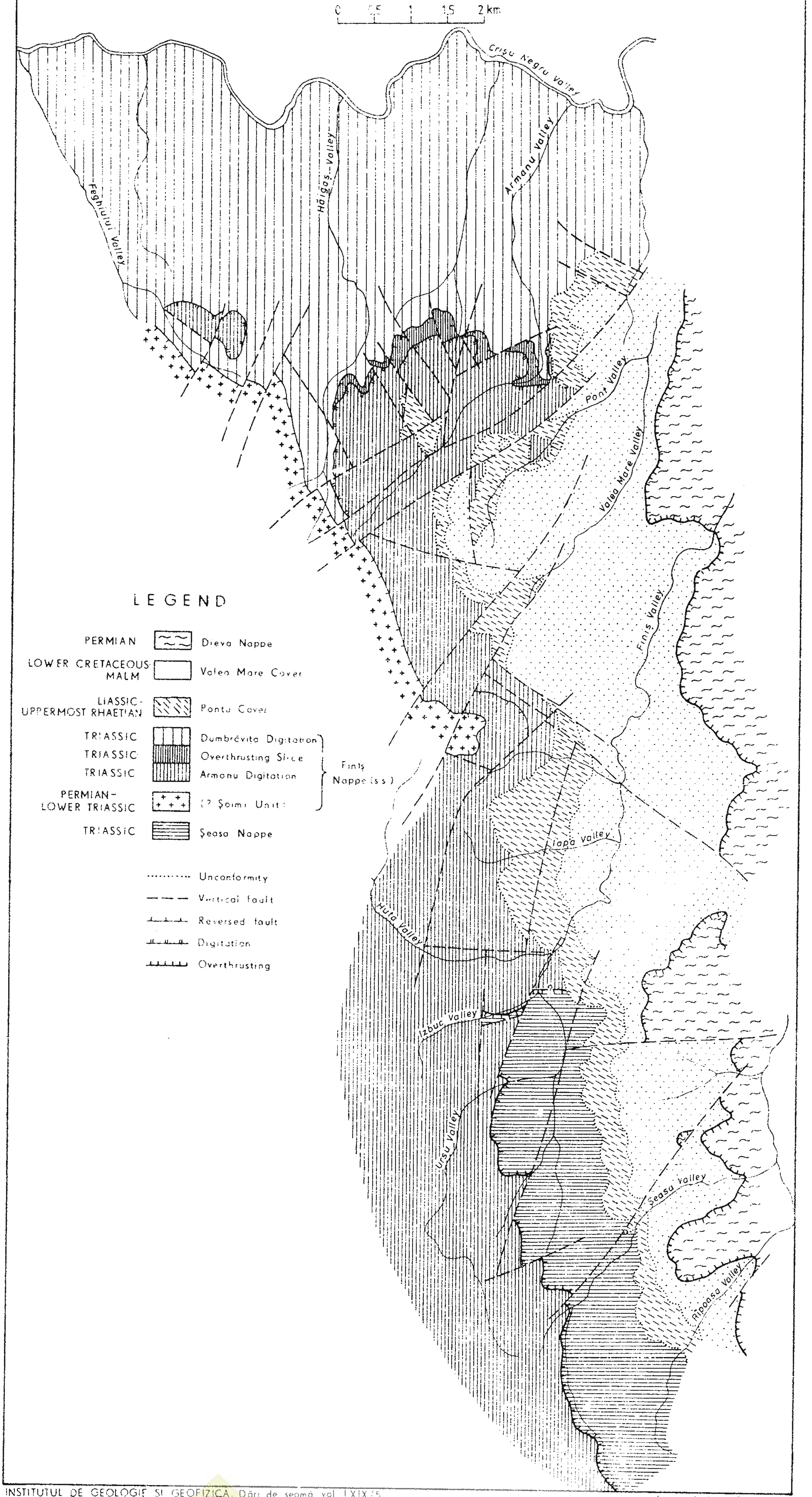
STRUCTURAL SIMPLIFIED MAP OF THE CRIŞU NEGRU - RÎPOASA VALLEYS AREA
(NORTH APUSENI MOUNTAINS)



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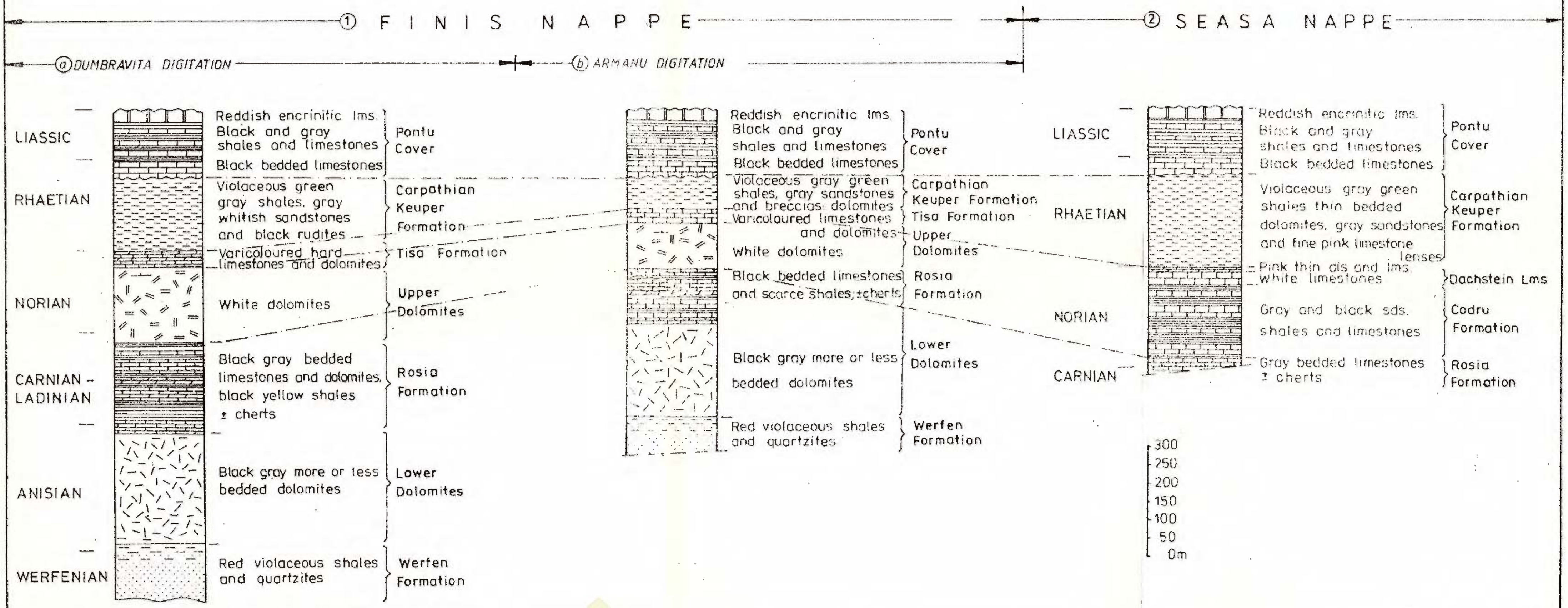
LITHOSTRATIGRAPHIC COLUMNS OF THE TRIASSIC-LIASSIC DEPOSITS OUTCROPPING BETWEEN CRIŞU NEGRU AND RÎPOASA VALLEYS



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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

STRATIMETRY AND STRUCTURE OF THE MATEIAŞ LIMESTONE [1]

BY

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Organogenous limestones. Upper Jurassic. Kimmeridgian. Tithonian. Stratimetry. Lithostratigraphy. Faults ; East Carpathians — Crystalline-Mesozoic Zone — Braşov Mountains.

Sommaire

Stratimétrie et structure du calcaire de Mateiaş. La note présente les trois composantes lithologiques qui ont pu être séparées cartographiquement dans la zone d'affleurement du calcaire de Mateiaş. Grâce à ces informations lithostratigraphiques détaillées on a mis en évidence une série de fractures verticales. A mentionner que quelques fractures ont joué des sens différents pendant diverses étapes de l'évolution de la région.

Introduction and Evolution of Researches

The calcareous formation, cropping out east of Cîmpulung-Muscel in the Mateiaş, Hula and Măgura summits, at Piatra and in the Albina Summit (Pl.), represents a strong relief, generally in contrast with the neighbouring one, which offered easy possibilities of its economic turning to account as limestone and building stone. Its dominant geographic position in the region, as well as the economic interest has drawn the geologists' attention to the Mateiaş Limestone, since the end of the last century. The results of the researches were presented in papers and reports by several authors, as follows : Ştefănescu (1884), Popovici-Hatzeg (1896-1897), Popescu-Voiteşti (1909), Ştefănescu et al. (1965), Botez (1967), Patrulius et al. (1968 b).

All the above-mentioned authors regard the Mateiaş Limestone as a massive rock displaying a quiet tabular structure. Our researches,

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however, pointed out that lithostratigraphic divisions can be made within the Mateiaş Limestone in the outcropping area between the spring zone of the Argeş Valley and the Dîmboviţa Valley. The minute following of those separations led to a structural image different from the previous ones. The aim of the paper is to describe both the lithostratigraphic units and the tectonics including the Mateiaş Limestone.

Stratigraphic Evidence

Substratum of the Mateiaş Limestone

The Mateiaş Limestone unconformably overlies crystalline formations both on its northwestern border and on its eastern one.

Mateiaş Limestone

The Mateiaş Limestone represents a calcareous formation, predominantly bioconstructed, of Upper Jurassic age (Kimmeridgian-Tithonian).

This type of formation crops out between the spring zone of the Argeşel Valley and the Dîmboviţa Valley (Pl.). It displays lithological variations on the vertical, which allow its division into three components named in stratigraphic order from α to γ (Fig.). Their description is based on the macroscopic studies, as well as on the preliminary analysis of the thin sections.

Component α. It consists of limestones, generally sparitic or micritic, massive, seldom stratified, white, yellowish, more rarely pink, which contain numerous coral colonies (particularly developed in the western part of the outcropping area), algae, bryozoa, gastropoda, and, in places, brachiopoda. Reefal breccia made up of a white, micritic matrix, including elements of bioconstructed limestone blocks, can be recognized in the Hula-Mateiaş zone. The thickness of component α varies from more than 300 m in the Hula and Mateiaş summits to 30 m in the area of the Piatra village or even to its disappearance at Colţii Albinei.

Component β. It is represented by biosparitic or pelsparitic limestones, yellowish, well stratified, occurring as beds of 5-25 cm thick. The presence of the yellow or brown, siliceous accidents, with irregular shapes is a further characteristic element of this component. The siliceous accidents evidently formed after the accumulation of their constitutive strata as their contours are independent both as against the limestone structure and the organic particles. In certain profiles the siliceous accidents are grouped into two distinct levels — at the lower and upper part of the packet under discussion — leaving between them an interval of about 10-15 m thick (east of the Mateiaş Summit), devoid of cherts and within which the stratification becomes more rare and lenses and beds of white limestones, in places with corals, are to be found. The thickness of the component β varies from 20 to 40 m.

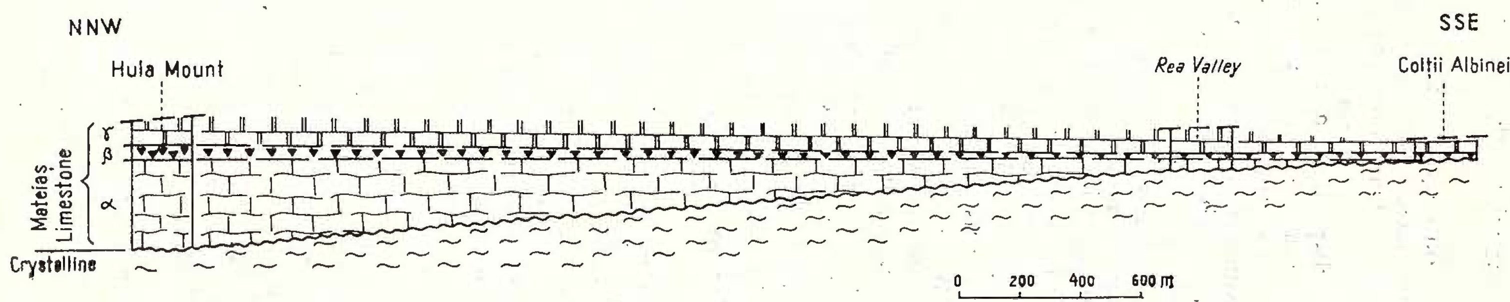

Fig. — Unfolded section of the Mateiaș Limestone between the Hula Mount and Colții Albinei.

Component γ. In its outcropping area it is mostly constituted of sparitic or micritic limestones, of a white, more rarely pink colour. On certain profiles a gradual transition from component β to component γ can be observed, in which cases the latter starts with a packet more stratified, in places displaying morphological aspects similar to mud-mounts. Isolated, arborescent corals have been found within this component in some places. The thickness of the component cannot be established accurately because of the combined effects of tectonization and erosion. However, it can be ascertained that it exceeds 100 m west of the Mateiaş Summit.

It is worth mentioning that the total thickness of the Mateiaş Limestone becomes gradually smaller from west to east, as indicated especially by the narrowing up to disappearance of component α (Fig.) as well as the thinning of component β. This type of development of the Mateiaş Limestone component indicates that east of its present outcropping area, there was an uplift zone and therefore the Mateiaş Limestone and the synchronous limestones of Piatra Craiului-Dîmbovicioara represent a calcareous ramp developed on the western flank of the Leaota massif. Concomitantly, on the eastern flank of the same massif, which subsided uninterruptedly during the Kimmeridgian-Tithonian time-span, there was another ramp where the limestones cropping out in the Bucegi massif nowadays had accumulated. The two ramps were separated by an uplift zone where deposits did not accumulate (probably in the central part of the Leaota massif) or they accumulated but in a condensed facies (e.g. the Poliţie succession, Patrulius, 1969), protected by reefal bars developed both eastwards and westwards.

The limestones cropping out in the Mateiaş Mountain and at Piatra were generally assigned by previous authors to the Upper Jurassic. An exception is Ştefănescu (1884) who considered them of Middle Jurassic age. The attempts to specify the age of the Mateiaş Limestone started in 1909 when Popescu-Voiteşti assigned them only to the Tithonian. Later studies did not bring further arguments so that on the Geological Map of Romania, scale 1 : 200 000, sheet Tîrgovişte (Patrulius et al., 1968 a) the Mateiaş limestones are also referred to the Upper Jurassic although on the neighbouring sheet (Braşov, Patrulius et al., 1967) the tantamount limestones were considered of Kimmeridgian-Tithonian age.

The study of the thin sections carried out in the whole pile of limestones brought information useful in stratigraphic respect, especially from the limestones of component β. Among the numerous forms recognized[3] in the mentioned limestones only those of stratigraphic importance will be mentioned : *Clypeina jurassica* Favre and *Macroporella pygmaea* (Gümbel) Pia, which indicate the Kimmeridgian-Tithonian time-span. Therefore these forms are in favour of the Upper Jurassic age of the limestones, moreover specifying that at least from compartment β upwards starts the Kimmeridgian-Tithonian interval. In our opinion compartment α would, however, belong to the Kimmeridgian, as well.

In different points of the outcropping zone of the Mateiaş Limestone (north of the Măgura Summit, east of the Mateiaş Summit, on the territory of the locality of Piatra) dolines with different diametres can easily be recognized, witnesses of a fairly intense phenomena of the limestone carstification during the Quaternary. Nevertheless, it is worth mentioning that carstification phenomena also occurred before the Upper Cretaceous and the Oligocene. This statement is based on the observations made in the fronts of the Hula Mountains quarry, where crevices usually filled up with Upper Cretaceous, grey-greenish, siltic marls and more rarely with Oligocene brown pelites with fish scales are to be found.

Cover of the Mateiaş Limestone

Upper Cretaceous. Since the Geological Map of Romania, scale 1 : 500 000, the Jurassic limestones in the study region are overlain by Upper Cretaceous deposits. Our evidence points out that both the Jurassic limestones and the crystalline schists are unconformably overlain by a mostly siltitic series resembling the Dumbrăvioara Series (Popescu, 1954). In places it starts with conglomerates, more or less calcareous, usually directly overlain by a packet of sandstones and grey siltic marls which passes gradually into grey siltites with remains of ammonites, inocerami and aucelinae, a fauna which made it possible to establish the Upper Vraconian-Cenomanian age of the deposits under discussion.

Oligocene-Lower Miocene. West of the main ridge formed by them, the limestones come into tectonic contact with the Oligocene-Lower Miocene pelitic deposits developed in the Valea Caselor Facies (Ştefănescu, 1976). Among the three members of the mentioned facies only the highest one — the upper horizon of the dysodilic schists — unconformably overlies the Mateiaş Limestones only in one point (in the old quarry north of "Monument").

Tectonics

The accurate mapping facilitated the finding out of vertical fractures, so that the structural image of the Piatra-Mateiaş zone becomes similar to that of the Dîmbovicioara Culoir (Patrulius, 1969), its southeastern extension (Popescu-Voiteşti, 1909).

Except for the Geological Map of the Zones of Oil Interest (1957) and of Botez' map (1967), on which a fault is drawn up between the Mateiaş Limestone and the deposits west of it, on all the other graphic representations no tectonic accidents occur in the Upper Jurassic formation area. The Mateiaş Limestone, beside its basement and cover, is affected by a large number of vertical faults with different trendings : NNW-SSE, NNE-SSW, E-W. Their throws vary from some metres to tens of meters. Among all the detected faults we consider it necessary to deal with some of them which are of greater interest.

The Piatra and Mateiaş limestones are limited to the west by a fascicle of faults approximately trending NNW-SSE. They generate a step-like sinking of the deposits towards WSW. It is worth mentioning that some of them had different directions of movement in different stages of the region evolution. Thus the upper limit of the crystalline schists of Valea Rea (left tributary of the Stoeneşti Valley, upstream of its confluence with the Dîmboviţa Valley) is slightly shifted by a vertical fault with a NW-SE trend. But while in the northeastern compartment the crystalline schists are overlain by the Upper Jurassic limestones, in the southwestern one they directly underlie Upper Cretaceous deposits. This situation can be explained only by a higher pre-Vraconian position (therefore a positive movement) of the southwestern compartment which might have permitted the removal by erosion of the Upper Jurassic limestones. Later on it underwent a reversed movement resulting in the sinking of the Upper Vraconian-Cenomanian deposits. A similar movement also seems to have taken place on the NE-SW fault, situated northwest of the Măgurele Summit, as north of this fault the Upper Cretaceous deposits directly overlie the crystalline schists.

In the Măgura-Mateiaş-Hula Ridge limestones form an assymetrical syncline, with a low-dipping northwestern flank and a high-dipping southeastern flank, the latter being straightened out to the vertical southeast of the Mateiaş Summit. The axis of the syncline corresponds to a fault which separates the two flanks. Therefore, it is to be mentioned that the Upper Jurassic limestones, besides numerous fractures, have also been affected by folding in places.

[3] Thanks are due to C. Tomescu for the determination of the quoted forms.

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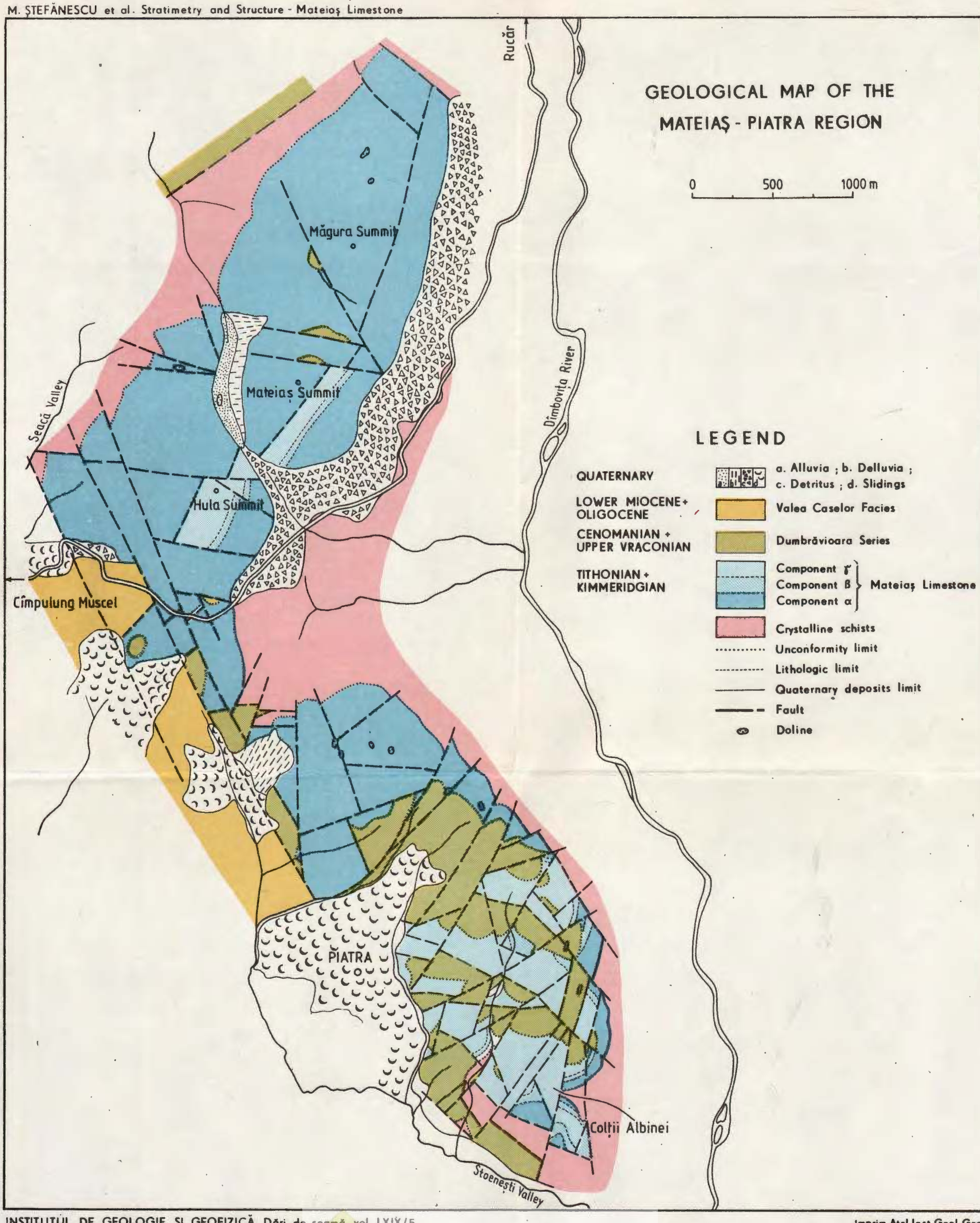

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

RECENZII — ANALYSES D'OUVRAGES

H. MARTIN, F. W. EDER (editors) : *Intracontinental Fold Belts.* Springer-Verlag Berlin, Heidelberg, New York, Tokyo, 1983, 945 p., 300 Figs., 24 Pls.

The volume "Intracontinental Fold Belts — Case Studies in the Variscan Belt of Europe and the Damara Belt in Namibia", edited by Prof. Dr. Henno Martin and Dr. Franz Wolfgang Eder, from the Universiy of Göttingen, represents the final report of the Gottingen Sonderforschungsbereich (SFB) 48 programme, sponsored by the Deutsche Forschungsgemeinschaft. Launched in 1969, the SFB 48 programme developed until 1980, enjoying the cooperation of German researchers in Earth sciences, some of whom representing outstanding personalities of the international scientific élite. Only three years after the conclusion of the researches, the widely known Springer Verlag offers the international public a rich volume in the English language, of almost 1000 pages, reuniting 41 contributions, written by one, two, or .... seven of the 48 participants to the programme. The SFB 48 Göttingen programme stands as a successful example of interdisciplinary cooperation, and the volume "Intracontinental Fold Belts" as a work of reference, vast and complex, amazing by the multitude of the data available, analytical to the extreme, but nonetheless full of interpretations, assumptions and reconstitutions, either of, sedimentary environments, tectonic evolutions, the physical conditions under which metamorphism occurred, or of the movement of some lithospheric plates. Thanks to the very high level of the authors, the contributions presented reveal and advanced professionalism, so that the lecture and especially the criticism of all the 41 papers suppose a geo-encyclopaedic knowledge quite difficult to meet.

Subjected to the concentric firing of prominent sedimentologists stratigraphers, tectonicians, geochemists, specialists in isotopic geology, geophysicians, mineralogists, field or experimental petrologists, the Variscan orogen in Germany and the Damara one in Namibia (South-West Africa) are shown dissected to their most intimate details, to be reconstituted thereafter under the form of evolutive models, characteristic of the topic of every particular paper, but also generalized, from a geodynamic standpoint, in the works ending each of the two cycles of articles. Another feature, common to the large majority of these reconstitutions — the present structure of the respective territories — is due to the important spatial changes occurred during hundreds of millions of years under the influence of the divergent or convergent drift of the lithospheric plates. Therefore although the terminology of the geosynclinal orogenesis is still surviving in a few papers, that of the global tectonics is very intensely used in case of events having taken place in the Upper Precambrian or the Paleozoic, numerous figures illustrating the views of the authors on riftogenesis, sea floor spreading, subduction, subfluence, delamination, etc. It results clearly that the programme, named at its launching, at the dawn of the revolutionizing of the geosciences by plate tectonics, "Evolution, composition and distinctive characteristics of the Earth's crust, particularly in geosynclinal regions", fully benefited from the innovating ideas of the '70s, adapting itself on the move to the mobilisic conceptions. With nuances dictated by the specific field of interest and the personality of the authors, the lecture of the volume reveals the fact that the Göttingen school is under a full process of reconsidering, from a dynamic and mobilistic standpoint, certain parts of folded belts so earnestly scrutinized as the Variscan belt in Europe, or more recently come into attention of the geologists — the Damara belt in South-West Africa. Warm congratulations, not only for this aspect, but also for the tremendous field, laboratory and library work, necessary to the successful matching of such puzzle pieces, having disseminations of hundreds of kilometres, and observed over hundreds of millions of years.

T. Berza

D. SHIMOZURU and I. YOKOYAMA (editors) : *Arc Volcanism and Tectonics.* Proceedings 1981 IAVCEI Symposium, Tokyo and Hakone, 1983, 263 p. Terra Comp. Tokyo and D. Reidell Comp. Dordrecht.

The proceedings of the symposium called "Arc Volcanism", held in Japan at Tokyo and Hakone, took place under the auspices of the Volcanological Society of Japan and of the IAVCEI. The book edited by the well-known volcanologists D. Shimozuru and I. Yakoyama includes 15 papers which were presented at the symposium. It is worth mentioning that this important international scientific event, that takes place every four years, was held also in Romania in 1973. The symposium in Japan debated mainly problems connected with the circumpacific "fire ring". These problems were approached in the light of the remarkable progress recorded in the last 20 years in the interpretation of the tectonic aspects of the arc volcanism. The present collection of articles is divided in two parts according to the subjects treated.

The first part comprises 7 papers dealing with problems of the earth physics applied to volcanology and problems of modelling the volcanic activity. The second part consists of 8 papers in which the authors refer to the magmatism-volcanism relation and the subduction processes. Each paper includes figures, tables and a list of references.

Some of the papers presented in the book are of a great general interest : A Model of Eruption Sequence and Magma Supply Rate for Polygenetic Volcanoes ; Thermal Activities of Volcanoes in the Japan Arc — A Nature and Geological Meaning ; A Theoretical Model for Dispersion of Tephra ; Volcano Spacing and Subduction ; Fire-Arc Volcanism and Cycles of Subduction.

By the problems of great present interest related to the theoretical and practical volcanological researches, this collection of papers is of use to the specialists in our country, allowing a more profound knowledge of the arc volcanism problems with their theoretical and practical implications on the one hand, and the knowledge of the way in which several problems are approached, such as physical modelling and volcanology with implications in the prediction of eruptions and the geothermal energy associated with the volcanism arc areas on the other hand.

S. Peltz

F. J. SAWKINS : *Metal Deposits in Relation to Plate Tectonics.* Springer-Verlag, Berlin-Heidelberg-New York-Tokyoo, 1984, 325 p., 173 Figs.

One of the main valences of the global tectonics is that it comprises all aspects of the geological sciences, the geoeconomic ones inclusively. This fact has been recognized earlier and the possibility of employing the new concepts in the interpretation of the genesis and the distribution of the metal deposits have been forseen since 1970. The accumulation of data made possible the appearance of volumes on this subject, e.g. the one edited by D. F. Strong in Canada in 1976, or that edited by Tarling in Great Britain in 1981. Such books include, however, papers dedicated to certain matters and, consequently do not offer a general view on the general subject as in case of the author papers. Among the latter there are only two in the World's literature, the one written by A. H. G. Mitchel and M. S. Garson, issued in 1983, and the present paper, issued a year later.

Sawkins' book is well-structured. It is divided into three parts : Part I — Convergent Plate Boundary Environments, Part II — Divergent Plate Boundary Environments, Part III — Collisional Environments and Other Matters. Each part is subdivided into several chapters e.g. Part I — (1) Principal arcs, (2) Inner sides of principal arcs, (3) Arc-related rifts, (4) Arc-related metallogeny ; Part II — (5) Ocean-type crust, (6) Intracontinental hotspots, (7) Early stages of continental rifting, (8) Advanced stages of rifting.

In each tectonic setting are included various deposits of the world, well-known and studied, which in classical books on metallogenesis have been grouped on different criteria (liquid-magmatic, pisometasomatic, hydrothermal, etc. deposits). This time they are closely related to the tectonic context, their association being made according to the dynamics of the crust in that place. For each type of deposits are given the petrographic characteristics, the genetic relationships between them and the metal deposits, as well as suggestions on the research work.

The global tectonics has not succeeded in clearing up why one or another metal accumulates within a certain dynamic context and which are the rules controlling their association but it is not a drawback of the global tectonics only, but of all the metallogenetic theories. However, the new system of knowledge has reached a better understanding of the geological and geochemical processes which might lead to economic concentrations of metals as well as to a full understanding of the dynamic evolution of the crust, of its chief tectonic subdivisions which proved to be fundamental for all the geological and geophysical aspects of the world.

Sawkins' book is a good example of how a discipline which seemed to have been fundamentally established can be entirely reversed and described with entirely new coordinates, may be less proved scientifically but more inciting for the new perspectives offered by it. It is a captivating book by its novelty and inciting by its suggestions.

M. Bleahu

K. FUCHS, K. von GEHLEN, H. MALZER, H. MURAWSKI, A. SEMMEL (editors) : *Plateau Uplift — the Rhenish Shield — A Case History*. Springer-Verlag, Heidelberg, New York, Tokyo, 1983, 441 p., 185 Figs.

This volume represents, as a matter of fact, a monography of the Rhenish Massif as it includes a multi-disciplinary research of this massif.

In the beginning, the regional geotectonic context is presented, the Rhenish Massif being regarded as an element in the foreland of the Alpine collision front and its relation to the tectonics of the Upper Rhine Graben, emphasis being placed on the interaction between uplift, rifting, volcanism and mantle heterogeneity.

Further to the study on Pre-Quaternary geological history the elements signifying old uplift movements, e.g. transgressions, are reviewed.

A special chapter is on the volcanic activity for which the Middle Cretaceous-Lower Miocene time span represents high intensities. The present-day features of the Rhenish Massif are analysed also in view of the seismic, gravimetric and geothermal results, the conclusion being inferred that at present, as during all its geologic evolution, this territory does not behave as a rigid and unitary block but also different fragmentary.

Geological investigations are studied in a chapter which deals with the structure of the crust and mantle, the physical properties, the relations between the geophysic and petrologic models of the upper mantle.

As a result of this combined study the authors infer the existence, within the Rhenish Massif zone, of a mantle diapir.

The paper is well structured based on investigations, sometimes of detail, and is intended to represent, in the plate tectonics epoch, a pleading for the epirogenesis reconsideration.

M. Lupu

FISCHER R. V., SCHMINCKE H. U. : *Pyroclastic Rocks*. Springer-Verlag, Berlin-Heidelberg-New York-Tokyo, 1984, 472 p.

The authors of this book are well-known specialists due to their contributions brought in recent years to the complex study of the products of explosive volcanism, to the improvement of the method of classification of clastic volcanic rocks.

The present work is divided into three parts : Part I (chapters 2-4) briefly presents the volcanics, volcanic rocks, physical properties of magmas, eruptive and clastic processes. This part represents a basis for the understanding of the origin of the deposits discussed in Part II (chapters 5-11, pyroclastic deposits, fragments, pyroclastic flows, submarine pyroclastic rocks and hydroclastic, lahars). This part also includes a chapter dealing with the volcanic glass alteration, the main constituent of the tephra (chapter 12). Part III (chapters 13, 14) describes the stratigraphic relationships and the tectonic implications related to the emplacement of the pyroclastic rocks.

The book refers not only to proper pyroclastic rocks as it includes discussions on rocks varying from lahars, formed by the remobilization of non-cemented superficial fragments, to breccias of pillow lavas occurring as a result of the fragmentation of pillow lavas coming from submarine eruptions. Eruptive and clastic materials formed under the earth crust — e.g. breccified intrusions and epiclastic materials coming from alteration and erosion of preexistent volcanic rocks.

The present book represents the first monography dealing, in a modern and comprehensive manner, with pyroclastic rocks, their classification and way of formation. The topics are discussed in view of the plate tectonics theory and the achievements in volcanology, sedimentology and magmatic petrology.

Magmatic volatiles, pyroclastic processes as well as magma-water interactions are studied in detail. Most of the book deals with the wide range of pyroclastic processes which took place at surface and underwater as well as a result of different mechanisms of flow and emanations. Diagenetic processes by means of which pyroclastic particles are altered into rocks are described in detail, too.

The authors illustrated the tectonic and stratigraphic significance of pyroclastic rocks.

The book has a very rich reference list and a subject index, very useful in the study of similar topics.

In the Romanian territory the pyroclastic rocks are widespread as a result of an explosive volcanic activity which developed during several periods of the geological history. In numerous situations pyroclastics participate in the geological constitution of areas of interest for different metalliferous and non-metalliferous mineral resources. Therefore the thorough study of volcanoclastic products — the present book being included — is of interest for Romanian specialists in volcanology as well as in metallogeny and in prospection, exploration and mining works.

S. Peltz

Geoscientific Aspects of the Radioactive Waste Deposition. Deutsche geologische Gesellschaft, Enke Verlag, 1980, 219 p.

The book, as a matter of fact an issue of the German Geological Society (Zeitschrift der deutschen geologischen Gesellschaft, vol. 131, p. 339-340) also printed as a volume, comprises the papers and discussions of a symposium of the society on the geological problem of the radioactive waste, held in Braunschweig, on November 1979.

Two inaugural speeches, made by the chairman of the German Geological Society and the chairman of the Federal Institute for Geosciences and Raw Materials, point out the importance of the scientific session. The first paper deals with the notion of radioactive waste and the source of the radioactive waste : from fission products in the thermo-nuclear station, from industrial, medical and scientific research uses. The waste is concentrated and immobilized in final products which have to constitute multiple barriers against the return of the radioactive nuclides in the biosphere. With this aim in view they must be deposited in certain geological formations ; the author of this paper states the conditions which are to be met by these formations considering the radioactivity intensity time of its loss and the heat which might be released by the deposit and which might bring serious perturbations to the environment. The second paper presents the principles according to which the waste deposition should be made and puts forward to discussion the principles on the basis of which the materials with warm activity (without alpha radiations and less than 1000 years of isolation) and those with hot activity (with emission of alpha radiations and with an annihilation time of 10^5-10^6 years) have to be deposited. From another point of view there are separated the deposits which might be formed in the continental crust — cavities dug into salt, anhydrite, clayey fields, garnet, basalt and tuffs — and in ocean floor deposits, in which one has to choose zones with reduced biologic activity, high tectonic and climatic stability, and without mineral resources whose extraction should be prevented in the future. The deposits may also be regarded within the political views of the respective countries or they may be regarded on a world scale. In the latter case deserted regions are chosen, remoted from any human dwelling, where the deposits may be formed in the continental crust. However, it seems that the best solution for the future huge deposits is that of burying them in the ocean floor in special containers. It is considered their location in the northern part of the Pacific plate, where the conditions of stability and isolation are best realized.

The other seven papers deal with the possibility of formation of deposits in the salt formations from West Germany. There are discussed the conditions of tectonic stability, the geochemical processes which might be generated by the waste, the changes brought about by the temperature rise as well as the technical difficulties implied. An important factor is the diapir form of the salt formations and the seismic field of the area.

All the questions discussed in the papers are of great importance taking into account the increasing development of the nuclear industry and especially the ever increasing stress laid on the nuclear energy. The main idea inferred from this book is that the deposition of the radioactive waste is first of all related to geology and only on the basis of the geological factors technical solutions can be adapted (nature of the smaller immobilization, kind of containers, depth of deposition, etc.).

M. Bleahu

ALLAN R. ROBINSON (coordinator) : *Eddies in Marine Science.* Springer-Verlag, 1983, 609 p.

In the central oceanic zones water dynamics is governed by eddies. They revolve round a quasivertical axis, affecting the whole water column. Eddies possess transport capacities and generate the mixture of the water bodies. Their sizes are of the order of tens or hundreds of kilometers, and their revolving period lasts weaks or even months.

Eddies obviously represent a highly important factor for the marine environment. However, in certain zones eddies are not studied and their general understanding is incomplete. The volume "Eddies in Marine Science", published by Springer-Verlag, coordinated by Professor A. R. Robinson, represents a synthesis of the knowledge on this major marine factor. It has 609 pages, numerous figures and tables, an ample reference list as well as a subject index. The book is divided into several parts : introduction, regional data (kinematic, dynamic and statistic), models of eddies, and a final part referring to effects and applications. Except for introduction (by A. R. Robinson) each part represents a gathering of papers worded by several authors.

The "Regional Kinematics, Dynamics and Statistics" — the largest part — presents researches carried out in the Pacific, Atlantic and Indian oceans, as well as the Antarctic. The last two contributions represent syntheses and global comparisons of knowledges on eddies.

The second part of the book ("Models") includes two contributions which represent numerical, periodical and regional models of the eddies.

The final part ("Effects and Applications") is of great importance. It comprises subjects on the effects of eddies on climate, littoral zone, on the spreading and mixture of marine water bodies, as well as on marine biologic and acoustic processes.

The volume "Eddies in Marine Science" deals with the modern knowledges of this domain, synthetized by the Working Group on the marine internal dynamics, constituted within the Scientific Committee on Oceanic Research (SCOR).

Publishing this volume the Springer-Verlag put at the disposal of those interested in the marine field a valuable synthesis, which constitutes the basis of understanding and utilization of the eddies.

D. Jipa

IDA K. and IWASAKI T. (editors) : *Tsunamis : Their Science and Engineering.* Terra Scientific Company/Tokyo; D. Reidel Publishing Company, Dordrecht, Boston, London, 1983, 563 p.

This book comprises the most important papers presented at the International Tsunami Symposium held in Japan in 1981.

The main object of the editors of this book was to give complex and up-to-date information on the degree of knowledge and prevision of the tsunamis reached by the scientific research as well as on the degree of improvement of the engineering technique and methodology of prevention of tsunami hazards.

The papers were arranged in seven major topics :
1. The Tsunami Impact on Society (2 papers) ; 2. Tsunami Source and Earthquake (6 papers) ; 3. Historical and Statistical Studies of Tsunamis (9 papers) ; 4. Tsunami Generation and Propagation (4 papers) ; 5. Topographic Effects on Tsunami Waves (7 papers) ; 6. Sea Walls and Breakwaters (4 papers) ; 7. Tsunami Run-up (7 papers), followed by the symposium programs, list of participants, author index, subject index and preceded by editors' preface and welcome and opening addresses.

The first three chapters include studies carried out with modern technique of calculation, using statistical and spectral analyses, which permit an exact determination of location and parameters of marine seisms as well as a rapid estimation of their tsunamic potential with a view to a fast and efficient tsunami warning of coastal inhabitants.

The knowledge of the tsunami dynamics, presented in chapters 4 and 5 as mathematic models of the process of their generation and propagation, pointing out the implications of tectonics and topography of the continental shelf in this process, leads to the possibility of foreseeing the manifestation of these waves offshore and on coastal zones.

Estimation by numerical and experimental methods of the characteristics of propagation of the tsunami wave on land and in shallow waters, measurements of warning and their efficiency are presented in detail in the last two chapters. The systematic measurements of tsunami in the open ocean for research purposes and for reliable tsunami warning is still not accomplished. Further technical improvements are for fast and efficient tsunami warning, and tsunami risc estimations are required for the prevention of tsunami hazards to coastal inhabitants and important construction in coastal regions of the Pacific and other oceans.

This book will contribute to increase international development of scientific and technical knowledge of tsunamis and to serve to protect mankind from the ravages of destructive tsunamis.

This book is interesting for specialists in seismology, hydrology and constructions.

E. Spânoche

HAROLD D. FOSTER : *Disaster Planning. The Preservation of Life and Property.* Springer-Verlag, New York, Heidelberg, Berlin, 275 p. Figs. and 15 photographs.

Newspapers contain more and more information concerning events dangerous for people's life and property, which are called disasters. Such events are generally accepted as unforeseeable and are impossibly to fight against, both in case of natural events (earthquakes, earth slidings, cyclones or tsunami) and in case of man made events (accidents in nuclear plants, high mercury pollution, building collapses etc.).

Actually, any human decision implies a risk and therefore a disaster probability. This is obvious both for everything built by man, and for the relationship between man and environment, decisions taken in a certain natural context being drastically modified by natural events which cannot be controlled by man. Generally, in case of a disaster, emergency measures, which are always off-hand, are taken, as well as improvisations and efficient solutions, imposed by local and momentary conditions. The result is not always satisfactory as expenses are too high the time loss is large and efficiency is rather low. In order to improve this situation, H. D. Foster gives several solutions, showing that in spite of the fact that disasters are unforeseeable, measures to be taken can be planned. Thus we need a preliminary detailed study and to set up an organization of efficient fighting against effects. This is the author's aim, which he achieves in eight chapters.

The first chapter contains general elements : identification of zones which are endangered by catastrophic events (hazards as they are called in the Anglo-Saxon literature), tolerable levels of risk, evaluation of the relationship between expense and benefit in order to overpass these levels, potential strategies to be adopted and adoption criteria of alternative strategies. After setting up these principles, there follows the analysis of elements of an intervention programme in case of disasters. The first element is that of classifying the risk (identification of hazard zones of high frequence and intensity of events, establishment of some risk standards and classification of the total risk). The risk classification can be made on independent elements (e.g. maps of seismic risk) or totally, for all possible hazards. Thus, there are 42 identified possible accidents, of which half are natural (avalanches, earthquakes, floods, forest fires, hurricanes, tsunami, volcanic eruptions etc.), and the other half is because of man (directly or indirectly, as chemical contaminations, dam breakings, intense water or air pollution, railway catastrophes, tanker sinkings, mining disasters etc.). The probability of such events can be evaluated for every place in the world, considering if it is possible, improbable, with low, middle or high probability or certain. By summing up these data, it is obtained a total risk value which is the basis for maps of isolines of total risk on which there can be elaborated the security plans.

The next chapter deals with disaster foresight. As a branch of the future science, the disaster foresight is treated according to its common techniques and mainly with the elaboration of some models (on scale, analogous or mathematic — on computer), the establishment of alternative scenarios, the Delphi technique etc. An important part in actions of fighting against disaster effects is played by alarm systems, which are dealt with in a special chapter. It speaks about the advance that can be taken, the danger of false alarms, duration of alarm, the implied legislation and educational problems, the factors leading to the decision to alarm, transmission of alarm messages. Finally, the most important part is that of disaster planning, a chapter presenting different scenarios, organization of operations according to scenario, organization of commandments, of assistance (sanitary, medical, supplies, shelters), restoring of communal services, the way of improvement (for example a new town planning formula in a destroyed town), problems of aesthetics, ethics and sociology implied by resuming the normal life.

As it can be seen, the book is a very general guide where, by eliminating details (even the disaster cause) it is discussed the element common for all events, having in view only the framework in which efficient and economical measures can be planned. Undoubtfully, no country can consider itself out of hazard risks, this being valuable as well for smaller communities (districts, towns, plants, mines etc.). Therefore, the attentive examination of this book is recommended for everybody in order to think in due time about measures to save life and properties with lower efforts and expenses.

M. Bleahu

Geomorphology of Europe — General editor : Clifford Embleton Verlag Chemie, 1984

This book has been compiled and written by members of the Commission on Geomorphological Survey and Mapping of the International Geographical Union.

The regional division of Europe employed in this book is based mainly on structure and to a less extent on relief. The book represents a considerable effort of international cooperation, and shows above all how much remains to be achieved in understanding the geomorphology of what is probably the world's most complex continent as regards the relief. The various chapters that comprise Part I are concerned with broad aspects of the relief and structural patterns, the evolution of these patterns through time and the formative processes controlling their evolution. In Part I a considerable attention is devoted to morphostructure. This term, introduced by Gerasimov (1946) refers to a structural unit expressed in the relief, modelled by denudation and/or sedimentation. The degree of modelling depends on the tectonic activity of the area and on its climate. A hierarchical system of morphostructures is recognized ranging from the largest 'megamorphostructures', such as the Fennoscandinavian Shield and Russian Platform, to micromorphostructures, such as a fault-controlled valley perhaps only a few tens or hundreds of metres in size. Three major elements are considered : (1) the differentiation of relief classes according to relief amplitude, (2) the differentiation of so-called 'special relief forms' of mainly exogenic origin.

The characteristics of the ocean floor relief which borders Europe are presented in a separate chapter, taking into account the modern theories which separate morphostructure on the world scale. The particularities of the submarine continental margins, the ocean floor relief, transition zones and medio-oceanic crests as well as the main features of the ocean floor are also presented.

The first part of the paper ends with the basic ideas on the classification of the exogenic landforms, relationships between fossil and contemporary morphoclimatic zonation, spatial distribution of exogenic forms and zonation of weathering processes, the mechanism generating the carstic, glacial, antropogenic fluviatile relief and the exogenic processes operating on ocean and sea floors and on coasts.

The main part of the paper is concerned with great morphostructural units. Using the papers written by well-known authors (Embleton, Joly, Aseev, Dimitrashko, Demek, Sala, Sestini, etc.) the geomorphology of several areas is treated in detail : Iceland, Fennoscandinavian Shield, Russian Platform, Caledonian Highlands, West and Central European Lowlands, Hercynian Europe, the Alps, Pyrenees and Ebro Basin Complex, Iberian Massif, Baetic Cordillera and Guadalquivir Basin, Appennines and Sicily, Carpathians, Balkan Peninsula, Northern Black Sea Lowlands and Crimea, Caucasian Mountains and Armenian Highlands, Ural Mountains. Likewise the main features of the submarine morphology around Europe are also described.

The book is richly illustrated with sketches, block diagrams, morphologic profiles, maps, photographs and is accompanied by a rich reference list.

C. Ghenea

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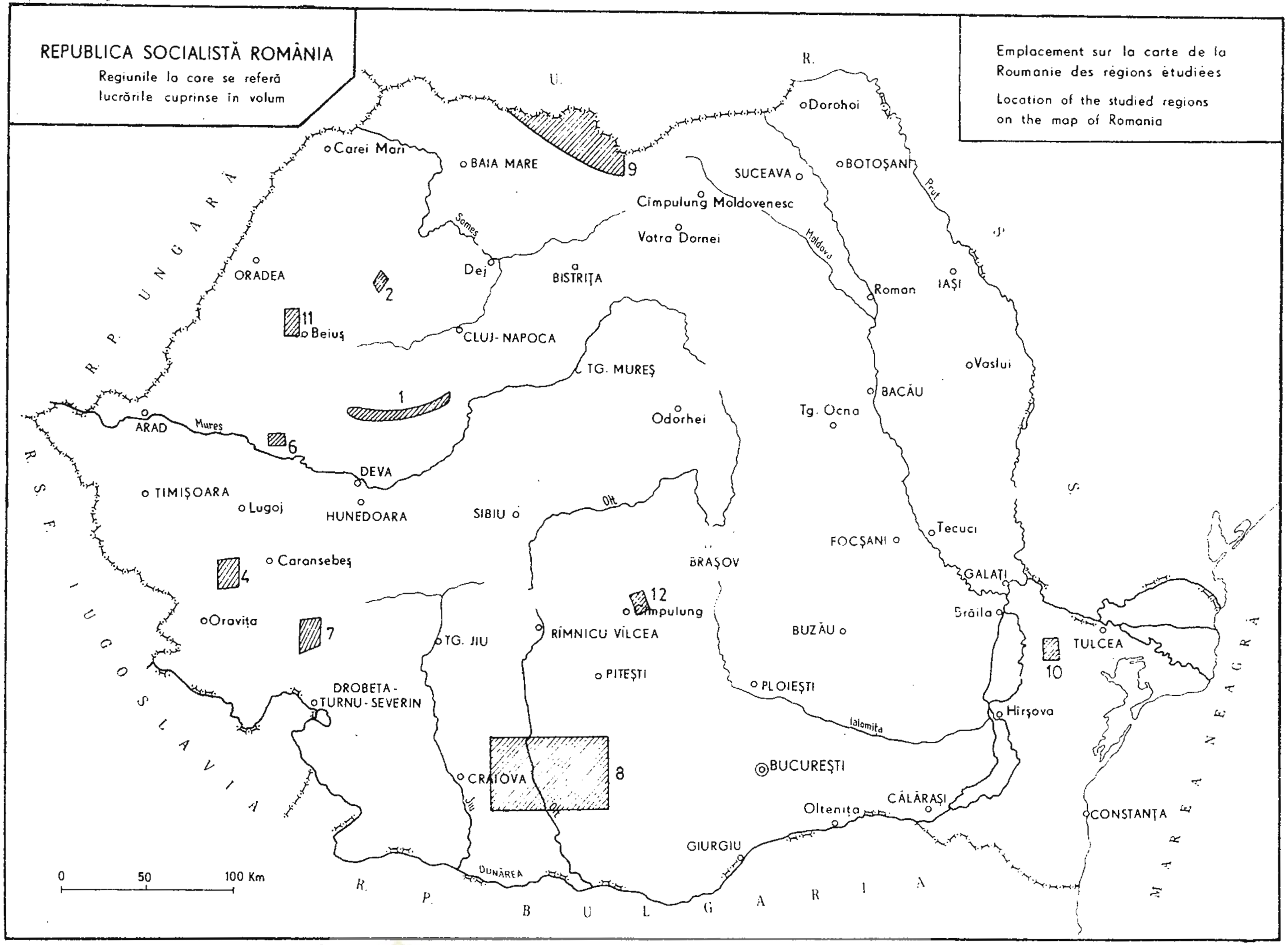
REPUBLICA SOCIALISTĂ ROMÂNIA
Regiunile la care se referă lucrările cuprinse în volum
Emplacement sur la carte de la Roumanie des régions étudiées
Location of the studied regions on the map of Romania
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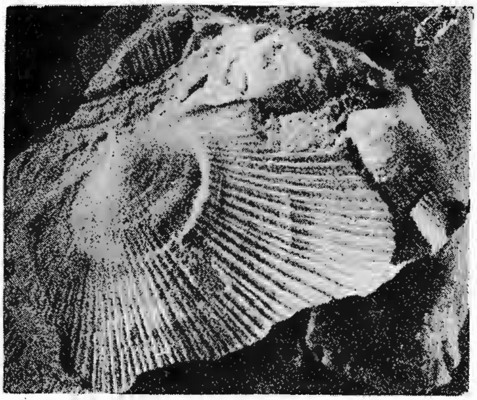

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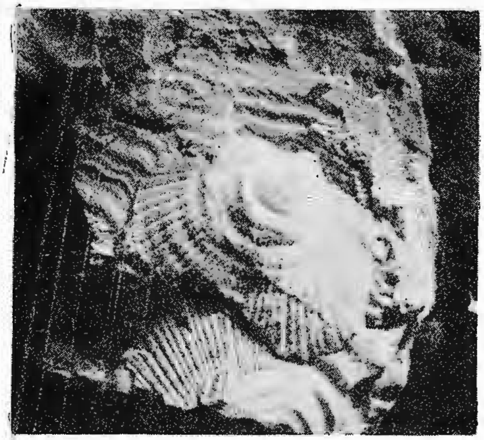

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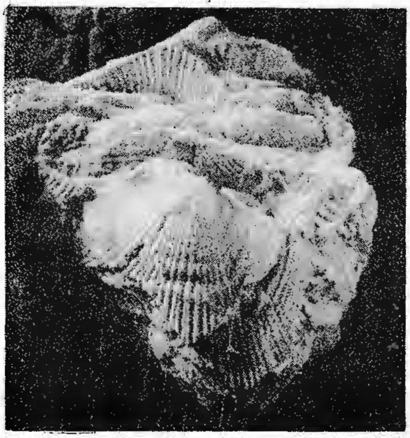

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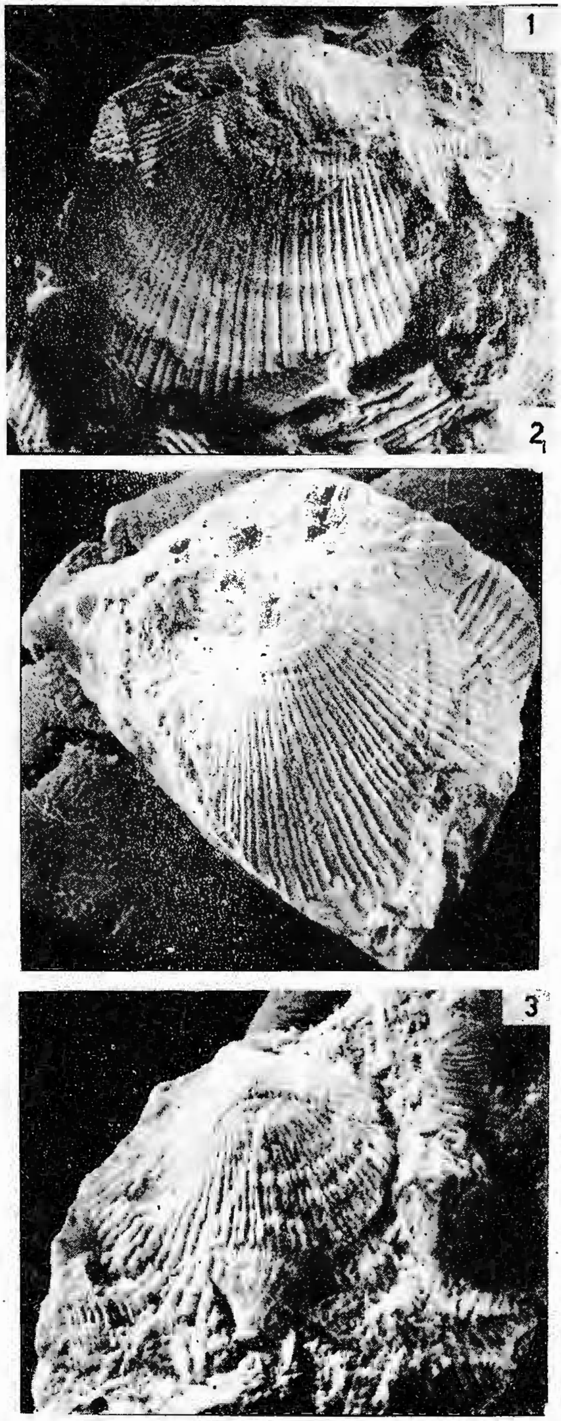
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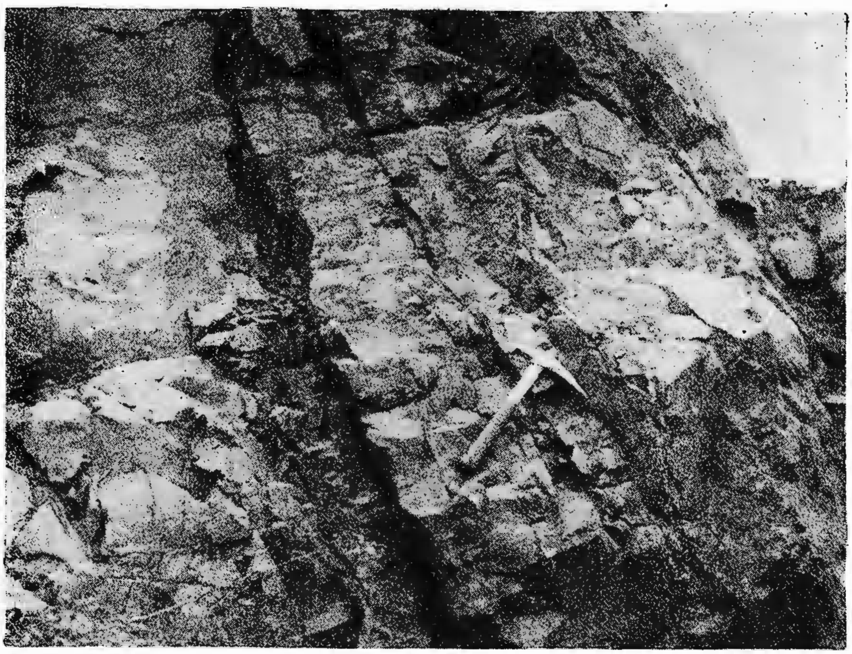

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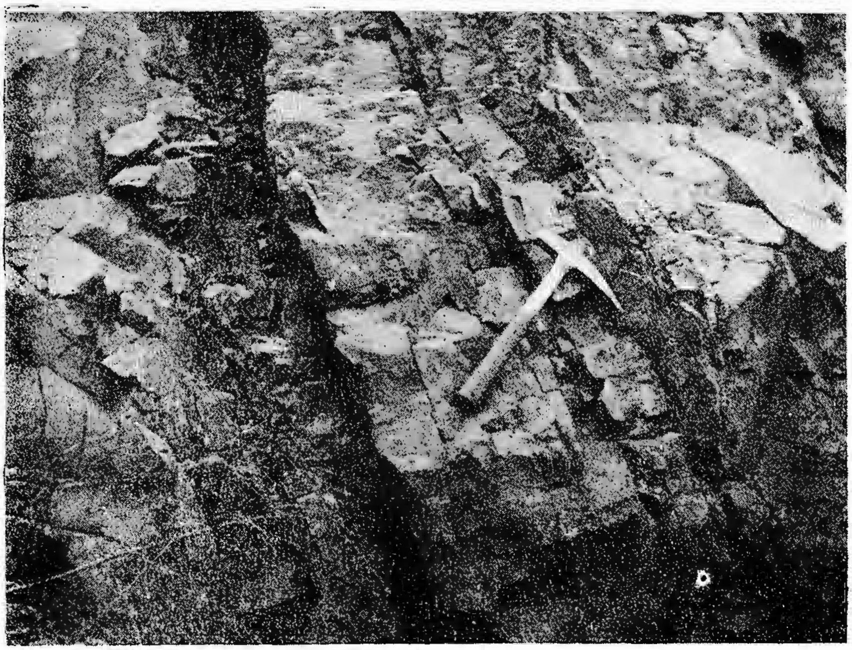

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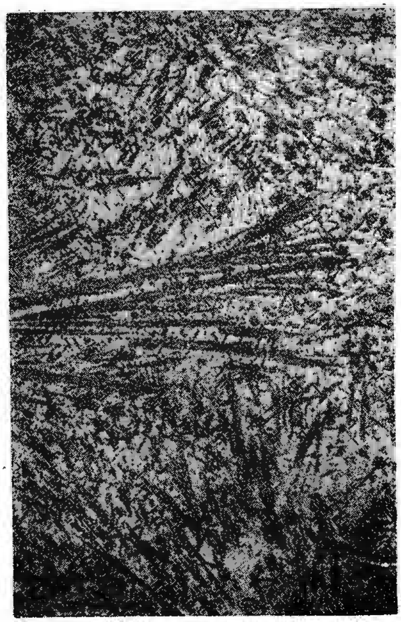

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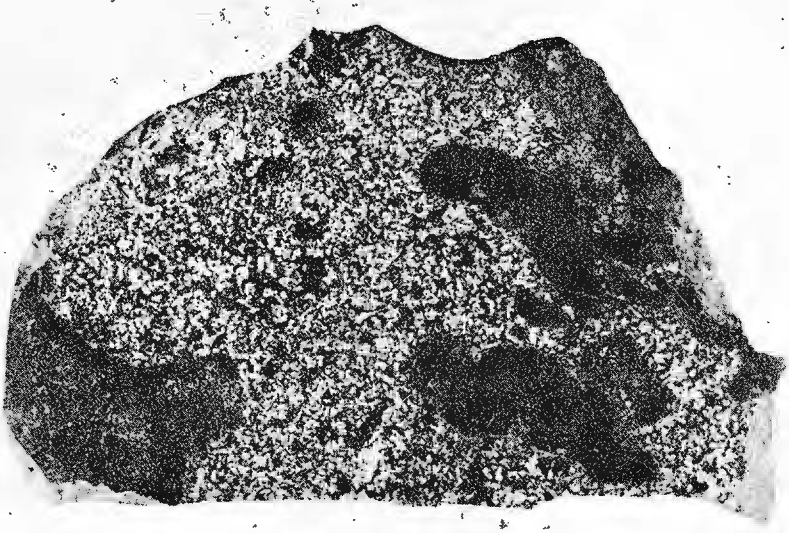

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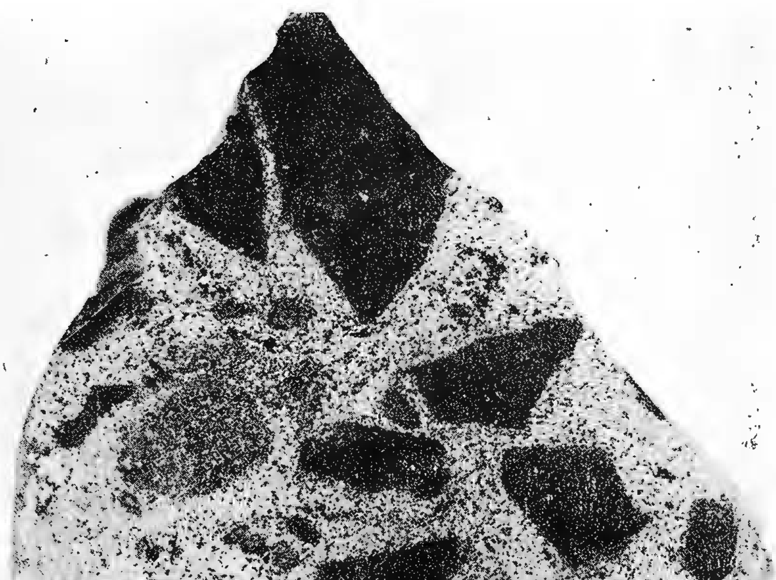

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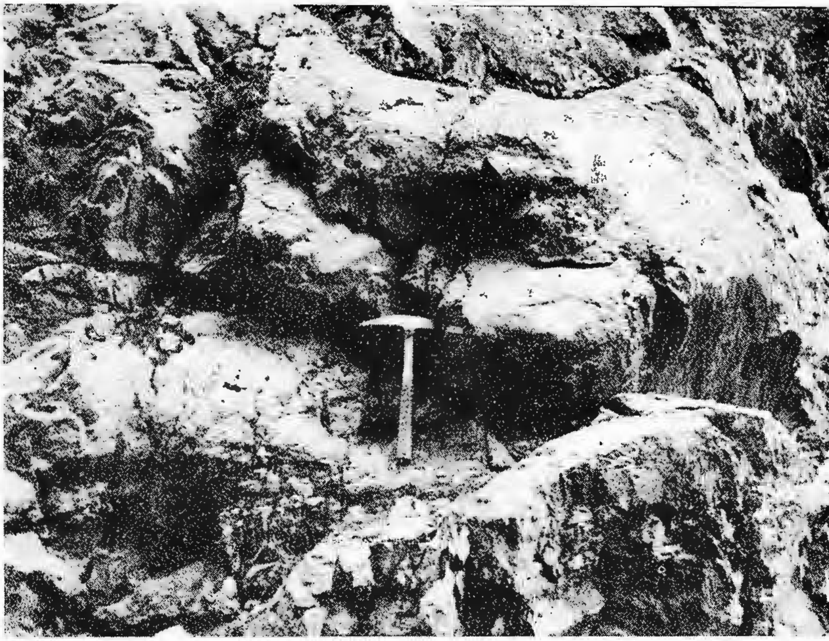

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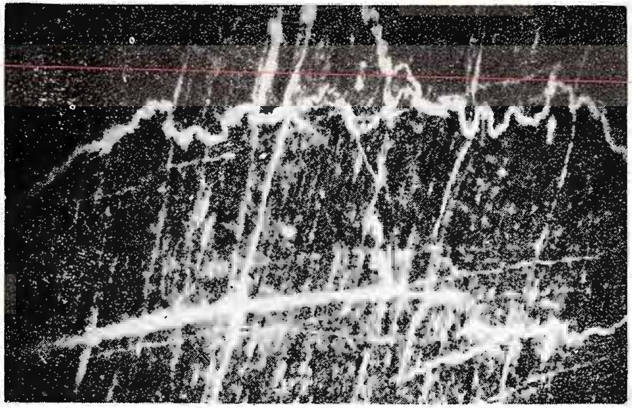

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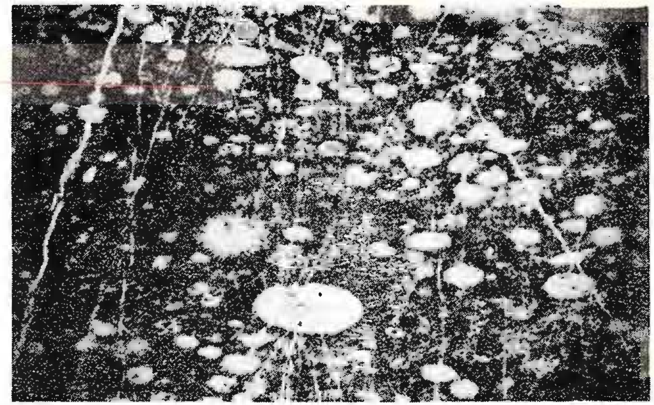

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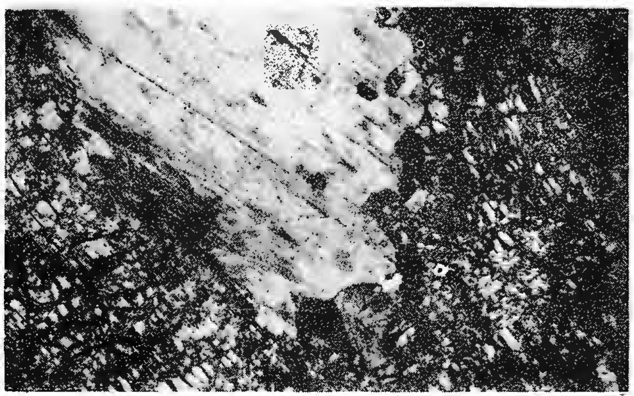

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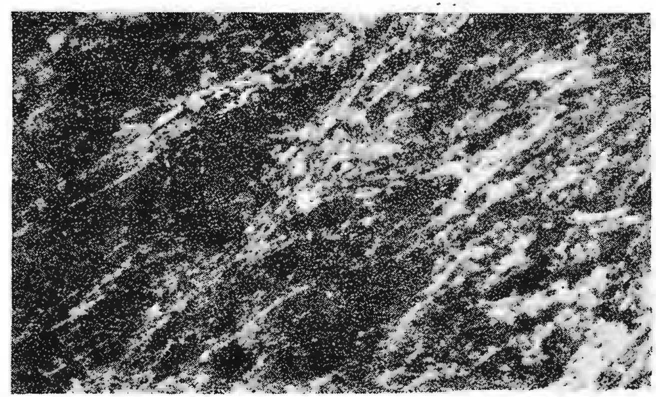

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