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LES FORMATIONS CRISTALLINES DES MONTS DU PETREANU ET ŢARCU SEPTENTRIONAUX (CARPATHES MERIDIONALES)

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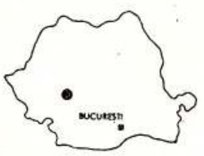


Key words: Danubian. Crystalline series. Granitoid rocks. Tectonic units. South Carpathians. Petreanu-Ţarcu mountains.

Abstract: *The Crystalline Formations of the Petreanu and Ţarcu Mts (South Carpathians).* The Lower Danubian (Petreanu-Rof Unit) consists of the Nisipoasa and Bodu crystalline Series, intruded by Petreanu metagranitoids and the Vârful Pietrii granitic massif, as Precambrian basement, covered by the Mesozoic low-grade Schela (Vidra) and Oslea (Poleatcu) Formations. The Upper Danubian nappes consist of the Poiana Mărului Unit (Godenele Crystalline Series), the Muntele Mic Unit (Măgura Marga Crystalline Series and Muntele Mic granitoid massif) and the Măru Unit (Zeicani Crystalline Series, covered by similar Mesozoic low-grade formations). The Getic Nappe is represented only by Lotru medium-grade metamorphics.

Les travaux géologiques entrepris par un des auteurs dans la partie septentrionale des monts Retezat et Petreanu (Dimitrescu, 1986) ont été consignés sur la feuille 1:50000 Retezat (Berza et al., 1989). Le prolongement occidental des structures de la région de Râul Mare (Berza et al., 1983; Dimitrescu, 1986) a demandé la continuation de ces études, utilisées à l'élaboration de la feuille 1:50000 Petreanu; leurs résultats sont éxposés dans la présente note.

I. Danubien inférieur

1. Unité de Petreanu-Rof

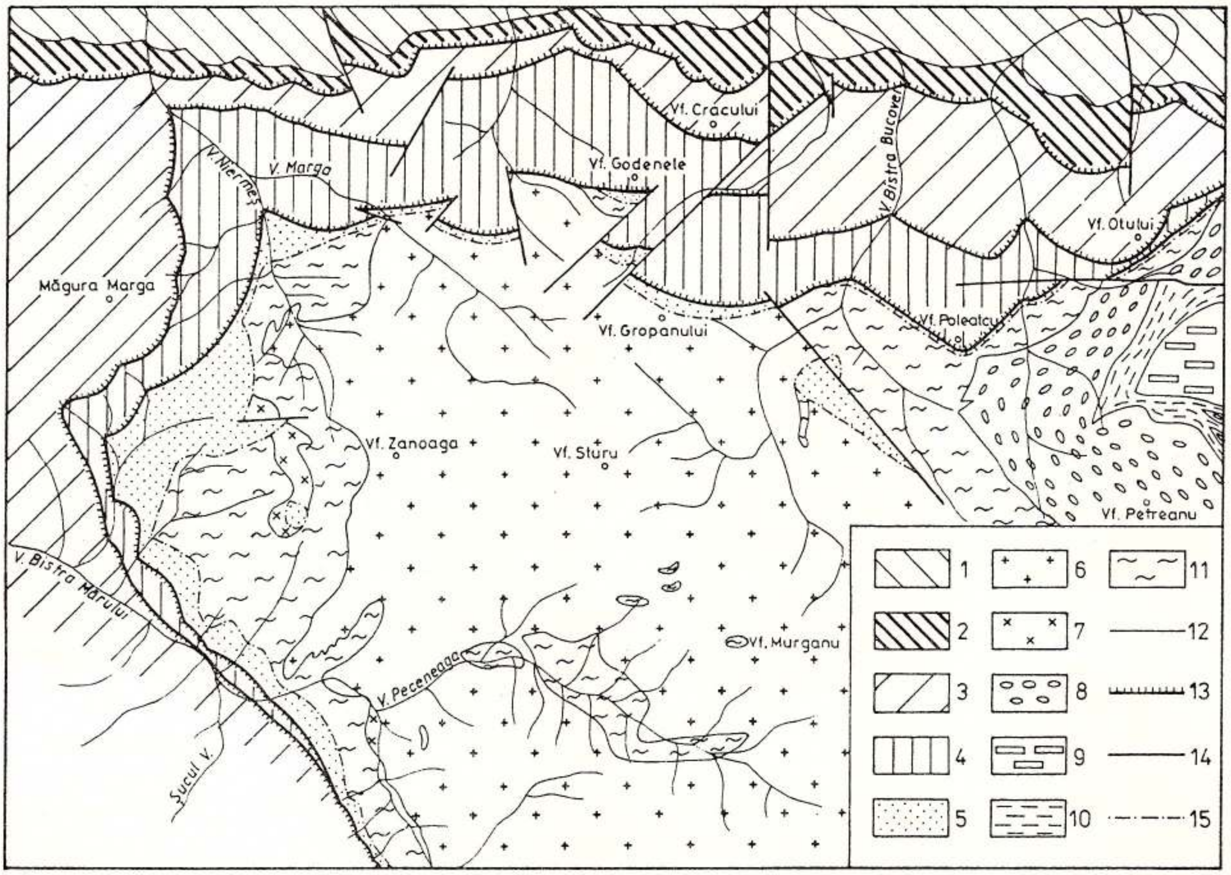
a) La Série de Nisipoasa occupe une bande de terrain arquée au Nord du sommet du Petreanu, dans le bassin supérieur du Râuşorul Hobiţei, et sépare les métagranitoïdes de Petreanu de ceux de Furcătura. La largeur de cette bande est de quelques centaines de mètres. La série est constituée par des schistes quartzeux à biotite et à grenats menus, avec une intercalation d'amphibolites. Dans ces schistes est intercalée à Pietriceaua Albă une lentille de quartz. Dans le sommet Coposu affleurent des phyllonites grises séricitéuses à biotite, dérivées de la même série. Par son contenu lithologique, cette série métadétritique ressemble bien à celle de Bodu; d'après T. Berza ces séries sont même identiques.

b) La Série de Bodu représente l'enveloppe schisteuse du massif granitique de Vârful Pietrii, ainsi que le toit des métagranitoïdes de Petreanu. Elle affleure dans les bassins supérieurs du Niermeş, de la Bistra Bucovei et du Zeicani, sur le Râuşorul Hobiţei et sur le Bodu, ainsi qu'à l'Est et au Nord de Poiana Mărului, dans le bassin supérieur de la Bistra Mărului.

Dans la constitution de la Série de Bodu entrent un fond de paragneiss à biotite et grenat, de micaschistes et de quartzites parfois à muscovite et des intercalations d'amphibolites et de gneiss amphiboliques. Au Nord de Poiana Mărului on remarque des intercalations de roches à fines aiguilles de hornblende. De rares lentilles de calcaires cristallins à trémolite apparaissent aussi. Des ultrabasites serpentinisées traversent les roches mentionnées ci-dessus.

La Série de Bodu se retrouve sous la forme de septums, ou d'intercalations de l'ordre de grandeur de centaines de mètres, à l'intérieur des corps granitiques de Petreanu et de Vârful Pietrii. Sur un de ces septums de gneiss à biotite et hornlende on a détérminé les âges K/Ar de 667 ± 20 et 538 ± 20 Ma (Grünenfelder et al., 1983).

Sous les calcaires du sommet de Poleatcu, à la partie



Esquisse géologique des Monts Petreanu-Ţarcu septentrionaux.

Série de Zeicani: 1, membre supérieur; 2, membre inférieur; 3, Série de Măgura Marga; Danubien superieur: 4, Série de Godenele; Danubien inférieur: 5, Formations de Schela et Oslea; Jurassique-Crétacé inférieur: 6, Granites de Vârful Pietrii; 7, Diorites quartziferes de Bistra Mărului; 8, Métagranitoïdes de Petreanu; 9, Métagranitoïdes de Furcătura; 10, Série de Nisipoasa; 11, Série de Bodu; 12, limite géologique; 13, nappe de charriage; 14, faille; 15, limite de décollement tectonique.

terminale de la Série de Bodu, affleurent des phyllonites séériciteuses à biotite, de couleur gris-noirâtre, similaires à celles du sommet Coposu.

c) Les métagranitoïdes (orthogneiss) de Petreanu ont dans leur ensemble la forme d'un corps stratoïde, avec la Série de Nisipoasa dans leur mur à l'Est et avec la Série de Bodu dans leur toit à l'Ouest, leur épaisseur maximum étant d'environ 5 km. Etant soumise à au moins deux phases de métamorphisme et de plissement, cette intrusion plus ancienne se présente sous l'aspect de gneiss granitoïdes et de gneiss rubanés (lit-par-lit). L'intrusion du magma a été accompagnée de procésus de migmatisation métasomatique (Gherasi, Dimitrescu, 1968, 1970), visible aussi bien à la périphérie du massif, dans les schistes cristallins des Séries de Nisipoasa et de Bodu, qu'à son intérieur, en liaison avec les intercalations de schistes quartzeux, biotitiques ou amphiboliques. Dans la Série de Nisipoasa, sur la crête Petreanu-Furcătura Clopotivei, apparaissent des apophyses du massif intrusif. Dans la Série de Bodu se développent des gneiss ocellaires et perlés. En dehors de la migmatisation métablastique, mise en évidence aussi par des porphyroblastes lenticulaires de microcline, longues jusqu'à 3 cm, dans l'orthogneiss lui-même une migmatisation métatectique (diadisytes, embréchites, stromatites, phlébites) apparaît aussi.

Les métagranitoïdes de Petreanu ont été décrits par Gherasi (1937) et Dimitrescu (1986). Des études géochimiques détaillées ont été entreprises sur ce massif par Andăr (1991), qui montre que les roches correspondent à des granites et surtout à des granodiorites.

Il éxiste aussi des opinions selon lesquelles le corps entier de métagranitoïdes de Petreanu a été engendré par la granitisation métasomatique de la Série de Bodu (Tatu, 1992, rapport pour la feuille Petreanu, archives de l'I. G. R.).

d) Une seconde phase d'intrusions granitiques est réprésentée par le massif de Vârful Pietrii, qui constitue la limite méridionale de la zone en discussion. Cette limite traverse la crête Vârful Pietrii-Petreanu, en intersectant ensuite, vers l'Ouest, les vallées de Bistra Bucovei et de Marga, et arrive jusqu'au meridien de Poiana Mărului. Le granite, très uniforme, est de couleur blanche, les éléments mafiques faisant défaut. Sa texture est toujours légèrement orientée, la zone marginale, de maximum 500 m de largeur, ayant la texture foliée plus prononcée.

Sa pétrographie étant décrite pour la première fois par Gherasi (1937) et ensuite par Gherasi et al. (1968 a), le massif a fait, ces dernières années, l'objet des études géochimiques detaillées dûes à Andăr (1991) et à Gandrabura et Şabliovschi (1987, 1988 a, 1988 b). D'après Andăr (1991), la structure de l'intrusion est concentrique, le procéssus de différentiation ayant lieu par une cristallisation procédant de la périphérie vers l'intérieur. Le magma granitique a une origine anatectique lithogène, étant de type S. D'après Gandrabura et Şabliovschi (1987, 1988 b), des types de monzogranite, syénogranite et de granite alcalin apparaissent aussi, les deux derniers représentant des produits de métasomatose du premier.

Les diorites quartzifères de Bistra Mărului, antérieures à la mise en place du massif granitique, étudiées au point de vue géochimique par Andăr (1991) et par Gandrabura et Şabliovschi (1988 a), font partie, d'après ces derniers, d'une série trondhjémite-tonalite-diorite-gabbro.

L'âge K-Ar de 287 ± 9 Ma obtenu pour ce massif (Grünenfelder et al., 1983) est très probablement rajeuni.

e) La Formation de Schela (Vidra) a été séparée dans la région sous le nom de Formation de Schela (Lias) par Gherasi (1937), puis décrite par Al. Codarcea et N. Gherasi (1944, rapport, archives de l'I. G. R.) sous le nom de "Formation de Vidra". Cette séquence peut être suivie de manière continue depuis Poiana Mărului, par Valea Mare, dans les bassins du Niermeş et de la Marga; plus loin vers l'Est, elle se prolonge en traversant les sources de la Bucoviţa et de Valea Lupului et disparaît immédiatement à l'Ouest de la Bistra Bucovei. L'épaisseur de cette formation varie dans les monts Petreanu-Ţarcu entre quelques dizaines de mètres et cca 1000 m (Valea Mare).

La Formation de Schela (Vidra) est constituée de: métaconglomérats à galets de quartz (à ciment formé par du quartz, du feldspath et de la séricite), quartzites noirâtres à séricite, métaarkoses sériciteuses jaunâtres, schistes quartzitiques à séricite et "graphite" (métaanthracite), calcaires noirâtres et calcschistes "graphiteux" (Valée de Roşia, à Poiana Mărului), phyllades sériciteuses-"graphiteuses", contenant parfois du chloritoïde.

Le substratum de cette formation est représenté soit par la Série de Bodu, soit par le granite de Vârful Pietrii. La limite avec la première est visible sur le chemin forestier du versant droit de la Vidra; aussi bien les gneiss ocellaires du mur que les métaconglomerats du toit sont affectés, sur quelques dizaines de mètres, par une forte lamination dûe au décollement de la couverture sur son socle; sur ces quelques dizaines de mètres, les deux formations, au moins à l'oeil nu, ne peuvent plus être distinguées. Le contact primordial de transgression est donc à présent masqué par ce décollement. Dans le toit de la formation de couverture se trouve la Série de Godenele du Danubien supérieur, toujours en position tectonique.

Les calcaires du ruisseau de Roşia contiennent des piéces de crinoïdes et des fragments d'échinodermes (Gherasi et al., 1968 a). Des schistes à chloritoïde de Valea Mare, Zimmerman a récolté en 1968 un fragment de céphalopode, attribué par M. Iordan aux genres de Michelinoceras ou Cornulites, d'âge Ordovicien moyen-Trias (Gherasi et al., 1975).

Les déterminations palynologiques éffectuées sur les phyllades et les calcschistes de Valea Mare et de Roşia ont mis en évidence des formes caractéristiques pour le Dévonien, à côté d'autres formes qui se continuent au Carbonifère. L'ensemble de ces données a détérminé Gherasi et al. (1975) d'attribuer la formation de Vidra au Dévonien, et son métamorphisme au cycle hercynien.

D'autre part, Berza et Seghedi (1993), prenant en considération les ressemblances frappantes avec les séquences attribuées à la formation de Schela dans les monts Vulcan, Parâng et Retezat, attribuent la Formation de Vidra de Codarcea-Gherasi au Lias et son métamorphisme au cycle alpin.

Cette formation représente dans les deux hypothèses une couverture, soit paléozoïque, soit mésozoïque, du Précambrien de l'unité danubienne inférieure.

f) la place de ce que Codarcea et Gherasi ont appelé la Formation de Vidra est prise brusquement, à partir de la Bistra Bucovei vers l'Est, par une bande de calcaires cristallins de quelques dizaines de mètres d'épaisseur, qui passe par le sommet de Poleatcu, la vallée de Zeicani, l'ensellement de Cracu, et qui s'éffile immédiatement à l'Ouest de Râuşorul Hobiţei. Tout comme la Formation de Schela, ces calcaires ont la Série de Bodu du Danubien inférieur dans leur mur et la Série de Godenele du Danubien supérieur dans leur toit.

Gherasi et al. (1968 a) ont attribué les calcaires de Poleatcu au soubassement cristallin, puis (Gherasi et al., 1975) à la Formation dévonienne de Vidra. Nous les considérons d'âge Jurassique supérieur-Crétacé inférieur tant par leur position dans l'édifice structural, que par leur ressemblance avec les calcaires d'Oslea qui affleurent plus à l'Est dans le massif de Retezat (Berza et al., 1988).

II. Danubien supérieur

1. Unité de Poiana Mărului

a) La Série de Godenele peut être suivie à partir de Poiana Mărului par le versant Ouest de la Valea Mare, la Valée du Niermeş, le sommet de Godenele, Valea Lupului, Bistra Bucovei, l'ensellement de Cracu, jusqu'au Râuşorul Hobiţei. Cette série est intensément rétromorphe, se présentant surtout sous la forme de schistes à chlorite et muscovite, affectés de mylonitisation; la chlorite s'est formée aux dépens de la biotite. Les intercalations d'amphibolites sont aussi affectées par la chloritisation, avec formation d'épidote et de calcite. D'autres intercalations sont constituées de gneiss blancs quartzo-feldspatiques (en partie, des granites mylonitisés), de rares quartzites vitreux, gris, et de schistes sériciteux à graphite (vallée de Marga). Au Nord du sommet Poleatcu apparaît une ultrabasite sérpentinisée.

La Série de Godenele recouvre la Formation de Schela ou les calcaires de Oslea et, plus à l'Est, directement la Série de Bodu et les métagranitoïdes de Petreanu du Danubien inférieur. Dans son toit se trouve la Série de Măgura Marga. Elle a été décrite par Gherasi et al. (1968 a), par Gherasi et Savu (1969) et par Dimitrescu (1986) sous le nom de "Série de Bărniţa", comme étant constituée par des schistes épimétamorphiques chlorito-sériciteux ± biotite, chlorito-épidotiques à albite et chlorito-actinolitiques; les mêmes auteurs considéraient qu'elle se retrouve dans les "schistes verts de Baicu-Râu Şes" (Gherasi et al., 1973). Le lever de la vallée Bărniţa (affluent droit de Bistra Mărului) nous a montré que la série respective n'affleure pas sur cette vallée, ce qui nous a détérminé à changer son nom sur la feuille 1:50.000 Retezat (Berza et al., 1983) en "Série de Godenele". Elle diffère aussi bien des "schistes verts de Baicu-Rau Şes", que de la Série de Zeicani, avec laquelle elle a été mise en parallèle (Gherasi et al., 1968 a, 1973; Berza, Seghedi, 1983).

2. Unité de Muntele Mic

a) La Série de Măgura Marga affleure dans une bande de terrain à direction Est-Ouest, à partir de Măgura Marga jusque dans le Râuşorul Hobiţei et plus loin, sur le versant Nord de la Furcătura Clopotivei. Dans son mur se trouve la Série de Godenele et dans son toit la Série de Zeicani. Définie initialement (Gherasi et al., 1968 a; Gherasi, Savu, 1969) comme épimétamorphique (faciès schistes verts), elle représente en fait une toute autre chose et notamment une série migmatisée de type Lainici-Păiuş. Le fond de cette série est constitué par des plagiogneiss (paragneiss) blancs à muscovite et par des schistes quartzeux micacés à albite, en passant parfois à des leptynites, et alternant avec de nombreuses amphibolites. Sur ce fond se développent des phénomènes de migmatisation, allant jusqu'à la granitisation. La migmatisation peut être de type homogéne, à formation de gneiss lenticulaires, ocellaires ou oeillés, à porphyroblastes de microcline, ou de type hétérogène, consistant en un réseau de filons de composition granitique ou plagiogranitique, fréquemment affectés par une schistosité mylonitique. Des lentilles pegmatoïdes, ainsi que des nids boudinés de biotites (restites), apparaissent aussi. Entre les amphibolites et les gneiss micacés, la transition s'effectue parfois par l'intermédiaire de gneiss amphiboliques. La diaphtorése de la Série de Măgura Marga se manifeste par endroits par la chloritisation de l'amphibole ou de la biotite.

Plusieurs corps d'ultrabasites recoupent cette série dans la Măgura Marga elle-même, mais aussi plus à l'Est, par exemple dans la crête de Lupul; ces roches sont presque totalement serpentinisées (à formation d'antigorite), mais parfois elles se transforment en un aggrégat de bastite, de chlorite et d'épidote.

b) Entre la vallée de la Bistra Mărului et son affluent droit nommé la Bolvaşniţa affleure l'extrémité nordique du massif granitoïde de Muntele Mic, représenté par des diorites-granodiorites à biotite ± hornblende et des granites à biotite. Leur contact nordique est marqué par une faille importante (Gherasi, Savu, 1969), tandis que celui méridional présente une zone de feldspatisation de l'encaissant, attribuée à la Série de Măgura Marga.

3. Unité de Măru

a) La Série de Zeicani affleure dans une bande de terrain large de 1 à 2 km, le long de la bordure du Nord des monts Petreanu-Ţarcu, entre les localités Marga et Zeicani. De même que sur les feuilles 1:50.000 Lupeni (Berza et al., 1986) et Retezat (Berza et al., 1989), nous avons séparé dans cette série un membre inférieur, représenté par un fond d'amphibolites et de gneiss amphiboliques, à intercalations de gneiss ocellaires, et un membre supérieur, formé par un fond de plagiogneiss (paragneiss) micacés, à intercalations d'amphibolites, de gneiss ocellaires et de leptynites. Les deux membres sont fortement rétromorphes, les amphibolites étant transformées en schistes à chlorite, albite, épidote et calcite ou en schistes actinolitiques, et les plagiogneiss micacés en schistes quartziteux à albite et muscovite ou en schistes séricito-chloriteux.

Dans le soubassement de la Série de Zeicani se trouve la Série de Măgura Marga; elle est recouverte par endroits par des grés noirâtres ou des calcaires cristallins représentant sa couverture mésozoïque, ou directement par la Série de Lotru de la Nappe Gétique.

Dans les environs de Marga et Bucova (vallée de Brăila et de Bucoviţa), la Série de Zeicani est traversée par des roches ultrabasiques sérpentinisées, contenant des concentrations de talc. Dans la carrière de Marga (Valea Mânzului), deux corps principaux sont séparés par un niveau d'amphibolites rubanées; ces roches sont à leur tour traversées par un filon de plagiogranite de 300 m de longueur et de quelques mètres d'épaisseur, orienté Est-Ouest; il est accompagné par des lentilles pegmatitiques à phlogopite. Au contact des roches acides avec les ultrabasites se sont formées des zones de réaction à roches amphibolitiques à structure de Garbenschiefer, à chloritites et à poches de talc avec des nids et des filonnets de quartz et de carbonates.

L'âge Précambrien supérieur (éventuellement jusqu'au Cambrien inférieur) a été attribué à la Série de Zeicani par Gherasi et al. (1973), sur la base de certaines indications palynologiques. Les données géochronologiques (Grünenfelder et al., 1983) attestent en tout cas un intense rajeunissement alpin, dû aux charriages.

b) Entre les vallées de la Bistra et du Niermeş, une bande discontinue de grés et schistes argileux noirs, suivis de calcaires cristallins gris ou blancs, a été remarquée par Gherasi et al. (1975) et attribuée à la Formation dévonienne de Vidra. Ces lambeaux peuvent être attribués au Lias (les roches détritiques) et au Jurassique supérieur–Crétacé inférieur (les calcaires cristallins) et ressemblent parfaitement aux roches constituant les Formations de Schela et Oslea.

4. Nappe Gétique

a) La Série de Lotru du cristallin gétique n'occupe qu'une aire réduite dans l'extrémité Nord de la région. Elle est représentée par des paragneiss (plagiogneiss) micacés et par des micaschistes, à petites lentilles de pegmatites. Le métamorphisme barrovien de cette série a eu lieu dans les conditions du faciès amphibolite à almandin.

III. Structure

Outre le charriage de la Nappe Gétique, le principal élément de la région septentrionale des monts Petreanu-Ţarcu est le charriage du Danubien supérieur sur le Danubien inférieur. La semelle du premier (Série de Godenele) chevauche les formations anchi- ou épimétamorphiques de la couverture du Danubien inférieur, en provoquant en même temps leur partiel décollement de son soubassement normal. Cette couverture a été attribuée, comme nous l'avons vu plus haut, soit au Paléozoïque, soit au Mésozoïque; l'âge du charriage est en tout cas alpin, comme il résulte tant de la structure du versant Sud du Lăpuşnic, situé un peu plus à l'Est (Gherasi et al., 1986), où des formations acceptées plus aisément comme mésozoïques sont impliquées, que des âges K/Ar, rajeunis à environ 100 Ma, jalonnant ce contact (Grünenfelder et al., 1983).

Dans le cadre du Danubien supérieur, on a vu que chacune des trois séries cristallines est attribuée à une unité tectonique indépendante, conformément au modèle élaboré par Berza et al. (1983) et utilisé aussi par Dimitrescu (1986) et par Berza et al. (1989). Nous croyons pouvoir argumenter qu'il ne s'agit pas d'une simple succession stratigraphique (comme l'admettaient Gherasi et al., 1974, rapport, archives de l'I.G.R.), par les épaisseurs apparemment inconstantes des deux subdivisions de la Série de Zeicani ainsi que par celles des autres deux séries, de Măgura Marga et de Godenele. D'autre part, nous ne sommes plus convaincus de l'âge alpin de toutes ces unités (nappes), aucune formation pouvant être considerée comme mésozoïque n'étant pas impliquée.

Le Danubien inférieur affleure ainsi dans une fenêtre tectonique qui se ferme de manière périclinale au Nord-Ouest de Poiana Mărului. Un rôle important est joué dans le cadre de cette fenetre par l'antiforme Vârfu Pietrii-Rof, à direction Est-Ouest et à plongement axial vers l'Ouest.

La Nappe Gétique n'affleure que dans l'extrémité septentrionale du massif montagneux, sous la forme de quelques lambeaux constitués par les micaschistes et plagiogneiss micacés de la Série de Lotru. La limite entre le soubassement cristallin et le couloir post-tectonique de la Bistra, rempli de formations sédimentaires (maastrichtiennes ? paléogènes ?) est représentée par une faille normale ou de décrochement à direction Est-Ouest, remarquée plus à l'Est déjà par E. de Martonne.

Dans le cadre de la tectonique disjonctive, on remarque un système de failles NE-SO d'âge post-charriage, mais pré-couverture sédimentaire tertiaire et un second système de failles qui affecte aussi la couverture post-tectonique (par exemple à Zeicani), en décrochant la grande faille E-O mentionée plus en haut.

En ce qui concerne la mésostructure du Danubien inférieur, en adaptant à notre région le schéma élaboré par un des auteurs (Dimitrescu, 1986) pour la région adjacente à l'Est, la succesion des événements tectoniques serait la suivante:

1) une phase de plissement F1 asociée au dé-

veloppement d'une schistosité S1 dans les Séries de Nisipoasa et de Bodu, synchrone à leur métamorphisme en faciès amphibolite;

1a) intrusion, probablement cadomienne ancienne, du massif granitoïde de Petreanu;

2) plissement F2 à direction NNE-SSO des formations cristallines et des granitoïdes, avec le développement d'une schistosité S2, qui transpose la foliation S1, étant en même temps la premiére foliation qui pénètre les granitoïdes;

3) plissement F3 à direction ENE-OSO de toutes les formations mentionnées, associé à l'apparition d'un système de linéations L3, rétromorphisme en zone de la biotite;

4) intrusion du massif granitique de Vârful Pietrii;

4a) déposition de la Formation de Schela;

5) développement d'une schistosité S4 qui transpose la foliation S3 dans le soubassement, mais qui est la première schistosité affectant la Formation de Schela; métamorphisme en zone à chlorite, progressif pour celle-ci, mais régressif pour le soubassement cristallin;

5a) charriage alpin à métamorphisme dynamique, concentré sur les surfaces de chevauchement ou de décollement de la Formation de Schela sur son mur Danubien inférieur, des calcaires de Oslea et des depôts sédimentaires équivalents réposant sur la Série de Zeicani et recouverts par la Nappe Gétique;

6) bombement de l'antiforme Vârful Pietrii-Rof conduisant à "l'émersion" de la fenêtre d'érosion du Danubien inférieur de sous sa couverture danubienne supérieure.

Dans le Danubien supérieur, les éléments planaires dominants sont les schistosités S2 à caractère de foliation planaxiale orientés E-O/30-70N, associés à des linéations L2 à direction E-O; vers l'Ouest, les foliations s'infléchissent vers la direction NE/O. La schistosité S1 décrit des plis mésoscopiques trés aigus de l'ordre des centimètres.

Bibliographie

Andăr, P. (1991) Geochemistry of the granitoid rocks in the Ţarcu Mountains. *An. Inst. Geol. Geof.*, 68, p. 15–114, Bucureşti.

Berza, T., Kräutner, H., Dimitrescu, R. (1983) Nappe structure in the Danubian window of the Central South Carpathians. *An. Inst. Geol. Geof.*, 60, p. 31–39, Bucureşti.

— , Seghedi, A. (1983) The crystalline basement of the Danubian units in the Central South Carpathians. *An. Inst. Geol. Geof.*, 61, p. 15–22, Bucureşti.

— , Seghedi, A., Pop, Gr., Szasz, L., Hârtopanu, I., Săbău, G., Moisescu, V. (1986) Harta geologică a României, sc. 1:50.000, foaia Lupeni, Inst. Geol. Geof., Bucureşti.

— , Dimitrescu, R., Dimitrescu, M., Tatu, M., Strucsiewicz, R., Pop, Gr., Szasz, L., Popescu, G., Bandrabur, T. (1989) Carte géologique de la Roumanie, sc. 1:50.000, feuille Retezat, Inst. Geol. Geof., Bucureşti.

Dimitrescu, R. (1986) Le Danubien des monts de Petreanu et de Retezat dans la région Râul Mare. *D. S. Inst. Geol. Geof.*, 70–71/1, p. 269–289, Bucureşti.

Gandrabura, Em., Şabliovski, V. (1987) Petrochemical data on the southern part of the Vârful Pietrii granitic massif. *An. St. Univ. "Al. I. Cuza", sect. II b (Geol.-Geogr.)*, 33, p. 11–12, Iaşi.

— , Şabliovski, V. (1988 a) The Bistra Mărului "diorites", a gabbro-diorite-tonalite-trondhjemite series in the Ţarcu Mountains. *An. St. Univ. "Al. I. Cuza", sect. II b (Geol.-Geogr.)*, 34, p. 7–11, Iaşi.

— , Şabliovski, V. (1988 b) New petrochemical data on the Vârful Pietrii granitic massif. *An. St. Univ. "Al. I. Cuza", sect. II b (Geol.-Geogr.)*, 34, p. 12–16, Iaşi.

Gherasi, N. (1937) Étude pétrographique et géologique dans les monts Godeanu et Ţarcu. *An. Inst. Geol.*, 18, p. 1–78, Bucureşti.

— , Mureşan, M., Mureşan, G., Kräutner, H., Kräutner, F., Lupu, M., Marinescu, F., Savu, H., Drăgulescu, A. (1968) Carte géologique de la Roumanie au 1:200.000, feuille Deva. Inst. Geol., Bucureşti.

— , Zimmerman, P., Zimmerman, V. (1968) Structura şi petrografia şisturilor cristaline din partea de N a munţilor Ţarcu. *D. S. Inst. Geol.*, 64/1, p. 50–80, Bucureşti.

— , Dimitrescu, R. (1968) Contribuţii petrotectonice la structura Cristalinului Danubian în partea nordică a munţilor Retezat şi Petreanu. *An. St. Univ. "Al. I. Cuza", sect. II b (Geol.-Geogr.)*, 14, p. 29–38, Iaşi.

— , Savu, H. (1969) Structura masivului granitoid Muntele Mic. *D. S. Inst. Geol. Geof.*, 54/3, p. 55–82, Bucureşti.

— , Dimitrescu, R. (1970) Anticlinalul Rof şi rolul lui în structura părţii nordice a munţilor Retezat şi Petreanu. *An. St. Univ. "Al. I. Cuza", sect. II b (Geol.)*, 16, p. 22–55, Iaşi.

— , Visarion, A., Zimmerman, P. (1973) Consideraţii asupra vîrstei unor şisturi cristaline şi depozite sedimentare din autohtonul danubian, situate la marginea de nord a munţilor Godeanu. *St. Cerc. Geol.*, 18/2, p. 303–310, Bucureşti.

— , Visarion, A., Zimmerman, P., Iordan, M. (1975) Asupra vîrstei paleozoice (Devonian) a formaţiunii de Vidra din partea de nord a munţilor Ţarcu. *D. S. Inst. Geol. Geof.*, 61/4, p. 3–9, Bucureşti.

— , **Berza, T., Seghedi, A., Stepan, M., Iancu, V. (1986)** Structura geologică a părţii nordice a masivului Godeanu. *D. S. Inst. Geol. Geof.*, 70-71/5, p. 63-85, Bucureşti.

Grünenfelder, M., Popescu, Gh., Soroiu, M., Berza, T. (1983) K-Ar and U-Pb dating of the metamorphic formations and the associated igneous bodies of the Central South Carpathians. *An. Inst. Geol. Geof.*, 61, p. 37-45, Bucureşti.

Kräutner, H., Năstăseanu, S., Berza, T., Stănoiu, I., Iancu, V. (1981) Metamorphosed Paleozoic in the South Carpathians. *Assoc. Geol. Carp.-Balc., Congr. XII, Guide to Excursion A1*, p. 1-116, Bucureşti.

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SUR LA TERMINAISON SEPTENTRIONALE DES SYSTEMES DE NAPPES ALPINES DANS LES MONTS DE GILĂU

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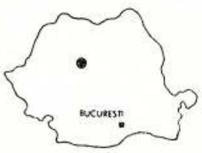

Key words: Nappes. Alpine-type. Metamorphic rocks. Thrust fault. Stratigraphic units. Apuseni Mts.

Abstract: *On the Northern Ending of the Alpine Nappe Systems in the Gilău Mountains.* East of the Muntele Mare granitic massif, the limit between the Someş Series (micaschists, leptynites) and the Biharia Series (amphibolites, chlorite schists, dolomitic limestones) is of tectonic nature. Along this line, black ultramylonites, green and white mylonites and breccias appear.

En 1957, l'un des auteurs a fait pour la première fois la distinction dans le cadre du soubassement préalpin des Monts Apuseni, entre un "Cristallin du Gilău", à faciès monotones, ayant une large répartition aréale (Séries de Someş et d'Arada), et un "Cristallin de l'Arieş", beaucoup plus varié (Séries de Codru-Mădrizeşti, Biharia, Muncel, Baia de Arieş). Dans son ensemble, le second (avec ses complications internes), était considéré comme charrié sur le premier dans le Bihor et le Gilău méridional (Bleahu, Dimitrescu, 1957).

Dix ans plus tard (Dimitrescu, 1966), je considérais encore que les charriages ne paraissaient pas se prolonger dans la partie nord-est du Gilău, opinion partagée par Mureşan (1980).

Après encore dix ans, déjà la conception générale avait évolué dans le sens de l'acceptation du prolongement des systèmes de nappes dans la région citée, mais les données de détail manquaient: pour preuve, la différence entre le tracé de ces nappes figuré sur la carte géologique annexée à la monographie des Monts Apuseni (Ianovici et al., 1976) et celui figuré sur l'esquisse tectonique du meme ouvrage.

Nos levers entrepris en vue de la rédaction de la feuille 1:50.000 Călăţele nous a mis, en 1990, en situation de résoudre ce problème.

L'extrémité orientale de la feuille constitue une partie de la région étudiée en thèse de doctorat par Mureşan (1971, 1980), la carte ayant été publiée séparément (Mureşan, 1973). Comme nous l'avons déjà mentionné, sur cette carte, à séparations pétrographiques extrêmement détaillées, aucune ligne de charriage n'apparait.

À l'Est du massif granitique du Muntele Mare, dans les bassins hydrographiques du Căpuş et de l'Agârbiciu Mureşan (1973, 1980), tout comme Hanomolo, Hanomolo (1962) - dont les données ont constitué la base de la feuille 1:200.000 Cluj (Saulea et al., 1967) - font une distinction non justifiée entre la Série ("mésométamorphique") de Someş et des parties de la même série affectées de diaphtorèse, attribuées ensemble avec d'autres termes au "complexe I de la Série épimétamorphique"; celui-ci équivalant, d'après l'auteur à la "Série de Bistra" (Giuşcă et al., 1967) et d'après d'autres géologues de l'IPEG-Cluj, à la Série d'Arada.

Nos levers ont montré qu'à l'Est de Bedeciu, la Série de Someş du Cristallin de Gilău, constituée par des micaschistes, des paragneiss micacés (parfois à grenat et à staurotide) et des gneiss quartzo-feldspathiques (léptynites), diaphthoritiques ou non, s'étend jusqu'à une ligne bien marquée orientée NW-SE qui coupe la vallée de Căpuş immédiatement en aval de son confluent avec le Dingău et la vallée de l'Agârbici (Soponii), immédiatement en aval de son confluent avec le Gruiul de Sus. À l'est de cette ligne apparaît la Série de Biharia du Cristallin de l'Arieş, formée par des amphibolites, des schistes séricito-chloriteux, des chlorito-schistes à épidote et albite, des calcaires et des dolo-cristallines, des calcschistes, des gneiss albitiques (métagranites trondhjémitiques) et des schistes graphiteux. Mais ces derniers n'apparaissent que dans la vallée du Căpuş, au confluent avec Parâul Plaiului et un peu en aval, vers son confluent avec Pâraul Băii: toutes les autres roches définies par Mureşan (1973, 1980) comme schistes graphiteux ne sont en réalité que des ultramylonites variées. Elles sont accompagnées par de fréquentes mylonites vertes (provenant d'amphibolites) et de mylonites blanches (provenant de gneiss). Des mylonites apparaissent aussi dans la Série de Someş, vers le contact tectonique. En plus, l'auteur cité (Mureşan, 1973) lui-même, a séparé dans ce sécteur une série de bréches - autres que celles du confluent des deux Someş - sans leur attribuer une signification quelconque; elles sont intercalées dans les schists cristallins de la Série de Biharia dans les bassins du Căpuş et de l'Agărbiciu et leur caractère tectonique est clair dans les affleurements, sans aucun doute possible.

De ce que nous avons montré, on peut déduire que:

a) La distinction lithostratigraphique entre les séries de Someş et de Biharia est évidente.

b) Entre les deux séries s'interpose une importante surface de charriage à vergence Ouest, marquée par une zone mylonitique qui se manifeste au-dessous et, surtout, au-dessus du plan tectonique.

c) D'après le caractère destructif du métamorphisme dynamique exercité par le charriage, l'âge de celui-ci est très probablement alpin, des récristallisations blasto-mylonitiques qui accompagnent les charriages plus anciens n'apparaissant pas.

Cette ligne tectonique ainsi identifiée par nous, représente le prolongement vers le Nord de celle figurée par Hârtopanu et al. (1982) sur la feuille 1:50.000 Valea Ierii. Plus au Nord, au-delà de la vallée du Căpuş, cette ligne est cachée par les formations paléogènes du bassin de Huedin.

Dans la même zone apparaissent des minéralisations sporadiques, parfois à texture brécheuse, à sulfures, Au et Ag, par exemple à Pârâul Băii (Mureşan, 1980); elles sont analogues à celle de Gilău (Dealul Mieilor et Valea Seacă) (Luca, 1936; Laţiu, 1941; Mureşan, 1980). Elles ne sont pas liées aux banatites et appartiennent au groupe des gîtes tectogéniques (Udubaşa, Hann, 1988).

À notre avis, il n'existe pas de motifs pour séparer dans cette région une apparition de la Série d'Arada.

En conclusion, la limite entre le Cristallin du Gilău et celui de l'Arieş est, partout dans les Monts Apuseni, de nature tectonique. L'identification des charriages alpins dans le secteur Sud et Est des monts de Gilău ne peut être réalisé que sur la base de la distinction entre les faciès des formations métamorphiques, étant donné que des formations sédimentaires ne s'interposent pas entre les diverses unités.

Bibliographie

Bleahu, M., Dimitrescu, R. (1957) Stratigrafia şi tectonica Munţilor Apuseni. *An. Rom.-Sov. (Geol.-Geogr.)*, 2, p. 29–42, Bucureşti.

Dimitrescu, R. (1966) Muntele Mare. Studiu geologic şi petrografic. *An. Com. Geol.*, 35, p. 165–239, Bucureşti.

Giuşcă, D., Savu, H., Borcoş, M. (1967) Asupra stratigrafiei şisturilor cristaline din Munţii Apuseni. *St. cerc. geol.*, 12/1, p. 41–56, Bucureşti.

Hanomolo, I., Hanomolo, A. (1962) Geologia şi petrografia regiunii Someş-Muntele Rece-Mânăstireni-Căpuş. *D. S. Inst. Geol.*, 47 (1959-1960), p. 97–113, Bucureşti.

Hârtopanu, I., Balintoni, I., Borcoş, M., Rusu, A., Lupu, M. (1982) Harta geologică a României, sc. 1:50.000, foaia Valea Ierii. Inst. Geol. Geofiz., Bucureşti.

Ianovici, V., Borcoş, M., Bleahu, M., Patrulius, D., Lupu, M., Dimitrescu, R., Savu, H. (1976) Geologia Munţilor Apuseni. 632 p., Edit. Academiei, Bucureşti.

Laţiu, V. (1941) Contribuţiuni la studiul calcografic al mineralizărilor apomagmatice din Munţii Gilăului. *Bul. Soc. Politehn. Timişoara*, 10, p. 3–4, Timişoara.

Lucca, V. (1936) Regiunea auriferă de la Someşul Rece. *Rev. Muz. Geol. Miner. Univ. Cluj*, 6/1-2, p. 155-164, Cluj.

Mureşan, I. (1971) Studiul mineralogic şi petrografic al formaţiunilor dintre Someşul Cald şi valea Căpuşului. Rezum. tezei de doctorat, p. 1-36, Cluj.

— **(1973)** Rocile epimetamorfice şi metagabbrourile dintre valea Finişelului Mare şi valea Căpuşului. *Stud. Univ. "Babeş-Bolyai", seria Geol.-Min.*, 18/2, p. 13–22, Cluj.

— **(1980)** Geologia şi petrografia bordurii de nord-est a munţilor Gilău, p. 1–128, Edit. Academiei, Bucureşti.

Saulea, Em., Dumitrescu, I., Bombiţă, Gh., Marinescu, Fl., Borcoş, M., Stancu, J. (1967) Harta geologică a României, sc. 1:200.000, foaia Cluj. Inst. Geol., Bucureşti.

Udubaşa, Gh., Hann, H. (1988) A Shear-Zone Related Cu-Au Ore Occurrence: Valea lui Stan, South Carpathians. *D. S. Inst. Geol. Geofiz.*, 72-73/2 (1985, 1986), p. 259–282, Bucureşti.

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A NEW HYPOTHESIS ON THE ORIGIN OF THE ŞUCU BRECCIA (ŢARCU MOUNTAINS)

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Key words: Diatremes. Breccia. Dyke swarms. South Carpathians.

Abstract: The authors reached the conclusion that the breccia exposed in the flood plain of the Şucu and Şuculeţu valleys (Ţarcu Mts) constitutes a diatreme. The breccia pipe is closely connected to a swarm of porphyritic dykes. The breccia elements - gabbros and tonalites, would come from close vicinity, namely from the Şucu diorites and Şucu granitoids massifs. The breccia is affected by Alpine lamination planes. The Baicu conglomerates, situated more eastwards, are not necessarily included here.

In the Ţarcu-Bloju Mountains as well as within the adjacent Râu Mare and Bistra Mărului hydrographic basins a type of rocks with conspicuous features appears. They are breccias and conglomerates with gabbroic blocks, described by Schafarzik (1899) and, in detail, by Gherasi (1937). After the Second World War such rocks as well as similar occurrences were mentioned by Morariu (1976), Morariu and Morariu (1982), Kräutner et al. (1981), Berza et al. (1983), Mărunţiu and Seghedi (1983), Năstăseanu et al. (1988), Iancu et al. (1990).

These breccias and conglomerate rocks carrying gabbroic blocks cluster in two main exposures: along the Şucu and Şuculeţu valleys and their adjacent walls, and in the vicinity of the Baicu and Bistricioara peaks. Furthermore, Gherasi (1937) discriminated the "Şucu-Poiana Nedeii breccia and conglomerate" from the "Baicu conglomerate". Morariu and Morariu (1982) used the comprehensive term "Baicu-Şucu metaconglomerate".

Early authors assumed an autochthonous location of this formation, whereas during the last 15 years it has been considered as thrust over the crystalline schists basement (e. g. the Şucu Unit, thrust over the Poiana Mărului Unit, acc. to Berza et al., 1983, or the Baicu Nappe, Iancu et al., 1990).

The following review will consider almost exclusively those breccias exposed 3 km along forestry roads bordering the Şucu and Şuculeţu valleys and investigated in underground through small adits and shallow drillings. It is worth mentioning that the Şucu schistose diorite massif with restricted gabbroic differentiation and pyrrhotite mineralization is found in their close vicinity.

The most proper name of the rocks cropping out in the Şucu basin is, anyway, that of breccia. They consist of gabbro, subordinate granite-tonalite, even hornblendite, serpentinite and gabbroic pegmatite blocks of angular shape; in places, rounded or oval-shaped blocks are found. They show a massive or even schistose structure. No grading or compositional layering of the fragments is noticed. Their dominant size is coarse, up to 2 m in diameter.

The matrix corresponds roughly to the gabbroic composition and exhibits frequent moderately penetrative mylonitization/lamination planes of Alpine age, striking NE-SW and dipping towards SE. The lack of matrix is conspicuous in numerous exposures.

Talcization is interpreted as a dynamic product resulted along planes of ruptural deformation.

Sets of centimetric to decimetric microgranite and white porphyric dikes striking NE-SW and dipping towards SE intersect the Şucu breccia especially near its margins and exhibit lamination in places.

Field evidence led us to propose a new explanation of the Şucu breccia formation and setting. We suggest that this occurrence is a diatreme, irregularly shaped in plane and seeming to narrow downwards into a subvertical conduit filled with gabbroic material. Somewhat similar breccia pipe structures are common to specific metallogenic provinces wherein they show intimate association with swarms of porphyritic dykes.

The suggested Şucu-Şuculeţu intrusion of pre-Laramian age exposes therefore its brecciated position, whereas the apical part is eroded and the root is not known. Fracturation initiated during the magmatic passage controlled the heterolithic breccia occurrence enclosing gabbroic blocks and wall rock xenoliths, especially of Şucu granites and Şucu diorites; blocks derived from unknown or less developed formations in the area are expected to occur at depth. Wall fracturation promoted repeated brecciation with initiation of roundness. The magmatic activity characteristic of this subsequent event is represented by acidic dike emplacement.

We do not intend to take into account the conglomerate with gabbroic blocks occurring in the Baicu crest. This rock type contains a larger amount of matrix and frequent rounded blocks as compared to the Şucu breccia type. It is, however, likely that such a pattern leads to a "pebble breccia" resulted during polyascendant tectonics (multistage brecciation and refragmentation). It would also be possible that the Şucu diatreme might have fed a stratigraphical term of the Brustur Formation (Iancu et al., 1990), precisely the Baicu Conglomerates.

References

Berza, T., Kräutner, H., Dimitrescu, R. (1983) Nappe Structure in the Danubian Window of the Central South Carpathians. *An. Inst. Geol. Geofiz.*, 60, p. 31-39, Bucureşti.

Gherasi, N. (1937) Étude pétrographique et géologique dans les Monts Godeanu et Ţarcu. *An. Inst. Geol. Rom.*, 18, p. 1-78, Bucureşti.

Iancu, V., Seghedi, A., Măruntiu, M., Strusiewicz, R. (1990) The Structural Background of the Brustur Formation in the Inner Danubian Nappes. *D. S. Inst. Geol. Geofiz.*, 74/5 (1987), p. 61-80, Bucureşti.

Kräutner, H., Năstăseanu, S., Berza, T., Stănoiu, I., Iancu, V. (1983) Metamorphosed Paleozoic in the South Carpathians. *Guide to Excursion A 1. KBGA 12-th Congr.*, p. 1-146, Bucureşti.

Măruntiu, M., Seghedi, A. (1983) New Data Concerning the Metamorphic Rocks and Metamorphic Processes in the eastern Almăj Mountains. *Rev. Roum. Geol.*, 27, p. 29-35, Bucureşti.

Morariu, D. (1976) Date noi asupra litostratigrafiei seriei epimetamorfice de Zeicani şi metaconglomeratelor de Baicu din bazinul superior al Bistrei Mărului. *D. S. Inst. Geol. Geofiz.*, 62 (1974-1975), p. 23-36, Bucureşti.

— , **Morariu, A. (1982)** A New Structural model of the Ţarcu-Râul Mare Region. *D. S. Inst. Geol. Geofiz.*, 67/5 (1979-1980), p. 95-108, Bucureşti.

Năstăseanu, S., Iancu, V., Russo-Săndulescu, D. (1988) Schema preliminară a unităţilor structurale din bazinele văilor Bela Reca şi Râul Rece. *D. S. Inst. Geol. Geofiz.*, 72-73/5, p. 171-176, Bucureşti.

Schafarzik, Fr. (1899) Die geologischen Verhältnisse der Umgebung von Borlova und Poiana Mărului. *Jahresber. d. k. ung. geol. R. A. f. 1899*, Budapest.

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CENTRAL SOUTH CARPATHIANS – PETROLOGIC AND STRUCTURAL INVESTIGATIONS IN THE AREA OF THE OLT VALLEY

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Key words: Dynamic metamorphism. Sibişel Lithogroup. Nappe pile. Olt Valley. Central South Carpathians.

Abstract: The region of the Olt Valley (central part of the South Carpathians) exhibits along a relative narrow area the inner structure of a whole nappe pile. From west towards east, that is from a lower to an upper position, the following nappes are to be found: Lotru Laramian Unit consisting of Getic Nappe and Măgura, Uria, Câineni (with Balota scale) Nappes of Austrian age; Boia Laramian Unit, built up of Argeş and Moldoveanu Nappes. The Olt fault separates the E–W directed Făgăraş structures from the N–S directed Lotru structures. The metamorphic rocks forming the nappes are represented by schists that underwent earlier a Proterozoic regional metamorphism of medium grade and were subsequently more or less affected by a mainly dynamic metamorphism of Alpine and possibly Variscan age. The Sebeş-Lotru Lithogroup builts up the Getic Nappe, the Cumpăna lithogroup the Măgura, Câineni and Argeş Nappes. The Sibişel Lithogroup builts up the Uria Nappe and the Făgăraş Lithogroup (Suru Lithozone, respectively Suru Formation), the Moldoveanu Nappe; they show obvious correlation patterns. The Sibişel Lithogroup exhibits a complex metamorphic evolution, involving a retrograde metamorphism giving rise to large areas of complete alteration.

1. Evolution of Knowledge

As the Olt Valley represents a deep cross-section through the central part of the South Carpathians, numerous geologists have been attracted to study it from the very beginning of the geological activity in Romania.

Mrazec and Murgoci (1898), Reinhard (1906) and later Popescu-Voiteşti (1918) took into account the petrography and structure of this region. Thus, after deciphering and grouping the sedimentary and metamorphic rocks, this assemblage was structurally-tectonically analysed too. Crystalline Group I was separated by Mrazec (1897) and ascribed by Murgoci (1907) to the Getic Nappe. It was delineated from the Bucegi Conglomerate Nappe to the east, along the Olt Valley (Popescu-Voiteşti, 1918). Schmidt (1931) noticed the strike variation of the crystalline schists of "Cumpăna gneiss and Făgăraş Crystalline" from E–W towards NNW–SSE. He also observed that "Lotru Crystalline" vanishes eastwards near the Olt Valley, beneath the crystalline schists of the Făgăraş Massif. The same author stated that Crystalline Group I (Getic Nappe, Murgoci, 1907) does not represent a unique tectonic unit, but comprises two-three nappes. Significant were his considerations about the polymetamorphic character of the Făgăraş Crystalline, too.

Using his own data and also Schmidt's contribution, Streckeisen (1934) separated the plane between the Getic Nappe and "Upper tectonic units" (Făgăraş, Leaota, Cozia) based on mylonites between Răşinari-Sadu in the north and Valea lui Stan in the south. Ghika-Budeşti (1939) subsequently denied the existence of both a Supragetic line and the tectonic contact between Lotru and Făgăraş crystalline realms, admitting only local dislocations (e. g. Valea lui Stan, left bank of Călineşti Valley). He only recognized a sequence of schist piles located in anticlines and synclines, advocating therefore the belonging of both Cumpăna-Cozia Făgăraş and Sebeş-Lotru crystalline realms to a common Getic Nappe, characterized by perfect internal continuity.

Dessila-Codarcea (1961, 1962 a, b, c, 1965, 1967, 1969 a, b) provided a significant contribution to structural, petrographic and age knowledge concerning crystalline rocks of the area under discussion. She introduced the term "Sibişel Series", which was mapped and described in detail between Răşinari and Sadu Valley. According to Dessila-Codarcea, the Sibişel Series has an epimetamorphic character and lies transgressively and discordantly above the mesometamorphic schists of the Sebeş-Lotru Series. On the other hand, the Sibişel Series is overlain by the Olt Unit that represents the western prolongation of the Făgăraş and Cozia Crystalline.

Codarcea et al. (1967) suggests that the Getic Nappe is tectonically covered by a superior unit, namely the Supragetic Unit, which contains the Făgăraş and Cozia massifs. The thrusting event is considered of Austrian age.

The Sibişel Series was prospected and mapped in detail by Dragomir, Arsenescu (1961, unpubl. data), Arbore et al. (1965, unpubl. data), Trifulescu (1972, unpubl. data), Dragomir et al. (1973, unpubl. data), Dragomir, Arbore (1974, unpubl. data), Mânecan, Mânecan (1974, unpubl. data), Mânecan et al. (1978-1979, unpubl. data). Investigations of Mânecan were accomplished by papers about structure and petrography of the Sibişel Series (1977, 1984, 1989), as well as Fe mineralization found in this formation (1979). Mânecan points along some sectors the Supragetic tectonic plane overlying the Sibişel Series, delineates a segment of the Olt fault, north of Căineni, and describes the petrography of the Sibişel Series, considering it a Paleozoic epimetamorphic unit which overlies discordantly and transgressively the Sebeş-Lotru Series. Hann and Szász (1981) investigate breccias formed on augen gneisses and paragneisses from the Cozia massif and farther north, in the Călineşti-Cornetu monastery area, proving that these breccias are of various genetic types, mostly of tectonic origin.

The way of understanding various problems connected with petrography and structure of the region changed in 1984, when Hann and Szász showed that Sibişel Series schists are not epimetamorphic but retrograde mesometamorphic rocks. They are not in transgressive and discordant position above the Sebeş-Lotru Series, but overlie it tectonically along a thrust plane. These data are partly found in the Titeşti map, scale 1:50.000 (Hann, in Ştefănescu et al., 1982). Hann and Szász provide also explanation of the way of acting of the Olt fault (N-S) simultaneously with the Cozia Fault (E-W).

Balintoni et al. (1986) separated two major thrust nappes in the area of the Olt Valley, built up of several pre-Laramian units and emplaced during Intra-Maastrichtian, that is Laramian time. The upper nappe overlies the Vasilatu Formation (Szász, 1976) which belongs to the Coniacian-Santonian Lotru Nappe and is transgressively covered by the Brezoi Formation of Campanian-Maastrichtian age. This upper nappe was named the Boia Nappe. It encloses the whole crystalline assemblage found east of the Olt, wherein the structures strike mainly E-W.

Simultaneously, the crystalline classification of Balintoni (1984) was adopted suggesting that the Proterozoic crystalline of the Getic-Supragetic realm may be divided into two major types: one rich in crystalline limestones, amphibolites, micaschists, paragneisses, quartzites, that lacks migmatites, representing the Negoi metalithofacies; the other relatively depleted of carbonatic and amphibolite rocks, mainly gneissic and highly migmatized, representing the Sebeş-Lotru metalithofacies. The crystalline of the Măgura Caineni and Argeş Getic Nappes are related to Sebeş-Lotru metalithofacies, whereas the crystalline of the Moldoveanu and Uria Nappes to the Negoi metalithofacies.

Hann and Balintoni (1988) show that, besides the Getic, Uria and Căineni Nappes, among the pre-Laramian units of the Lotru Nappe one may enclose also the Măgura Nappe, which develops southwards, but up to a place situated N-W of Căineni. The authors also mention that the Sibişel Lithogroup is homogeneous as concerning the rock assemblages. It represents a Precambrian crystalline early metamorphosed in the almandine amphibolite facies, later highly retrograde and laminated. Metamorphites of these series are correlated with the Suru Lithozone and therefore represent a typical subdivision of the Negoi metalithofacies.

Cu-Au ores of Valea lui Stan type, that is of tectonic origin, were recently described by Udubaşa and Hann (1988), who also exposed a genetic model. Fe ore of the Sibişel Lithogroup was described by Hann, Szász (1984), Hann et al. (1989, unpubl. data), arguing for the role played by the retrograde metamorphism as metallogenic factor. Mânecan (1989) argued in detail for the epimetamorphic biphasic character of the Sibişel Series and separated three formations: the basal initial migmatite formation, the middle terrigenous formation and the upper carbonatic-graphitic formation.

Gheuca (1988) described the lithostratigraphy and the tectonics of the western part of the Făgăraş Mts. and treated tangentially a segment of the Olt Valley (between Călineşti and Lotrioara Valley) in a different conception. Accordingly, he denied the existence of the Boia Nappe plane, accepting only partly the plane of the Olt Fault ("Boiţa-Grableşti fault").

Hann et al. (1989, unpubl. data) presented a study of the Sibişel series between Tălmaci and Brezoi, with an enclosed regional at the map scale 1:50,000. These data were taken from the guide "Precambrian Metamorphics in the South Carpathians" (Balintoni et al., 1989), two synthesis reports (Balintoni et al., 1990 a, b, unpubl. data) and represent the basis of the considerations listed below.

The structure, petrography and lithostratigraphy of the Cozia Massif were recently discussed by Hann (1990).

2. The Structure of the Region

West of the river Olt a few kms wide sequence consists of several tectonic units of various thickness, sometimes reduced in places up to complete vanishing. The thinner tectonic zone is at the same time the most important one and occurs in the southern part of the investigated area, along Valea lui Stan. Similar situations are to be found in the north too, e.g. basins of Călineşti and Balota Valley, a tributary of Vlad Valley, on the Iacob creek, right tributary of it.

Looking at the structural architecture of the region from down upwards it is to be mentioned that various tectonic units are geometrically superposed from the west towards the east.

Following the inner structure of the nappe pile from the right bank of the river Olt, there is to be remarked that the Getic Nappe is located in lower as well as western position and consists of mesometamorphic rocks of the Sebeş-Lotru Lithogroup. Above it occurs the Măgura Nappe – a tectonic unit of rather restricted character: west of Câineni (north of Coasta Câinenilor) this nappe vanishes beneath the Uria Nappe. Petrographically, the Măgura Nappe consists of paragneisses and augen gneisses, rocks that are similar to the Cumpăna-Cozia type. This reseamblance permitted the correlation between those two petrotypes.

The nappe sequence is continued by the Uria Nappe which covers both the Măgura Nappe between Sadu Valley and NW of Câineni, and the Getic Nappe, NW of Câineni to Valea lui Stan, where the whole structure is covered transgressively and discordantly by Upper Cretaceous rocks of Coniacian age. It is to mention that the Uria Nappe has some discontinuity in its occurrence, as in the basins of Balota and Valea lui Stan Valleys, whereas the right bank of the Călineşti Valley exposes an outlier. The noteworthy tectonic window formed by schists of the Uria Nappe in the basin of the Balota Valley was delineated by Mânecan (1984) and is found beneath the Câineni Nappe. Its thrust plane represents in fact the initial Supragetic plane of Streckeisen (1934). This tectonic unit is built up of medium-grade metamorphic rocks, rather similar to those of the Cumpăna Lithogroup: paragneisses, augen gneisses, amphibolites and quartz-feldspar gneisses. North of Râul Vadului Valley augen gneisses prevail, whereas south of it paragneisses with interbedded amphibolites or quartz-feldspar gneisses are common.

West of Balota, within the Câineni Nappe a tectonic complication was noticed, namely the Balota scale. This was described as "Călineşti Unit" by Hann and Szász (1984), developing up to Valea lui Stan, but was subsequently included in the "Caineni Unit" by Hann and Balintoni (1988). NW of Robeşti and above the Caineni Nappe an outlier occurs, represented by augen gneisses alternating with paragneisses characteristic of the Cumpăna lithogroup and belonging to the Argeş tectonic unit. This outlier partly covers the Uria Nappe. North of the Vadului river, the Câineni Nappe is bordered towards east by the Olt Fault, a fault that crossed along the river Olt this segment of the Central South Carpathians.

In the Brezoi region, in the left bank of the Vasilatu Valley (Valea Boului), beneath the thrust plane of the Caineni Nappe, Middle-Upper Cenomanian sediments (Szász, in Savu et al., 1977) are found. The same plane is covered south of Valea lui Stan by undisturbed Coniacian sediments, suggesting a Turonian, that is Mediterranean age. This situation is interpreted by Balintoni et al. (1989) as a minor local reworking of an older plane.

East of the Olt Fault, facing from south to north, the Argeş Nappe occurs first. It comprises micaschists of Măgura Câinenilor type (Cumpăna Lithogroup) and then from the left bank of Râul Vadului northwards the Moldoveanu Nappe occurs, built up of mesometamorphic rocks of Suru Lithozone (Făgăraş Lithogroup of the Negoi metalithofacies). The Olt Fault has its eastern wall downthrown, acting simultaneously with the Cozia Fault whose northern wall is also downthrown (1000 m). This compartment underwent a 1 km westward slip, too. North and east of Câineni, in the southern bank of the river Olt, the thrust plane of the Boia Nappe (Hann, Szász, 1984) occurs in downthrown position, too. From the systematic and age of thrusting viewpoints, Balintoni et al. (1986), Hann and Balintoni respectively (1988), discriminate two major tectonic units, that is the lower Laramian unit, named Lotru Nappe, and the upper Laramian unit, named Boia Nappe, each of them containing several pre-Laramian (Austrian) units. Thus, the Lotru Nappe comprises the Austrian: Getic, Măgura, Uria and Caineni Nappes and the Boia Nappe includes the pre-Laramian Austrian units of the Argeş and Moldoveanu. The thrust planes of the Austrian Nappes are transgressively and discordantly covered by Coniacian-Santonian sediments. The Laramian thrust plane separating the Boia and Lotru Nappes is transgressively and discordantly covered by Campanian-Maastrichtian sediments.

According to Hann and Balintoni (1988), the Uria Unit, that is Sibişel Lithogroup, and the Măgura Unit (augen gneisses of Cumpăna-Cozia type) are to be found in the north of the Sebeş massif, in the northern downthrown wall of a reverse fault.

West of Sadu, near Răşinari, the Măgura, Uria and Căineni Nappes occur again due to a N-S fault. In this area the thrust plane of the Caineni Nappe is covered by Cenomanian sediments.

According to Balintoni et al. (1989), the above-mentioned units were also recognized still farther west, in the Semenic and Poiana Ruscă massifs. The Bocşa Unit has been correlated with the Uria Nappe and the Tilva Drenii Unit with the Măgura Nappe.

3. Lithostratigraphy

The informal lithostratigraphic classification proposed by Balintoni, Berza and Hann, in Balintoni et al. (1989) was adopted in the present paper.

Concerning the Getic Nappe, a sequence from the Sebeş-Lotru Lithogroup, assigned to the median paragneisses complex, occurs in the right bank of the Olt Valley. This complex thickness eastwards and vanishes beneath Supragetic structures. Consequently eastward dips increase and, due to southward twisting, the lithologic levels become N-S oriented. As a result the tabular character noticed westward in the area of Cindrel-Negovan-Sterpu massifs disappeared.

The Cumpăna Lithogroup is found in the Măgura, Căineni and Argeş Nappes. Măgura Nappe is built up of augen gneisses, banded gneisses and paragneisses, therefore containing a typical sequence of the Cozia (Vâlsan) Complex, located at the basal part of the Cumpăna Lithogroup. Căineni Unit is also composed of schists characteristic of the Cozia (Vâlsan) Complex; anyhow, it contains a larger sequence of the respective schist pile. Thus, besides augen gneisses developed especially north of Căineni, there are to be noticed quartz-feldspar white gneisses, amphibolites, micaschists distributed on paragneisses background. It is likely that part of the paragneisses and micaschists belongs to the lithozone situated above Măgura Căinenilor micaschists and paragneisses.

Argeş Nappe is composed of Cozia (Valsan) Complex and Măgura Căinenilor Lithozone from the Cumpăna Lithogroup. Somehow N and NE of Căineni micaschists and paragneisses of Măgura Căinenilor Lithozone prevail, whereas southwards paragneisses and augen gneisses of the Cozia (Vâlsan) Complex are common. This complex is recognized in the Cozia massif too; both the river Olt banks contain augen gneisses with subordinate amounts of amphibolites, paragneisses described by Hann (1990) as "augen gneisses formation", with a lower position in the Cozia massif.

Sibişel Lithogroup builts up the Uria Nappe and can be correlated with the Făgăraş Lithogroup, precisely with Suru Lithozone, partly with Sambăta Lithozone of the Moldoveanu Nappe. Within the Sibişel Lithogroup, the basal part consists of Leptino-amphibolitic Lithozone (amphibolite+crystalline limestones, quartz+feldspar white gneisses on paragneissic background) and the upper part of Gneissic Lithozone, formed of a prevailingly paragneissic pile with intercalated amphibolites and quartz-feldspar gneisses.

Făgăraş Lithogroup, represented by Suru Lithozone, makes up the Moldoveanu Nappe, bounded westwards by the Olt Fault. It is characteristic of alterning paragneisses, amphibolites and crystalline limestones, developed in parallel with metamorphic rocks of the Argeş Nappe, situated in the south, and strike E-W. Worth mentioning is the contrasting aspect of the above-listed rocks as compared to lithologic levels of the Căineni, Sibişel and Măgura lithozones of N-S strike.

4. Age Problems

Palynologic studies carried out by Naumova et al. (1964), Dessila-Codarcea and Iliescu (1969) based on samples of a restricted area around Răşinari made us consider the Sibişel Lithogroup as Upper Proterozoic. In recent times Vaida and Hann (1995) were able to use samples for the whole Sibişel Lithogroup, the Acritarche association indicating again the Upper Proterozoic. These authors consider that Sibişel Lithogroup schists were formed during the Cadomian orogenesis (the same as Suru Lithozone metamorphic rocks, specific to the Upper Proterozoic, too).

Sibişel Lithogroup and Suru Lithozone resulted from a different geotectonic realm as age and evolution comparing with that one of the crystalline schists of the Sebeş-Lotru and Cumpăna Lithogroups, which are considered earlier metamorphosed, during the Grenville orogenesis.

5. Metamorphism and Deformation

All the above-discussed schists underwent an initial medium-grade metamorphism during Proterozoic time. Metamorphic evolutions were generally similar, if we consider that the two metamorphic events – M1 and M2 – performed in almost identical thermodynamic conditions through the whole rock piles.

Subsequently a M3 metamorphic event acted too, but its character was unhomogeneous and retrograde-dynamic. New minerals formed along S3 foliation do not exceed the chlorite zone level. This retrogression, presumably Variscan (Balintoni et al., 1989), affected especially schists of the Sibişel Lithogroup from the Uria Nappe. Following this one, an Alpine dynamic metamorphism acted with a predominant cataclastic character (Hann, in Berza et al., 1984). Along alpine thrust planes significant mylonites occurred with slight recrystallization. This is probably due to water lack at the superficial level where alpine shearing acted. The schist piles of the Uria Nappe (Sibişel Lithogroup) were significantly affected by this type of metamorphism.

Lotru Lithogroup metamorphic rocks of the Getic Nappe contain the following index minerals: kyanite, staurolite, sillimanite. S1 and S2 foliations may be rarely observed in the confine of a single outcrop.

Schists of the Cumpăna Lithogroup belonging to Măgura, Căineni and Argeş units were also affected by M1 and M2 metamorphism and exhibit frequently an augen migmatic structure, characteristic of the Cumpăna-Cozia gneisses. Among index minerals, staurolite, subordinate kyanite and sillimanite are to be mentioned. The whole rock complex underwent subsequent Alpine laminations.

Sibişel Lithogroup of the Uria Nappe exhibits a complex metamorphic evolution, delimiting it therefore from schists that built up the other tectonic units. First of all these rocks underwent a medium-grade metamorphism during two metamorphic events (Hann et al., 1989, unpubl. data). Following this, the schist pile underwent a retrograde mesometamorphism, mainly dynamic, but unhomogeneous, since a lot of sequences were only slightly influenced or remained unaffected. As mentioned above, the age of this metamorphism seems to be Variscan. If the Variscan metamorphism represents more or less a hypothesis, the dynamic Alpine metamorphism is a certainty. This event took place during thrusting and left significant evidence. The field exposures show repeated metamorphic phases with retrograde character. The early ones, that may be Variscan, vehiculated more water, yielding new mineral occurrences. But anyhow, the identification of such metamorphic events is difficult. Along the whole process resulted both zones with partial alterations and crystalline terrains, initially mesometamorphic, converted into typical epimetamorphic schists. The initial metamorphism, linked to M1 and M2 events, is represented by index minerals, namely staurolite, garnet, biotite. These minerals occur as relicts in the highly retrograded schists. They are replaced by sericite, clorite, albite, magnetite. Biotite, garnet, staurolite-bearing schists in undisturbed sequences of "shadow-pression" zones are noticed between Sadu and Lungşoara Valleys and in the basin of the Mogoş Valley. Practically the main problem was how to follow along strike certain characteristic lithological levels or associations, no matter how great the intensity of alterations was. It can be concluded that through the whole area the pile of the Sibişel Lithogroup is homogeneous. This situation is similar to the one mentioned by Hann and Balintoni (1989) in the Răşinari region, situated westwards.

Worth mentioning is the blastomylonitic microblastic gneiss occurrence near the limit of the thrust plane where they form discontinuous levels or lenses in the Uria Nappe, that is in the Sibişel Lithogroup. These rocks may exhibit a schistose as well as a granoblastic character and were generated upon paragneiss and amphibolite background. Two biotite generations were noticed, with the second one and minute, marking two oblique "S" planes. Garnet, plagioclase and quartz recrystallize beside biotite, too. Microblastic gneisses formed near the thrust plane during cataclasis and deformation, immediately followed by intensive recrystallization due to dynamic metamorphism. Blastomylonitic recrystallization of biotite and garnet indicates thermodynamic conditions characteristic of the medium-grade metamorphism. As a result, these rocks were generated at deeper levels and are not so typical of phenomena linked to the Alpine tectogeneses. It is likely that Uria Nappe thrusting plane was formed during several stages, the first one during pre-Alpine times. Anyhow, field evidence strongly suggests that micro-blastic gneisses underwent Alpine deformations, even if they belong to a late stage.

Significant is also the fact that the crystalline limestones of the Sibişel Lithogroup underwent alterations near the thrust plane. The dynamic metamorphism dissolved them partly, then allowed their recrystallization. These rocks generally behaved in a plastic way, explaining why mechanic enclaves of mylonitized paragneiss or amphibolite composition are enclosed by limestones. Such situations are well exposed along the Sadu Valley, and were also described in the Căpăţânii Mts. near the tectonic plane of the Ursu scale from the Getic Nappe (Hann et al., 1987).

The effects of the dynamic metamorphism are obvious along the Olt Fault, too. Quartz lenses with breccia texture of tectonic origin are spread along the fault plane and were formed during the ruptural activity. Such quartz lenses are found in the left bank of the Olt Valley, between Căineni and Robeşti and also north of Căineni.

Conclusions

This paper is an attempt to present a new view concerning the structure, petrology and litho-stratigraphy of this area located in the central part of the South Carpathians.

The Olt Valley has been intensively studied as compared to other regions of the South Carpathians. It deeply eroded this montain range, yielding a characteristic cross section through a significant nappe pile.

From the west toward the east, along a few km, and from bottom to top, the Getic Nappe, followed by Măgura, Uria and Căineni Nappes of Austrian age, forming the Laramian Lotru Nappe, are to be recognized. Farther eastwards and in Upper position, the Laramian Boia Nappe is encountered. This is developed especially on the western bank of the river Olt and contain northwards the Austrian Moldoveanu Nappe and southwards the Austrian Argeş Nappe.

A significant fault along the river Olt striking N-S and exhibiting a downwards displacement of the eastern block. It acted simultaneously with another major fault, that is the E-W striking Cozia Fault.

The Cozia Fault moved 1000 m upwards the Cozia gneiss massif that represents its southern block. The Moldoveanu and Argeş Nappes enclose the whole bulk of crystalline rocks striking E-W (that is Făgăraş and Cozia Massifs). The Getic, Măgura, Uria and Căineni Nappes contain N-S striking structures of the Cibin and Lotru massifs.

The Getic, Măgura, Căineni and Argeş Nappes consist of Proterozoic crystalline rocks, represented especially by gneisses and highly migmatised formations belonging to the Sebeş-Cibin Lithogroup. Măgura, Căineni and Argeş Nappes are built up of gneisses belonging to the Cumpăna Lithogroup, with common augen or ribbon character.

The Sibişel Lithogroup is a component of the Uria Nappe. This group as well as the Suru Lithozone of the Moldoveanu Nappe located almost completely east of the river Olt, in the Făgăraş massif, belongs to the Negoi metalithofacies. Such facies comprises alternations of crystalline limestones, amphibolites, micaschists and paragneisses. The Sibişel Lithogroup is to be correlated with the Suru Lithozone based on lithologic criteria, palynology (the age is Upper Proterozoic, that is Cadomian orogeny) and metamorphic degree.

The above discussed schists underwent early metamorphism of medium-grade during Proterozoic times, namely two separate metamorphic events similar from the thermodynamic view point. The Sibişel Lithogroup is characterized by a complex metamorphic evolution. Above the medium-grade metamorphism of presumable Grenvillian age is superimpressed a retrogression of prevailingly dynamic and inhomogeneous character and Alpine age.

Occurrences of microblastic schists with blastomylonitic character located within the metamorphic pile of the Sibişel Lithogroup, indicate the polystadial formation of the Uria Nappe in pre-Alpine times.

References

Balintoni, I., Hann, H. P., Gheuca, I., Nedelcu, L., Conovici, M., Gridan, T., Dumitraşcu, G. (1986) Considerations on a Preliminary Structural Model of the South Carpathians Crystalline East of the Olt River. *D. S. Inst. Geol. Geofiz.*, 70–71/5 (1983, 1984), p. 5–21, Bucureşti.

— **, Berza, T., Hann, H. P., Iancu, V., Kräutner, H. G., Udubaşa, G. (1989)** Precambrian Metamorphics in the South Carpathians. *Guide to Excursions*, Inst. Geol. Geofiz., p. 1–83, Bucureşti.

Berza, T., Iancu, V., Hann, H. P., Seghedi, A. (1984) Dynamic and Retrograde Metamorphism Examples from the Romanian South Carpathians. *An. Inst. Geol. Geofiz.*, LXIV, p. 25–32, Bucureşti.

Codarcea, Al., Lupu, M., Dessila-Codarcea, M., Lupu, D. (1967) Unitatea supragetică în Carpaţii Meridionali. *St. cerc. geol., geofiz., geogr., (Geol.)*, 12, 2, p. 387–392, Bucureşti.

Dessila-Codarcea, M. (1961) Contribuţii la stratonomia şi tectonica şisturilor cristaline ale Carpaţilor Meridionali centrali la vest de Olt. *St. cerc. geol., geofiz., Acad. R. P. R.*, VI/3, p. 553–560, Bucureşti.

— **(1962a)** Incercare de reconstituire paleogeografică şi orogenetică a Carpaţilor Meridionali centrali. *St. cerc. geol., Acad. R. P. R.*, VII/3–4, p. 569–576, Bucureşti.

— **(1962b)** Incercare de reconstituire paleogeografică şi orogenetică a Carpaţilor Meridionali centrali. *St. cerc. geol., Acad. R. P. R.*, VII/3–4, p. 569–576, Bucureşti.

— **(1962c)** Contribuţii la cunoaşterea structurii formaţiunilor metamorfice din regiunea Brezoi-Călineşti-Robeşti, Comun. Acad. R. P. R., XII/5, p. 583–588, Bucureşti.

— **(1965)** Studiul geologic şi petrografic al regiunii Răşinari, Cisnădioara-Sadu. *Memorii*, VI, Inst. Geol., 96 p., Bucureşti.

— **, Iliescu, V. (1967)** Asupra prezenţei depozitelor metamorfice ale Paleozoicului inferior în Carpaţii Meridionali centrali (regiunea Răşinari, Cisnădioara, Sadu). *St. cerc. geol., geofiz., geogr., (Geol.), Acad. R. S. România*, 12/2, p. 311–319, Bucureşti.

— **, Iliescu, V. (1969a)** Noi date microfloristice asupra vârstei complexului calcaros al seriei de Sibişel. *St. cerc. geol., geofiz., geogr., (Geol.), Acad. R. S. România*, 14/1, p. 279–282, Bucureşti.

— **, Iliescu, V. (1969b)** Asupra vârstei metaflişului intracarpatic. *St. cerc. geol., Acad. R. S. România*, 14/1, p. 13–19, Bucureşti.

Gheuca, I. (1988) Versantul sudic al munţilor Făgăraş – litostratigrafie şi tectonică. *D. S. Inst. Geol. Geofiz.*, 72–73/5, p. 93–116, Bucureşti.

Ghika-Budeşti, St. (1940) Les Carpates méridionales centrales (recherches pétrographiques et géologiques entre le Parâng et le Negoi). *An. Inst. Geol. Rom.*, XX, p. 175–220, Bucureşti.

Hann, H. P., Szász, L. (1981) Originea breciilor din masivul cristalin Cozia-Năruţiu (Brezoi-Vâlcea, Carpaţii Meridionali). *St. cerc. geol., geofiz., geogr., (Geol.)*, 26/2, p. 249–261, Bucureşti.

— , **Szász, L. (1984)** Geological structure of the Olt Valley between Câineni and Brezoi (South Carpathians). *D. S. Inst. Geol. Geofiz.*, LXVIII, 5, p. 23–39, Bucureşti.

— , **Balintoni, I. (1988)** Geological structure of the Olt Valley between Câineni and Răşinari (South Carpathians). *D. S. Inst. Geol. Geofiz.*, 72–73/5, p. 119–130, Bucureşti.

— , **Ricman, C., Pană, D., Săbău, G., Bindea, G., Tatu, M. (1988)** Structural and petrographic study of the Getic Nappe metamorphics in the Căpăţâna Mountains (South Carpathians). *D. S. Inst. Geol. Geofiz.*, 72–73/5 (1985–1986), p. 131–143, Bucureşti.

— **(1990)** Cozia Gneiss (South Carpathians): Petrography, Structure, Genesis. *D. S. Inst. Geol. Geofiz.*, 74/5, p. 41–57, Bucureşti.

Kräutner, H. G. (1980) Lithostratigraphic Correlation of the Precambrian of the Romanian Carpathians. *An. Inst. Geol. Geofiz.*, LVII, p. 229–297, Bucureşti.

Lupu, M., Popescu, B., Szasz, L., Hann, H., Gheuca, I., Dumitrică, P., Popescu, G. (1978) Harta geologică a României, sc. 1:50.000, 126a, foaia Vânturariţa (Olăneşti), Inst. Geol. Geofiz., Bucureşti.

Mânecan, C. D. (1977) Asupra prezenţei unor roci metamorfice acide în seria de Sibişel (Munţii Cibin). *St. cerc. geol., geofiz., geogr., (Geol.), Acad. R. S. România*, 22, p. 131–137, Bucureşti.

— , **Bordea, R., Mânecan, R. (1979)** Mineralizaţia de fier din seria de Sibişel (Valea Uria - munţii Cibin). *St. cerc. geol., geofiz., geogr., (Geol.)*, 24, p. 75–80, Bucureşti.

— **(1984)** Litostratigrafia şi tectonica formaţiunilor metamorfice din regiunea de la vest de Olt, între V. Vadului şi valea Călineşti (Carpaţii Meridionali). *D. S. Inst. Geol. Geofiz.*, LXVIII, 5, p. 39–63, Bucureşti.

— **(1989)** Consideraţii asupra petrogenezei seriei de Sibişel. *St. cerc. geol., geofiz., geogr., (Geol.)*, 34, p. 65–75, Bucureşti.

Mrazec, L. (1897) Essai d'une classification des roches cristallines de la zone centrale des Carpathes roumaines. *Bull. Soc. Sci.*, VI, p. 190–193, Bucureşti.

— , **Murgoci, G. (1898)** Dare de seamă asupra cercetărilor geologice din vara anului 1897, III, Munţii Lotrului (Extras).

Murgoci, C. M. (1907) La grande nappe des Carpathes Meridionales. Contributions á la tectonique de Carpathes Meridionales. *C. R. Acad. Sci. Paris*, 3, VIII (1905).

Naumova, S. N., Codarcea-Dessila, M., Iliescu, V. (1964) Resturi de plante din complexele de şisturi cristaline din Carpaţii Meridionali centrali. *St. cerc. geol., geofiz., geogr., (Geol.)*, 22, p. 131–137, Bucureşti.

Popescu-Voiteşti, I. (1918) La nappe du conglomerate des Bucegi dans la vallée de l'Oltu avec des nouvelles données sur la structure de cette vallée dans la region des Carpathes Meridionales. *An. Inst. Geol. Roum. (1914)*, VIII, p. 57–116, Bucureşti.

Popescu, B., Szász, L., Hann, H., Schuster, A. C. (1977) Harta geologică a României, sc. 1:50.000, 126b, foaia Călimăneşti, Inst. Geol. Geofiz., Bucureşti.

Reinhard, M. (1906) Der Coziagneiszug in den rumänischen Karpathen. *Bul. Soc. Ştiinţe*, A, XVI, 3, 4, p. 1–103, (Extras), Bucureşti.

Savu, H., Schuster, A. C., Szász, L. (1977) Harta geologică a României, sc. 1:50.000, 108c, foaia Mălaia, Inst. Geol. Geofiz., Bucureşti.

Schmidt, O. (1930) Scurtă expunere asupra rezultatelor cercetărilor geologice făcute în regiunile cristaline ale Carpaţilor Meridionali. *D. S. Inst. Geol.*, XVII, p. 80–99, Bucureşti.

Ştefănescu, M., Hann, H. P., Gheuca, I., Szász, L., Ştefănescu, M. Harta geologică a R. S. România, sc. 1:50.000, 108d, foaia Titeşti, Inst. Geol. Geofiz., Bucureşti.

Streckeisen, A. (1934) Sur la tectonique des Carpates Méridionales. *An. Inst. Geol. Roum.*, XVI, p. 327–417, Bucureşti.

Szász, L. (1976) Biostratigrafia şi paleontologia cretacicului superior din bazinul Brezoi (Carpaţii Meridionali). *D. S. Inst. Geol. Geofiz.*, LXII/4, p. 189–220, Bucureşti.

Udubaşa, Gh., Hann, H. P. (1988) A shear zone related Cu-Au ore occurrence Valea lui Stan, South Carpathians. *D. S. Inst. Geol. Geofiz.*, 72–73/2, p. 259–282, Bucureşti.

Vaida, M., Hann, H. P. (1995) New palynological data on the Upper Precambrian of the Sibişel Series (Central South Carpathians). *Rom. J. Paleontology*, 76, p. 1–7, Bucureşti.

Zoubek, Vl., Cogné, J., Kozhoukarov, D., Kräutner, G. H. (1988) Precambrian in Younger Fold Belts. European Variscides, the Carpathians and Balkans. John Wiley and Sons, Chichester New-York, Toronto, Singapore.

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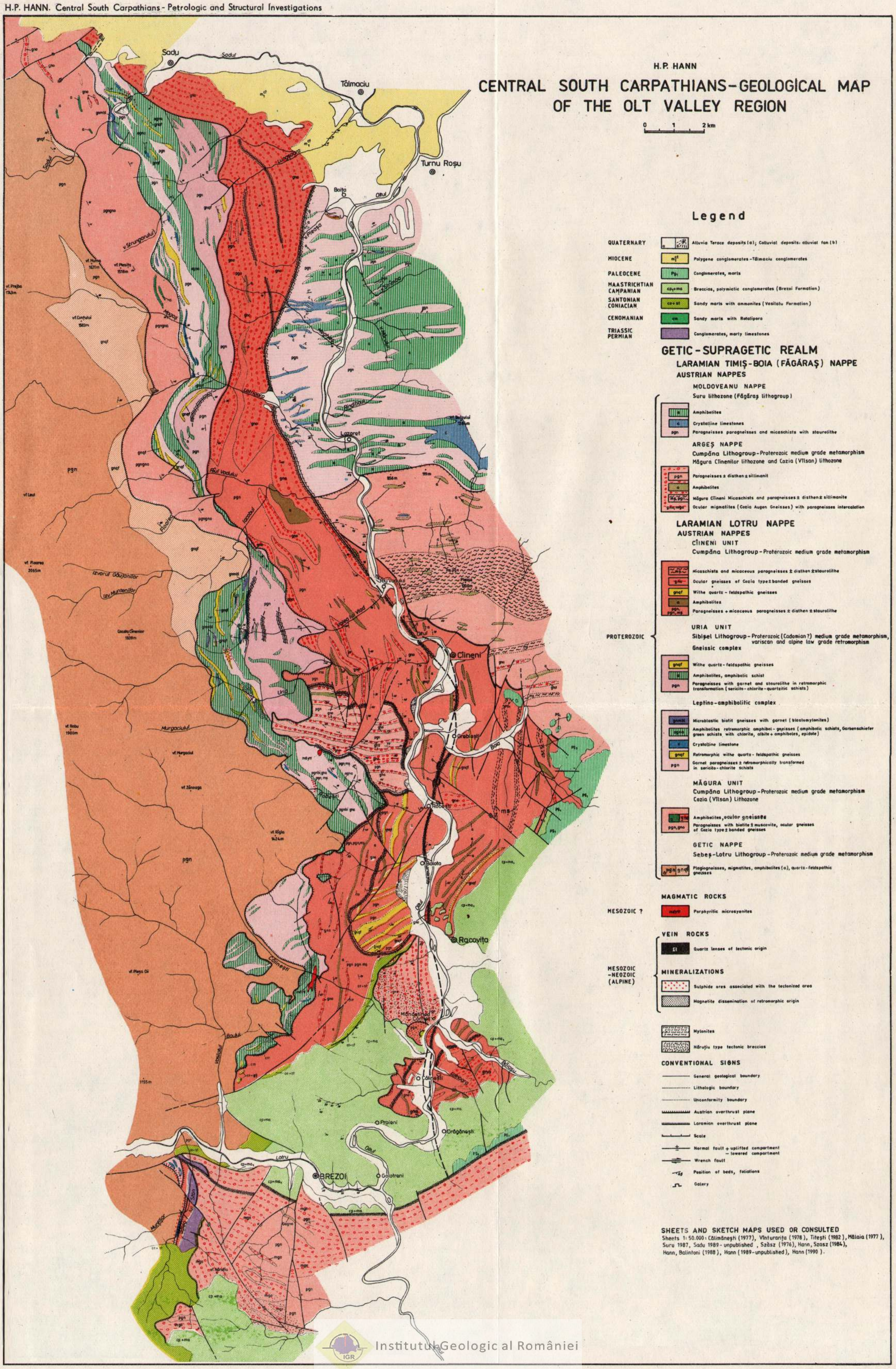
H.P. HANN
CENTRAL SOUTH CARPATHIANS - GEOLOGICAL MAP OF THE OLT VALLEY REGION
0 1 2 km
Sadu
Tălmaciu
Turnu Roşu
Boita
Lazaret
Clineni
Racoviţa
BREZOI
Legend
QUATERNARY
Alluvia Terace deposits (a); Colluvial deposits: alluvial fan (b)
MIOCENE
Polygene conglomerates - Tălmaciu conglomerates
PALEOCENE
Conglomerates, marls
MAASTRICHTIAN CAMPANIAN
Breccias, polymictic conglomerates (Brezoi Formation)
SANTONIAN CONIACIAN
Sandy marls with ammonites (Vasilatu Formation)
CENOMANIAN
Sandy marls with Rotalipora
TRIASSIC PERMIAN
Conglomerates, marly limestones
GETIC - SUPRAGETIC REALM
LARAMIAN TIMIŞ - BOIA (FĂGĂRAŞ) NAPPE
AUSTRIAN NAPPES
MOLDOVEANU NAPPE
Suru lithozone (Făgăraş lithogroup)
Amphibolites
Crystalline limestones
Paragneisses paragneisses and micaschists with staurolithe
ARGEŞ NAPPE
Cumpăna Lithogroup - Proterozoic medium grade metamorphism
Măgura Cîinenilor lithozone and Cozia (Vîlsan) lithozone
Paragneisses ± disthen ± sillimanit
Amphibolites
Măgura Cîineni Micaschists and paragneisses ± disthen ± sillimanite
Ocular migmatites (Cozia Augen Gneisses) with paragneisses intercalation
LARAMIAN LOTRU NAPPE
AUSTRIAN NAPPES
CÎINENI UNIT
Cumpăna Lithogroup - Proterozoic medium grade metamorphism
Micaschists and micaceous paragneisses ± disthen ± staurolithe
Ocular gneisses of Cozia type ± banded gneisses
Withe quartz - feldspathic gneisses
Amphibolites
Paragneisses + micaceous paragneisses ± disthen ± staurolithe
PROTEROZOIC
URIA UNIT
Sibişel Lithogroup - Proterozoic (Cadomian ?) medium grade metamorphism, variscan and alpine low grade retromorphism
Gneissic complex
Withe quartz - feldspathic gneisses
Amphibolites, amphibolic schist
Paragneisses with garnet and staurolithe in retromorphic transformation (sericito - chlorite - quartzitic schists)
Leptino - amphibolitic complex
Microblastic biotit gneisses with garnet (blastomylonites)
Amphibolites retromorphic amphibol - gneisses (amphibolic schists, Garbenschiefer green schists with chlorite, albite + amphiboles, epidote)
Crystalline limestone
Retromorphic withe quartz - feldspathic gneisses
Garnet paragneisses ± retromorphically transformed in sericito - chlorite schists
MĂGURA UNIT
Cumpăna Lithogroup - Proterozoic medium grade metamorphism
Cozia (Vîlsan) Lithozone
Amphibolites, ocular gneisses
Paragneisses with biotite ± muscovite, ocular gneisses of Cozia type ± banded gneisses
GETIC NAPPE
Sebeş - Lotru Lithogroup - Proterozoic medium grade metamorphism
Plagiogneisses, migmatites, amphibolites (a), quartz - feldspathic gneisses
MAGMATIC ROCKS
MESOZOIC ?
Porphyritic microsyenites
VEIN ROCKS
Quartz lenses of tectonic origin
MESOZOIC - NEOZOIC (ALPINE)
MINERALIZATIONS
Sulphide ores associated with the tectonized area
Magnetite dissemination of retromorphic origin
Mylonites
Năruţiu type tectonic breccias
CONVENTIONAL SIGNS
General geological boundary
Lithologic boundary
Unconformity boundary
Austrian overthrust plane
Laramian overthrust plane
Scale
Normal fault + uplifted compartment
− lowered compartment
Wrench fault
Position of beds, foliations
Galery
SHEETS AND SKETCH MAPS USED OR CONSULTED
Sheets 1: 50.000 : Călimăneşti (1977), Vînturariţa (1978), Titeşti (1982), Mălaia (1977), Suru 1987, Sadu 1989 - unpublished, Szàsz (1976), Hann, Szasz (1984), Hann, Balintoni (1988), Hann (1989 - unpublished), Hann (1990).

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COMMENT ON THE AGE OF SOME OPHIOLITES FROM THE NORTH DROCEA MTS

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Key words: Ophiolite. Magmatic arc. Microflora. Radiolarian association. Apuseni Mts.

Abstract: In the present paper paleontological arguments (palynological and radiolarian analyses) are brought in favour of a more recent Neojurassic age of the tholeiitic series from the Căpâlnaş-Techereu Unit in the Drocea Mts. On the other hand the existence of some tholeiitic basalts in the Tithonian flysch of the Criş Nappe indicates a certain synchronism of the tholeiitic and the calc-alkaline series; the transition from one complex to another took place during an unique process. Although the calc-alkaline products are prevalent in the upper part of the ophiolitic succession, the transition shows frequent reccurrences. The nature of the succession (which is) interbedded between the tholeiitic basalts, pleads rather in favour of its formation in a basin occurring within a magmatic arc, than in an oceanic basin.

The northern part of the Drocea Mts in the Valea Zeldişului springs area consists mainly of basalts, anamesites and dolerites which are part of the first stage ophiolitic magmatites. In subsequent interpretations they have been equivalated to ocean floor ophiolitic rocks formed in a first stage, coinciding with the spreading phase of the oceanic zone (Savu, 1980).

In the geological and structural maps of the Mureş area from a more recent paper (Savu, 1983), both the sheeted dyke complex (O_2) and the upper basaltic complex (O_1) are figured in this zone, south of the overthrust line of the Căpâlnaş-Techereu Unit overlying the Criş Unit. In the same area on the mentioned maps, Neocomian and Upper Jurassic jaspers are figured as less developed interbedded with the magmatites of the second stage, consisting of basalts, andesites, spilite pyroclasts (Savu et al., 1970) or so-called "second stage initials" containing pyroclastites and basaltic andesite flows, associated with jaspers (Savu, 1979, in the sheet Roşia Nouă, 1:50,000). They might belong to the IAV_1 complex (Savu, 1983). According to the radiometric age data, the Lower-Middle Jurassic age, namely 180 Ma has been assigned to the tholeiitic series, whereas according to Savu et al. (1979) the Oxfordian-Neocomian age was assignable to the upper series, namely the calc-alkaline one.

During the study of the sedimentary formations from the Drocea Mts and the related eruptive rocks, the authors of the present paper have accomplished some researches in the mentioned area. The Valea Huţului profile has been pointed out from the analysed sections, as the association of ophiolites and sedimentary sequences is largely revealed, and the palynological analysis allowed an accurate age determination of the succession.

Petrographically and chemically typical rocks of the tholeiitic series are encountered in the area which belongs to the Căpâlnaş-Techereu Nappe starting from the contact with the subjacent Criş Nappe, up to the analysed zone. Petrographically (Pl. I, II) they consist of basalts with intersertal and intergrained structures, locally having anamesitic trends. Albitized plagioclase rocks, which are no more than 0.5 mm but frequently 0.2–0.3 mm in length, float in a glassy ground mass or host clynopyroxene grains. Chlorite amygdales occur in places; the presence of some minerals, such as laumonite and prehnite has been established in some samples (Rx) indicating a metamorphism of ophiolites at the level of the zeolitic facies and the beginning of the prehnite-pumpellite facies. These minerals develop both in basaltic flows and in breccias partly constituted of basic rocks of equal type. Unlike the large surfaces occupied by the magmatites belonging to the calc-alkaline series, as figured on the sheet Roşia Nouă, 1:50,000, a spilitized andesite occurs in fact only in one place, containing a cryptocrystalline ground mass in which albitized plagioclase and smaller biotite (locally hornblende) phenocrysts pierce the basalts of the tholeiitic complex. The first stage ophiolites (sensu Giuşcă et al., 1963) do not represent ocean floor magmatites, as mentioned by H. Savu in these papers, but island arc tholeiites, according to other proofs (Cioflică et al., 1980; Cioflică, Nicolae, 1981; Nicolae, 1983; Lupu, 1983).

The authors present chemical analysis of the samples from the tholeiitic complex in the area which is figured as such on the sheet Roşia Nouă, 1:50,000, indicating the following results (analysed by R. Ianc for major elements and I. Bratosin for minor elements):

SiO_2 - 48,96	FeO - 9,70	K_2O - 0,23
TiO_2 - 2,15	MnO - 0,16	Na_2O - 4,59
Al_2O_3 - 14,29	MgO - 6,68	P_2O_5 - 0,23
Fe_2O_3 - 4,96	CaO - 5,28	H_2O - 2,70
S - 0,11	Ni - 14	N - 10
Pb - 2	Co - 23	Yl - 4,6
Cu - 27	Cr - 7	Y - 27
Zn - 20	V - 300	La - 30
Sn - 2	Se - 24	Sr - 83
Ca -21	Zr - 45	B - 14

The sample plotted in the Ti/Cr diagram (Figure) occurs in the field of the weak-potassium tholeiites and not in the oceanic-crust type basalts.

In spite of the relatively large surfaces represented by the first stage magmatites belonging to both the calc-alkaline and alkaline series on the sheet Roşia Nouă 1:50,000 in Valea Huţului we have in fact located only one outcrop with a cryptocrystalline basement namely a spilitized andesite in which albitized plagioclase phenocrystals and small biotites (sporadically hornblende) pierce the tholeiitic complex basalts and therefore being assigned to this series.

According to both the microscopical aspects (Pl. I and II) and the chemical analysis mentioned above, it is obvious that the sedimentary sequence from Valea Huţului is intercalated within tholeiitic-type intersectal basalts, being recently described by Savu (1980, 1983) as occurring only in the ophiolitic complex, namely in the ocean floor basaltic complex (01).

According to some of our previous interpretations (Cioflică, Nicolae, 1981; Cioflică et al., 1980) they might be assigned to the prevalent tholeiitic complex which represents immature island arc products.

The sedimentary succession which is interbedded with the above mentioned basalts consists of siltitic quartzous purple breccias, displayed in beds with thicknesses ranging from 5-50 cm to 2 m. The thin beds show frequent parallel and especially oblique current lamination. Compact greyish-purple clayey schists are associated with breccias having thicknesses exceeding 1 m. Levels of marly breccias with a greenish, in places purple matrix are also observable, being similar to sandstones containing variolitic basalt elements. Some tholeiitic basalt are intercalated within the succession.

The sedimentary succession has a homocline structure with a slight SE oriented inclination and a thickness of 300 m, being followed by similar-type magmatites.

Samples have been taken for radiolarian and palynological analyses in order to establish the ages of the succession and implicitly the ophiolitic magmatites among which the succession is interspersed.

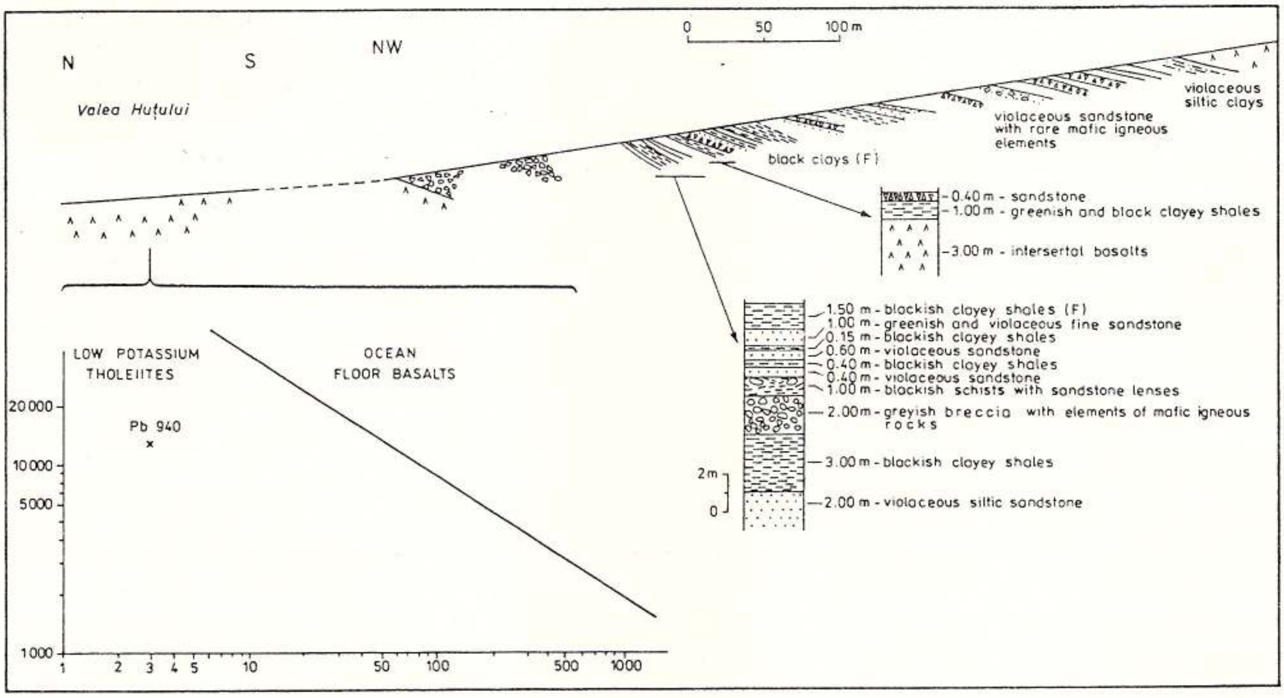
The palynological association from the level 2A107, 2A60 is relatively rich in genera and species and consists of spores, pollen, vegetal tissue fragments, dinoflagellates, acritarches. The vegetal material is strongly altered, turned into black coal – the specimens are dark brown-black coloured – and most of them are broken. Therefore their determinations have become considerably difficult, most of the elements have been only generically established (some of them constituting question marks). From this point of view, the association represents a typical example of a palynological assemblage whose components cannot be entirely established as on the one hand the preservation stage does not allow it and on the other hand some genera and species cannot be traced, although they had been initially deposited and subsequently destroyed. Palynomorphs were not observable in other levels (Figure=F), only palynofacies consisting of vegetal tissues being completely turned into black coal.

The microflora consists of:

– Continental microflora: *Dictyophyllidites* sp., *Deltoidospora* sp., *Auritulinasporites* sp., *Gleichenidites* sp., *Lycopodiumsporites* sp., *Murospora* sp., cf. *Perinopollenites* sp., *Callialasporites* sp., *Cycadopites* sp., *Classopollites torosus* (REISSINGER) BALME, 1957, *emend.* MORBEY, 1975; *Vitreisporites pallidus* (REISSINGER, NILSON, 1958; *Alisporites* sp.

– Marine microflora: cf. *Adnatospaeridium* sp., *Chlamidophorella raritubulae* DODEKOVA, 1975; *Chytroeispaeridia* cf. *chytroeides* (SARJEANT) DOWNIE & SARJEANT, 1965 emend, DAVEY, 1979; cf. *Ctenidodinium* sp.; *Dimidiadinium* sp.; *Escharispaeridia* n. sp.; *Gonyaulacysta* cf. *jurassica (Deflandre)* NORRIS & SARJEANT, 1965; cf. *Gonyaulacysta* sp. 1, cf. *Gonyaulacysta* sp. 2; cf. *Hystrichogonyaulax* sp.; cf. *Meiourogonyaulax* sp.; cf. *Scriniocassis dictyotus* (COOKSON & EISENACK) BEJU, 1971; *Sentusidinium pilosum* (EHRENBERG) SARJEANT & STOVER, 1978, emend ERKMEN & SARJEANT, 1980; *Stephanelytron scarburghense* (SARJEANT) STOVER, SARJEANT & DRUGG, 1977; cf. *Stephanelytron scarburghense*; cf. *Tubotuberella* sp.; *Micrhystridium* sp.

The palynofacies can be characterized by quantitative mostly brown-darkish vegetal tissues. Among the



Section of the Sedimentary Rocks Sequence Interbedded in the Tholeiites from Valea Zeldişului Springs

continental elements, such as spores and pollen, both the Classopollis type and the simple pollen are observable, whereas the remaining genera and species are quantitatively subordinated.

The dinoflagellates are quantitatively subordinated to the continental microflora. Prevalent are the specimens assignable to *Stephanelytron scarburghense* (the specimens belonging to *S. scarburghense* are less numerous) besides the specimens assignable to the *Chlamidophorella raritubulae* and those of type *Sentusidinium* and *Chytroeisphaeridia* cf. *chytroeides*. The remaining genera and species are represented in only one or two cases.

From the point of view of the stratigaphical distribution the most important are the genera *Stephanelytron* and *Chytroeisphaeridia* cf. *chytroeides*. Their stratigraphical distribution is known since the Callovian-Lower Kimmeridgian time span and Bajocian-Tithonic respectively, having an acmezone in the Oxfordian (Stover, Sarjeant & Drugg, 1977; Davey, 1979). Apparently *Gonyaulacysta jurassica* does not exist beyond the Kimmeridgian, whereas *Scriniocassis dictyotus* has been known since the Oxfordian-Tithonic time span (Lentin & Williams, 1981). *Chlamidophorella raritubullae* has been described from the Upper Bathonian from Bulgaria (Dodekova, 1975).

Some taxonomic remarks. The specimens ascribed to the *S. scarburghense* are similar with the respective species in their general carriage, but they lack the particular feature of the "corona" genus. Probably either the membrane building up the corona was destroyed by the diagenesis and the specimens could be assigned to *S. scarburghense* as well, or they might be assignable to another genus, most likely to *Chlamidoporella raritubulae*, which at first sight might indicate several similar characteristics in their general carriage. The specimens assigned to *Chlamidoporella raritubulae* correspond to the particular description. Nevertheless the membrane which embodies the distal processed was not observable in our researches and was probably destroyed. Dodekova (1975) mentions that the ectophragma of the original specimens is in places destroyed. The specimens assignable only to *Chytroeisphaeridia* - species *chytroeides* - are generally dorsoventrally flattened and the precingular archeopyle is observable; these specimens are likely to belong to *C. pericompus*, as both species are morphologically close related (*C. pericompus* has a thinner test).

The microflora from the sedimentary interbed from the ophiolites in the region of the Valea Zeldişului springs seem to be relevant for the age determination of the eruptions, although many of the assemblage parts are difficult to establish. Based on both the association of the *Stephanelytron, Chytroeisphaeridia, Chlamidoporella, Dimidiadinium* and the other dinoflagellate species, a Lower Jurassic age can be ruled out, as the above mentioned genera have generally occurred worldwide since the Middle Jurassic. Both the Aalenian and the Bajocian indicate palynological associations in which the *Nannoceratopsis* specimens are quantitatively prevalent, whereas during the Bathonian in Romania the *Ctenidodinium* specimens were quantitatively prevalent (Beju, 1971; Drugg, 1978) in the present case they occur only sporadically, so that the microflora can be assigned to the Upper Bathonian-Lower Kimmeridgian stratigraphical range, according to the stratigraphical distribution of the species *Stephanelytron scarburghense*, *Chytroeisphaeridia chytroeides* and *Chlamidoporella raritubulae*. Considering that *Stephanelytron* and *Chytroeisphaeridia* have the largest quantitative distribution within the Oxfordian (Stover, Sarjeant & Drugg, 1977; Davey, 1979), the sedimentary intercalations from the ophiolites in the region might be assigned to the Oxfordian. Nevertheless, it is to point out that the assemblage could not supply a more accurate determination than the Upper Bathonian-Lower Kimmeridgian from the point of view of the "qualitative" stratigraphical distribution of its genera and species.

The prevalence of the continental over the marine vegetales from the previously mentioned deposits indicate their sedimentation in a marine environment where the contribution of the continental material was very active. Taking into account that the deposit is not a turbiditic one (in which case the continental material is likely to be driven off shore), the sedimentation took probably place in the neritic area.

Therefore we might ask ourselves if the age of the tholeiitic series which was so far considered as Lower-Middle Jurassic, does not include a larger range for this zone. The radiolarian association, besides jaspers interbedded in the tholeiitic series, located by P. Dumitrică in Tara Valley and Valea Nucului, as right tributaries of the Tănăşeşti Valley, are arguments in favour of this idea.

Specimens such as *Acaenyclile diaphorogona* FOREMAN, *Triacblakei (*PESSAGNO*), Acanthocirus variabilis (Squinabol), Paronaella mulleri* PESSAGNO, *P. bandyi* PESSAGNO, *Tritrabs casmaliaensis* (PESSAGNO), *T. ewingi* (PESSAGNO), *T. hayi* (PESSAGNO), *Tetratrabs gratiosa* BAUMGARTNER, *Angulobrachia purisimaensis* (PESSAGNO), *Higumastra inflata* BAUMGARTNER, *H. imbricata* (OZVOLDOVA), *Hagiastrum plenum* RUST, *Emiluvia antigua* (RUST), *Eucyrtis ? dicera* BAUMGARTNER, *Andromeda podbielensis* (OZVOLDOVA), *Podobursa helvetica* (RUST), *Hsuum maxwelli* PESSAGNO, *Setoccapsa* cf. *carpatica* DUMITRICA, *Parvicingula altissima* (RUST), *Mirifusus guadalupensis* PESSAGNO, *Perispyridium* cf. *dettermani* PESSAGNO & BLOME, *Protunuma costata* (HEITZER), *Obesacapsula morroensis* PESSAGNO, *Sethocapsa pyriformis* (HEITZER), *Tricolocapsa conexa* MATSUOKA, *Gongylothorax sakawaensis* MATSUOKA, *Stilokapsa* KOCHER, *Protunuma* spp. can be most likely assigned to the Upper Callovian and related to biozone A established by Baumgartner et al., 1980, for the Tethys area.

Without considering our elements as relevant markers in order to assign to the tholeiitic series an exclusively Upper Jurassic age, we might consider they justify these considerations.

Furthermore, the location of tholeiitic basalts in the tithonic flysch from the Criş Nappe (Lupu et al., in press) indicates the synchronism (pro parte) of both the tholeiitic and the calc-alkaline series. In this case the ophiolitic zone which has been so far assigned only to the Căpălnaş-Techereu Nappe, includes in fact several structural units, some of their tholeiites being contemporary with the calc-alkaline magmatites of the others.

Finally, the nature of the sedimentary succession interbedded between the tholeiitic basalts pleads rather in favour of a sedimentation within a basin belonging to a magmatic arc, than in an ocean floor basin.

As regards the transition from one magmatic complex to another, this has been possible within an unique process, even if in the upper part of the ophiolite succession calc-alkaline products are prevalent, the transition takes place with frequent reccurrences (Cioflică, Nicolae, 1981).

References

Baumgartner, P. O., De Wevrer, P., Kocher, R. (1980) Correlation of Tethyan Late Jurassic-Early Cretaceous Radiolarian Events. *Cahiers de Micropaléontologie*, 2, p. 23-72, Paris.

Beju, D. (1971) Jurassic Microplanckton from the Carpathian foreland of Romania. *Ann. Inst. Geol. Publ. Hung.*, 54/2, p. 275-317, Budapest.

Cioflica, G., Nicolae, I. (1981) The origin, evolution and tectonic setting of the alpine ophiolites from the South Apuseni Mts. *Rev. Roum. Géol., Géophys., Géogr., (Géol.)*, 25, Bucureşti.

Davey, R. J. (1979) A Re-Appraisal of the genus Chytroeisphaeridia Sarjeant, 1962. *Palynology*, 3, p. 209-218, Madrid.

Dodekova, L. (1975) New Upper Bathonian Dinoflagellata Cysts from Northeastern Bulgaria. *Palaeontology, Stratigraphy and Lithology*, 2, p. 17-34, Sofia.

Drugg, W. S. (1978) Some Jurassic Dinoflagellate Cysts from England, France and Germany. *Palaeontographica*, B 168, p. 1-3, 61-79, Stuttgart.

Herz, N., Jones, L. M., Savu, H., Walker, R. L. (1974) Strontium isotope compossition of ophiolitic and relates rocks, Drocea Mts., Romania. *Bull. Volc.*, XXXVIII/4, 1110-1124, Napoli.

Lentin, J. K., Williams, G. L. (1981) Fossil dinoflagellates Index to genera and species, 1981 edition.

Bedford Inst. of Oceanography Report Series/B1-R-81-12 Aug., 300 p., Darmouth.

Savu, H. (1980) Genesis of the Alpine Cycle Ophiolites from Romania and Their Associated Calc-Alcaline and Alcaline Volcanics. *An. Inst. Geol. Geofiz.*, LVI, p. 55–77, Bucureşti.

— **(1983)** Geotectonic and magmatic evolution of the Mureş zone (Apuseni Mountains) - Romania. *An. Inst. Geol. Geofiz.*, LXI, p. 253–262, Bucureşti.

— **, Vasiliu, C., Udrescu, C. (1970)** Geochimia şi petrologia ofiolitelor din prima etapă de evoluţie a magmatismului iniţial alpin din masivul Drocea (Munţii Apuseni). *D. S. Inst. Geol. Geofiz.*, LVI/1, p. 219–252, Bucureşti.

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Plate I

Fig. 1 – Intersertal basalt with rare clynopyroxene crystals, N+, 35 x
Fig. 2 – Basalt with intergrained texture, N+, 35 x

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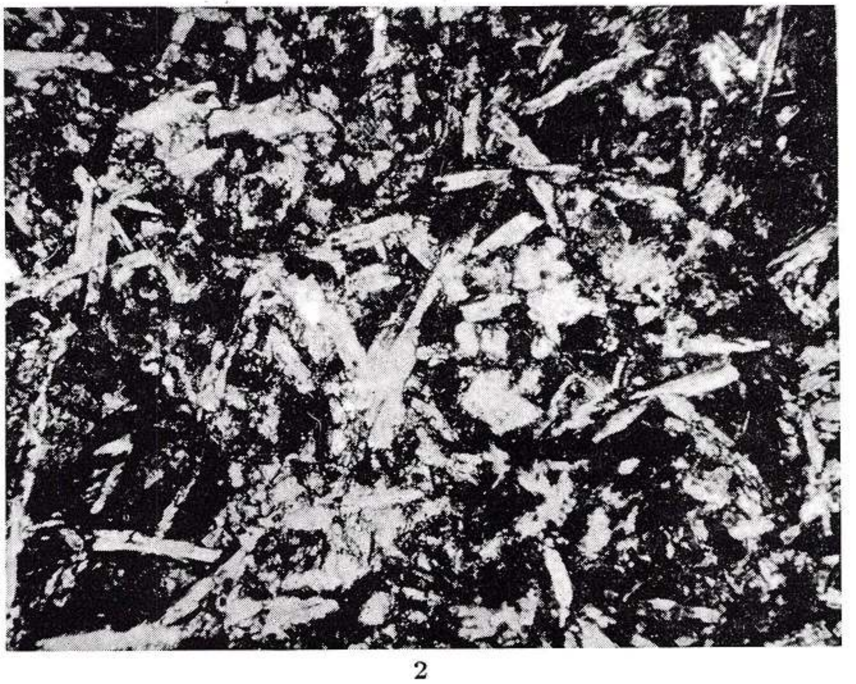

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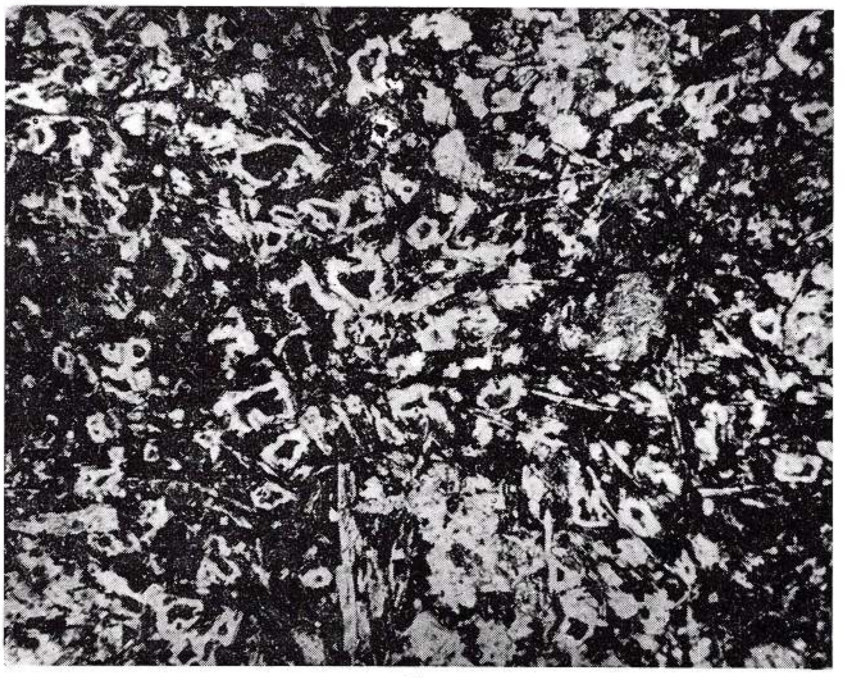
Plate II

Fig. 1 – Intersertal basalt with small clorite amygdales, NII, 35 x
Fig. 2 – Basalt with intergrained texture. N+, 35 x



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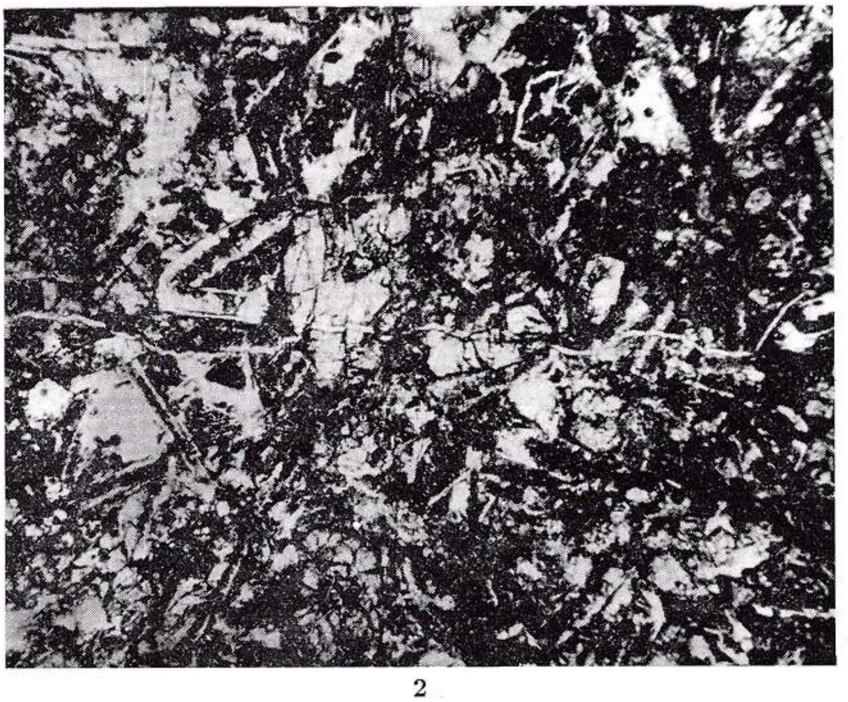

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Plate III

Elements from the sedimentary rocks microflora interbedded in the tholeiites from Valea Zeldişului springs. Oxfordian (?)

Fig. 1 – *Dictyophyllidites* sp. Slide 2A105/1; 121,3/48,8; 30μ, f. 1A68.
Fig. 2 – *Gleicheniidites* sp. Slide 2A105/2; 124,9/46,2; 25μ, f. 1A68.
Fig. 3 – *Lycopodiumsporites* sp. Slide 2A105/5; 111,5/58; 23μ, f. 1A68.
Fig. 4 – *Vitreisporites pallidus* (REISSINGER) NILSSON, 1958. Slide 2A60/16; 111,7/105,2; 18μ; f. 1A65.
Fig. 5 – *Alisporites* sp. Slide 2A60/14; 4,1/110,6; 45μ, f. 1A65.
Fig. 6 – *Cycadopites* sp. Slide 2A105/7; 113,6/67,8; 58μ, f. 1A69.
Fig. 7 – *Classopollis torosus* (REISSINGER) BALME, 1957 emend. MORBEY, 1975.
Slide 2A60/22; 5,5/91,5; 21μ, f. 1A66.
Fig. 8 – cf. *Gonyaulacysta* sp. 1. Slide 2A60/17; 8,8/111; 40 μ, 1A64.
Fig. 9 – *Chlamidoporella raritubulae* DODEKOVA, 1975. Slide 2A105/2; 105,7/52,1; 40μ, f. 1A68.
Fig. 10 – cf. *Stephanelytron* sp. Slide 2A60/15; 126,2/49,9; 35μ, f. 1A65.
Fig. 11 – *Stephanelytron scarburgense* (SARJEANT) STOVER, SARJEANT & DRUGG, 1977.
Slide 2A105/2; 124,9/51,6; 49μ, f. 1A66.

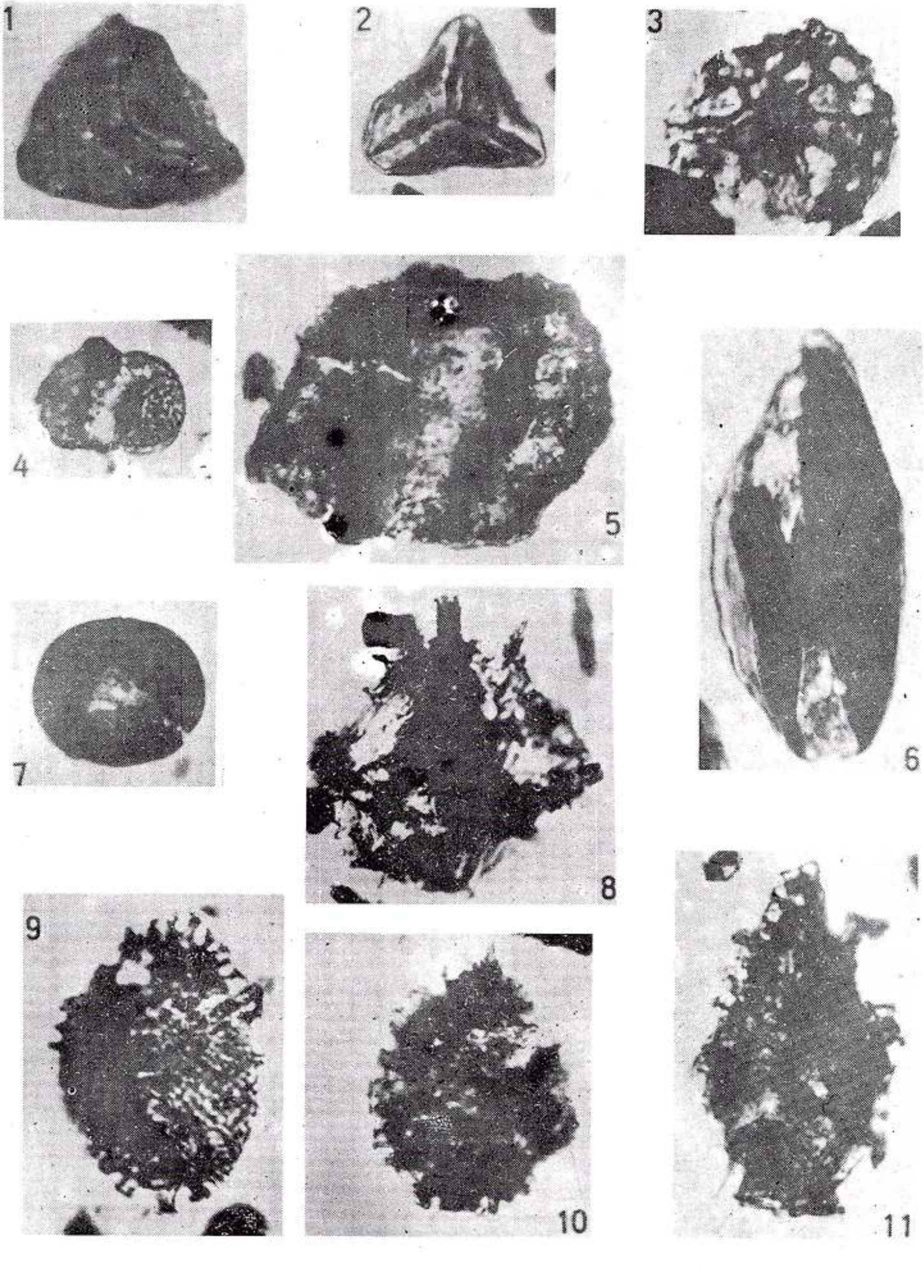
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Plate IV

Elements from the sedimentary rocks microflora interbedded in the tholeiites from Valea Zeldişului springs. Oxfordian (?).

Figs. 1, 2 – *Chytroeisphaeridia* cf. *chytroeides* (SARJEANT) DOWNIE & SARJEANT, 1965 emend. DAVEY, 1979. (Fig. 1 – slide 2A60/27; 107/64,9; 40μ, f. 1A65. Fig. 2 – slide 2A105/18; 114,5/51,6; 35μ, f. 1A68.)

Fig. 3 – *Dimidiadinium* sp. Slide 2A105/5; 115,9/45,2; 63μ, f. 1A67.

Fig. 4 – *Sentusidinium pilosum* (EHRENBERG) SARJEANT & STOVER, 1978 emend. ERKMEN & SARJEANT, 1980. Slide 2A107/1; 118/60; 40μ, f. 1A64.

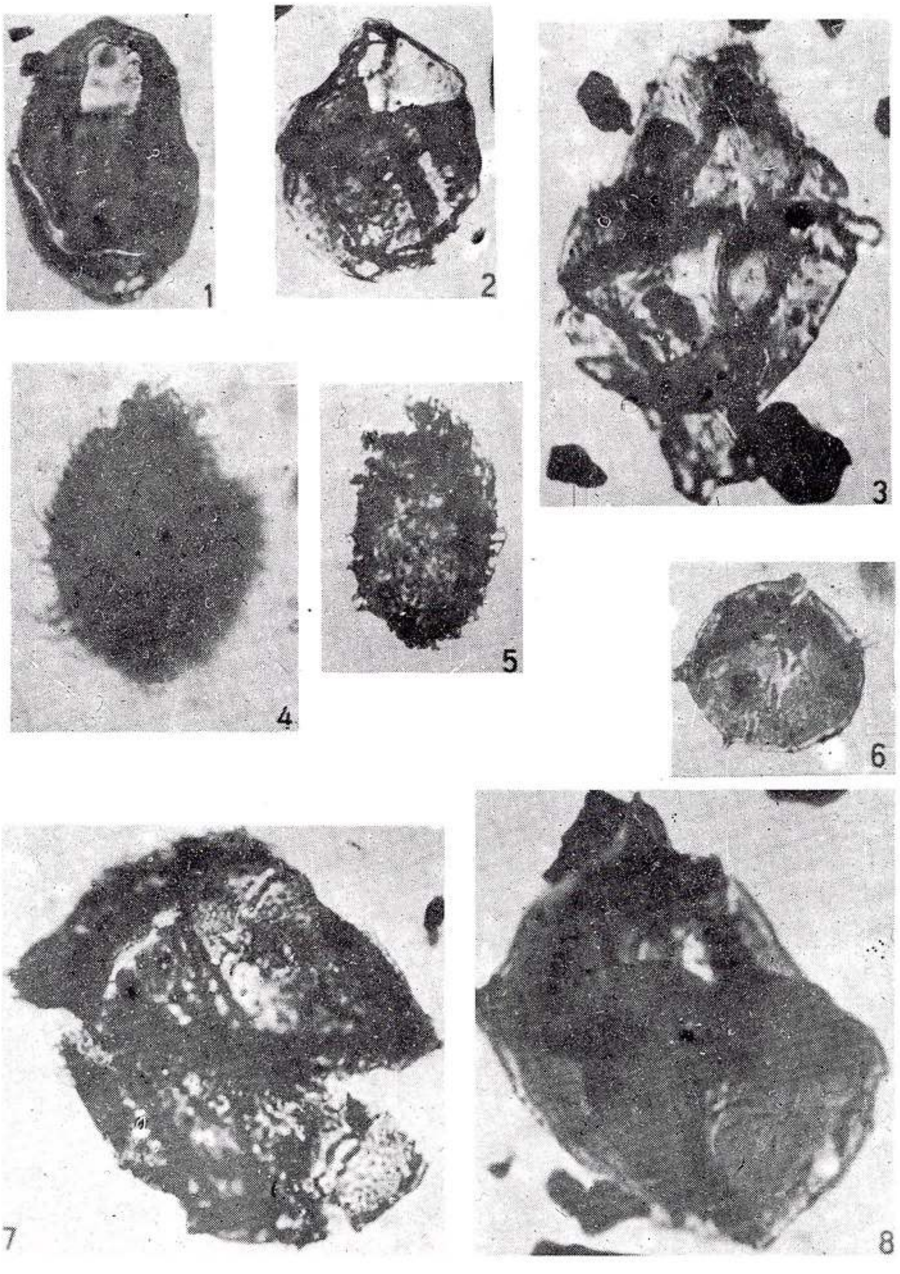
Fig. 5 – *Chlamidoporella raritubulae* DODEKOVA, 1975. Slide 2A60/17; 111,8/71,9; 35μ, f. 1A64.

Fig. 6 – *Micrhystidium* sp. Slide 2A60/15; 115,2/58,8; 25μ, f. 1A65.

Fig. 7 – cf. *Ctenidodinium* sp. Slide 2A107/3; 114,8/64,6; 78μ, f. 1A73.

Fig. 8 – cf. *Gonyaulacysta* sp. 2. Slide 2A107/14; 123,3/59,2; 58μ, f. 1A63.

The coordinates of the specimens are taken from microscope I.O.R. MC 7, seria 0325–0035.
The palynological products contain elements figured in the collection of the palynological lab of the I.G.R.



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TECTONIC SETTING OF THE OPHIOLITES FROM THE SOUTH APUSENI MOUNTAINS : MAGMATIC ARC AND MARGINAL BASIN

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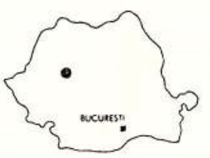

Key words: Ophiolites. Island arc. Magmatic arc. Marginal basin. Ocean floor magmatites. Tholeiites. Calc-alkaline rocks.

Abstract: The ophiolites from the South Apuseni Mts are assigned to two geotectonic settings: magmatic arc and marginal basin. The most complete development of the magmatic arc ophiolites is to be found in the Căpâlnaş-Techereu Nappe. Products of a tholeiitic series are distinguished in the base, being overlain in some sectors by rocks of a calc-alkaline series; Early Cretaceous intermediate acid intrusions were emplaced in the final evolution stage of this magmatic arc. The ophiolites from the Rimetea Nappe, similar to the magmatic arc ones, consist of a poorly represented tholeiitic series. The transition to the calc-alkaline series took place here earlier (in the Middle Jurassic) than in the Căpâlnaş-Techereu Nappe, where, according to some paleontological and K-Ar isotopic ages, the transition took place in the Upper Jurassic (post-Oxfordian). Keratophyres (Oxfordian-Lower Tithonian) occur at the top of these ophiolites, being overlain by Upper Tithonian-Berriasian limestones. The Bedeleu and Hospea Nappes are marked by the prevalence of keratophyres, while the basic rock dikes-basalts-anamesites-dolerites occur subordinately. The magmatic activity in the Curechiu-Stănija Nappe is reduced, being represented by flows and pyroclastics of calk-alkaline porphyric basalts. The marginal basin type ophiolites are to be found in the Criş and Feneş Nappes. In the Criş marginal basin basaltic intercalations occur in prevailingly flysch sediments at various levels in the Callovian-Tithonian interval, followed by a less extended phase of andesite-dacite-rhyolite±albitized, chloritized rocks. In the Feneş Basin, which probably opened in the Upper Jurassic and functioned until the Lower Aptian, beside an olistostrome type volcano-sedimentary formation, prevailingly volcanic magmatic products consisting of spilites occur.
The paper pesents the results of 42 analyses of some rocks from the tholeiitic series and eleven analyses from the basic dykes crossing the keratophyres of the Bedeleu and Hospea Nappes.

Introduction

The ophiolitic rocks from the South Apuseni Mts, which show a great development between the Pătarş locality to the SW and the Turda locality to the NE, have been asigned to three evolution phases (Giuşcă et al., 1963).

After the concept of plate tectonics appeared, the petrological studies became more numerous, which made possible a better knowledge of these rocks and enabled the establishment of the tectonic setting to which they belong.

The ophiolitic rocks were considered of ocean floor type by Rădulescu, Săndulescu (1973); Bleahu (1974); Herz, Savu (1974). Subsequently they were separated by Savu (1980, 1983) and Savu et al. (1982, 1985) as follows: ophiolites of ocean floor type (equivalent to the first phase sensu Giuşcă et al., 1963) and island arc rocks (equivalent to the second and third phase sensu Giuşcă et al., 1963).

Savu (1983) distinguishes within the ophiolitic rocks from the Mureş Zone a lower sheeted dyke complex (O_2) and an upper ocean floor basalt complex (O_1). Over this edifice there are island arc products, that occurred as a result of the bilateral subduction as two branches, a north-western one and a south-eastern one. The former branch is represented by a magmatism considered bimodal (Savu et al., 1986), and the latter, marked by numerous volcanic structures covered by a reef barrier, continues to the north-eastern extremity of the South Apuseni Mts in the Trascău Mts (Savu, 1983). In a synthetic column the mentioned author distinguished within the island arc volcanics of the south-eastern branch a lower andesite-basaltic complex (IAV_1) followed by an upper leucocratic volcanic one (IAV_2) and at the upper part, in the Trascău Mountains Zone, Albian flysch and volcanics deposits of the spilite-keratophyre assemblage.

Other authors, using the ophiolite term in a wider sense, admitting therefore that they may form in various tectonic settings, distinguish in the South Apuseni Mountains island arc ophiolites (Cioflica et al., 1980; Cioflica, Nicolae, 1981) or magmatic arc ophiolites (Lupu, 1983; Lupu et al., 1993) assigned to two series, a tholeiitic one and a calc-alkaline one, equivalent to the first two evolutionary phases, as well as other ophiolites of marginal basin type, equivalent to the third phase sensu Giuşcă et al., 1963 (spilitic complex – Cioflica, Nicolae, 1981; Cioflica et al., 1980; Lupu, 1983; Nicolae, 1983, 1985).

1. Participation of the Ophiolitic Rocks in the Various Tectonic Units of the South Apuseni Mts

The South Apuseni Mts show a very complicated tectonics, numerous units being pointed out, which consisist also of ophiolitic rocks (Lupu, 1976; Lupu in Bleahu et al., 1981). The differences in the petrographic constitution of the ophiolitic rocks from various tectonic units will be further presented.

1.1. Magmatic Arc Type Ophiolites

They are to be found in the Căpâlnaş-Techereu Nappe, which is the largest of all the tectonic units and consists almost exclusively of ophiolitic rocks of a great petrographic variety, the Rimetea Nappe, the Bedeleu Nappe, the Hospea Nappe and the Curechiu-Stănija Nappe. Two series can be distinguished: a tholeiitic one and a calc-alkaline one.

1.1.1. Tholeiitic Series

Situated in the lower part of the magmatic arc ophiolites, the products of the tholeiitic series are largely developed in the Căpâlnaş-Techereu Nappe (Figs. 1, 2), especially in the western part, where there develop mainly tholeiitic, aphyric basalt flows, with intersertal textures, often in pillow lava facies, anamesites, dolerites, small gabbroic intrusions, more rarely peridotitic separations. Pyroclastics occur only subordinately. In the eastern part of a sinuous line connecting the Zam and Basarabasa localities, there are smaller areas in which the rocks of the tholeiitic series crop out, usually from under the products of the suprajacent calc-alkaline series, such as the Visca, Buneşti-Poiana zone or, in the southern part, on the Mureş Valley, between Zam and Valea Lungă.

In the Rimetea Nappe (Fig. 2), which lies between the Turzii Gorges to the north and Poiana Aiudului to the south, being crossed in the median part by the Arieş Valley, between Buru and Moldoveneşti, the products of the tholeiitic series occur on very restricted areas on the Podeni Valley, Părâul Porcului and Pârâul Dracului, small tributaries of the Pietroasa Valley. Their tholeiitic character was pointed out by Gandrabura (1981). They are represented by tholeiitic basalts, locally in pillow lava facies, subordinately by dolerites and microgabbros, cropping out from under the products of the calc-alkaline series, greatly developed in this unit.

1.1.2. Calc-alkaline Series

The magmatites of the calc-alkaline series develop in the central-eastern part of the Căpâlnaş-Techereu Nappe, where various products are to be found such as pyroclastics and porphyritic basalt flows, basaltic andesites, andesites, dacites, rhyolites, rarely trachyandesites, orthophyre and oligophyre dykes, subvolcanic or plutonic intrusive rocks of andesites, dacites, rhyolites, (micro)diorites-(micro)granodiorites-tonalites that occur as bodies or apophyses of some rooted bodies, the age of which has been demonstrated through the K-Ar isotopic method as being Early Cretaceous (Lemne et al., 1983; Savu et al., 1986 b). Most of these intrusions have been previously considered to be banatitic, as in the case of those at Cerbia, Săvârşin, Căzăneşti-Pietroasele. The Early Cretaceous intrusions appear both in the west of the region, where the rocks of the tholeiitic series largely develop, and in the central and eastern parts, as in the case of the bodies at Vorţa, and Gliganul Hill-Dealul Mare or Poiana-Techereu. They are products typical of the final phases of the evolution of a magmatic arc.

In the Rimetea Nappe (Figs. 1, 2) the magmatites of the calc-alkaline series are represented by pyroclastics and porphyritic basalt flows, basaltic andesites, andesites, dacites, rhyolites, trachyandesites, latiandesites, dykes of orthophyre and oligophyre, porphyritic microdiorites; at the upper part there occur pyroclastics as well as keratophyre and rhyolite flows overlain by Upper Tithonian-Berriasian limestones (Lupu, 1972) that develop from Cheile Turzii to south of the Arieş Valley, in the Valea Albă zone.

In the Bedeleu Nappe (Figs. 1, 2) pyroclastics and keratophyre flows are to be found, with thin cherty limestones interbeds towards the upper part, locally

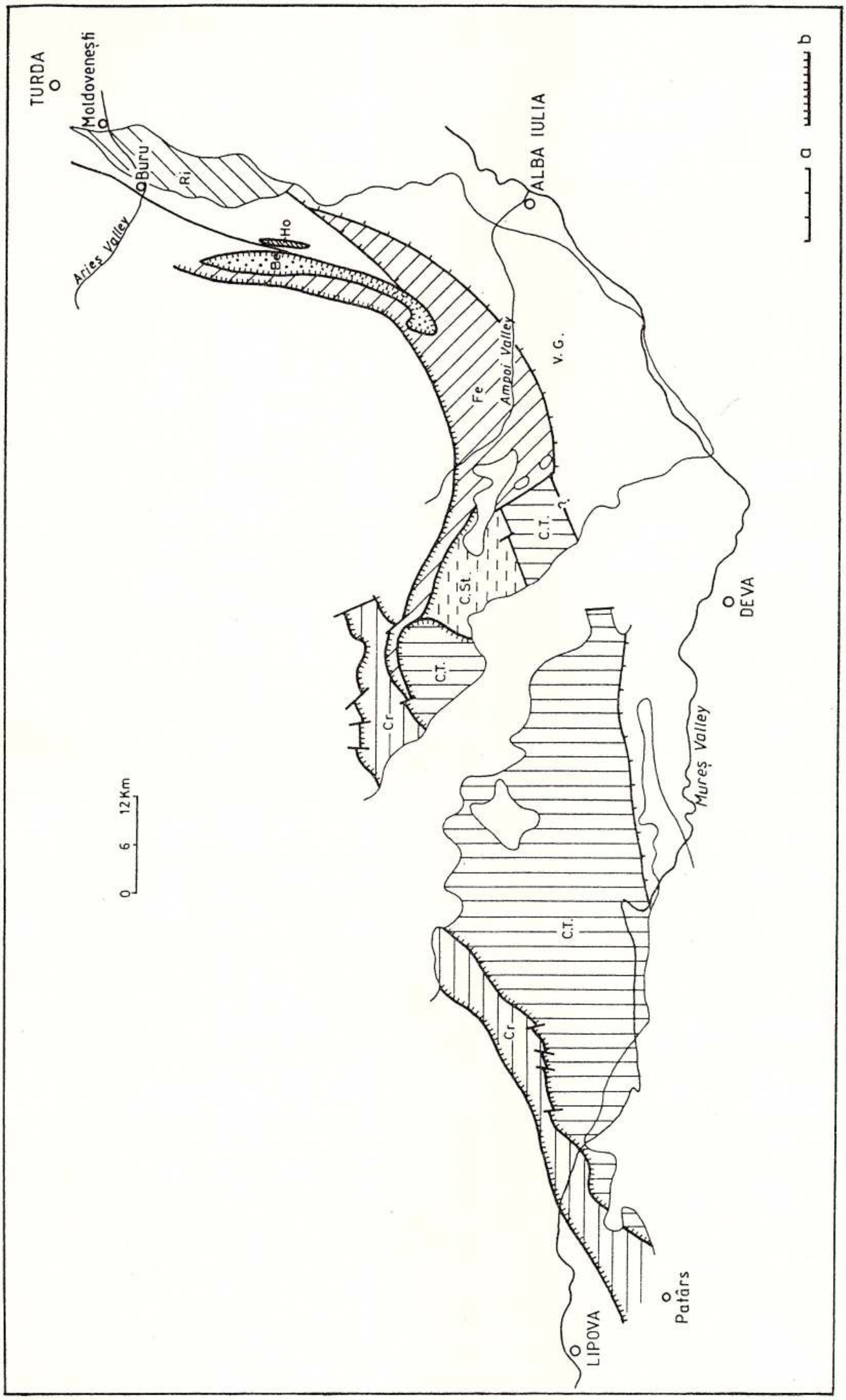

Fig. 1 – Tectonic Units with Ophiolitic Rocks from the South Apuseni Mountains (according to M. Lupu. 1981, simplified).

CT, Căpâlnaș-Techereu Nappe; Ri, Rimetea Nappe; Be, Bedeleu Nappe; Ho, Hospea Nappe; C-St, Curechiu-Stănija Nappe; Fe, Feneș Nappe; Cr, Criș Nappe; V-G, Valea Mică-Galda Nappe; a, reverse fault; b, nappe.

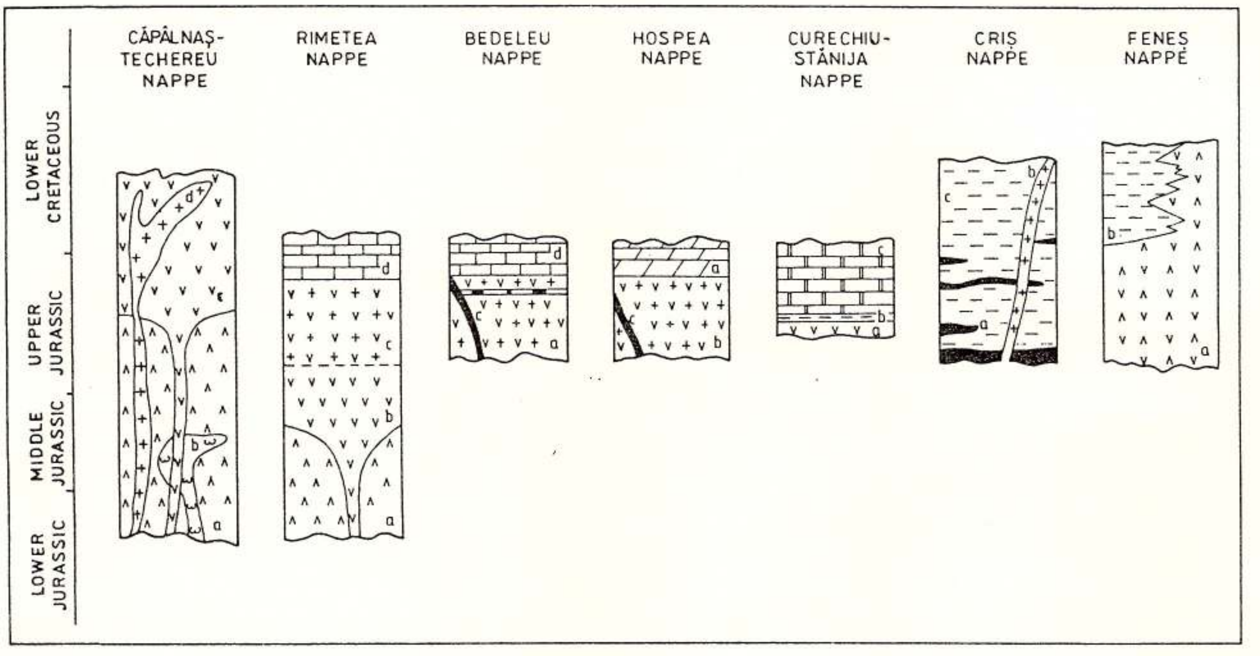

Fig. 2 – Synthetic Columns of the Main Tectonic Units with Ophiolitic Rocks.

1. Căpâlnaş-Techereu Nappe:
Tholeiitic series: basalts, anamesites, dolerites (a); gabbros (b).
Calc-alkaline series: Pyroclastics and flows of porphyritic basalts, basaltic andesite, andesite, dacite, rhyolite, rarely trachyandesite, orthophyre and oligophyre dykes (c);
Early-Cretaceous intrusive magmatites: andesite, dacite, rhyolite, (micro)diorite-(micro)granodiorite-tonalite (d).
2. Rimetea Nappe:
Tholeiitic series: basalts, in places in pillow lava facies, subordinately dolerites, microgabbros (a).
Calc-alkaline series: pyroclastics and flows of porphyritic basalts, basaltic andesite, andesite, dacite, rhyolite, trachyte, latiandesite; orthophyre, oligophyre, porphyritic microdiorite dykes (b); pyroclastics and flows of keratophyre and rhyolite (c);
Upper Tithonian-Berriasian limestones (d).
3. Bedeleu Nappe:
Keratophyres (a) with cherty limestones intercalations (b); basic rock-basalt, anamesite, dolerite dykes (c).
4. Hospea Nappe:
Aptychus Beds (Tithonian-Neocomian) (a); keratophyres (b); quartz basalt-anamesite dykes (c).
5. Curechiu-Stănija Nappe:
Porphyritic basalt flows and pyroclastics (a); Kimmeridgian-Tithonian jaspers (b); grey-greenish micritic limestones with Tithonian-Neocomian red clay intercalations (c).
6. Criş Nappe: Flows, rarely pyroclastics of basalt at various stratigraphic levels in the Callovian-Tithonian time-span (a); dykes of andesites, dacites, rhyolites ± albitized, chloritized (b); prevailingly flyschoid sediments (c).
7. Feneş Nappe:
Spilites: prevailingly porphyritic basalts, andesites, anamesites, dolerites, microgabbro (a); sedimentary and volcano-sedimentary rocks (b).

crossed by basic, basalt-anamesite-doleritic rock dykes. As in the Rimetea Nappe, the keratophyres are overlain by Middle Tithonian-Berriasian limestones (Lupu, 1972).

In the Hospea Nappe, subunit of the "Bedeleu Nappes System" (Lupu, in Bleahu et al., 1981) keratophyre flows and pyroclastics are to be found, associated with the Tithonian-Neocomian Aptychus Beds (Lupu, 1972). Keratophyres are seldom crossed by quartz basalt-anamesite dykes.

In the Curechiu-Stănija Unit the products of the calc-alkaline series are represented by basalt-andesitic pyroclastic products.

A great similarity is noticed between the ophiolites of the Căpâlnaş-Techereu Nappe and those from the Rimetea Nappe, the latter differing from the former by the smaller surface occupied by tholeiitic rocks; also, an extremely reduced participation of the rocks belonging to the Early Cretaceous intrusive phase is noticed. Probably only the oligophyre and porphyritic microdiorite dykes may be assigned to this phase. On the other hand, the keratophyres and limestones from the Bedeleu Nappe, except for the basic rock dykes, are identical with the upper sequence of the Rimetea Nappe from which they are likely to originate.

A few considerations on the age of the magmatites from the tholeiitic series will be further presented. H. Savu has often quoted the age of 180 Ma for the tholeiitic rocks (Herz et al., 1974). In fact the paper provides only presumed ages and not Rb-Sr isotopic age determinations. There are no data available on the beginning of the magmatic activity. A few K-Ar isotopic age determinations that indicated ages ranging between 138.9±6 and 155.7±5.7 Ma carried out in a laboratory in Grenoble by M. Bonhomme corroborated with some palynological data (Antonescu, in Lupu et al., in press), and with some radiolarian determinations carried out by P. Dumitrică attest that in the western sector the tholeiitic series might have formed up to the Callovian-Oxfordian inclusive (Nicolae et al., 1992).

On the other hand, several K-Ar isotopic age determinations were carried out (Bleahu et al., 1981; Nicolae et al., 1987). Except for the obviously rejuvenated ages obtained, the oldest age is of 160±5 Ma. The keratophyres are considered Oxfordian-Lower Tithonian in age based on micropaleontological evidence provided by micrite intercalations (Lupu, in Cioflica et al., 1981). As keratophyres from the Rimetea Nappe are the final products from a thick volcanic succession, it may be concluded that the transition from the products of the tholeiitic series to the calc-alkaline one took place earlier in this unit, probably in the Middle Jurassic.

1.2. Marginal Basin Type Ophiolites

The ophiolites from this category are those from the Feneş Beds, considered Barremian-Lower Aptian (Lupu et al., 1980) as well as those from the Criş Nappe, which have been recently defined of ensialic marginal basin type (Lupu et al., 1993).

The Feneş Nappe. The Feneş Beds represent an olistostrome type volcano-sedimentary formation (Lupu et al., 1980) in which the ophiolitic rocks are represented by porphyritic basalts, andesites, more rarely anamesites, dolerites and microgabbros of spilitic composition. Considered equivalents of the third phase, sensu Giuşcă et al., 1963, they were described as the "spilitic complex" being assigned to the marginal basin type (Cioflica et al., 1980; Cioflica, Nicolae, 1981). Taking into consideration a K-Ar isotopic age determination of 139±5 Ma (Nicolae et al., 1987), the Feneş marginal basin seems to have opened through a magmatic activity since the Upper Jurassic.

The Criş Nappe. This unit is also marked by the presence of marginal basin type ophiolites (Lupu, 1983; Lupu et al., 1993). Sedimentation of the basin began in the Callovian, showing an early calcareous followed by a flysch character along the internal flank and a flysch character towards its central part. Eruptive episodes were pointed out at various levels that manifested in the Callovian-Tithonian time-span and led to the prevailing occurrence of some flows, subordinately basaltic tuffs (Lupu et al., 1993). Unlike the Feneş marginal basin, the subsidence seems to have been less ample, and the magmatic activity more reduced. The appearance of some more or less altered (albitized, chloritized) intermediate-acid rock dykes of andesites (rarely) - dacites, rhyolites which are not characteristic of a marginal basin, might be explained through the direct or indirect influence of the sial from the initial basement of the region (Lupu et al., 1993).

2. New Petrochemical Data and Tectonic Setting

31 new chemical analyses (Tabs. 1, 2) of the tholeiitic rocks, 11 unpublished analyses (Cioflica et al., 1984) as well as six new analyses of the basic rocks from the dykes that cross the keratophyres from the Trascău Mts are available; they are going to be processed together with the five pre-existent ones (Nicolae, 1985). Our observations regard only these two rock types as there are divergent opinions on them according to the earlier or more recent interpretations.

Beside the different opinions on the interpretation of the tholeiitic rocks (as ocean floor type and magmatic arc type magmatites, respectively), Savu et al. (1992 a, b) have recently interpreted as ocean floor ophiolites included in an ophiolitic mélange what we considered to be basic rocks dykes crossing the keratophyres (Nicolae, 1985).

In order to establish the tectonic setting the following diagrams will be used:

In the Zr/Y–Ti/Y diagram (Fig. 3) with one exception the analyses are plotted in the marginal-plate basalt field.

In the Ti/Cr diagram (Fig. 4) 24 tholeiitic rock analyses are plotted in the low potassium tholeiites field (island arc) and 18 in the ocean floor basalts field. In a previous paper (Nicolae, 1983) out of another 70 analyses 66% plotted in the island arc tholeiite field. Of the basic dykes analyses eight are plotted in the island arc field and three in the ocean floor field.

In the Ti/100–Zr-3Y diagram (Fig. 5) nine analyses of the tholeiitic series are plotted outside the diagram, two in the D field of the "within plate basalts", five in the A and C (island arc) fields and 24 in the B field

Table 1

Chemical Analyses (rocks of tholeiitic series)

No.	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CuO	K_2O	Na_2O	H_2O^-	P_2O_5	H_2O^+	CO_2	S	Total
1	41.17	1.06	16.76	3.88	3.69	0.12	6.20	16.10	0.15	2.30	2.03	0.08	3.18	2.81	0.24	99.77
2	41.47	1.33	15.16	9.89	13.96	0.27	7.81	0.26	0.00	0.79	n.d.	0.08	8.14	0.00	0.27	99.66
3	42.61	0.58	16.22	6.57	2.46	0.15	4.70	10.99	0.25	4.97	0.58	0.07	5.34	4.26	0.01	99.76
4	43.18	1.03	15.60	3.02	5.42	0.18	7.53	10.08	0.35	2.81	n.d.	0.06	4.54	6.38	0.06	100.24
5	45.05	1.54	16.51	1.16	7.96	0.14	7.70	4.34	0.80	6.25	0.95	0.20	5.09	2.37	0.17	100.23
6	45.32	0.95	18.51	1.01	6.48	0.16	6.40	10.64	0.00	5.30	0.86	0.16	0.69	2.86	0.25	99.59
7	45.62	0.65	19.21	0.78	4.86	0.11	9.10	8.54	0.90	3.15	0.89	0.09	2.60	2.94	0.16	99.60
8	46.53	0.63	18.34	4.48	5.53	0.10	6.80	8.85	0.27	3.95	0.26	0.12	3.16	0.61	0.00	99.63
9	46.57	1.19	13.85	5.92	2.95	0.11	5.75	11.74	1.00	3.95	n.d.	0.20	2.90	3.68	0.22	99.93
10	46.64	1.60	12.24	1.48	9.29	0.18	7.00	10.50	4.10	3.75	0.67	0.17	2.53	0.00	0.17	100.32
11	46.66	1.57	13.70	7.80	2.00	0.14	4.14	12.54	0.96	4.01	n.d.	0.14	1.56	4.05	0.18	99.55
12	46.74	1.42	10.27	1.50	9.44	0.20	7.00	11.48	5.30	3.85	0.44	0.17	2.28	0.00	0.21	100.31
13	46.88	1.81	16.83	2.12	8.26	0.23	7.60	9.38	0.25	2.95	0.33	0.17	1.63	0.96	0.15	99.55
14	47.09	0.54	21.59	1.15	6.78	0.14	5.80	5.60	0.60	5.00	0.72	0.09	2.72	1.54	0.16	99.52
15	47.28	1.30	18.58	2.57	5.60	0.18	7.40	9.60	0.00	2.80	1.16	0.19	2.67	0.00	0.21	99.60
16	47.54	1.05	14.95	4.51	5.27	0.15	7.77	12.31	0.08	2.30	n.d.	0.08	3.94	0.00	0.22	100.17
17	47.60	2.20	13.95	7.29	6.27	0.19	7.29	7.11	0.43	3.84	n.d.	0.23	2.44	0.42	0.26	99.52
18	47.87	1.25	13.26	3.42	5.48	0.18	6.65	10.22	0.13	3.68	n.d.	0.11	3.01	3.83	0.84	99.93
19	48.11	1.52	11.72	2.95	5.16	0.15	6.20	12.18	0.15	2.25	0.88	0.12	6.63	1.45	0.19	99.66
20	48.51	1.68	20.74	4.27	6.64	0.11	3.90	7.56	0.00	2.85	0.90	0.32	2.05	0.00	0.17	99.70
21	49.37	2.45	15.89	5.31	6.64	0.20	5.20	7.00	0.35	4.55	0.60	0.27	1.63	0.00	0.24	99.70
22	50.53	0.32	15.49	2.33	3.48	0.09	10.53	12.04	0.15	2.10	n.d.	0.00	2.77	0.00	0.20	100.03
23	50.76	2.28	13.39	5.63	8.46	0.22	5.43	5.28	0.10	4.46	n.d.	0.31	3.15	0.00	0.23	99.70
24	51.07	1.97	13.88	6.63	6.85	0.17	5.27	4.77	0.02	6.00	n.d.	0.31	2.71	0.00	0.24	99.66
25	51.11	1.21	14.57	4.22	4.60	0.16	7.03	8.79	0.13	4.15	n.d.	0.11	2.69	0.78	0.20	99.71
26	51.21	0.49	17.26	5.62	5.86	0.14	7.05	2.74	1.00	3.95	0.51	0.07	3.36	0.52	0.01	99.88
27	51.68	0.84	20.45	0.95	6.78	0.18	6.00	7.00	0.40	4.10	0.19	0.19	0.56	0.00	0.33	99.65
28	52.19	0.63	15.03	7.07	4.57	0.16	5.80	3.15	0.27	4.72	1.35	0.09	3.73	0.94	0.01	99.71
29	53.52	2.45	11.85	6.58	7.30	0.26	4.61	5.46	0.02	4.36	n.d.	0.43	2.58	0.00	0.24	99.66
30	53.95	1.51	17.31	5.38	5.36	0.17	2.77	5.19	0.78	4.71	n.d.	0.26	2.05	0.00	0.24	99.62
31	54.60	2.06	15.60	5.23	5.40	0.10	2.70	6.86	0.15	5.45	0.27	0.40	0.46	0.00	0.21	99.49
	Chemical analyses (dykes of basic rocks)															
32	40.19	1.08	13.99	3.79	7.91	0.48	3.99	11.45	0.13	3.81	n.d.	0.14	4.59	8.35	0.05	99.95
33	43.53	3.38	13.96	6.08	3.97	0.20	3.50	10.90	0.75	4.00	n.d.	0.39	2.75	6.01	0.15	99.57
34	49.73	0.67	18.98	4.25	4.71	0.11	5.40	7.24	0.75	4.15	0.46	0.13	2.81	0.61	0.00	99.75
35	49.80	0.69	21.23	4.52	4.64	0.12	2.20	7.84	1.00	4.37	0.41	0.10	2.43	0.59	0.00	99.94
36	54.88	1.53	15.74	5.38	4.68	0.15	3.26	4.98	1.02	4.59	n.d.	0.25	2.43	0.60	0.06	99.55
37	64.67	0.44	14.44	4.45	1.84	0.06	1.80	1.13	3.05	5.12	0.26	0.17	2.09	0.22	0.01	99.75

n.d. – not determined

Table 2

Trace Elements (rocks of tholeiitic series)

No.	Pb	Cu	Zn	Sn	Ga	Ni	Co	Cr	V	Sc	Zr	Yb	Y	Sr	Ba
1	<2	85	50	<2	15	125	48	240	240	40	50	2	33	160	10
2	3.5	>3000	n.d.	<2	13	210	88	230	115	25	90	0.7	11	10	18
3	<2	30	50	<2	15	98	53	420	230	32	70	4.3	33	280	50
4	6	18	n.d.	2	14	145	31	300	190	29	47	2.3	14	180	75
5	<2	125	74	<2	25	78	50	160	430	45	190	5	53	65	26
6	<2	67	56	2.5	20	130	52	360	350	40	135	5.2	50	190	38
7	<2	150	50	8.5	15	210	50	600	250	37	60	2.6	30	140	40
8	<2	75	48	<2	26	50	35	150	250	34	76	4.5	30	400	53
9	<2	30	73	<2	9.5	80	18	380	210	33	95	3.2	30	135	44
10	<2	55	55	3	23	80	48	85	550	46	120	4.8	50	93	10
11	<2	55	55	<2	17	46	42	170	300	36	125	5.5	40	260	28
12	<2	75	90	2	24	60	50	75	350	38	160	6	53	95	15
13	32	28	46	<2	22	75	42	160	360	44	145	5.4	48	70	17
14	<2	90	50	<2	25	42	37	45	480	37	55	2.3	24	120	56
15	<2	145	65	<2	26	65	47	125	350	38	140	5.4	50	170	20
16	<2	55	57	<2	12	46	30	80	420	38	65	3	32	140	40
17	<2	2.5	60	5	14	58	23	170	650	44	120	4.5	50	125	30
18	<2	19	105	<2	13	50	32	100	290	38	78	3.6	38	80	<10
19	<2	90	70	3	24	88	40	240	300	38	120	4.2	43	130	15
20	<2	23	40	<2	33	24	37	13	380	32	240	8.5	70	110	15
21	<2	34	100	<2	24	32	36	14	360	26	200	7.5	60	130	22
22	<2	5	<30	<2	10	46	28	190	185	50	34	1.6	21	95	<10
23	<2	13	76	3.5	16	38	34	47	380	40	145	6	60	48	20
24	<2	17	60	2.5	16	17	23	8.5	280	25	130	5.3	44	32	15
25	<2	115	70	<2	17	55	40	220	330	44	80	2.3	30	140	60
26	<2	85	37	<2	23	72	53	80	236	40	76	2.6	38	190	190
27	<2	160	50	2.5	26	53	40	100	240	30	170	6	54	180	33
28	<2	5.5	38	<2	17	13	33	8.5	350	32	48	2.9	18	340	45
29	<2	32	140	<2	16	24	25	8.5	320	54	280	8	80	12	<10
30	<2	53	70	<2	16	8	18	2	190	23	115	2.8	28	240	95
31	<2	17	37	<2	27	23	23	18	150	27	480	13	95	84	10
Trace elements (dykes of basic rocks)															
32	8.5	23	n.d.	<2	17	10	17	19	150	21	100	2	22	280	550
33	5	28	n.d.	22	20	20	25	46	280	35	300	4.6	42	310	200
34	8	26	40	3	26	40	36	240	220	35	175	5.3	36	500	240
35	53	23	40	2	17	39	40	200	220	32	165	4.7	37	630	340
36	7.5	13	n.d.	3.5	22	7	16	1	210	21	120	2.9	24	340	215
37	12	9	45	4.5	22	3	9	4	25	2	240	7	42	185	65

n.d. – not determined

Annex to Tables 1 and 2

Rock type and location: 1, Basalt, Mureş Valley (Toc); 2, Basalt, Valea Porcului; 3, Basalt (pillow lava), Podeni Valley; 4, Basalt, Valea Porcului; 5, Basalt, Valea Bozului; 6, Basalt (pillow lava), Tămăşeşti Valley; 7, Basalt (pillow lava), Visca Valley; 8, Dolerite, Podeni Valley; 9, Amygdaloidal basalt (pillow lava), Pârneşti Valley; 10, Gabbro, Mureş Valley (N. Bălcescu); 11, Basaltic breccia, Dumeşti Valley; 12, Dolerite, Bătuţa Quarry; 13, Basalt, Vărădia de Mureş; 14, Anamesite, Zam; 15, Anamesite, Troaş Valley; 16, Anamesite, Cerboaia Valley; 17, Microgabbro, Furu Valley; 18, Microgabbro, Ponor Valley; 19, Basalt, Zăbalţ; 20, Anamesite, Juliţa quarry; 21, Dolerite, Pătârş; 22, Gabbro, Tătăroaia Valley; 23, Dolerite, Pârneşti Valley; 24, Dolerite, Valea Gomilelor; 25, Dolerite, Pârneşti Valley; 26, Anamesite, Fornădia; 27, Dolerite, Valea Mică (Curcubăta); 28, Basalt, Pietroasa Valley; 29, Anamesite, Pârneşti Valley; 30, Gabbro, Furu Valley; 31, Basalt, Bătuta quarry; 32, Basalt dyke, Izvoarele Valley; 33, Anamesite dyke, Rogoaze; 34, Dolerite dyke, Valea Muntelui; 35, Dolerite dyke, Valea Muntelui; 36, Intersertal quartz basalt dyke, Valea Mânăstirii; 37, Quartz basalt dyke, Hospea.

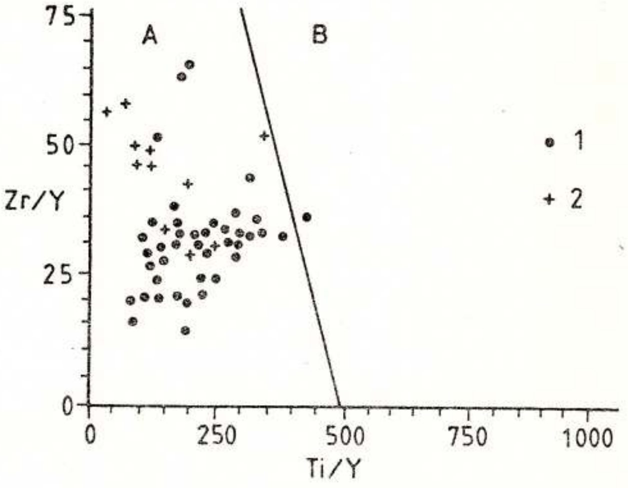

Fig. 3 – Zr/Y–Ti/Y Diagram (acc. to Pearce and Gale, 1977). A, field of margin-plate basalts; B, field of within-plate basalts; 1, tholeiitic rocks; 2, basic dykes.

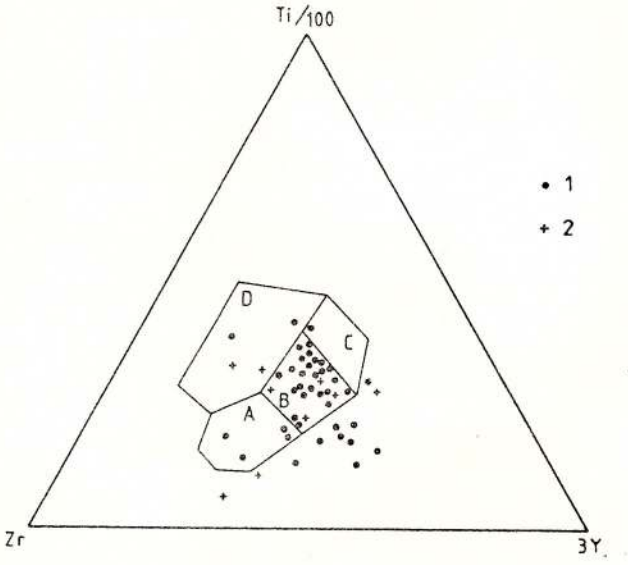

Fig. 5 – Ti/100–Zr–3Y Diagram (acc. to Pearce and Cann, 1973). A, field of low potassium tholeiites; B, field of ocean floor basalts, low potassium tholeiites or calc-alkaline basalts; C, field of calc-alkaline basalts; D, field of ocean island or continental basalts (within plate basalts).

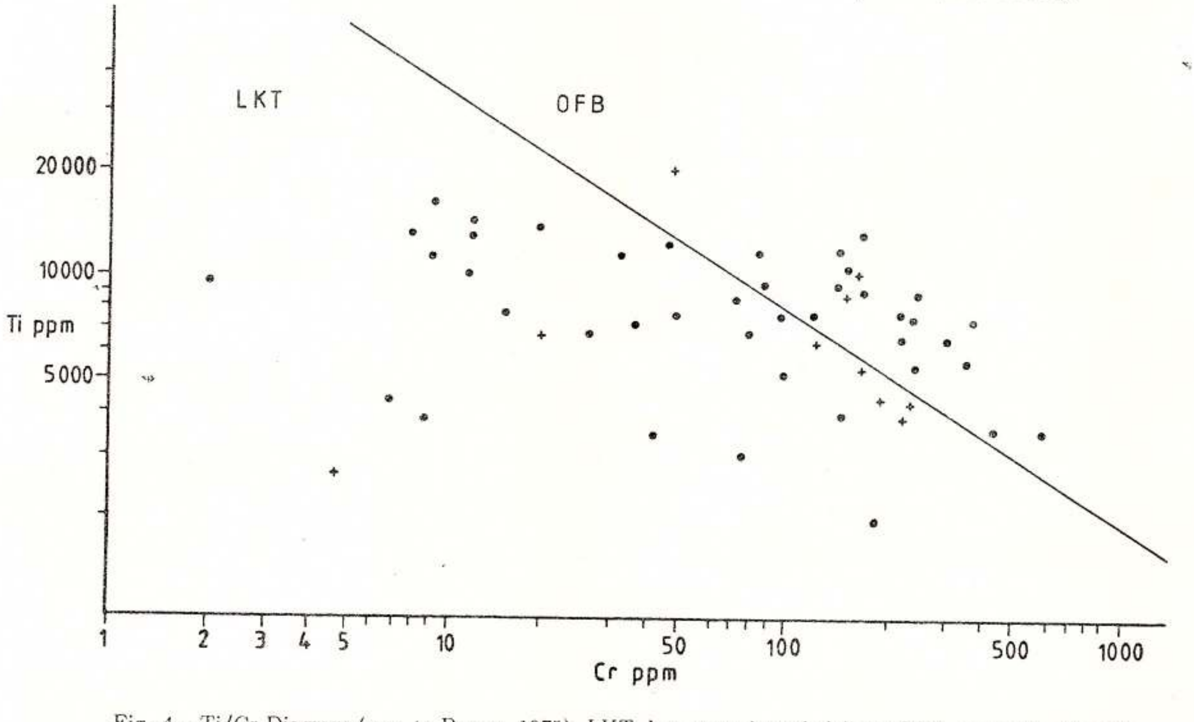

Fig. 4 – Ti/Cr Diagram (acc. to Pearce, 1975). LKT, low potassium tholeiites; OFB, ocean floor basalts.

which is common to the ocean floor and island arc basalts. The analyses of the dykes are plotted as follows: two in the D field, two in the A (island arc) field, four in the B field and three outside the diagram.

In the Ti-Zr diagram (Fig. 6) the tholeiite analyses are plotted as follows: eleven in the A and C island arc fields, six in the B field, common to the island arc and ocean floor rocks, fourteen in the D field (ocean floor) and eleven outside the diagram. The analyses of the dykes are plotted as follows: two in the D field, four in the C field, two in the B field and three outside the diagram.

In the Ti/Cr–Ni diagram (Fig. 7) tholeiites are plotted as follows: fourteen in the island arc tholeiite field, twenty-four in the ocean floor tholeiite one and four analyses outside the diagram. Of the basic rock dykes five are plotted in the island arc tholeiite field, three in the ocean floor tholeiite one and three outside the diagram.

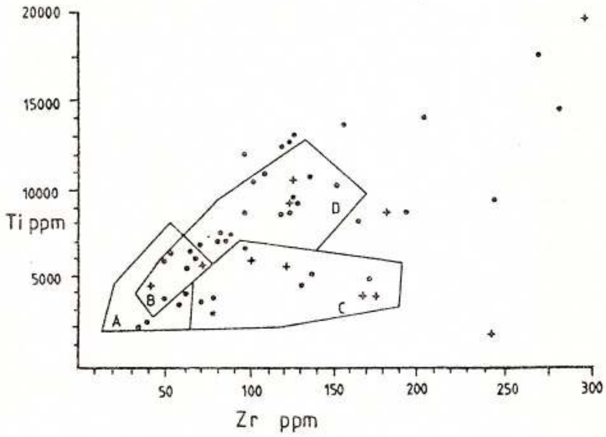

Fig. 6 – Ti–Zr Diagram (acc. to Pearce and Cann, 1973). A, field of low potassium tholeiites; B, field of ocean basalts, low potassium tholeiites or calc-alkaline basalts; C, field of calc-alkaline basalts; D, field of ocean floor basalts.

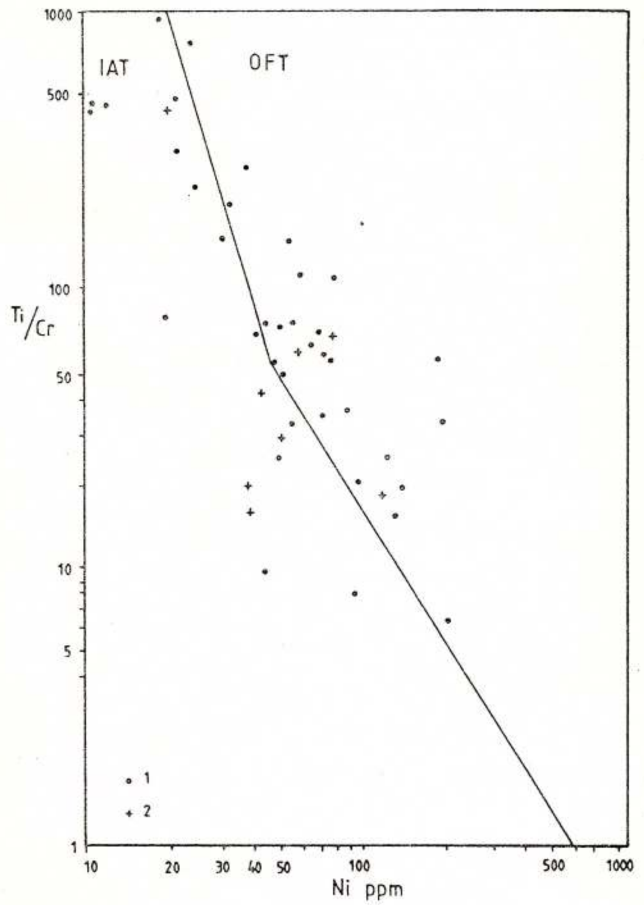

Fig. 7 – Ti/Cr–Ni Diagram (acc. to Beccaluva et al. (1979). IAT, field of island arc tholeiites; OFT, field of ocean floor tholeiites.

In the V–Ti/1000 diagram (Fig. 8) twenty-eight tholeiites are plotted in the field with the value of the ratio ranging between twenty and fifty, a zone in which the mid-ocean ridge basalts (MORB) are situated, and fourteen analyses in three neighbouring fields. Of the dyke analyses seven analyses are plotted in the field with the value of the ratio ranging between twenty and fifty, characteristic of the ocean floor rocks and another four analyses in two neighbouring fields.

If one takes into consideration all the present diagrams several disagreements are noticed, a lot of analyses are plotted in fields differing from one diagram to another, numerous other analyses plot outside the diagrams. The alternation phenomenon of the rocks cannot be taken into consideration as in the case of Pearce and Cann (1973) or Pearce and Gale (1977) diagrams, for instance, stable elements are used, which do not change during the subsequent alterations, not even during the metamorphism. In my opinion, the above-mentioned disagreements can be explained by the complexity of the magma generation phenomenon on the one hand, a process taking place at the boundary of some microplates, of which the one under which subduction takes place has a thin continental type crust, while the "pattern" tectonic settings on the basis of which the used diagrams were built, are situated within the limits of some major oceanic plates.

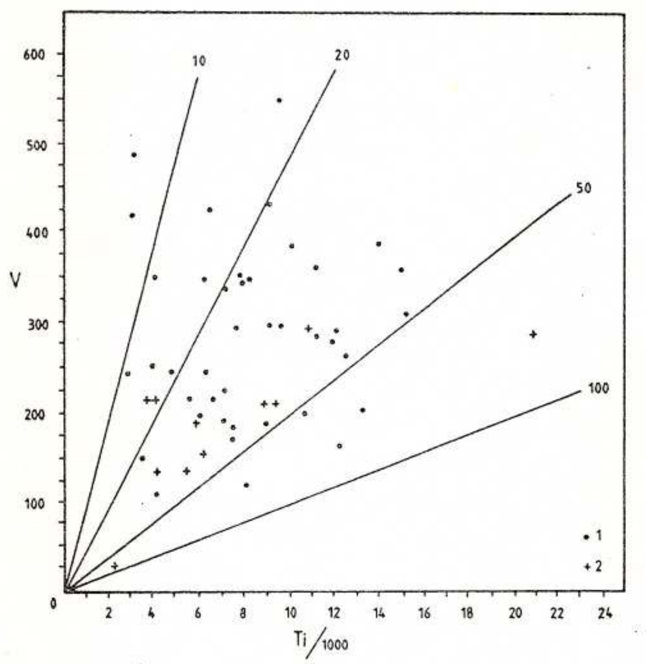

Fig. 8 – V–Ti/1000 Diagram (acc. to Shervais, 1982).

As far as the basic rocks dykes are concerned, beside the deposit form, easy to observe on the field, great geochemical differences are also noticed (Tabs. 3, 4).

Table 3

Average values and variation limits for Rb, Sr, Y, Zr, Nb, Rb/Sr
(Analyst G. Grabari)

Rocks of tholeiitic series

Average	Rb	Sr	Y	Zr	Nb	Rb/Sr
14 samples	12.8	160.5	32	114	11.7	0.109
Variation limits	5-24.9	90-315	15-80.6	50-369	8-21	0.014-0.150

Rocks from the basic dykes within keratophyres

Average						
3 samples	41.66	293.3	30	176.7	<20	0.023
Variation limits	25-60	120-395	25-40	145-230	<20	0.07-0.50

Table 4

Average values and variation limits for U, Th, K %, Th/U
(analyst A. Tănăsescu)

Rocks of tholeiitic series

Average	U	Th	K%	Th/U
19 samples	0.84	1.06	0.32	4.35
Variation limits	0.1-5.6	0.3-2.3	0.02-1.06	0.1-12

Rocks from the basic dykes within keratophyres

Average				
3 samples	2.73	7.13	1.6	2.86
Variation limits	2.0-4.0	6.9-7.3	0.73-3.03	1.80-3.45

These elements will not be insisted upon, constituting the object of some other communications, but differences are noticed especially in the Rb, Sr, Rb/Sr, U, Th, K values, which are much higher in the rocks of the basic dykes than in those of the tholeiitic series, the former probably undergoing a contamination as a result of the contact with the keratophyric rocks in which they were intruded.

3. Origin of the Ophiolitic Magmas

The ophiolitic magmas may form:

- at divergent plate boundaries, in spreading zones of the ocean floor, through the partial melting of the upper mantle, where similar ophiolites as regards the structure and the geochemical characters to the rocks making up the oceanic crust occurs;
- at convergent plate boundaries, in the subduction zones; here the phenomenon is more complicated as the subducted plate may have a certain amount of sediments which may geochemically influence the formed magmas. These magmas may occur after the melting of the upper part of the subducted plate as well as through the melting of the mantle above the subducted plate due to the conditions created through the dehydration of the plate. Also, intermediary magmatic basins may appear in which the magmas are contamined through the contact with the host rocks;
- behind the island arcs (or magmatic arcs) that form as a result of subduction, active marginal basins may individualize; these are zones in which the generation of the magmas takes place in a similar way to that from the active mid-ocean ridge, therefore also as a result of the spreading process.

It is difficult to accept the scheme presented by H. Savu for the South Apuseni Mts, that is a spreading phenomenon with the generation of the ocean floor (tholeiitic) magmatites, followed by a bilateral subduction phenomenon with the formation of the associated island arc magmatites if one takes into consideration:

- the age of the tholeiitic series, which reaches the Callovian-Oxfordian in the western part, while in the eastern part (Trascău Mts) the transition from the tholeiitic series to the calc-alkaline one took place since the Middle Jurassic. It is difficult to suppose that a distensive tectonic regime coexists along the same plate boundary, in which the ocean floor magmatites form in the ocean floor spreading with a compressive tectonic regime – a subduction followed by the generation of the calc-alkaline series;
- the at least contradictory chemical characteristics, if one takes into consideration all the available diagrams, a situation in which disagreements occur among the various diagrams, supports the genesis of the ophiolitic magmas in a more complex process than that taking place in the ocean floor magmatism;
- in an ocean floor magmatism the values of the $^{87}Sr/^{86}Sr$ isotopic ratio should be more homogeneous and not vary within such wide limits (0.700–0.704, Savu et al., 1994). Faure (1986) shows that in the subduction process there are greater posibilities of contamination with sialic material, which leads to the inhomogeneity of the resulted rocks.

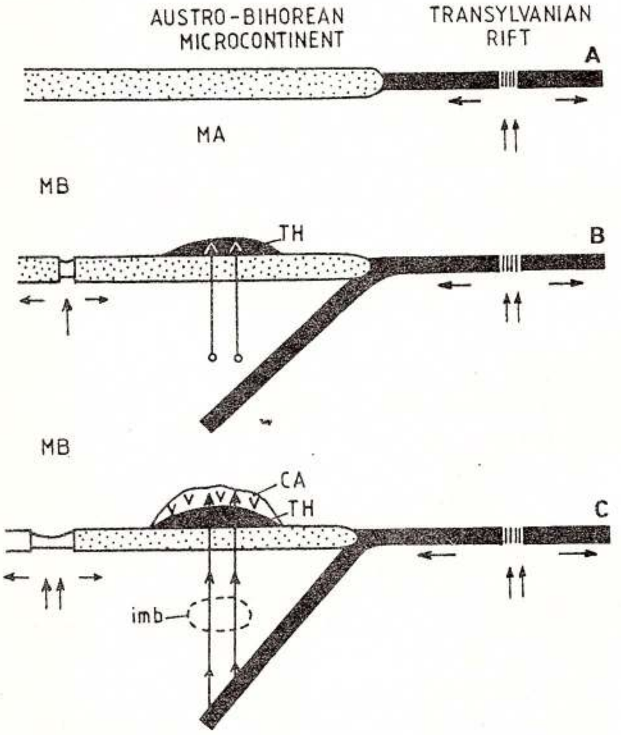

Fig. 9 – Sketch of the Generation Mode of the Ophiolites from the South Apuseni Mountains. In the Triassic-Lower Jurassic (?) the Transylvanian rift forms generating the oceanic crust (A). In the Lower or Middle Jurassic a subduction starts, leading to the formation of a magmatic arc (MA) behind which a marginal basin (MB) forms. Initially magmas are generated through the melting of the mantle above the subduction plate, giving rise to the tholeiitic series TH-(B). This process takes place in some zones until the Oxfordian inclusive and in some other zones until the Middle Jurassic. Further through the melting of the subducted plate the magmas move directly forwards the surface or linger in an intermediate magmatic basin (imb) leading to the formation of the rocks of the calc-alkaline (CA) series (C); these products may occur in places since the Middle Jurassic or may form until the Neocomian. The ophiolites from the marginal basin generally form between the Callovian and Neocomian.

That is why I support the formation of the ophiolites belonging to the tholeiitic and calc-alkaline series from the South Apuseni Mountains within a single subduction process (Fig. 9). Thus through the partial melting of the mantle above the subducted plate the rocks of the tholeiitic series were furnished and through the melting of the upper part of the subducted plate ± sediments the magmas giving rise to the calc-alkaline rocks were furnished. In this way it can be explained how the rocks of the tholeiitic series formed in this case at the end of the Middle Jurassic and in the Callovian-Oxfordian in the western part of the South Apuseni Mountains, and the rocks of the calc-alkaline rocks in the eastern part (Trascău Mountains).

In the Upper Jurassic a marginal basin (or marginal basins) individualized behind the magmatic arc, showing a magmatic activity of variable intensities, as result from the differences existing between the magmatites from the Feneş Nappe and those from the Criş Nappe.

Finally it should be mentioned that in the South Apuseni Mountains ophiolites are also known as sedimentary klippen in more recent formations (Lupu et al., 1980) such as the Meteş Beds (Upper Aptian-Albian) or the Valea Mică–Galda Beds (Senonian) which have not been the object of this study.

Conclusions

In my opinion, the ophiolites from the South Apuseni Mts belong to two geotectonic settings; magmatic arc and marginal basin.

The magmatic arc ophiolites (Lower Jurassic ?–Lower Cretaceous) are underlain by volcanic and intrusive tholeiitic arc rocks (tholeiitic series), and overlain by petrographically varied volcanics (calc-alkaline series), the activity ending with Early Cretaceous intermediate-acid subvolcanic or plutonic intrusive rocks.

The marginal basin ophiolites formed in the Callovian-Tithonian time-span, consisting mainly of basalts which are little developed in the Criş Basin. An intermediate-acid, probably Early Cretaceous intrusive phase of smaller extension that the previous one manifests here. In the Feneş Basin, ophiolites consist mainly of spilitic basalts. The magmatic activity was more intense and probably took place in the Upper Jurassic–Lower Aptian time-span.

The Mesocretaceous and Neocretaceous tectogeneses led to the tectonic dismembering of the ophiolitic rocks determining the differences in the participation of the petrographic types beside the initial petrographic variations in the different tectonic units.

References

Beccaluva, L., Ohnenstetter, D., Ohnenstetter, M. (1979) Geochemical discrimination between ocean-floor and island-arc tholeiites-application to some ophiolites. *Can. J. Earth. Sci.*, 16, p. 1874–1882, Ottawa.

Bleahu, M. (1974) Zone de subducţie în Carpaţii româneşti. *D. S. Inst. Geol. Geofiz.*, LX/5, p. 5–25, Bucureşti.

— **, Lupu, M., Patrulius, D., Bordea, S., Ştefan, A., Panin, S. (1981)** The structures of the Apuseni Mountains. *Guide to Excursion, B_3 – Carp.- Balk. Assoc, XIIth Congress*, Bucureşti.

— **, Soroiu, M., Catilina, R. (1984)** On the Cretaceous tectonomagmatic evolution of the Apuseni Mountains as revealed by K-Ar dating. *Rev. roum. phys.*, 29, 1, p. 123–130, Bucureşti.

Cioflica, G., Lupu, M., Nicolae, I., Vlad, S. (1980) Alpine Ophiolites of Romania: Tectonic Setting, Magmatism and Metallogenesis. *An. Inst. Géol. Géophys.*, LVI, p. 79–95, Bucureşti.

— **, Nicolae, I. (1981)** The Origin, Evolution and Tectonic Setting of the Alpine Ophiolites from the South Apuseni Mountains (Romania). *Rev. roum. géol., géophys., géogr., (Géol.)*, 25, p. 19–29, Bucureşti.

— **, Savu, H., Nicolae, I., Lupu, M., Vlad, S. (1981)** Alpine Ophiolites Complexes in South Carpathians and South Apuseni Mountains. *Guide to Excursion A_3, Carp.- Balk. Assoc., XIIth Congress*, Bucureşti.

— **, Andrei, J., Borcoş, M., Ciucur, E., Calotă, C., Duma, N., Ioane, D., Ivăşcanu, E., Ivăşcanu, S., Libovski, L., Neştianu, T., Şerbănescu, A., Szabo, E., Vulpescu, G. (1984)** Repport, *Arch. Geol. Inst. of Romania*, Bucureşti.

Giuşcă, D., Cioflica, G., Savu, H. (1963) Vulcanismul mesozoic din masivul Drocea (Munţii Apuseni). *Asoc. Geol. Carp.- Balk., Vth Congress, Com. St. II*, p. 31–44, Bucureşti.

Herz, N., Savu, H. (1974) Plate Tectonics History of Romania. *Geol. Soc. Am. Bull.*, 85, p. 1429–1444, Boulder, Colorado.

— **, Jones, M. L. , Savu, H., Walker, L. R. (1974)** Strontium Isotope Composition of Ophiolitic and Related Rocks, Drocea Mountains, Romania. *Bull. volcanol.*, XXXVIII-4, p. 1110–1124, Napoli.

Faure, G. (1986) Principle of Isotope Geology. John Willey and Sons Ed., 589 p.

Gandrabura, Em. (1981) Studiul mineralogic, petrografic şi geochimic al eruptivului mesozoic din Munţii Trascău. *Ann. Inst. Géol. Géophys.*, LVIII, p. 5–121, Bucureşti.

Lemne, M., Vâjdea, E., Borcoş, M., Tănăsescu, A., Romanescu, O. (1983) Des datations K–Ar concernant surtout les magmatites subséquentes alpines des Monts Apuseni. *Ann. Inst. Géol. Géophys.*, LXI, p. 375–386, Bucureşti.

Lupu, M. (1972) Stratigrafia şi structura formaţiunilor mesozoice din Munţii Trascău. Rezumatul tezei de doctorat, Univ. Bucureşti.

— **(1976)** The Main Tectonic Features of the South Apuseni Mountains. *Rev. roum. géol., géophys., géogr., (Géol.)*, 20, 1, p. 21–25, Bucureşti.

— (1983) The Mesozoic History of the South Apuseni Mountains. *An. Inst. Geol. Geophys.*, LX, p. 115–124, Bucureşti.

— , **Avram, E., Antonescu, E., Lupu, D., Nicolae, I. (1993)** The Neojurassic and the Eocretaceous of the Drocea Mountains. The Stratigraphy and Structure of an Ensialic Marginal Basin. *Rom. J. Tect. & Reg. Geol.*, 75, p. 53–66, Bucureşti.

— , **Antonescu, E., Avram, E., Dumitrică, P., Nicolae, I. (1995)** Comments concerning the age of some ophiolites in the northern part of the Drocea Mountains (Romania). *Rom. J. Tect. & Reg. Geol.*, 76, p. 21–25, Bucureşti.

Nicolae, I. (1983) Consideraţii pe marginea interpretării cadrului tectonic al ofiolitelor din Munţii Apuseni de Sud. *St. cerc. geol., geofiz., geogr., (Geol.)*, 28, p. 35–45, Bucureşti.

— **(1985)** Ophiolites of the Trascău Mountains (South Apuseni Mountains). *An. Inst. Geol. Geofiz.*, LXV, p. 143–205, Bucureşti.

— , **Cuna, S., Soroiu, M. (1987)** Preliminary K–Ar Investigations of ophiolites from the South Apuseni Mountains (Romania). *St. cerc. geol., geofiz., geogr., (Geofiz.)*, 25, p. 43–49, Bucureşti.

— , **Soroiu, M., Bonhomme, G. M. (1992)** Ages K–Ar de quelques ophiolites des Monts Apuseni du Sud (Roumanie) et leur signification géologique. *Géologie Alpine*, 68, p. 77–83, Grenoble.

Pearce, J. A. (1975) Basalt geochemistry used to investigate past tectonic environments on Cyprus. *Tectonophysics*, 25, p. 41–67, Amsterdam.

— , **Cann, J. R. (1973)** Tectonic setting of basic volcanic rocks determined using trace elements analyses. *Earth Planet. Sci. Lett.*, 19, p. 290–300, Amsterdam.

— , **Gale, G. H. (1977)** Identification of Ore Deposition Environment from Trace Element Geochemistry of Associated Igneous Host Rocks. *The Geol. Soc. of London Proc., Joint Meet. Volc. Stud. Group of the G. L. F. and I. M. M.*, 21–22 Jan. 1976, London.

Rădulescu, D., Săndulescu, M. (1973) The Plate Tectonic Concept and the Geological Structures of the Carpathians. *Tectonophysics*, 16, p. 155–161, Amsterdam.

Shervais, J. W. (1982) Ti–V plots and the petrogenesis of modern and ophiolitic lavas. *Earth and Planet. Sci. Lett.*, 59, p. 101–118, Amsterdam.

Savu, H. (1980) Genesis of the Alpine Cycle Ophiolites from Romania and their Associated Calc-Alkaline and Alkaline Volcanics. *An. Inst. Geol. Geofiz.*, LVI, p. 55–77, Bucureşti.

— **(1983)** Geotectonic and Magmatic Evolution of the Mureş Zone (Apuseni Mountains) – Romania. *An. Inst. Geol. Geofiz.*, LXI, p. 253–262, Bucureşti.

— , **Udrescu, C., Neacşu, V. (1982)** Contributions to the Petrology and Metallogenesis of Alpine Initialites from the Vălişoara-Dumeşti Region (Metaliferi Mountains), with notes on the Pre-Alpine basement. *D. S. Inst. Geol. Geofiz.*, LXVII/1, p. 47–67, Bucureşti.

— , **Berbeleac, I., Udrescu, C., Neacşu, V. (1985)** The Study of Ophiolites and Island Arc Magmatites in the Băgara-Visca-Luncoi Region (Mureş Zone - Apuseni Mountains). *D. S. Inst. Geol. Geofiz.*, LXIX/1, p. 63–86, Bucureşti.

— , **Udrescu, C., Neacşu, V. (1986)** Bimodal Volcanism in the North-Western Island Arc of the Mureş Zone. *D. S. Inst. Geol. Geofiz.*, 70–71/1, p. 153–170, Bucureşti.

— , **Udrescu, C., Neacşu, V. (1992 a)** On the presence of ocean floor rocks (Liassic Ophiolites) in the Trascău Mountains (Mureş Zone). Their Petrology and Geochemistry. *Rom. J. Petrology*, 75, p. 53–62, Bucureşti.

— , **Udrescu, C., Neacşu, V. (1992 b)** Petrology and geochemistry of the Late Kimmerian leucocrate dyke swarm between Vărădia and Troaş (Mureş Zone). *Rom. J. Petrology*, 75, p. 63–74, Bucureşti.

— , **Stoian, M., Tiepac, I., Popescu, Gh., Grabari, G. (1994)** Petrological Significance of REE, U, Th, K, Rb, Sr, Hf and of the $^{87}Sr/^{86}Sr$ in the Basaltic Complex of the Mureş Zone. *Rom. J. Petrology*, 76, p. 77–84, Bucureşti.

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THE SUNKEN WESTERN PART OF THE EASTERN CARPATHIANS FORELAND

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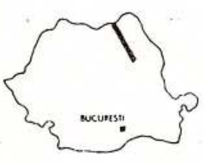

Key words: East Carpathian Foreland. Upper Jurassic. Cretaceous. Microfacies. Foraminifera. Ostracoda. Algae.

Abstract: The present paper deals with the western sunken part of the East-Carpathian Foreland, extending between the Suceava and Bistriţa Valleys, with peculiar emphasis laid upon the investigation of the Mesozoic sedimentary succession. The Mesozoic deposits, lying below the overthrust of the Lower Miocene Molasse and of the Eastern Carpathian Flysch, transgressively overlie the Vendian, Cambrian, Silurian or Devonian complexes and belong to the Jurassic (Upper Tithonian) and Cretaceous (Berriasian-Lower Valanginian, Upper Valanginian, Aptian, Upper Albian, Vraconian-Lower Cenomanian, Coniacian-Santonian, Lower and Upper Campanian).

Two major geological units can be distinguished at present within the Eastern Carpathians Foreland: the Moldavian Platform, epicarelian, an extension of the Eastern European Platform over the Romanian territory and the Central-European Platform, epipaleozoic. North of Romania's frontier, the boundary between the two units is represented by the Tornquist-Teysseire tectonic alignment, also including the Rava-Ruska Fault. In our country, this boundary is still debatable, but one may state that most of the deep, Mesozoic and older, sedimentary underlying the Lower Miocene molasse and the Paleogene flysch overthrust, is part of the Central European Platform.

During the last decades, geophysical, mainly seismic, prospectings, together with research drillings have been carried out within the Eastern Carpathian Foreland. Their main targets have been the Badenian and the Sarmatian, but a number of wells have equally penetrated older deposits, either Mesozoic or Paleozoic (even as far as the basement) the results on the geological constitution, the sedimentary conditions and the structure of some of the formations opened by drillings are presented in various papers. Nevertheless, fewer references have been made on the Mesozoic sedimentary; the present paper is an attempt to enrich these references as much as the information provided by the sunken south-western part of the Foreland between the Suceava and the Bistriţa valleys allow it.

The Deposit Succession

The Mesozoic transgressively overlies, with various terms, a whole range of complexes, belonging to the Vendian, Cambrian, Silurian or Devonian. The oldest Mesozoic deposits encountered are represented by the Upper Tithonian, the youngest ones by the Campanian. A Paleogene deposit sequence (Danian-Paleocene) can be equally added. Nevertheless, the succession is far from complete, but it is marked at various levels by larger or smaller stratigraphical gaps, testifying to the striking unstableness of the sedimentary area.

In order to better figure out the various Mesozoic complexes maps have been drawn out the present thickness variations (which might differ from the normal ones) and enclosed in the paper. For want of a better knowledge degree, these maps can be regarded as a first, incomplete but improvable, attempt.

JURASSIC

UPPER TITHONIAN

The Jurassic deposits belonging to the uppermost Tithonian have been identified by drillings in two sectors:

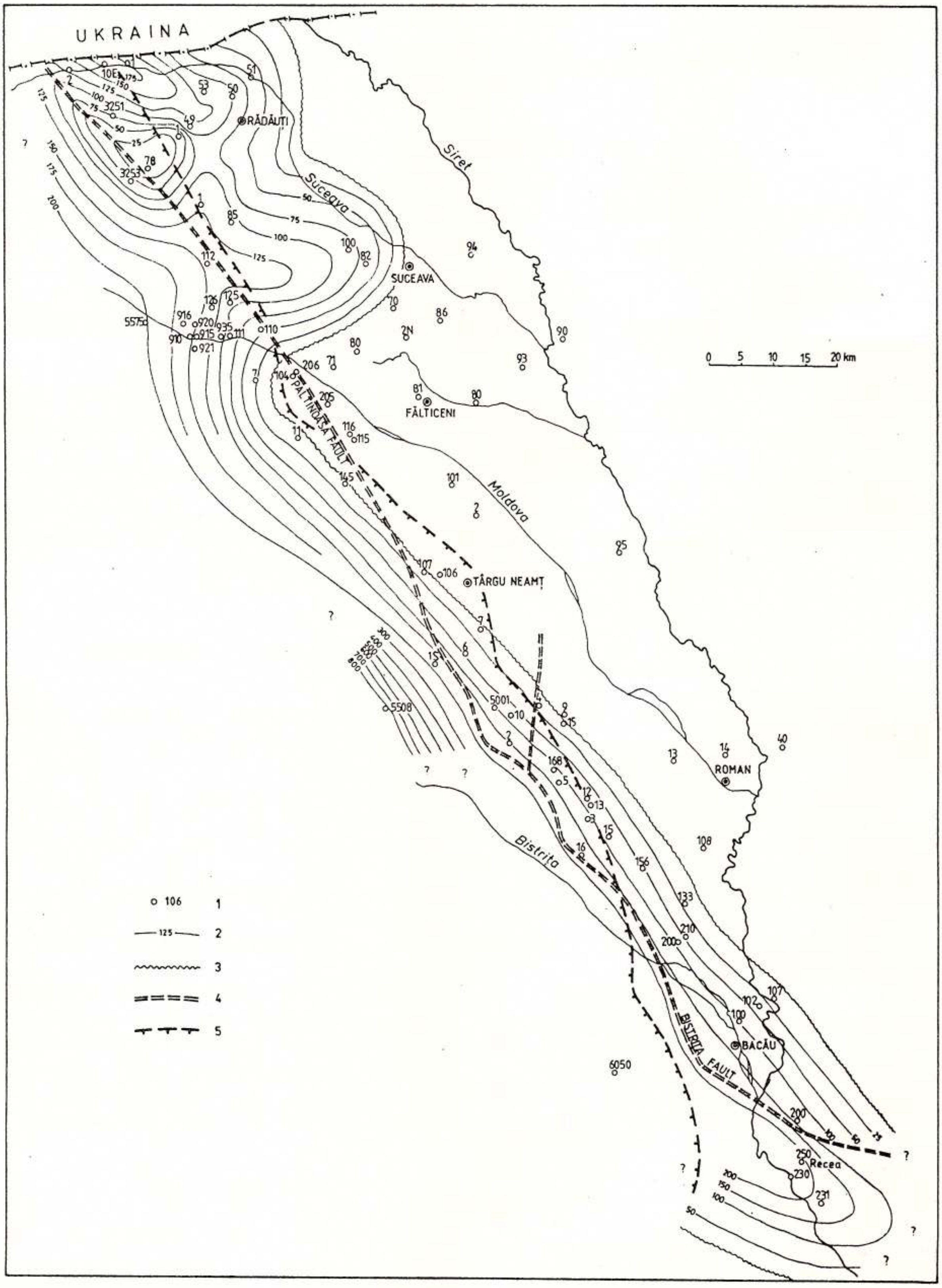

Fig. 1 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Upper Jurassic. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

– a northern one, between the Suceava and the Moldova valley, where it transgressively overlies the Paleozoic, and

– a southern one, in the Bistriţa valley, where the substratum is represented by the Vendian at Bodeşti, but stays unknown at Cuejdiu.

Deposits possibly belonging to the Tithonian have been also crossed by two wells in the Bacău area, but age assignment is still uncertain.

The respective sectors are likely to be interrelated, although nothing proves it within the areals separating them.

In the northern sector, the Tithonian deposits stand for the southern extension of the ones already known within the same geological unit of the Subcarpathian Ukraine and they prevailingly develop in detrital facies. The succession begins by mottled brecciae, followed by brick-brown or grey-greenish conglomerates (with elements of quartzites, phyllites and red ferruginous limestones) and by sandstones with siliceous, dolomitic, ferruginous or clayey-ferruginous cement. In between there get interbedded gritty calcarenites, gritty ferruginous marls, poorly aleuritic marly and brown-garnet red clays. Interbeddings also occur of micritic, biomicritic or microsparitic limestones, next to dolomicrites and dolomicrosparites and to microgritty calcareous or ferruginous dolomicrites. It is worth noticing the presence of anhydrites as enclaves and filmy interlayers or as centimetrically-thick layers.

The deposit thickness increases from east to west, maximally reaching 225 m. On the eastern border of the sedimentary area, thicker zones are remarked, in alternance with thinner ones (Fig. 1).

In the carbonate area at the uppermost part of the sequence in the 10 E Straja well there have been identified specimens of *Clypeina jurassica*, associated to *Salpingoporella annulata* and *Actinoporella podolica*, testifying to an Upper Tithonian age.

In the limestones from the well 5575 Frasin, the investigated microfauna is represented by radiolaria, ostracoda (the *Darwinula oblonga* species, common in the Clypeina jurassica zone) and algae (the *Salpingoporella annulata* and *S. appenninica* species, common throughout the Upper Tithonian).

In the southern sector, the Upper Tithonian was positively identified in the well 5508 Cuejdiu (Piatra-Neamţ), while loggings have estimated its presumable presence further east or south in the wells 5001 Boldeşti, 133 and 156 Secueni, 102 Bacău and 231 Turca Mare.

The deposits within this sector predominantly develop in carbonate facies (micritic, biomicritic and microsparitic grey-yellowish or tan limestones), sporadically associated to dolomicrites with anhydrite nests and films, grey calcareous sandstones or, more seldom, thin-bedded anhydrites.

The deposit thickness rapidly increases from 80 m in the well 5001 Bodeşti, up to over 500 m in the 5508 Cuejdiu well. Here, the thickness is obviously higher since the well stopped in the Tithonian. It was estimated at about 700 m on the thickness variation map (Fig. 1).

In the Bodeşti and Bacău wells, the Upper Tithonian overlies the Vendian. At Cuejdiu, the substratum is unknown, but the Upper Jurassic deposits are regarded as extending into the Bârlad Depression, where both the Malm and the Dogger are present.

The deposit microfacies in the southern sector prevails in foraminifera, sponges, hydrozoans, gastropoda, lamellibranchiata, ostracoda, echinidae, algae. Lituolidae and milliolidae, specific of the Neocomian boundary area, are worth noticing, taking into account the fact that the species *Pseudocyclammina irregularis, Everticyclammina virguliana* and *Kurnubia palastiniensis* never exceed the Jurassic. The algal assemblage is extremely rich, represented by dasycladaceae (*Clypeina jurassica, C. parvula, Acicularia elongata, Macroporella praturloni, Salpingoporella annulata*), codiaceae (*Cayeuxia moldavica, Arabicodium jurassicum, Lithocodium aggregatum*), solenoporaceae (*Picnoporidium lobatum*) and cyanophyceae (*Microoncoliths*) etc.

As a rule, the microfacies with all its components, can be drawn parallel to the Carpathian one from the Moesian Platform, the Black Sea Shelf a.s.o.

The Upper Tithonian deposits were generated by a different sedimentogenesis, namely an inner shelf one, possibly situated at the boundary with the outer marginal-litoral environment in the northern sector, and an outer shelf one, proximal to certain algal ridges in the southern sector. The ratio between the two sectors still needs further research.

CRETACEOUS

BERRIASIAN–LOWER VALANGINIAN

It is equally known within the two sectors, including the underlying Upper Tithonian.

In the northern sector, the deposits are prevailingly terrigenous, represented by conglomerates (mainly with red ferruginous calcareous elements with mollusc and ostracoda bioclasts), sandstones with calcareous-clayey cement, sandy calcareous or microgritty variegated marls, with anhydrite nests and mottled clays. Nor are there absent thin interlayers of poorly clayey, ferruginous, micritic and microsparitic limestones, of dolomicrosparites and dolomicrites or anhydrites.

The lithological deposit facies resembles that of the Upper Tithonian and suggests similar sedimentoge-

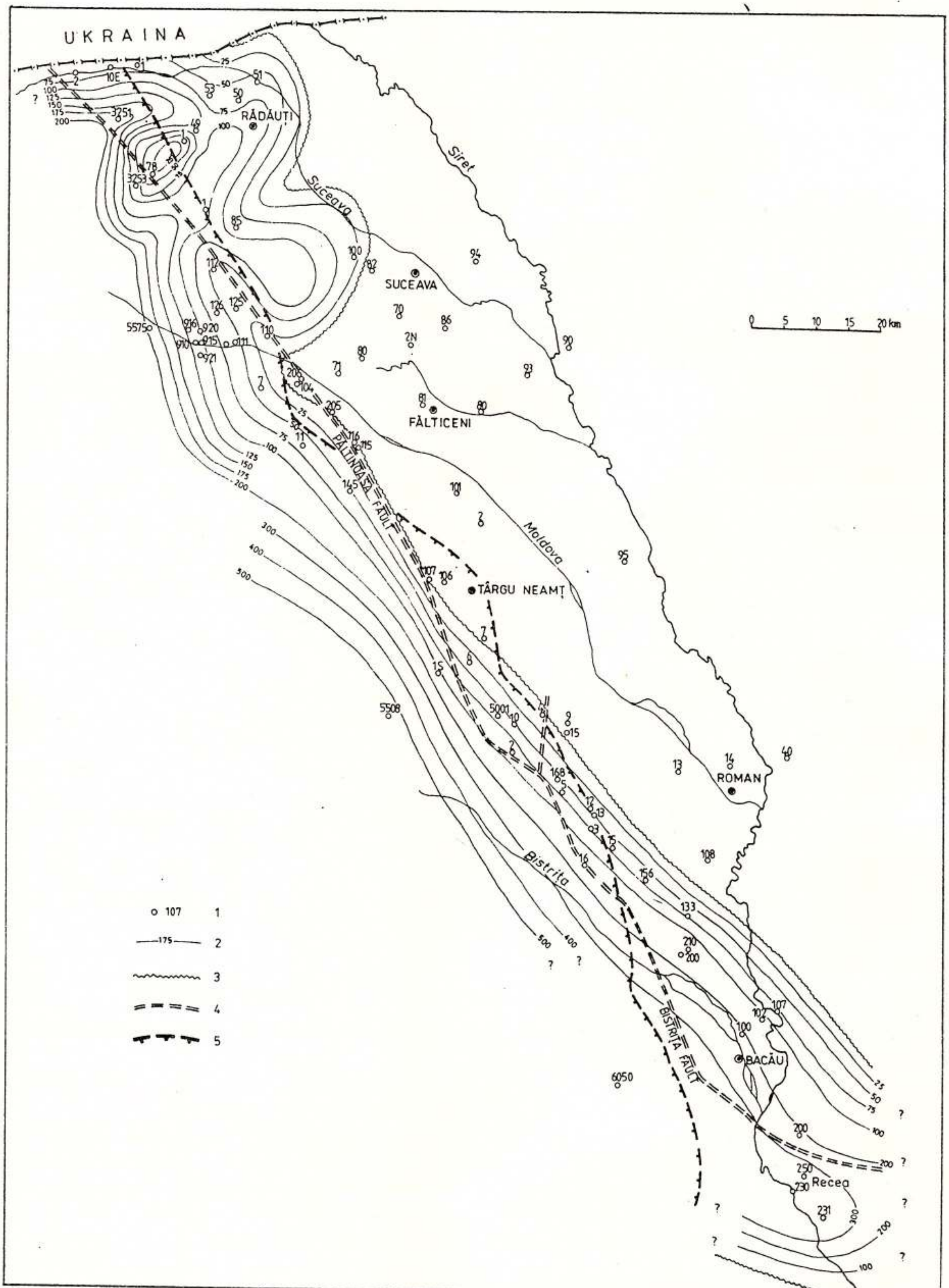

Fig. 2 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Berriasian-Lower Valanginian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

netic conditions. This resemblance, together with a rather unconclusive fossil content, have determined certain Subcarpathian Ukraine researchers (Voronova et al., 1969; Dulub, 1972; Prokopenko, Gorbunova, 1975) to comprise all of them into a Jurassic-Infravalanginian overstage.

Generally, the deposits are poor in life forms. In a few wells there have been however identified lituolidae, favreinae, ostracoda and Chara oogoans, which point to the Berriasian-Lower Valanginian and which in the well 10 E Straja directly overlie the Clypeina jurassica-bearing Upper Tithonian.

A Berriasian-Lower Valanginian assemblage, correlable to that in the 10 E Straja well, was equally encountered in the well 1 Vicov.

The deposit thickness increases from east to west (Fig. 2) but against this background, in the northwest of the sector, as well as in the Turonian, there are noticed thicker areas separated from thinner ones. The fauna and lithology suggest a lagoonal-continental basin with tidal events.

In the southern sector, the Berriasian-Lower Valanginian was paleontologically identified in the wells 5001 Bodeşti, 5508 Cuejdiu, 133 and 156 Secueni, 250 Recea and 6050 Tescani, while geo-electrical loggings have estimated it as presumably present in a few other drillings. The carbonate facies is prevalent within this sector (micritic and microsparitic, more or less diagenized, slightly clayey, often dolomitized limestones, in places with anhydrite nests), but calcareous or anhydritic sandstones also occur sporadically.

The deposit thickness rapidly increases from east to west, from 62 m in the Bodeşti well to 550 m in the Cuejdiu well.

The microfauna is relatively rich and there could be distinguished an *Anchispirocyclina lusitanica* assemblage in the base and an assemblage comprising favreinae, ostracoda (*Cypridea, Cytherella*), foraminifera (*Pfenderina neocominesis, Spirocyclinidae, Ophthalmidiidae*) and algae (*Cayeuxia anae, C. piae, Pseudoepimastopora cretacea, Actinoporella podolica, Thaumatoporella parvovesiculifera* etc) at the upper part. The micropaleontological assemblage quite obviously testifies to the Berriasian-Lower Valanginian age of the deposits and to sedimentogenetic conditions within an outer, shallow shelf with still aired waters.

UPPER VALANGINIAN

It has been identified within the same abovementioned sectors, only over a smaller areal. Lithologically, the deposits are practically similar in both sectors, predominantly carbonate (micritic and microsparitic limestones, calcarenites and calcirudites), with certain interlayers of calcareous sandstones and marls with foraminiferal tests, echinid plates and lamellibranchiata fragments.

The deposit thickness is generally restricted and increases from east to west up to maximally 90 m (the well 5508 Cuejdiu, Fig. 3).

In the carbonate rocks a microfauna has been identified which characterizes the Upper Valanginian, prevailingly consisting of foraminifera (*Pseudotextulariella salvensis, Quinqueloculina minima, Q. robusta, Vidalina bulloides, Trocholina alpina, T. elongata, T. gigantea, T. valdensis*), ostracoda and algae (*Radoicicella subtilis, Pseudoepimastopora, Cayeuxia anae, Bacinella irregularis* etc.).

The fauna and lithology of the deposit suggest sedimentation within a typical open shelf marine basin with shallow or average sheltered depths, poor hydrodynamic conditions, suitable for infralitoral and upper litoral-like stages.

APTIAN

It follows a corresponding Hauterivian-Barremian gap and was identified in the wells 100 Suceava, 100 and 102 Bacău and 6050 Tescani. It seems to have expanded over a wider area, but the subsequent erosion left behind only isolated patches.

Lithologically, the Aptian consists of detritogenous rocks at Suceava and Bacău and of carbonate rocks at Bacău and Tescani. The deposits reach up to 80 m in thickness in the well 6050 Tescani (Fig. 4).

The micropaleontological content testifying to its age is rich both quantitatively and qualitatively and consists of foraminifera, bryozoans, corals, echinidal remains, ostracoda and algae. Among foraminifera the species *Paleorbitolina conoidea, P. lenticularis, Choffatella decipiens, Cuneolina* sp., and among algae *Archaelithothamnium* are to be distinguished.

The whole assemblage suggests the platform-like sedimentogenesis at the boundary between the inner and the outer zones proximal to the algal ridge separating them.

UPPER ALBIAN

It has been solely identified in the Moldova Valley, where it overlies various Neocomian terms or the Aptian. Lithologically, it is predominantly detrital and represented by glauconitic sandstones with calcitic cement, quartzous sandstones with siliceous cement and gritty marls. However, there also occur interbeddings of grey clayey micritic limestones or of biomicritic and biosparitic silicified limestones.

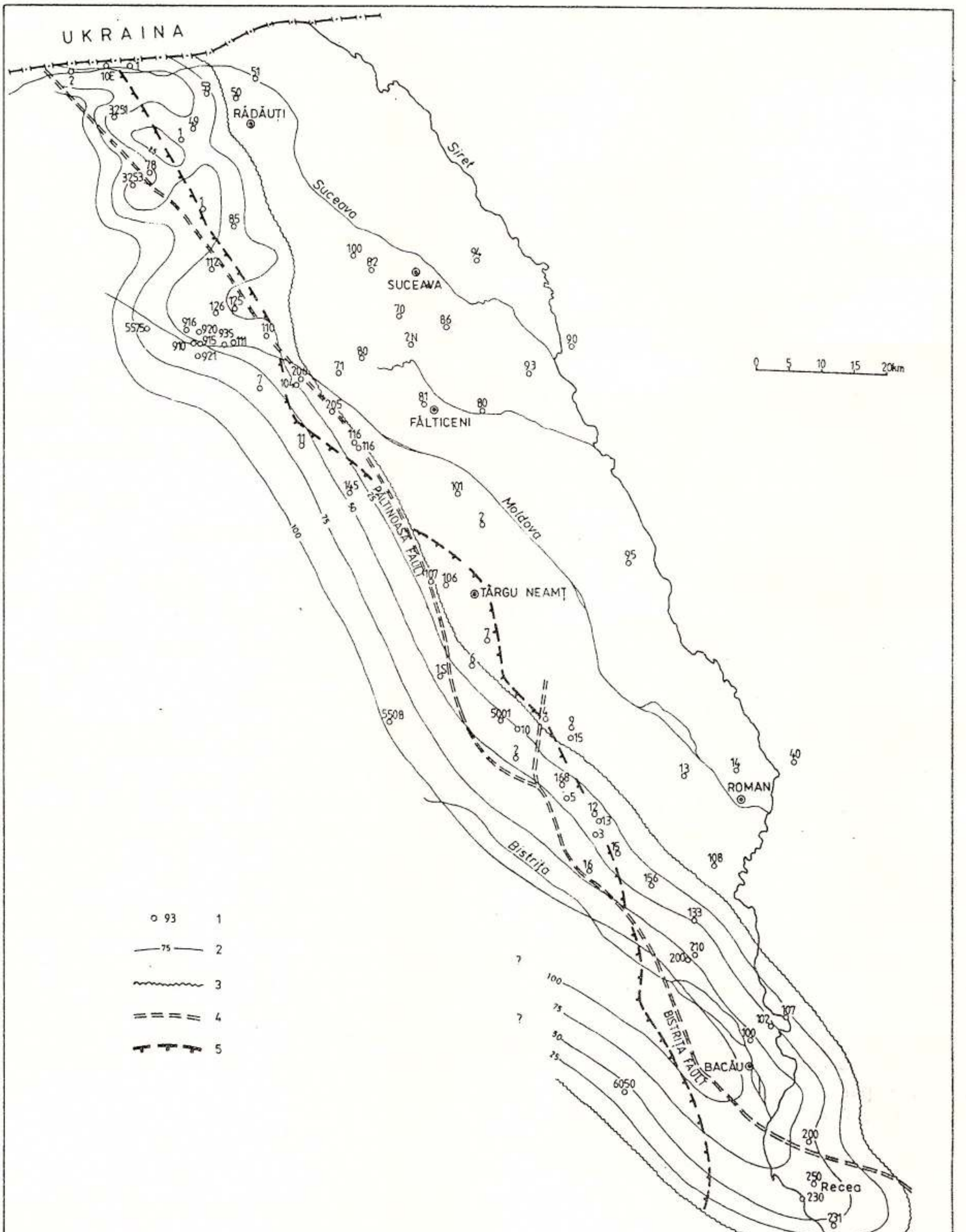

Fig. 3 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Upper Valanginian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

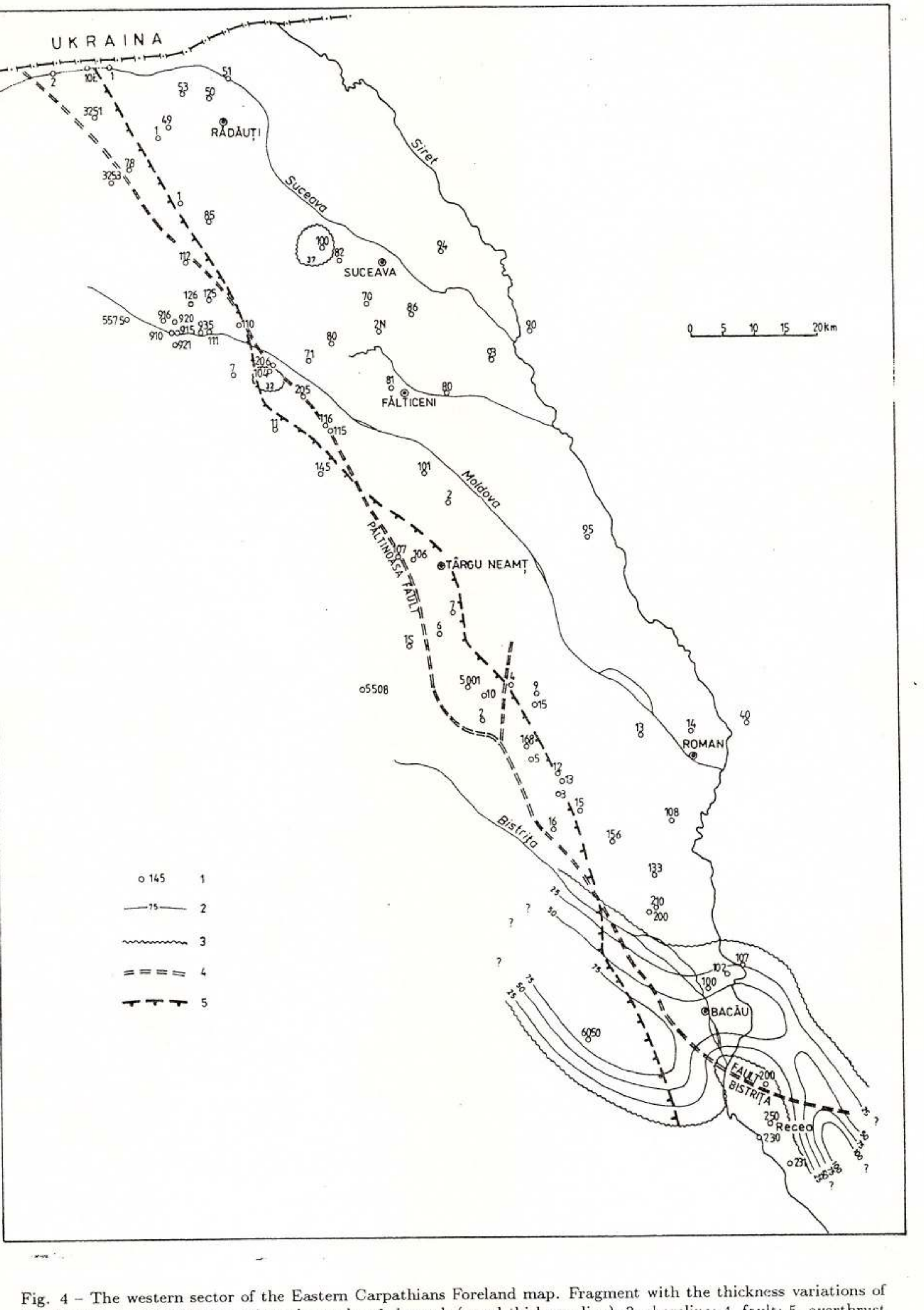

Fig. 4 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Aptian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

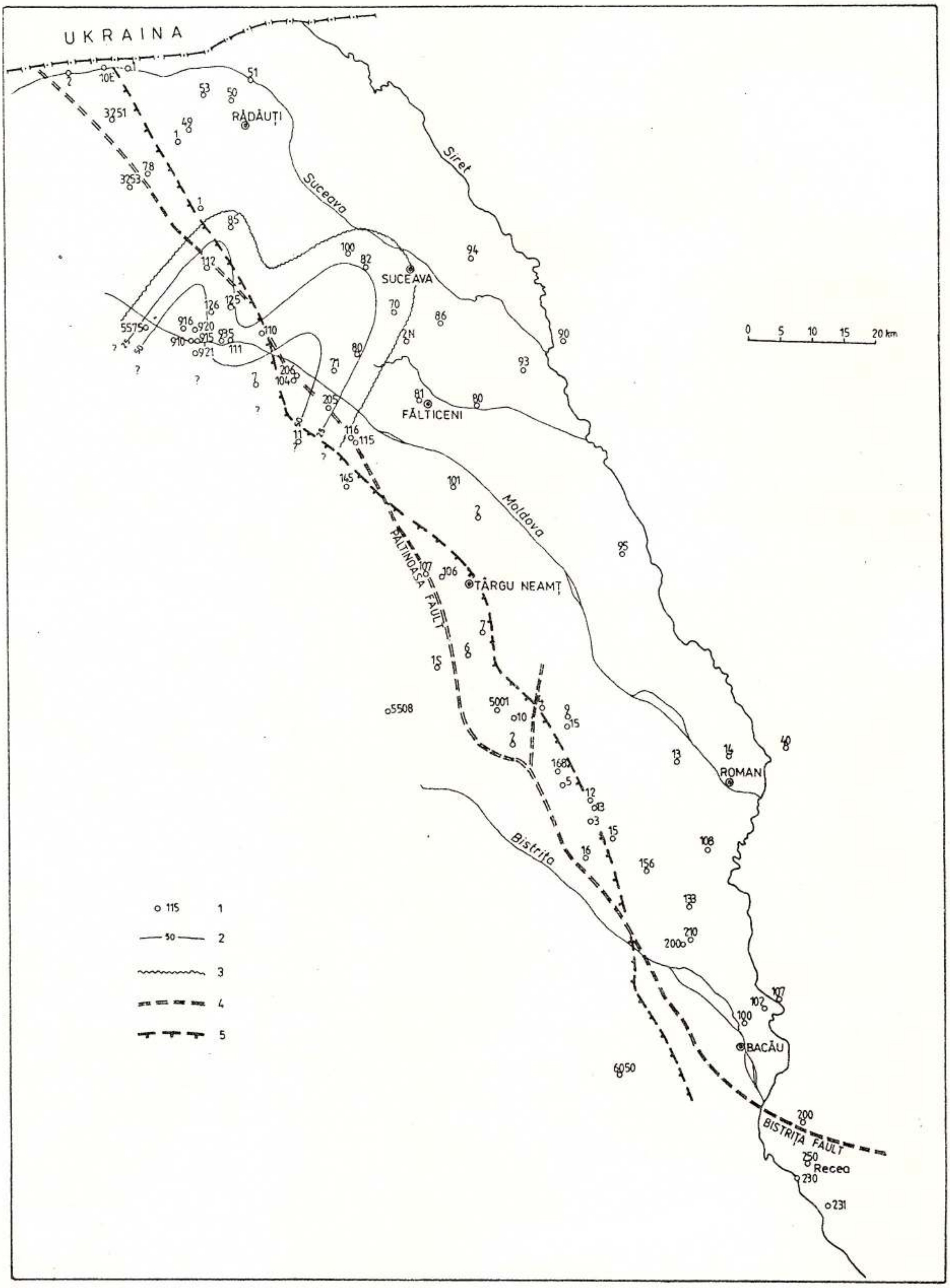

Fig. 5 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Upper Albian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

The deposits may reach a maximal (proven) thickness of 70 m (Fig. 5). Their age is testified by a microfaunal assemblage comprising *Hedbergella* and *Ticinella*. Foraminifera are prevalent. There also occur skeletal remains of sponges, inocerams, ostracoda, fish and coproliths. In the wells 100 Suceava and 910 Gura Humorului a nannoplankton assemblage was identified which equally characterizes the Upper Albian, while the well 104 Valea Seacă also provided with a typical spore-pollen and dinoflagellata assemblage.

As a rule, the lithology and fauna indicate an open, normal marine basin, corresponding to the extralitoral stage of the prelitoral domain.

VRACONIAN-UPPER CENOMANIAN

It stands for a particularly significant stage in the Eastern Carpathians Foreland geological history, when the sedimentary area extended farther eastwards as a result of a general negative displacement, within this framework, the deposits overlie a substratum ranging from the Vendian, Cambrian and Silurian to various older Cretaceous formations.

Lithologically, the Vraco-Cenomanian displays a variable composition. The carbonate rocks are represented by micritic, biomicritic and biosparitic limestones, in places gritty or clayey, with glauconite and silicifications, as well as by spongolitic limestones abundant in siliceous sponges.

The detrital rocks include sandstones with calcitic (partially silicified) cement, reddish, yellowish and grey, glauconite-bearing sandstones, quartzous sandstones, glauconitic sands, marls and clayey marls.

Highly specific are sillicolites, particularly expanded west of the Siret, where represented by spongolites (the well 80 Fălticeni) with opale or calcedonia background, numerous sponge spicules and spongolite-gneisses (wells 94 Lespezi and 100 Suceava) with the same siliceous, opaline or calcedonian background. The quartz, silicosponges, foraminifera, phosphatic pellets and lithological glauconite, the presence of the glauconite, phosphatic pellets and siliceous spicules, point to marine shelf environment sedimentation with depth ranging from 60 to 200 m.

Within the investigated area, the deposit thickness generally increases from east to west, while against this background the thickness variations also point to thicker or thinner areas generated as such by the burial of certain relief forms (Fig. 6). The distributional area of the Vraco-Cenomanian deposits south-wardly closes in the Cuejdiu-Turca Mare-Secueni sector.

The paleontological content of the deposits is extremely rich, represented by the Rotalipora and silicosponge assemblage. Micropaleontologically, they also comprise radiolaria and nannoplankton. Among foraminifera, an outstanding expansion is witnessed by *Ataxophragmidae*, *Nodosaridae*, *Heterohelicidae*, *Planomalinidae*, *Rotaliporidae*, *Anomalinidae*, *Pleurostomellidae* etc.

The vertical evolution of the above-mentioned assemblage testifies to its Vraconian-Lower Cenomanian age. The nannofossil assemblage, together with the macrofauna collected in the well 85 Botoşana, prove the same age.

From the Siret Valley eastwards, planktonic foraminifera become extinct and get replaced by siliceous organisms, mainly by sponge spicules which stay exclusive and are commonest.

The Upper Cenomanian and the Turonian lack throughout the investigated areal. Their presence, at least that of the Turonian somewhere in the area, is however suggested by certain typical foraminiferal complexes, reworked during the Lower Senonian.

SENONIAN

It transgressively overlies the Vraconian-Lower Cenomanian, practically (but for the Maastrichtian) being present all over. The facies is predominantly carbonate (biomicritic and micritic, clayey limestones, biosparitic limestones, clayey marls) with foraminifera, radiolaria, ostracoda, cocolithaceae, dinoflagellata and hystrichospaeridae, featuring sedimentation within an open marine basin benefitting from optimal bionomic conditions.

The *CONIACIAN-SANTONIAN*, primarily carbonate, equally comprises spongolitic limestones and sandstones with calcitic-clayey cement, containing inoceram prisms and numerous sponge skeletons.

The deposits are generally thin and decrease in thickness from east to west. Higher thicknesses, about 70–90 m, are recorded in the wells 200 and 210 Buhuşi and a western shore outlines west of the wells 5508 Cuejdiu and 6050 Tescani while the eastern one roughly follows the Suceava and Moldova streams. The areal closes right north of the well 200 Recea (Fig. 7).

A rich microfauna has been identified, which characterizes the *Globotruncana angusticarinata* and *Pithonella ovalis* assemblage.

The *LOWER CAMPANIAN* spreads over a wider surface than the Coniacian-Santonian, along the Siret stream north of Rădăuţi and north of the well 40 Sagna, from where the surface extends farther eastwards.

The sedimentary area of the Lower Campanian seems to have been much wider than now, since corresponding deposits encountered by certain drillings in Iaşi and Vaslui looked as erosional patches, while outcropping on the Prut bank at Crasnoleca, and east of the Prut River. Thus, the deposits grow in thickness

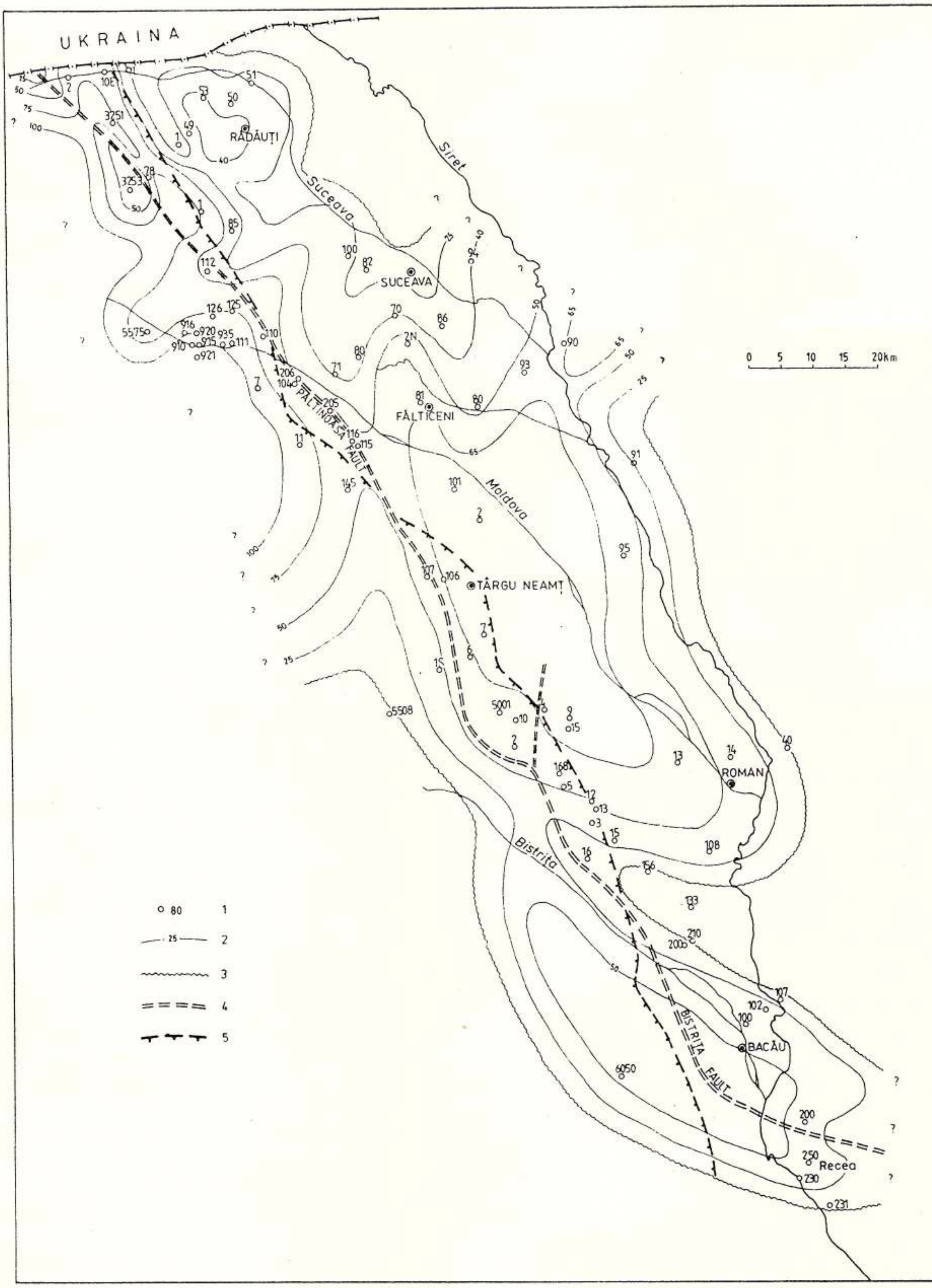

Fig. 6 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Vraconian-Lower Cenomanian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

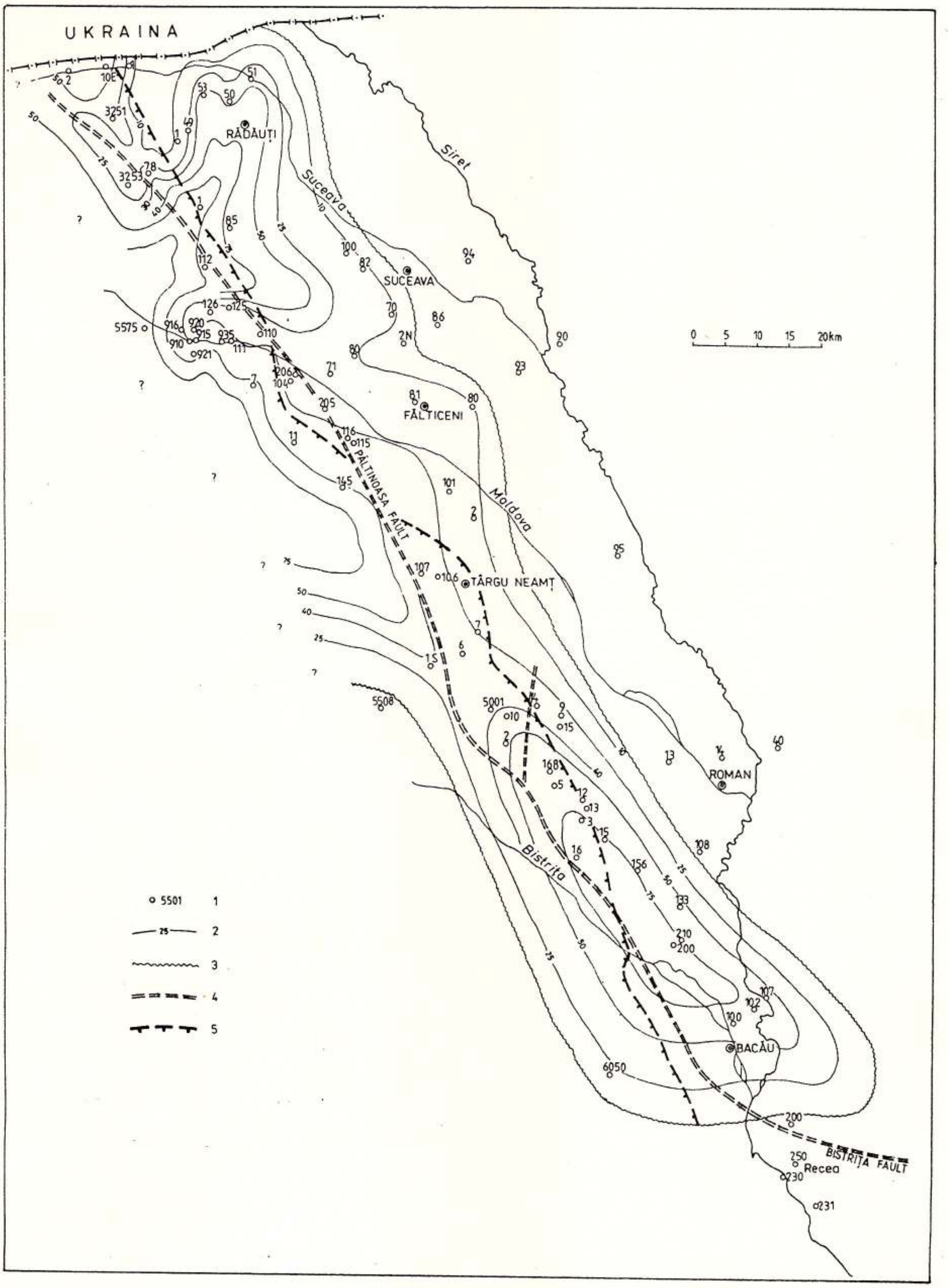

Fig. 7 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Coniacian-Santonian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

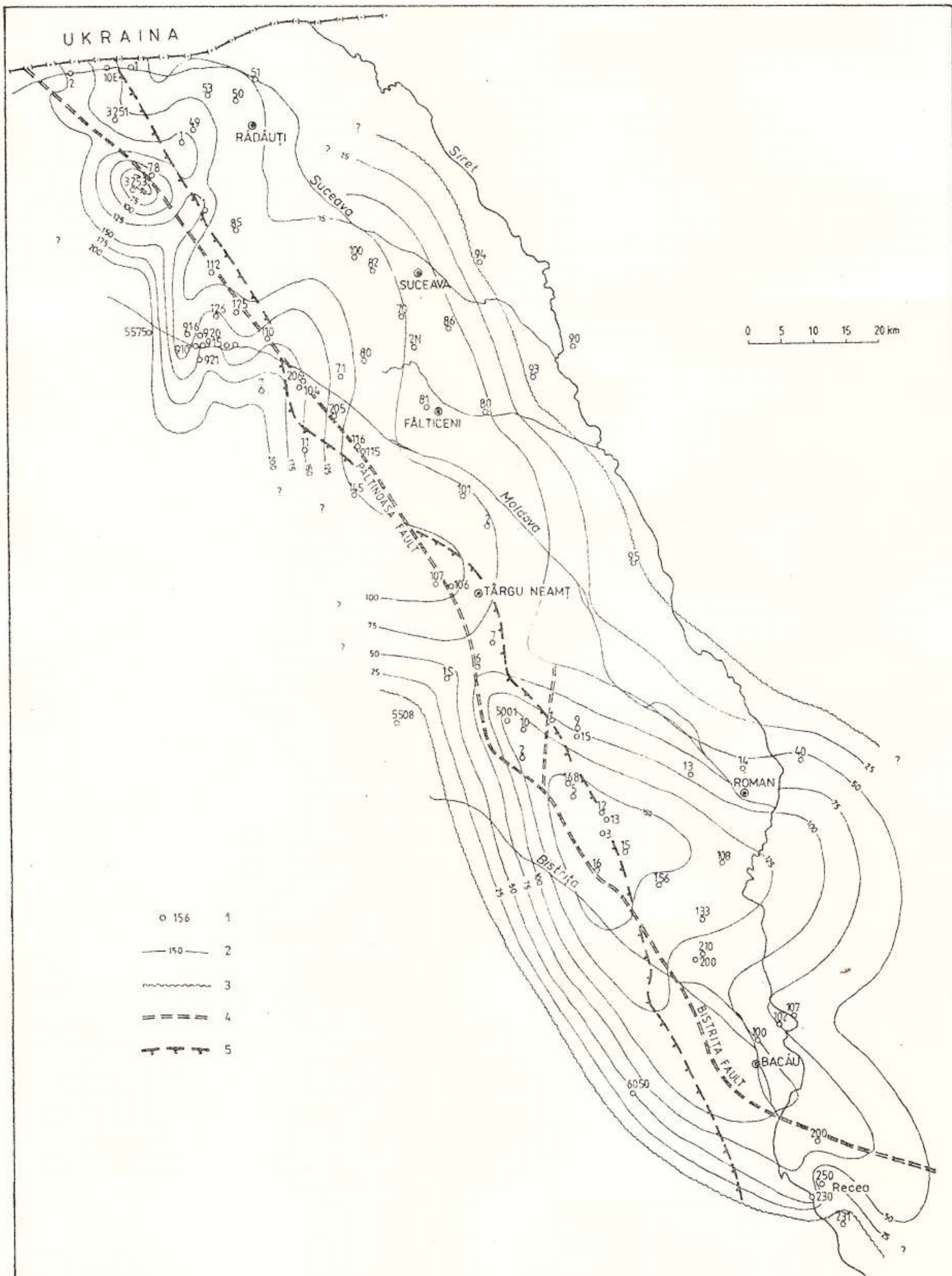

Fig. 8 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Lower Campanian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

from east to west, up to maximally 200 m. The thickness variations point to thinner areas, possibly ancient, buried relief forms or functional stratigraphical conditions. A western shore is outlined in the proximity of the well 5508 Cuejdiu well (Fig. 8).

The deposits are characterized by the *Globotruncana stuartiformis* and *Heterohelicidae* assemblage, mainly consisting of luxurious plankton, represented by foraminifera, nannoplankton, dinoflagellata, hystrichosphaerae.

The *UPPER CAMPANIAN* was solely identified west of the Siret Valley up to the north in the well 40 Sagna, from where it extends eastwards, roughly covering a surface equal to that of the underlying lower level. Besides the specific carbonate rocks, this level equally comprises micritic limestones with siliceous occurrences and locally even spongolites.

The deposit thickness increases from east to west up to maximally 20 m. The thickness variations reveal a buried preextant relief, also visible at the lower levels.

A western shore is outlined east of the wells 5508 Cuejdiu, 6050 Tescani and 231 Turca Mare (Fig. 9).

The deposits comprise an extremely rich microfaunal assemblage, with *Globotruncana confusa* and *Goesella carpathica*,where calcareous planktonic foraminifera are accompanied by numerous species with imperforated or even agglutinate calcareous test. The calcareous nannoplankton witnesses a qualitative and quantitative climax at this stratigraphical level.

DANIAN-PALEOCENE

The first Tertiary term, the Danian-Paleocene, overlies the Upper Campanian after a stratigraphical gap corresponding to the terminal Upper Campanian, the Maastrichtian and possibly to a segment in the Danian base.

The deposits are thin (never surpassing 30 m in thickness) and cover narrow surfaces, shaped as a north-southwardly extending bed, from the Moldova Valley to the Recea wells area. The western and eastern shores are shown on the thickness map (Fig. 10).

Lithologically, the deposits include both carbonate rocks (micritic and biomicritic limestones, gritty organogenous limestones) and terrigenous detrital rocks.

The fossil content corresponding to the *Globotruncana daubjergensis* microfaunal zone, mainly comprises foraminifera and , subordinately, lithothamnidae thalus fragments and lamellibranchiata shells, characteristic of open marine basinal sedimentation within a pelagic facies.

Brief Geological Considerations

The data presented in the previous chapter conspicuously emphasize the presence of a Mesozoic, platform-like sedimentary area, here denominated the Moldova Depression, which underlies the Lower Miocene molasse overthrust of the Carpathian Flysch within the Eastern Carpathians Foreland. It extends from the Suceava valley to the town of Bacău, from where it bends south-eastwards, gets out of the molasse overthrust and gets linked straightaway to the Bîrlad Depression. At all the stratigraphical levels encountered throughout the Mesozoic, the deposits making up these two depressions are perfectly similar, both sedimentologically and as to the faunal content. This proves that the deposits in question were generated under the same sedimentogenetic conditions and within the same sedimentary area. Furthermore, the Moldave Depression also extends northwards into the Subcarpathian Ukraine, displaying the already-mentioned features.

With the passing time (including the Sarmatian) the depression constantly and powerfully underwent the impact of typically ruptural tectonics; the latter generated a structure in steps and blocks separated by faults of various trendings, with prevalent NNW-SSE orientation. One of the most significant is the Păltinoasa Fault (also known as the Straja-Păltinoasa, Vicov-Păltinoasa or Solca Fault), seismically disclosed and obviously checked by drillings. It could be traced from the northern country frontier down to the town of Bacău (Fig. 11). It separates an eastern, uplifted Foreland step from a western, more sunken one, in places by 600-700 m.

This fault directly prolongates southwards the Kalus Fault from the Subcarpathian Ukraine, an equally significant tectonic accident, positively proven both by seismic and by drilling works. In its turn, the fault separates two structural steps (the western being more sunken) within the Central European Platform. Hence, most part of the Moldave Depression in our country is equally included within the Central European Platform, a fact which questions the boundary over Romanian territory between this large geological unit and the Eastern European Platform (the Moldavian Platform).

This problem was minutely approached by Săndulescu (1984), in relation with the geological data from the Subcarpathian Ukraine, where it could have been solved more readily. Viewpoints are even there contradictory.

According to certain researchers (Dickenstein et al., 1975, in Săndulescu, 1984), the boundary between the two units lies on the Rava-Ruska line, which could be part of an important structural alignment, the Tornquist-Teysseire line, represented by a fracture or

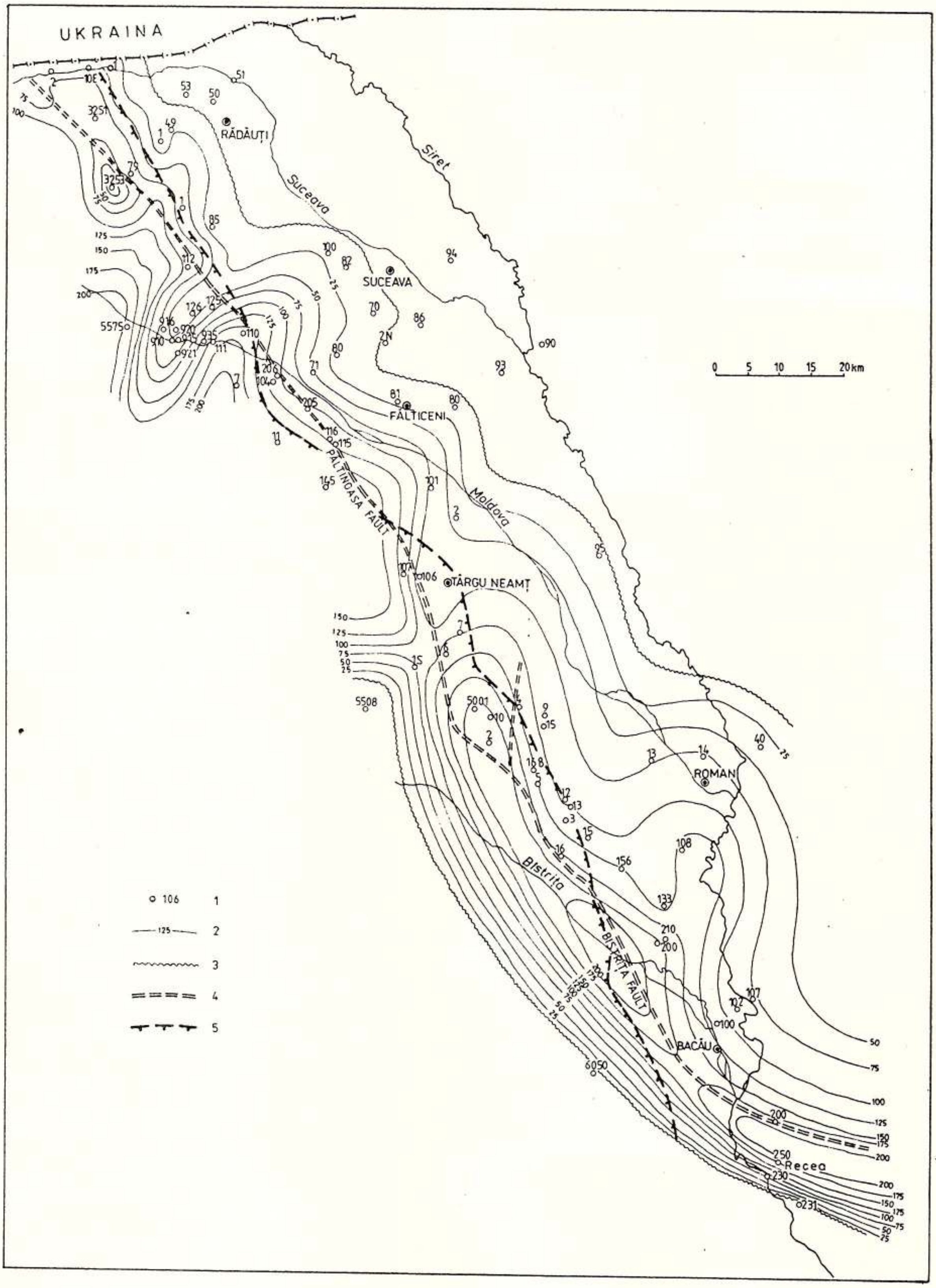

Fig. 9 – The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Upper Campanian. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

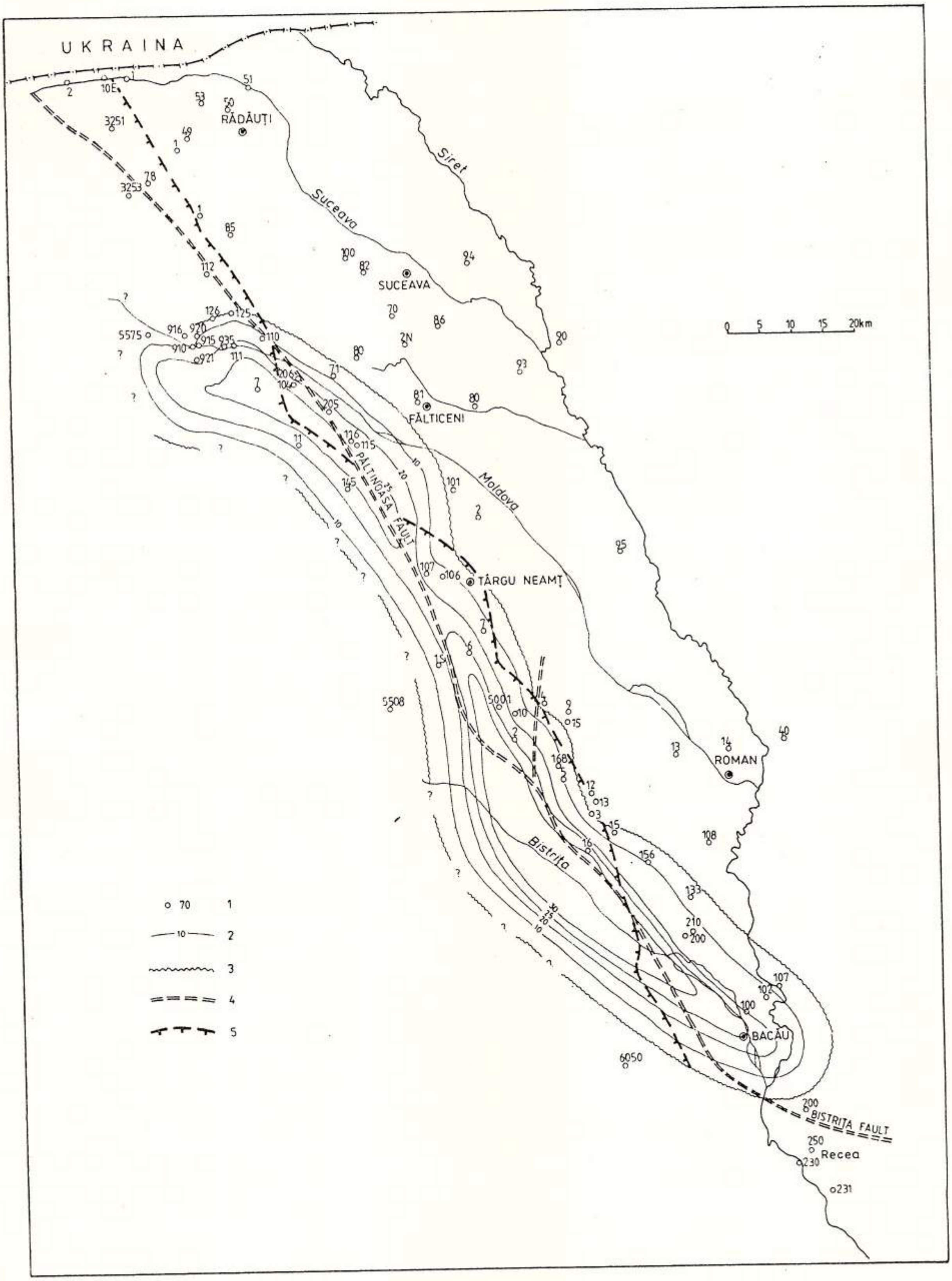

Fig. 10 - The western sector of the Eastern Carpathians Foreland map. Fragment with the thickness variations of the Danian-Paleocene. 1, well with investigated samples; 2, isopach (equal thickness line); 3, shoreline; 4, fault; 5, overthrust line.

a bundle of crustal, parallel and close, vertical fractures.

On the other hand, other researchers (Glusko et al., 1980, in Săndulescu, 1984) regard the Eastern European Platform margin as lying farther eastwards, over a deep fault line, the Ustilog-Rogatin Fault, situated on the eastern flank of the Lwow Depression. Between this fault and the Rava-Ruska one, the platform bottom might be epibaikalian (epicadomian) and not carelian.

While analyzing the extension of these two fault lines over the Romanian territory, Săndulescu (1984) estimates the Rava-Ruska Fault to be continued by the Solca Fault (i.e. the Păltinoasa one), and the Ustilog-Rogatin Fault by a fault following the Siret Valley (the Siret Fault). In between the two faults there lies a sunken tectonic block (the Rădăuţi-Paşcani Block). Judging by the Paleozoic features of the Vendian cover, the Central European Platform could extend even until the Solca Fault (the extension of the Rava-Ruska Fault), but according to the basement features, it could only extend until the Siret Fault (the prolongation of the Ustilog-Rogatin Fault). The Rădăuţi-Paşcani Block, separated in conformity with magnetometric and gravity data, cannot be accurately assigned.

Regarding this briefly sketched hypothesis (Săndulescu, 1984, p. 72), a few facts are worth signalling despite the lack or the contradictory character of the evidence:

- According to the geological map of the Ukraine (1964, sc. 1:1000000), the Ustilog-Rogatin Fault might continue along the Siret valley. Along it, a few fault fragments of a lesser importance have been identified east of Rădăuţi and east of Suceava, at the Neogene cover level, but neither seismic nor drilling data have been supplied by deeper levels.

- There is no evidence as to the south-eastern extension of the Solca Fault until the Roman town area, where it might join the Siret Fault. All present seismic and drilling data, without the least exception, point to its prolongation along the Păltinoasa-Tg. Neamţ-Bacău line (Fig. 11).

- The Vaslui Fault follows the Silurian-Cambrian boundary in the Mesozoic substratum (Fig. 11), passes north (and not south) of the Roman Town and joins the Păltinoasa Fault west of Fălticeni. Evidence on its presence is rather scarce.

- On the geological map of the Ukraine (1969, sc. 1:1000000) the Rava-Ruska Fault is figured out near the frontier with our country following a trajectory similar to that of the Kalus Fault, gets preserved also south of the Suceava valley (Fig. 11) but only over a restricted distance, since afterwards, it actually joins the Kalus Fault, south of the well 78 Suceviţa (Fig. 11). In depth, it separates the Silurian from the Cambrian.

In other papers (Burov et al., 1974), the Rava-Ruska Fault trajectory appears farther eastwards and reaches the frontier with Romania north of Rădăuţi. From there southwards, it might extend over the Fălticeni Fault, seismically identified at the Badenian level and checked by a few drillings. It is of a lesser importance and can be followed over the Rădăuţi-Suceava-Fălticeni-east Tg. Neamţ line until S-E of Piatra Neamţ, where it joins the Păltinoasa Fault (Fig. 11). In depth, it might separate two structural steps within the Paleozoic.

According to the above-presented information, the western margin of the Eastern European Platform (the Moldavian Platform) is still difficult to trace out, but if the features of the Paleozoic cover are taken into account, the Păltinoasa Fault northwardly extending over the Rave-Ruska Fault segment, several choices could sound truthful.

For the present, the southern border of the Moldavian Platform looks less questionable, since most researchers agree on its lying over a WNW-ESE trending fractural line, called the Bistriţa Fault. It separates the Moldavian Platform from the Bîrlad Depression (part of the Predobrudjan Depression), whose Jurassic-Cretaceous sedimentary is fairly and thoroughly developed, overlying a folded basement which belong to a younger platform, the Scythian Platform, now regarded as an equivalent of the Central European Platform.

The ruptural relief over the northern margin of the Barlad Depression has hindered drawing out seismic profiles and has generally provided little information. Actually, the fault has been traced according to certain gravity and magnetometric anomalies, as well as to certain drilling data. The presence of the fault is positive enough, but there is still room for certain, rather unessential viewpoint shifting as to its trajectory, basically in the area between the Bârlad and the Prut Valleys where the situation gets even more intricate because of further events. Săndulescu (1984) has allotted further expansion to the fault, considering it extends along the Bistriţa Valley, beyond Piatra Neamţ (hence its name). The present data do not testify to such an expansion. The fault stops south of the town of Bacău, where it joins the Păltinoasa Fault, thus allowing a straight-forward link between the Barlad Depression and the Moldave Depression in the north. On the other hand, joining the Păltinoasa Fault substantiates the idea that the western margin of the Moldavian Platform lies along this fault, including its prolongation over the Rava-Ruska northern sector, as previously pointed out.

On its southern flank, the Barlad Depression is bordered by the North-Dobrudjan Promontory (Fig. 11) including Paleozoic deposits in its axial area, overlain

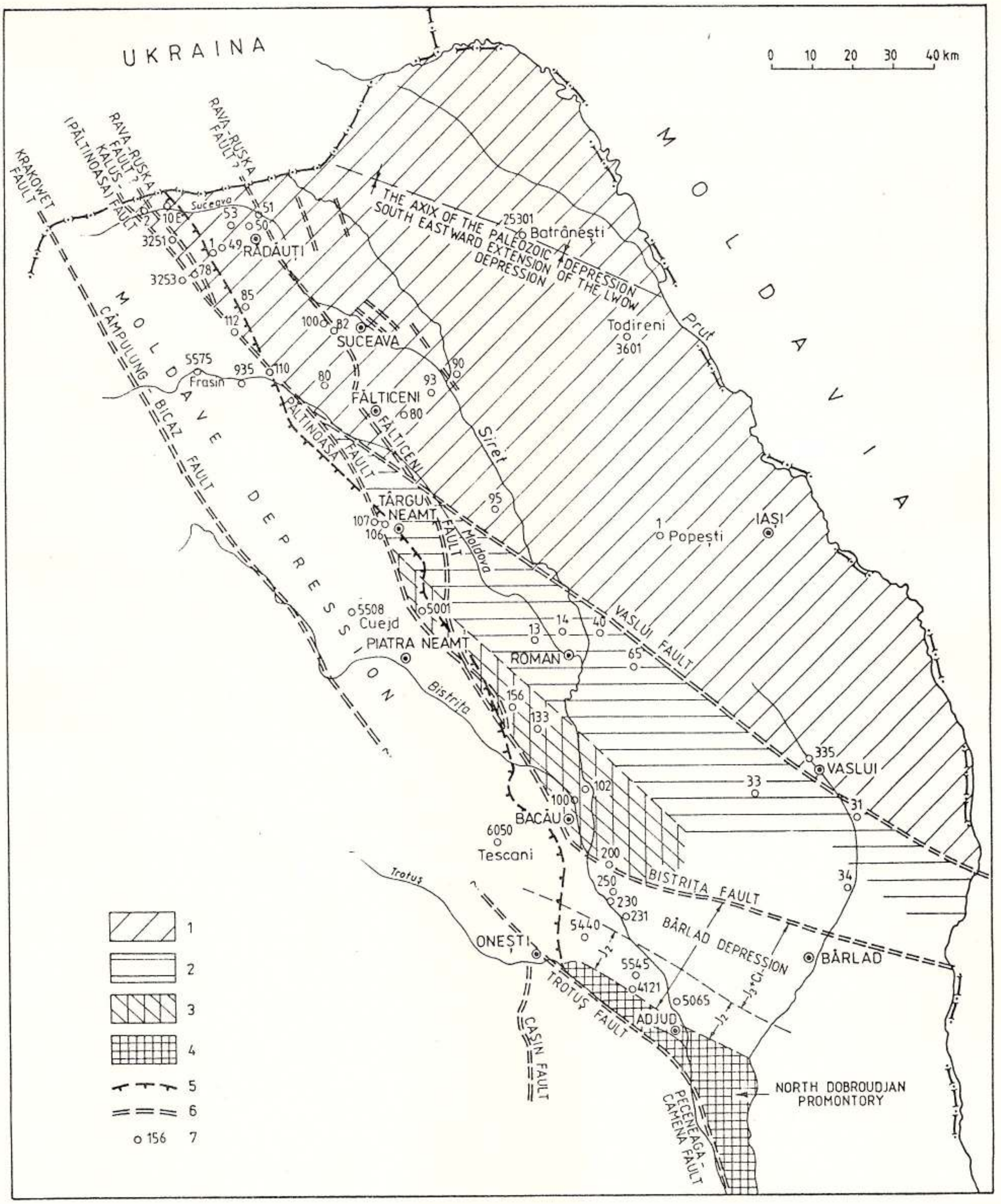

Fig. 11 – The Eastern Carpathians Foreland, the probable spreading of the Paleozoic and vendian deposits, below the mesozoic cover. 1, Silurian; 2, Cambrian; 3, Vendian; 4, undivided Paleozoic (the outer border of the North Dobrudjan promontory); 5, pericarpathian overthrust (at the surface); 6, fault; 7, well with investigated samples.

by the Middle Jurassic in the N-W and southwardly delimited by the Trotuş Fault, which actually represents the north-western prolongation of the Peceneaga-Camena Fault. From the Trotuş Valley towards the N-W this fault seems to extend, below the Carpathian Nappe complex, into the Bicaz-Câmpulung Moldovenesc Fault, the latter extending in its turn into the Krakovetz Fault from the Subcarpathian Ukraine (Fig. 11). The whole ensemble westwardly borders the Moldave Depression.

Mention must be made that west of the Krakovetz Fault deep drillings have proved that the Neogene directly overlies the Proterozoic greenschists (the Lejaisk Massif). This conclusion, related to others, is expected to stimulate geotectonic interpreters to provide a better understanding of the structure of the deep formations in the substratum of the Carpathian massifs.

References

Burov S. V., Mikita V. B., Sakiu A. V. (1974) O sovenosti stroenia razvitia Sventoksisko - Dobrujinskogo Soorujenia. *Sov. Gheol.*, 5, p. 139–144, Moskwa.

Costea I., Balteş N. (1974) Planktonic Organisms as Evolution Markers of the Cretaceous Sedimentary Basin in the Foreland of the Romanian East Carpathians. *Rev. Espan. Micropal.*, VI (3), p. 401–418, Madrid.

— , Comşa D., Motaş L., Voiculescu C., Şindilar V. (1991) Studiul sedimentologic şi stratigrafic al Mezozoicului de pe marginea de nord-vest a Platformei Moldoveneşti. Arhiva I. C. P. P. G. Tema 587, Bucureşti.

— , Vinogradov C., Comşa D., Bonig H. (1981) Studiul microfacial al depozitelor cretacice din Platforma Moldovenească şi Depresiunea Bârladului. *Stud. cerc. geol., geofiz., geogr., Geologie*, 26 (2), p. 275–292, Bucureşti.

Paraschiv D., Dicea O. (1988) Confirmarea prezenţei Devonianului în Platforma Moldovenească prin noi foraje. Lucr. Sem. Geol. "Gr. Cobălcescu", 2, Univ. "Al. I. Cuza" - Iaşi, Fac. Biol.-Geogr.-Geol., Secţia Geologie, p. 101–106, Iaşi.

Pătruţ I. (1982) Le Silurien de la Platforme Moldave et sa position dans le cadre géologique régional. Lucrările sesiunii ştiinţifice "Gr. Cobălcescu". Univ. "Al. I. Cuza" - Iaşi, Fac. Biol.-Geogr.-Geol., Secţia Geologie, p. 181–190, Iaşi.

— , Dăneţ Th. (1985) Le Pre-Cambrien (Vendien) et le Cambrien dans la Plate-forme Moldave. Comunicare în cadrul sesiunii "Gr. Cobălcescu", Univ. "Al. I. Cuza" - Iaşi, Fac. Biol.-Geogr.-Geol., Secţia Geologie, Iaşi.

Popescu M., Donos I., Antonescu F. (1985) Raport, arhiva I. P. G. G., Bucureşti.

Săndulescu M. (1984) Geotectonica României. Edit. tehn., Bucureşti.

Saraiman A., Costea I., Comşa D. (1968) Contribuţii la studiul Cretacicului din Platforma Moldovenească. *II Ştiinţe Naturale*, XIV, p. 49–60. Anal. St. Univ. "Al. I. Cuza", Iaşi.

Văscăuţeanu Th. (1923) Asupra Cretacicului superior din nordul Basarabiei. Acad. Rom. Mem. Sed. Ştiinţ., III/1 (13), p. 1–13, Bucureşti.

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EOCENE FORMATIONS IN THE CĂLATA REGION (NW TRANSYLVANIA): A CRITICAL REVIEW

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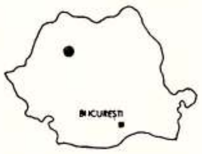

Key words: Eocene. Lithostratigraphic units. Lithostratigraphic nomenclature. Biostratigraphy. Gilău Area of NW Transylvanian Paleogene.

Abstract: The Călata region (south of Huedin town) in the Gilău sedimentary area has been selected due to its favourable research conditions of the Eocene formations, some of them deveoped here in their typical facies. The formal lithostratigraphic units from the area, for which a nomenclature according to the International Stratigraphic Guide (1979) has been adopted, are briefly characterized lithologically and biostratigraphically. These units are, as follows: Jibou Formation (Lutetian in age in the area of study) a continental alluvial red bed type formation, locally with a lacustrine limestone in its base, named Horlacea Limestone (with Lutetian palynological content); Călata Group (a new name for "Rakoczy" Group, this name being preserved only for Rakoczy Sandstone) – including the deposits of the first marine cycle, subdivided in Foidaş Formation (Upper Lutetian), Căpuşu Formation (Late Lutetian-Bartonian), Inucu Formation (Bartonian) with a calcareous member, namely Văleni Limestones, having its stratotype in the investigated area, Ciuleni Formation (Bartonian-Early Priabonian) (a new name for the upper part of the Mortănuşa Subgroup) with the description of the type section, and Viştea Limestone (Priabonian) (returning to the first valid formal name for the Lower Coarse Limestone Formation, subsequently named Leghia Limestone) indicating both the holostratotype and the parastratotypes from the type region. The continental red formation (Valea Nădaşului Formation), overlying Viştea Limestone, as well as the subsequent marine formations (Jebucu Formation and Cluj Limestone), which occur on the map area, are not described in the text. In order to avoid the same name for both a lithostratigraphical unit and a part of it, the name "Cluj" Group should be replaced with Turea Group. There is biostratigraphic evidence for tracing the Lutetian-Bartonian boundary, on the basis of calcareous nannoplankton just above the Pycnodonte brongniarti Level of the Căpuşu Formation, and the Bartonian-Priabonian boundary, on the basis of the nummulites fauna just above the Crassostrea orientalis Level in the Ciuleni Formation.

One hundred years since the issuing of the monography "Die Tertiärbildungen des Beckens der siebenbürgischen Landestheile. I Paläogene Abtheilung" (1894) our paper represents a homage paid to the late Anton Koch, professor of mineralogy and geology at the Cluj University. It is amazing to find out today that the lithostratigraphic subdivision of the Paleogene in NW Transylvania made by Koch based on personal researches and on those of the well-known geologist Karl Hofmann, contemporary to Koch, is still valid. About this Maxim (1959, p. 20) wrote: "Koch was a man of real facts, not of fancies, which he gathered and presented, and that is why his book is and will be a reference paper... and its author remains one of the great classics of stratigraphy". The truth of these lines will also be revealed by the stratigraphic description presented in this paper.

The studied region belongs to the so-called Gilău Area (Rusu, 1970), a sedimentary area within which Popescu (1984) distinguished three subareas: Iara, Călăţele and Aghireşu. The territory rendered on the map (Fig. 1) occupies the central part of the Călăţele subarea, in which the formations belonging to the "Rakoczy Group" (Hofmann, 1879) are exposed, some of them being developed in the most typical facies.

References to the region under study can be found in "Geologie Siebenbürgens" by Hauer and Stache (1863) as well as in Pávay's paper (1871); however, the stratigraphic deciphering of the area is due to Koch who carried out researches for the geological map, scale

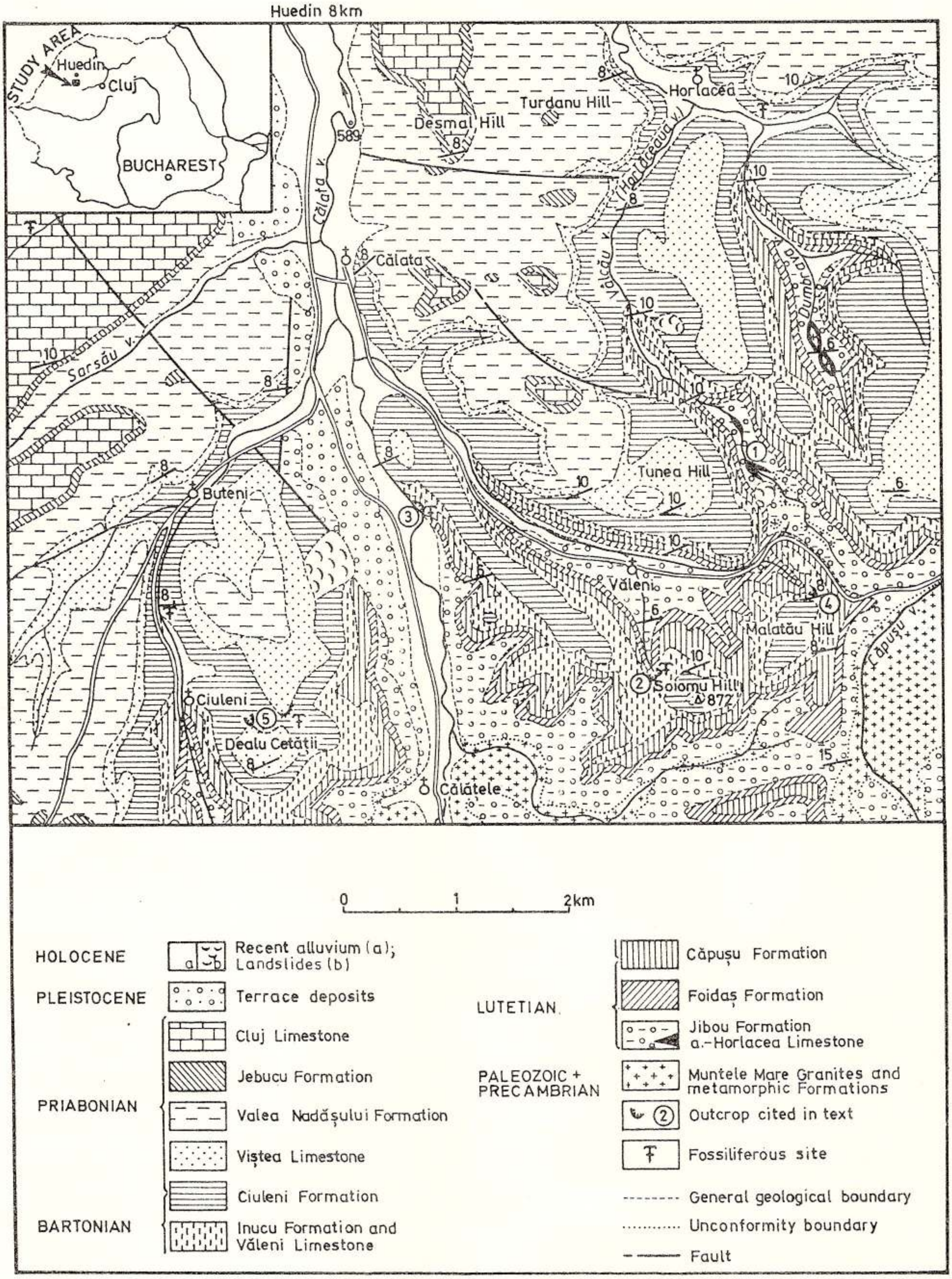

Fig. 1 – Geological map of the Călata region

1:75,000, Huedin sheet (Koch, Hofmann, 1887). Detailed descriptions were presented in the abovementioned monography. Later on, the papers written by Mateescu (1926) and Szádeczki-Kardoss (1930) brought no novelties on the stratigraphy of the Eocene in the Huedin Depression (acc. to Mateescu), the authors using Koch's scheme (1894) without modifications. In 1966, Dragoş made lithostratigraphic separations at rank of formation represented on a geological map, scale 1:50,000; he described in detail the most representative profiles and also estimated the age of the formations. Also in 1966 occurred a paper by Niţă Pion et al. in which the authors maintained almost the same lithostratigraphic subdivisions. These studies made on a restricted area in the Călăţele "Gulf" were based on the results obtained on large areas by Joja (1956), Răileanu and Saulea (1956), and Mészáros (1960).

In the present paper an attempt is made to include the Eocene formations in the Călata area into a modern unitary lithostratigraphic scheme, using a standard nomenclature according to the rules of the International Stratigraphic Guide (1979). This study is based on personal research and takes into consideration all the papers concerning the Paleogene lithostratigraphy in NW Transylvania.

The general lithological and paleontological characteristics of the formal lithostratigraphic units in the Călata area will be presented, starting with Jibou Formation – the first Eocene term in the studied area – and ending with Viştea Limestone – the most recent formation of the first marine cycle, a stratigraphic interval for which some critical remarks should be made.

JIBOU FORMATION (HOFMANN, 1879)

Gruppe der bunten Thone, Süsswasserkalke und Mergel von Zsibó (HOFMANN, 1879), p. 240)

Unter bunte Thonschichten + Unterer Horizont des Süsswasserkaltes (KOCH, 1894, p. 191, 203)

Seria argilelor vărgate inferioare (RĂILEANU, SAULEA, 1956, p. 274; VLAICU-TĂTĂRÎM, 1963, p. 27; NIŢĂ PION et al., 1966, p. 303)

Seria vărgată inferioară (MÉSZÁROS, 1960, p. 89)

Complex de argile continentale roşii (BOMBIŢĂ, 1963, p. 98)

Orizontul argilelor pestriţe inferioare (DRAGOŞ, 1966, p. 351)

Complexul vărgat inferior (RĂILEANU, MÉSZÁROS, 1966, p. 451)

Jibou Formation (POPESCU, 1978, p. 100; POPESCU et al., 1978, p. 298; POPESCU, 1984, p. 42)

Formation de Jibou (MÉSZÁROS, MOISESCU, 1991, p. 33)

The old basement constituted of Precambrian metamorphites and Paleozoic granites in the studied area is overlain by a continental formation of red beds-type, formed of red-violaceous or brick-red siltic clays, spotted or variegated, red sandy clays with pebbles, red or white-greenish sands, frequently with cross-bedding, gravels and polygenous conglomerates with red sandy-argillaceous matrix or red argillaceous sandstone, respectively. This formation is delimited by stratigraphic unconformity planes both at the lower part (where it covers a preexistent relief, in places accidented) and at the upper part (where it is cut off by the marine transgression) and it displays variable thicknesses ranging between 20 m and 50 m, extremely reduced in comparison with known in the type region (more than 1500 m).

Jibou Formation starts, in the studied area, with a thin level (0.1–1 m) of breccias with reworked components of substratum included in a red or greenish pelito-psamitic matrix; these breccias represent, at least partially, an eluvial deposit. No alteration crust of the type presented by Niţa Tătărîm (1963) in the Stolna-Hăşdate as "lateritic soils" can be found in this zone. These alteration crusts, formed quasi-continuously in the Paleocene-Ypresian time span and erroded intermittently, are preserved only in a few places on the border of the Apuseni Mountains. The remainder of the deposits, in fact representing the main part of the formation, are clearly alluvial as pointed out by many authors and recently proved by sedimentological observations made by Drăgănescu (in Rusu et al., 1993, unpubl. data).

Within Jibou Formation, lacustrine limestones are situated in the upper part (Koch, 1894, p. 207), at Călăţele, and in the base of the formation, south of Horlacea (Dragoş, 1966, p. 353, 355). As regards the latter, for which we used in a geological report from 1989 the name "Horlacea Limestone", we shall deal with them further on.

The final part of the Jibou Formation can be studied in the outcrop from the north-western part of the Şoiomu Hill at Văleni (no. 2 on the geological map, Fig. 1). Here, in the base of the outcrop (Fig. 3), red-violaceous siltic clays, which gradually pass to grey-greenish siltic clays (1m) overlain by the deposits of the Foidaş Formation, are observed. In our opinion the changing colour of the red clays from the top of the continental formation is due to the transformation of the Fe^{3+} into Fe^{2+} within a reducing medium beneath the water/sediment interface concomitantly with the Lutetian marine transgression. In almost all the sections in NW Transylvania this situation could be observed giving the impression of a sedimentary continuity even in areas where Jibou Formation supports directly the deposits of the Căpuşu Formation or even younger terms.

As regards the age of the Jibou Formation, it is still under discussion. Within the framework of the paleogeographic evolution of the "Transylvanian land", it is considered that the formation of the red continental deposits began at the end of the Maastrichtian. For the Jibou Formation, in the north-western part of the Transylvania, the only certain dating elements are represented by the Charophyte associations determined by Iva (in Bombiţă et al., 1979, unpubl. data) and the remains of Aligatodides evidenced by Grigorescu (in Rusu et al., 1988, unpubl. data) in the Rona Limestone (Ypresian), at Jibou. Today, in the studied area, biostratigraphic evidence pointing to the presence of the Lutetian in the Horlacea Limestone is also available. Considering all this, we can assume that Jibou Formation (Paleocene(?)-Ypresian-Lutetian in age) have been strongly heterochronous. For the studied area, one should admit that the sedimentation started only in the Lower Lutetian.

Horlacea Limestone (RUSU)

Towards the spring zone of the Horlăceaua Valley (or Vâlcău Valley) from Horlacea (Fig. 1, outcrop 1), over the basement of micaschists lies a sedimentary breccia (10–30 cm), which includes only crystalline material. It underlies (Fig. 2):

– 20 cm bank of grey-yellowish sandy dolomitic limestone, with microruditic elements in the base;

– 5–10 cm grey-blackish clayey interbed, with coaly material. The samples studied by Antonescu supplied a spore-pollen association considered to be not older than Lutetian age (Antonescu et al., 1988, unpubl. data);

– 1.5 m grey-bluish massive dolomitic limestones, with rare Chara fruits;

– 30 cm dark greenish siltic clays, which pass to a pile of red violaceous clays, typical of the Jibou Formation.

It would represent the type section of the Horlacea Limestone which is not exceeding 2 m in thickness. In the same valley the sequence of lacustrine limestones is also outcropping (first: 200 m downstream; second: 350 m more) comprising already in the base red and grey clays of the Jibou Formation.

At about 1.2 km to the northeast in the Dumbrava Valley (Valea Mare) at Horlacea locality, the Horlacea Limestone is found again, a bit changed in lithologic respect; it consists of a 3 m brownish, coarse-grained calcareous sandstones (lying directly over a granite basement) and 3.5 m grey-yellowish sandy dolomitic limestones (overlain by red-violaceous clays).

Here, Horlacea Limestone (Lutetian) represents, therefore, a local carbonatic member of the Jibou Formation, situated in the base of this formation.

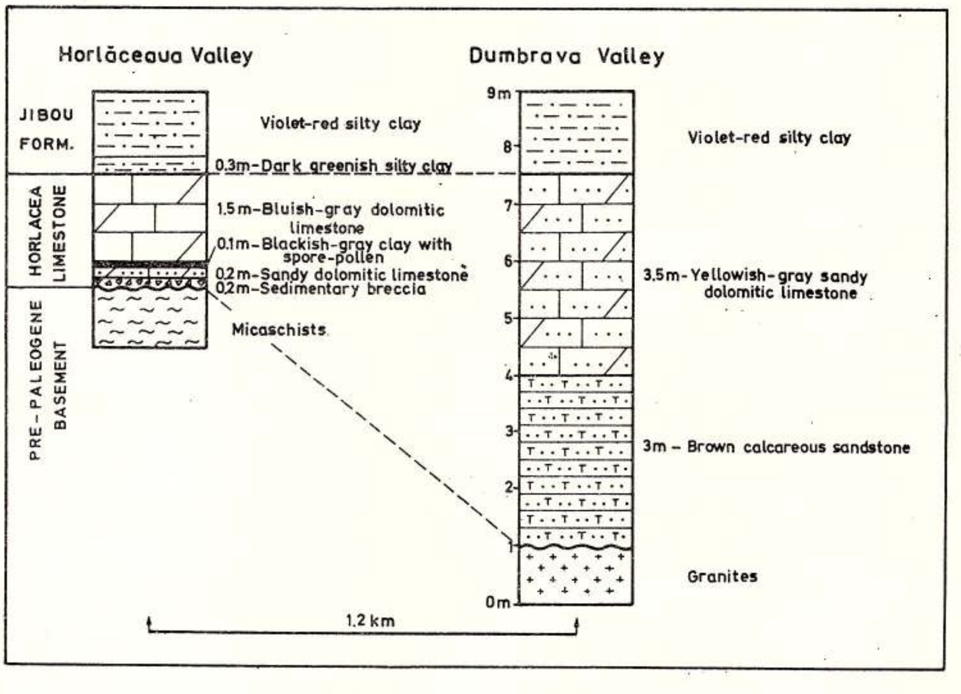

Fig. 2 – Stratigraphic sections in the Horlacea Limestone (Lutetian) at Horlacea

FOIDAŞ FORMATION (BOMBIŢĂ, 1984)

Perforata-Schichten (partim-Horizont der unteren Gypsbänke oder des Anomyenkalkes und Mergels)[1] (KOCH, 1894, p. 210, 222)
Orizontul gipsului inferior şi al marno-calcarelor cu Anomya (partim) (RĂILEANU, SAULEA, 1956, p. 278)
Orizontul marno-calcaros cu Anomii şi al gipsurilor inferioare (MÉSZÁROS, 1960, p. 90)
Orizontul gipsului inferior şi al marno-calcarelor cu Anomia (partim-Nivelul inferior) (VLAICU-TĂTĂRÎM, 1963, p. 37)
Orizontul marnelor şi al calcarelor cu Gryphaea eszterházyi şi Rostellaria (partim) (DRAGOŞ, 1966, p. 356)
Orizontul cu Anomia (partim) (NIŢĂ PION et al., 1966, p. 305)
Lower Gypsum Formation (POPESCU, 1978, p. 100; 1984, p. 45)
Lower Gypsum (POPESCU et al., 1978, p. 300)
Gypse de Foidaş (BOMBIŢĂ, 1984 a, p. 31)
Les Couches de Foidaş (MÉSZÁROS, MOISESCU, 1991, p. 33)

Foidaş Formation represents the first term of the Lutetian marine transgression in NW Transylvania. In the section, below the Şoiomu Hill (Fig. 3), this lithostratigraphic entity, mostly pelitic, is 8.5 m thick and consists of grey and grey-greenish clays with thin levels of laminated dolomitic claystones and grey-bluish siltic marly limestones. In the middle part of the succession a 50 cm thick dolomicritic bank of small-sized eurihaline molluscs (Cerithiids, Naticids, Corbulids, Lucinids, Cardiids, Venerids etc.), typical of the restrictive environment of the evaporitic facies, is intercalated. Both in the upper and in the lower part of the clayey sequence occur centimetric intercalations of coaly clays (indicating the proximity of the shore). The marly-clayey sequence (2 m thick) with tellinids and benthonic microforaminifera of polyhaline waters (nonionids, miliolids etc.) has been assigned to the Foidaş Formation, considering that this lithostratigraphic unit usually ends with a calcareous level with Anomia (also polyhaline) which here is replaced lithofacially. The miliolidic limestones with Anomia are also found in the neighbourhood, along a lineament situated about 1 km southwards. As a matter of fact, the facies variations are large within this formation, entirely calcareous-dolomitic south of Mănăstireni, where it is only 2–3 m thick. At Leghia, the same formation is pelitic and full of gypsums; the thickness exceeds 40 m. In the presented section at Văleni (Fig. 3), the gypsum is lacking, while in the neighbourhood it is found only as centimetric crusts.

In the lower part Foidaş Formation displays a sharp boundary of sedimentary discontinuity; here, deposits of marine-brackish facies rest on alluvial continental deposits. Locally, at the upper part, the boundary is less marked due to the sudden transition from brackish marine deposits, usually represented by the Anomia Level, to normal marine deposits of the next formation.

Foidaş Formation includes some eurihaline molluscs and ostracods as well as a poor nannoplankton assemblage (Popescu et al., 1978; Mészáros et al., 1987) attributed to the Lutetian.

CĂPUŞU FORMATION (POPESCU, 1978)

Perforata-Schichten (partim-Horizont der unteren Austerbank + Horizont der unteren Molluskenmergeles + Unterer Striata-Horizont + Horizont der Perforata-Bank) (KOCH, 1894, p. 210, 223, 224)
Orizontul gipsului inferior şi al marnocalcarelor cu Anomya (partim) + Bancul cu Nummulites perforatus (RĂILEANU, SAULEA, 1956, p. 278)
Orizontul cu Gryphaea eszterházyi + Orizontul cu Nummulites perforatus (MÉSZÁROS, 1960, p. 92–94)
Orizontul gipsului inferior şi al marnocalcarelor cu Anomya (partim - Nivelul superior) + Orizontul cu Nummulites perforatus (VLAICU-TĂTĂRÎM, 1963, p. 37, 38, 48).
Orizontul marnelor şi al calcarelor cu Gryphaea eszterhazyi şi Rostelaria (partim) + Orizontul cu Nummulites perforatus (DRAGOŞ, 1966, p. 356, 359)
Orizontul cu Anomya (Partim) + Orizontul cu Nummulites perforatus (NIŢĂ PION et al., 1966, p. 305, 307)
Căpuş Member (POPESCU, 1978, p. 101; POPESCU et al., 1978, p. 303)
Les Couches de Căpuşu (MÉSZÁROS, MOISESCU, 1991, p. 33)

Căpuşu Formation comprises the succession of typical marine deposits of marls with intercalations of bioclastic limestones which includes in the base the lumachelle level with *Pycnodonte brongniarti* and ends with the *Nummulites perforatus* bank. In the section under the Şoiomu Hill (Fig. 3), this formation is less than 15 m thick and consists of:

– 1.5 m grey-bluish bioclastic marly limestones, forming two less defined banks which here and there become lumachellic due to the frequency of the *Pycnodonte brongniarti* (BRONN) shells. The marly limestones contain euhaline foraminifera and ostracods, remains of echinoids and crabs; among molluscs, specimens of Globularia, Cardium and Pitar are found be-

[1] For a better correlation we mention in parentheses what the author separated as paleontological or lithological level, belonging to the respective lithostratigraphic unit.

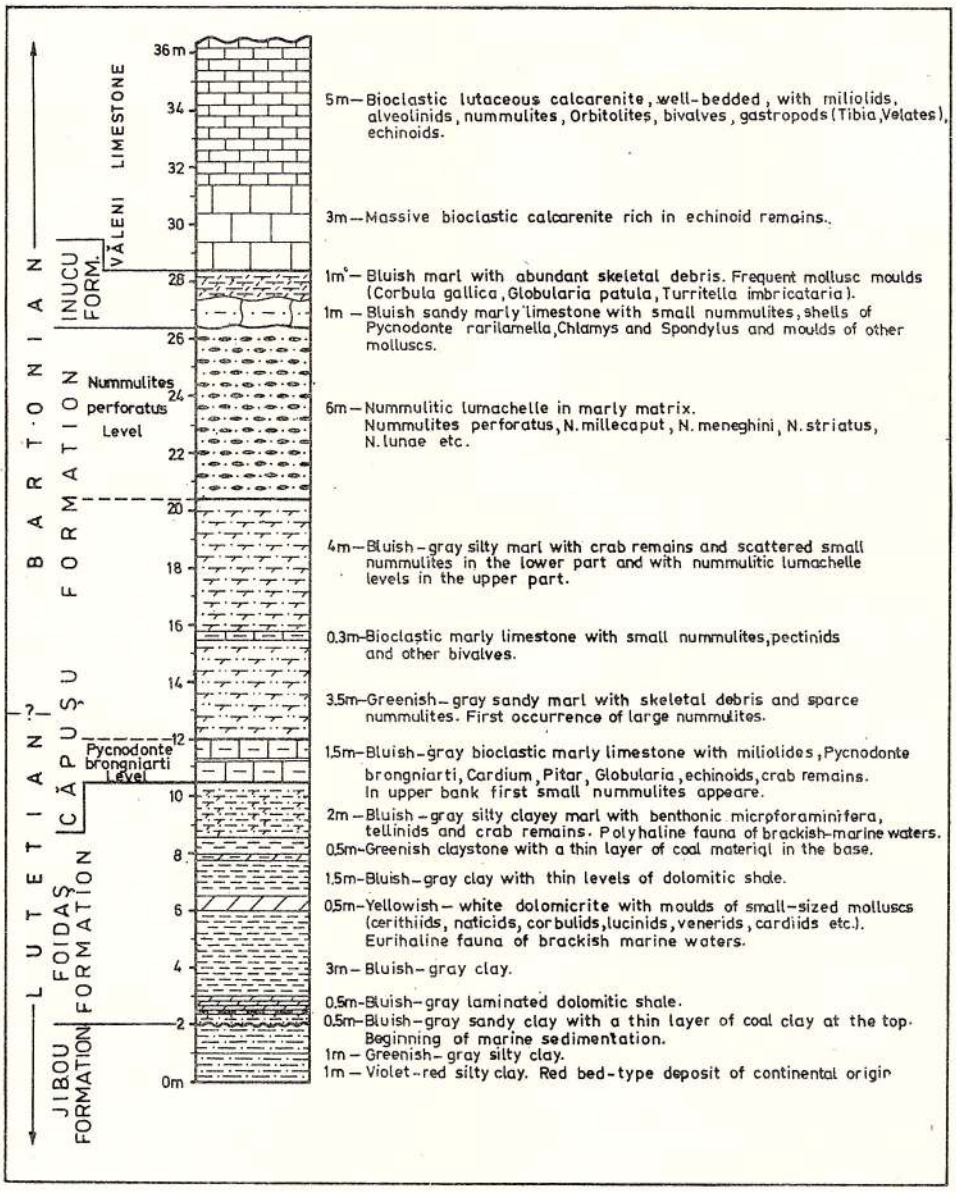

Fig. 3 Stratigraphic succession of the Eocene deposits cropping out in the Şoiomu Hill, at Văleni

side Pycnodonte. The upper bank also includes the first small-sized nummulites, such as *Nummulites striatus* (BRUG.) and *N. lunae* (BOMBIŢĂ) (see Bombiţă, 1984 a). This is the Pycnodonte brongniarti Level, the first biohorizon of the Căpuşu Formation.

– 3.5 m grey-greenish sandy marls with organogenous detritus and rare nummulite tests. At this level occur the first large-sized nummulites, such as *N. perforatus* (MONTF.);

– 0.3 m bioclastic marly limestone with small nummulites, shells of pectinides and moulds of other bivalves;

– 4.5 m grey-bluish siltic marls with rare specimens of small nummulites and remains of crabs in the lower part and lumachellic levels with nummulites in the upper part. In the neighbouring areas this marl pile frequently contains shells of *Sokolowia eszterhazyi* (PÁVAY) considered to be a good biomarker. Northeastwards, in the area of the type locality of the formation (Căpuşu Mic), the marls lying over the Pycnodonte brongniarti Level contain limonitic oolites and glauconite, constituting the known sedimentary iron ore deposit, nowadays depleted.

– 6 m lumachelle with large-sized nummulites in the marly matrix. This lumachelle is an excellent marker level named Nummulites perforatus Level, within which the index species, is accompanied by the assemblage with *Nummulites millecaput*, *N. menegh-*

inii, *N. obesus*, *N. crassus* and *N. uranensis* (see Bombiţă, 1984 b).

Therefore within the Căpuşu Formation there are three main biohorizons: the Pycnodonte brongniarti Level, the Sokolowia eszterhazyi Level, (both with slightly discontinuous developments), and the Nummulites perforatus Level, continuously developed in all the three depositional areas in NW Transylvania.

Until a decade ago the Upper Lutetian age of the Căpuşu Formation seemed solidly proved by the presence of a nummulite assemblage considered typically biarritzian (Bombiţă, 1984 a, b). According to the calcareous nannoplankton zonation carried out by Gheţa (1984) the local Reticulofenestra primitiva Zone in the Căpuşu Formation would be correlated with a part of the NP 16 standard zone (Martini, 1971); for this reason Gheţa assigned almost all this formation to the Bartonian. Recently, Mészáros (1992) pointed out in the "Lower Căpuşu Beds" biostratigraphic elements in favour of the presence of the NP 15 Zone of nannoplankton, assigning the whole formation to the Upper Lutetian. The nannoplanktonic studies carried out by Melinte (in Rusu et al., 1994, unpubl. data) on samples from the Gilău area showed that within the Pycnodonte brongniarti Level there is an assemblage poor in species, dominated by *Reticulofenestra tokodensis* BÁLDI-BEKE (which has its bloom). The species is known to proliferate in slightly hypohaline mediums and has a very reduced stratigraphic distribution, located according to Báldi-Beke (1984) in the middle of the NP 16 Zone, which belongs to the terminal Lutetian. The upper part of the NP 16 Zone, well represented in the studied sections, inclusively at Văleni, would belong to the Bartonian, pointing to the absolute age 41.2±2.1 m.a. obtained by Odin (1978) on glauconite from this level.

Therefore, the Căpuşu Formation would be Bartonian in age, excepting the extreme basal part (Pycnodonte brongniarti Level) which would correspond to the terminal Lutetian.

INUCU FORMATION (MÉSZÁROS, MOISESCU, 1991)

Perforata-Schichten (partim – Oberer Striata Horizont + Horizont des mittleren Molluskenmergels + Horizont der oberen Austernbank (KOCH, 1894, p. 1894, p. 210, 225, 226, 228)

Orizontul argilelor cenuşii (partim) (RĂILEANU, SAULEA, 1956, p. 282; VLAICU-TĂTĂRÎM, 1963, p. 51)

Orizontul marnelor şi calcarelor cu moluşte (partim) (MÉSZÁROS, 1960, p. 95)

Argile marnoase de Mortănuşa (partim) (BOMBIŢĂ, 1963, p. 102)

Orizontul marnelor şi calcarelor cu Velates şi Corbula (partim) (DRAGOŞ, 1966, p. 361)

Orizontul marnelor şi calcarelor cu Corbula gallica LAMK. = Orizontul argilelor cenuşii (partim) (NIŢĂ PION et al., 1966, p. 308)

Marnes supérieures á mollusques (BOMBIŢĂ et al., 1975, p. 164)

Mortănuşa Marls (partim) – Upper Molluscan Marls (POPESCU et al., 1978, p. 311, p. 312)

Les Couches de Inucu (MÉSZÁROS, MOISESCU, 1991, p. 34)

Inucu Formation is found for the first time as a lithostratigraphic entity in the description of the section at Leghia in a micropaleontological guidebook (Bombiţă et al., 1975, p. 164, Fig. 1). It was also described by Popescu et al. (1978) as a member of the "Mortănuşa Marls" and recently it was called Inucu Beds by Mészáros and Moisescu (1991).

This formation consists of a pile of marls, in places sandy marls, with intercalations of more or less bioclastic marly-limestones, with a rich fossiliferous content, characterized by the abundance of molluscan moulds. The marly packet is comprised between the *Nummulites perforatus* lumachelle, in the base (the last term of the Căpuşu Formation), and the Văleni Limestone, at the upper part; in the area of study its thickness ranges from 2 m in the Şoiomu Hill to 12 m in the Călata Valley in the outcrop situated between Călăţele and Călata (Fig. 1, outcrop 3). The significant thickness variations, also stressed out in the sections presented in Figure 4, are due to the replacement from offshore to onshore of the marly pelitic facies with the calcareous arenitic one. Therefore, if the lower boundary of the Inucu Formation is practically isochrone, the upper one is clearly heterochrone, being older nearshore and younger offshore.

The fauna of the Inucu Formation, rich and diverse, consists of microforaminifers, nummulites, ostracods, molluscs, echinoids etc., forming real lumachelles. The molluscs are preserved either as moulds (those with aragonitic shell), e.g. *Corbula gallica* LMK., *Panopea corrugata* DIX., *Globularia patula* (LMK.), *Turitella imbricataria* LMK., *Rimella fissurella* (LINNÉE), *Terebellum sopitum* (SOL.), *Cassidaria nodosa* (SOL.) etc., or as shells (the calcitic ones) such as ostreines, pectinids, spondylids etc. Among ostreines, *Pycnodonte rarilamella* (MELLEV.) occurs more frequently in the basal part of the formation while *Crassostrea bersonensis* (MATHERON) is concentrated in a lumachellic level in the upper part of the formation. This lumachelle, stressed out as a biomarker under the name Crassostrea bersonensis Level (Rusu, 1987), is well represented both at the spa of Leghia (Fig. 4) and at Inucu in the Fântâna cu Aluni Valley (Mogyoros kut) – where the type section of the formation is supposed by us.

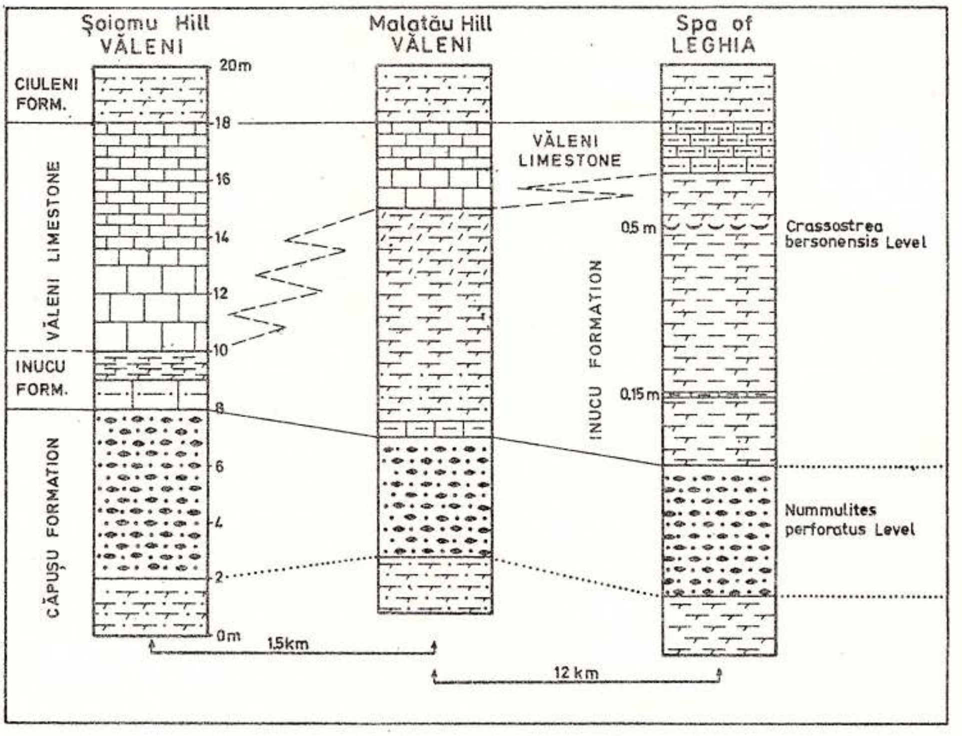

Fig. 4 – Correlation of the Inucu Formation and the Văleni Limestone (Bartonian) in the Văleni-Leghia area

In both places this level has 0.5 m in thickness and is situated at about 2 m below the Văleni Limestone.

The studies on calcareous nannoplankton evidenced the presence of the upper part of the NP 16 Zone (Gheţa, 1984; Mészáros, 1992), a doubtless biostratigraphic argument in favour of the Bartonian age of the Inucu Formation.

Văleni Limestone (RUSU, 1987)

Perforata-Schichten (partim – Horizont der gemischten Nummuliten oder des Molluskenkalkes und Mergels (KOCH, 1894, p. 228)

Orizontul argilelor cenuşii (partim) (RĂILEANU, SAULEA, 1956, p. 282; VLAICU-TĂTĂRÎM, 1963, p. 51)

Orizontul marnelor şi calcarelor cu moluşte (partim) (MÉSZÁROS, 1960, p. 95)

Argile marnoase de Mortănuşa (partim) (BOMBIŢĂ, 1963, p. 102)

Orizontul marnelor şi calcarelor cu Velates şi Corbula (partim) (DRAGOŞ, 1966, p. 361)

Orizontul marnelor şi calcarelor cu Corbula gallica LAMK. = Orizontul argilelor cenuşii (partim) (NIŢĂ PION et al., 1966, p. 308)

Calcaire "à Velates" (BOMBIŢĂ et al., 1975, p. 164)

Mortănuşa Marls (partim) – Velates Limestone (POPESCU et al., 1978, p. 311, 313)

Văleni Limestone (RUSU, 1987, p. 177)

Les Couches de Văleni (MÉSZÁROS, MOISESCU, 1991, p. 34)

This lithostratigraphic unit was separated as such for the first time by Bombiţă et al. (1975) in order to be characterized in detail litho- and biofacially by Popescu et al. (1978). In 1987 the term Văleni Limestone was introduced by us for the "Velates Limestone" "... which is well outcropped and reaches a maximum thickness at Văleni, a locality in the Călăţele "Gulf" (Gilău Area). Here, this calcareous member is 8 m thick and crops out on the ravine under the Şoimul Hill, proposed as type-section" (Rusu, 1987, p. 177). We considered the Văleni Limestone a member of the Mortănuşa Formation. Later researches convinced us that what we called "Mortănuşa Formation" includes in fact two formations: a carbonatic formation in the lower part – Inucu Formation, with the Văleni Limestone as a calcareous member – and a clayey-sandy formation in the upper part. Thus it was confirmed the significant boundary between the "Perforata Beds" and the "Lower Coarse Limestone Beds" (Koch, 1894), a lithostratigraphic unit which also includes the "Ostrea Clay Horizon".

Văleni Limestone, usually about 1–3 m thick on the Călata-Leghia-Luna de Sus lineament, has its maximum development in the type section below the Şoimul

Hill (or Şoiomu as known by the inhabitants) (Fig. 1, outcrop 2), designated as holostratotype and boundary stratotype (certainly for the lower boundary). The outcrop is situated 1 km south of the centre of the Văleni village, in the spring zone of the small valley which forms a slope of more than 30 m in the northwestern side of the Şoiomu Hill (871.78 m height). The section below the Malatău Hill, east of Văleni, has been proposed as parastratotype and at the same time as boundary stratotype for the next formation, namely Ciuleni Formation.

At the holostratotype (Fig. 3) the marls of the Inucu Formation are overlain by 3 m of massive limestones which, upwards, pass to 5 m fissile bedded limestones, separated as decimetric beds. The rock is a bioclastic calcarenite formed of tests of rotaliids, miliolids and other small-sized benthonitic foraminifers, of alveolinids and nummulites, valves and ostracods, fragments of bivalves, gastropods and echinoids, thalus of corallinaceans etc. Quartz grains occur rarely and lithoclasts only accidentally. The massive limestones are coarser and have a reduced biomicritic filling cement. The fossile limestones are finer and have a marly calcareous basal cement, within which the lutaceous terrigene material is more significant.

At 1.5 km ENE from the type sector, in the section below the Malatău Hill (Fig. 4), the calcareous pile is reduced to 3 m in thickness and the limestones are loaded with much more terrigene material, both lutaceous and arenitic, forming towards the upper part a sandy marly limestone. The limestones display numerous bioturbation traces.

The lower boundary of the Văleni Limestone can be sharp, as in the type locality, or it can be gradual, as in many sections; the upper boundary is always less clear due to a gradual and rapid transition from limestones through bioclastic marls to clays. It is to note that on the border of the basin, at Hodiş and southwest of Mărgău, the whole interval between the Nummulites perforatus Level and the clays of the Ciuleni Formation is constituted of fossiliferous calcareous sandstones (contaminated with microruditic material) with intercalations of sandy bioclastic limestones with *Alveolina*. Here, Inucu Formation is entirely replaced by an atypical facies of the Văleni Limestone, and consequently the deposits of this interval have been represented on the geological map sc. 1:50,000, Răchiţele sheet (Mantea et al., 1987) as a unitary formation, named "Văleni Limestone Horizon".

Within the exclusively marine fauna of the Văleni Limestone, the echinoids occupy a more important place than in the subjacent formation but they are represented by the same species, e.g. *Eupatagus multituberculatus* LMK., *Echinolampas calvimontanus* KLEIN, *Sismondia archiaci* COTT. etc. (see Popescu et al., 1978). Among molluscs, beside the common species (*Tibia goniophora* (BELL.), *Velates perversus* GMELIN, *Melongena subcarinata* (LMK.), *Spondylus podospideus* LMK. etc., also occur large forms, such as *Tibia ampla* (SOL.), *Campanile defrenatum* GREG., *Crassatella gigantica* KOCH and *Chama subgigas* d'ORB. Specimens of *Nautilus* and *Aturia*, characteristic of the warm seas, are found more rarely. The large foraminifera are represented by *Nummulites perforatus* and *N. striatus* as well as *Alveolina elongata* d'ORB. which proliferates.

Considering the sedimentologic character and the fossil content, the Văleni Limestone was formed in an inner shelf zone of a sea with euhaline marine waters, well oxigenated, with annual mean temperature of 21^0–25^0C, comparable to those existing now in the subtropical-warm zones (Rusu, 1994).

Confined by deposits with nannoplankton assemblages belonging to NP 16 and NP 17 zones (Gheţa, 1984; Mészáros, 1992) the Văleni Limestone is considered Bartonian in age.

CIULENI FORMATION (RUSU)

Untere Grobkalkschichten (partim – Ostreentegel-Horizont) (KOCH, 1894, p. 231, 239)

Orizontul argilelor cenuşii (partim) (RĂILEANU, SAULEA, 1956, p. 282; VLAICU-TĂTĂRÎM, 1963, p. 51)

Orizontul marnelor nisipoase cu Ostreide (MÉSZÁROS, 1960, p. 97)

Argile marnoase de Mortănuşa (partim) (BOMBIŢĂ, 1963, p. 102)

Orizontul marnelor cu Ostrei (DRAGOŞ, 1966, p. 363)

Orizontul marnelor şi calcarelor cu Corbula gallica LAMK. = Orizontul argilelor cenuşii (partim) (NIŢĂ PION et al., 1966, p. 308)

Marnes de Mortănuşa (BOMBIŢĂ et al., 1975, p. 165)

Mortănuşa Marls (partim) – Grey Clays and Marls (POPESCU et al., 1978, p. 311, 316)

Marnes de Mortănuşa (partim) (BOMBIŢĂ, 1984 b, p. 211, 212)

Mortănuşa Formation (Partim) (RUSU, 1987, p. 178)

Les Couches de Mortănuşa (MÉSZÁROS, MOISESCU, 1991, p. 35)

Mészáros (1960) is the first author who presented explicitly this formation as lithostratigraphic unit. Three years later Bombiţă (1963) introduced in the geological literature the term "Mortănuşa" (used by Saulea in a geological report for the year 1953) for the deposits situated between the Nummulites perforatus Level and the "Leghia Limestones" (name used in the same paper). Afterwards the term is used in two senses: either the larger, original one (mentioned above) or the restricted one denominating only the sandy clays overlying the Văleni Limestone. In our opinion, one has to preserve the original sense, but as a subgroup constituted of two formations: a lower one – the Inucu Formation with a calcareous member (Văleni Limestone) – and an upper one for which we propose the name Ciuleni Formation (a term which cannot be mistaken). The term "Mortănuşa" can still be useful in the most marginal zones in the Iara "Gulf" and nearby the Preluca crystalline, where the whole interval is invaded by coarse facies, either in the basinal zones (possibly in boreholes), where a continuous pelitic succession makes impossible the separation of the Ciuleni Formation from the Inucu Formation.

The name proposed by us comes from the village of Ciuleni (Cluj district), west of Călăţele, in whose neighbourhood is situated the best section of the clayey formation in the Gilău Area. Here, almost the whole succession is exposed in two very large outcrops north of Cetăţii Hill (east of Ciuleni) (Fig. 1, outcrop 5), located at about 300 m one from another. Out of the total thickness of the Ciuleni Formation, which in the study area is estimated at about 45 m, in the eastern outcrop, considered as holostratotype, are exposed more than 35 m deposits (Fig. 5). They are constituted of clays and marly-clays, gray or greenish-gray, more and more siltic and then more sandy up to clayey sandy towards the upper part, with intercalations of marly limestones with molluscan moulds, located in the base and with some intercalations of marls rich in oyster remains. The base of the formation can be observed on a ravine 150 m eastwards where there is a transition from the bioclastic limestones of the Văleni Limestone to marly clays with rare large-sized nummulites of *N. perforatus* type. There follows a hiatus of observation (5 m) corresponding to pelitic deposits. As boundary stratotype of the Ciuleni Formation we already mentioned the section below the Malatău Hill (Fig. 4) from Văleni.

At the upper part, Ciuleni Formation passes gradually to a suprajacent calcareous formation through a packet of sandy clays with sands and sandstones intercalations about 2–3 m thick in the studied area. At the holostratotype (Fig. 5) the upper boundary of the formation was arbitrarily situated in the base of a bioclastic calcarenite bank rich in oyster remains, with which the next formation begins.

Ciuleni Formation contains a diverse microfauna, quite abundant in the lower part and poorer towards the top, represented by benthonic microforaminifera belonging to the local Pararotalia subnimermis Zone, macroforaminifers (represented by nummulites) and ostracods, grouped into assemblages belonging to two local zones: Leguminocythereis striatopunctata Zone and Quatrocythere leghiensis Zone (see Popescu et al., 1978).

The molluscan fauna, generally poor, is dominated by ostreids and pectinids. In the upper third of the formation occur even lumachellic accumulations of oysters, forming a marker level with a regional significance – Crassostrea orientalis Level – widely spread in the Gilău Area (Rusu, 1987). Some microfaunal assemblages and the oyster lumachelles indicate brackish levels (polyhaline salinity) due to the supply of fresh waters in certain moments, in a common normal marine medium (euhaline salinity). Based on the paleontologic assemblage of the Ciuleni Formation, as well as the calcareous nannoplankton belonging to NP 17 and NP 18 Zones (Gheţa, 1984; Mészáros et al., 1987; Mészáros, 1992) this formation is considered of Bartonian age, excepting the terminal part which, considering the nummulites, would be Prianonian in age. Bombiţă (1984 b) stressed out a significant modification of the macroforaminiferal assemblage in the last sandy levels of the Motănuşa "Marls" where, beside *Nummulites perforatus* and *N. striatus*, occur the first representatives of the Priabonian: *N. garnieri* DE LA HARPE and primitive forms with affinities of *N. chavannesi* DE LA HARPE.

VIŞTEA LIMESTONE (RĂILEANU, SAULEA, 1956)

Untere Grobkalkschichten (partim-Horizont der Grobkalkbänke) (KOCH, 1894, p. 231, 241)
Orizontul calcarului grosier inferior sau al gresiei de Racoţi (RĂILEANU, SAULEA, 1956, p. 283)
Calcarul de Viştea (RĂILEANU, SAULEA, 1956, p. 284)
Orizontul calcarelor grosiere inferioare (MÉSZÁROS, 1960, p. 98; DRAGOŞ, 1966, p. 365)
Orizontul calcarului grosier inferior (VLAICU-TĂTĂRÎM, 1963, p. 58)
Calcarul de Leghia (BOMBIŢĂ, 1963, p. 102)
Orizontul calcarului grosier inferior = orizontul gresiei de Racoţi (NIŢĂ PION et al., 1966, p. 310)
Leghia Limestone (POPESCU et al., 1978, p. 320; RUSU, 1987, p. 178)
Calcaire de Leghia (BOMBIŢĂ, 1984 b, p. 211)
Les Couches de Leghia (MÉSZÁROS, MOISESCU, 1991, p. 35)

As regards the nomenclature of this formation, we consider that we have to come back to the first valid name "Viştea Limestone" proposed by Răileanu and Saulea (1956, p. 284), a name abandoned in favour of a more recent synonym "Leghia Limestone" introduced

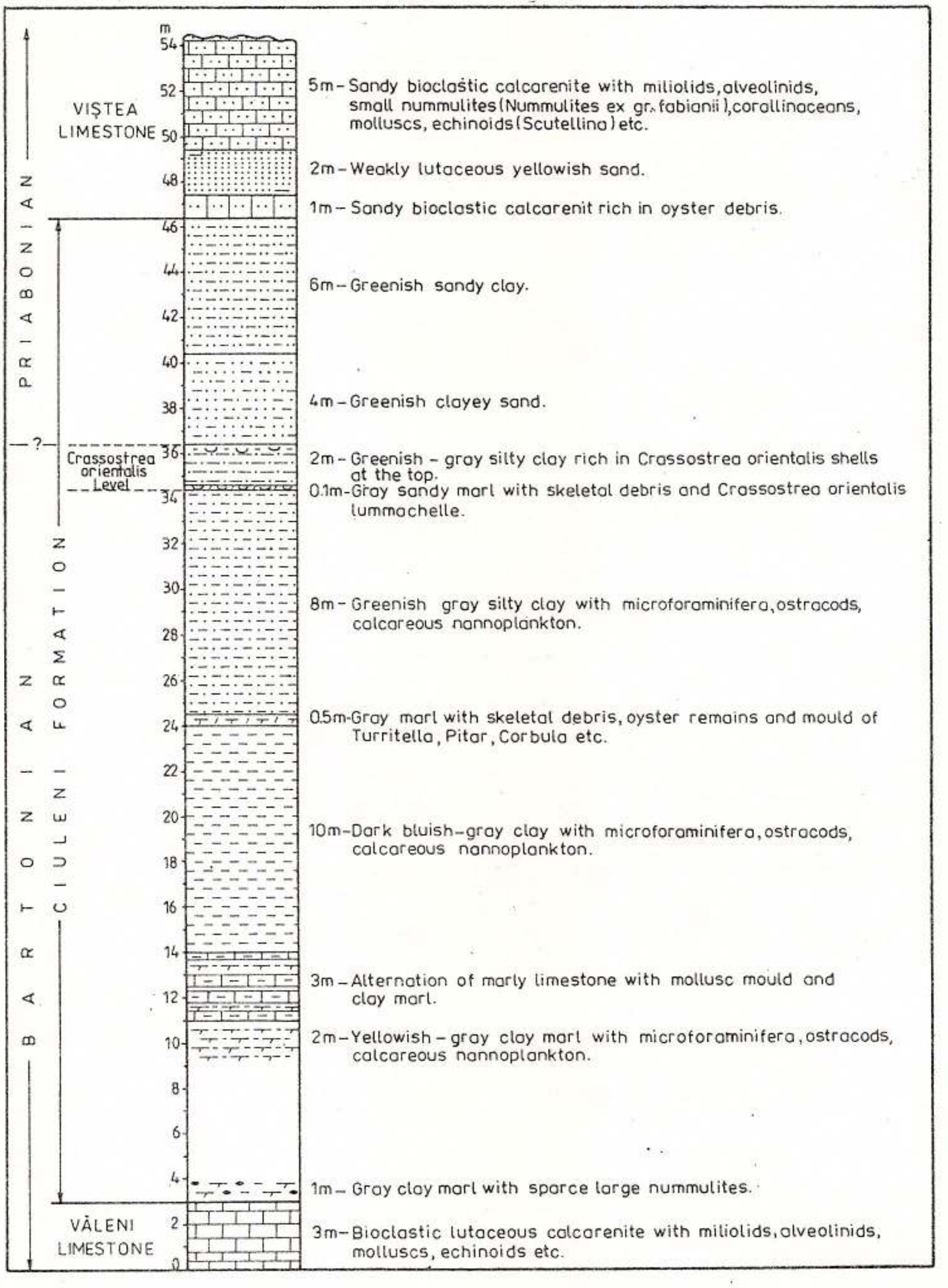

Fig. 5 – Stratigraphic succession of the Eocene deposits cropping out north of Dealul Cetăţii, at Ciuleni

by Bombiţă in 1963. As the limestone at Viştea is the best known building stone in this part of Transylvania since earlier times (see Koch, 1894, p. 234) and the Viştea outcrops (natural or artificial) are the most favourable for a type section than those of Leghia there is no objective reason to use the second name. Therefore, we propose as holostratotype of the formation the Golombodu quarry west of Viştea, where the limestones are exposed on 7 m out of the total thickness (8 m) and the boundary with the red clays of the overlying Valea Nadăşului Formation is quite visible. As parastratotypes can be used the quarry in the Pădurea Nemeşilor (NW of Gilău), where both limits are visible (boundary stratotype) and the section at the Spa of Leghia, below the Şesuri Hill.

In the study area, Viştea Limestone consists of a 6–9 m pile of sandy bioclastic limestones which, in places, includes in the base intercalations (up to 2 m) of weakly clayey sands, as one can see in the section at Ciuleni (Fig. 5). According to Popescu et al. (1978)

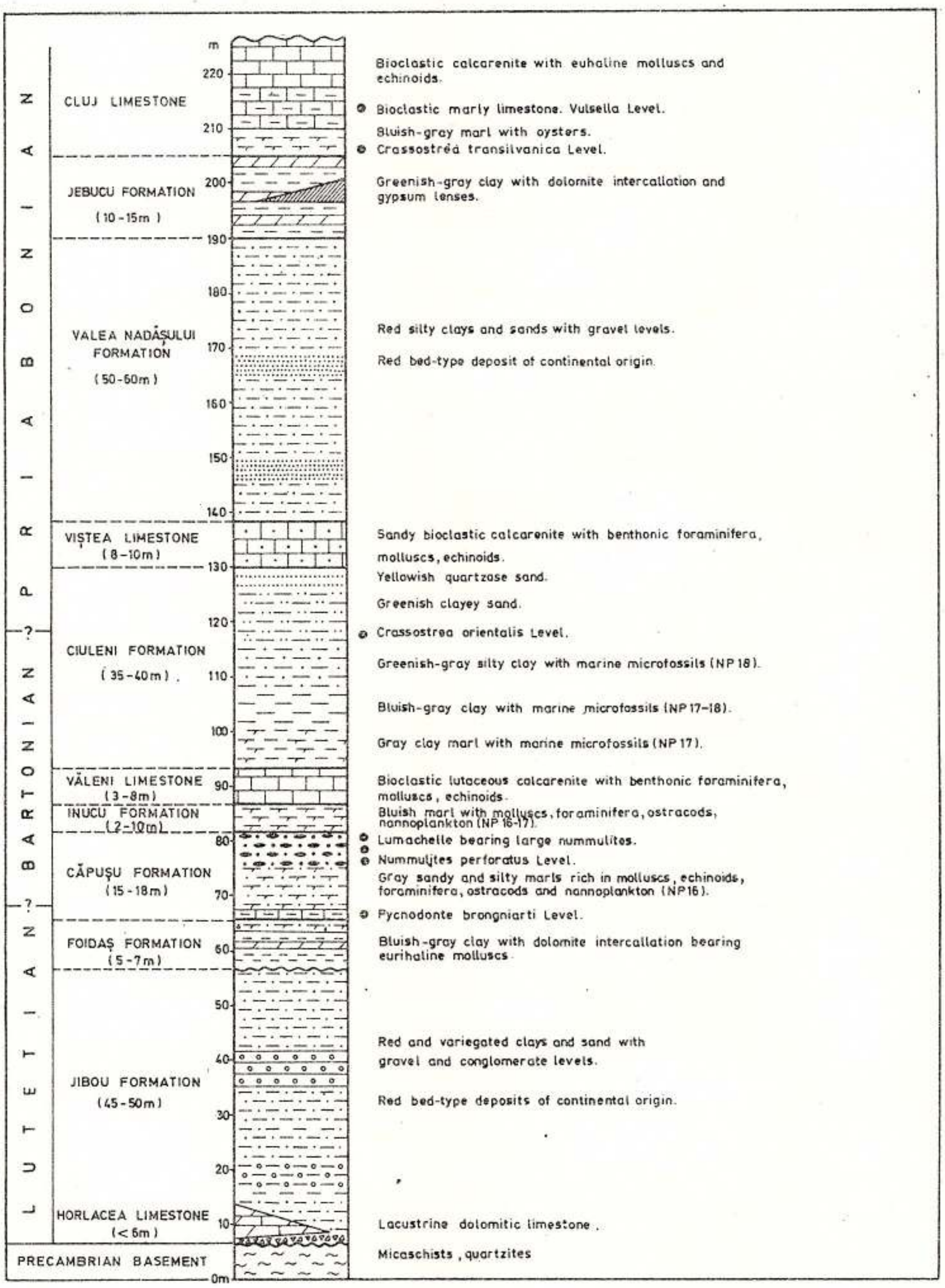

Fig. 6 - Generalized Eocene succession in the Horlacea-Văleni area

the limestones represent a skeletal packstone, as 0.5–1 m banks, with parallel lamination and they show numerous trace fossils. In certain zones the filled vertical shafts representing the tunnels of the Callianassa (decapod), considered a marker of the nearness to the shoreline, are quite frequently found. The bioclasts consist of tests of microforaminifers (especially miliolids and rotaliids), alveolines and small nummulites, thalus of corallinaceans and fragments of echinoids, molluscs, balanids etc.

The upper boundary of the Viştea Limestone, formation which ends the first marine cycle of the Paleogene in NW Transylvania, is clearly sharp, over the last calcareous bank or in places over a bioclastic marl with marine microfauna resting directly on red clays or alluvial sands of the Valea Nadăşului Formation.

Viştea Limestone contains an euhaline marine fauna characteristic of the inner shelf zones with sandy substratum. This formation is characterized by *Scutellina* (after Popescu et al., 1978), represented by three species: *Scutellina rotunda* FORB., *S. lenticularis* (LMK.) and *S. transylvanica* BARBU, DRAGOŞ) and for this reason Hauer and Stache (1863, p. 132) named it "Scutellinen Kalk". The assemblage with *Alveolina elongata* d'ORB., *Nummulites fabianii* (PERV.) (Bombiţă, 1984 b) forms here a concurrent-range zone. Among the mentioned fossils *N. fabianii* represents an undoubtful argument in favour of the Priabonian age for the Viştea Limestone.

The other lithostratigraphic units represented on the map (Fig. 1): Valea Nadăşului Formation (continental red clays formation), Jebucu Formation (marine evaporitic formation) and Cluj Limestone (marine calcareous formation) which also belong to the Priabonian (Fig. 6) do not involve discussion.

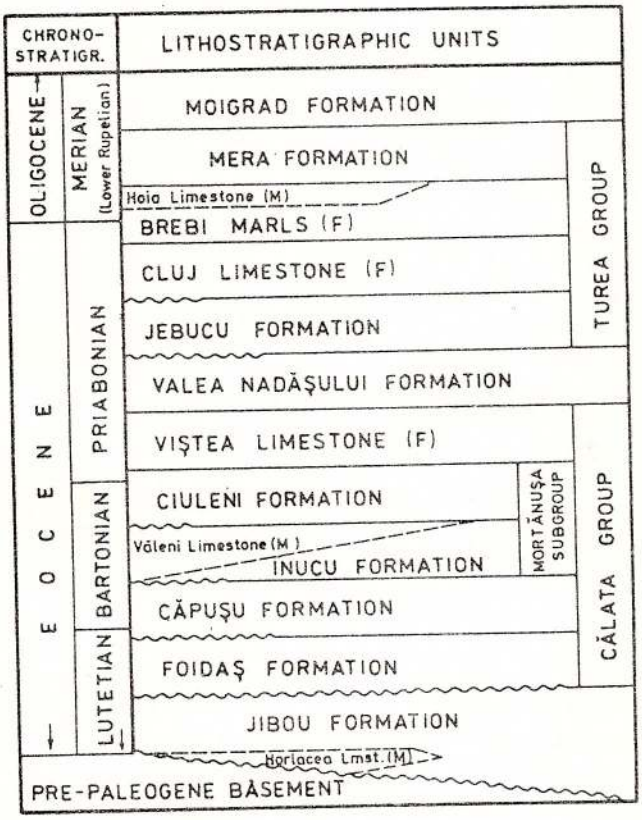

Fig. 7 - Eocene lithostratigraphic units in the Gilău Area (NW Transylvania). F, formation; M, member.

Prefigurating the actual lithostratigraphic classification Hofmann (1879) included the formations of the first marine cycle of the Transylvanian Paleogene in what he named "Rakoczy Gruppe", a unit for which the term "Lower Marine Series" was also used (Răileanu, Saulea, 1956). Taking into account the International Stratigraphic Guide (1979, p. 54) according to which "the same name should not be used for an entire unit and for a part of this unit" we propose the name Călata Group for Hofmann's group. The denomination Rakoczy is attributed only to the Rakoczy Sandstone (a sandy formation equivalent to the Viştea Limestone) of the same author, with the holostratotype in the Piscuiul Ronei ("Rakoczy Hill" in former times) between Jibou and Turbuţa.

We shall act similarly in case of the "Cluj Group", a name used by Popescu et al. (1978) for the "Upper Marine Series" acc. to Răileanu and Saulea (1956). For this unit we propose the name Turea Group (locality in the Cluj district) around which the formations of this group are typically represented and well exposed - see Popescu et al., 1978; Popescu, 1984), including the whole succession of deposits of the marine cycle, namely: Jebucu Formation, Cluj Limestone, Brebi Marls (with Hoia Limestone as calcareous member) and Mera Formation.

In our opinion, the lithostratigraphic scheme of the Eocene in the Gilău area would be that presented in Figure 7, in which all the lithostratigraphic units are formal and their rank is specified.

References

Báldi-Beke, M. (1984) The nannoplankton of the Transdanubian Paleogene formations. *Geol. Hung., Paleont.*, 43, 307 p., Budapest.

Bombiţă, G. (1963) Poziţia Ilerdianului şi a Biarritzianului în România. Observaţii şi propuneri cu privire la noua împărţire a Eocenului. *Asoc. Geol. Carp.-Balc., Congr. V, Com. St.*, III/1, p. 97–105, Bucureşti.

— **(1984 a)** Sur quelques nummulites évoluées de Roumanie. *D. S. Inst. Geol. Geofiz.*, LXVIII/3, p. 23–37, Bucureşti.

— **(1984 b)** Le Napocien, vingt ans après sa définition. *Rev. Paléobiol.*, 3, 2, p. 209–217, Genève.

— **, Gheţa, N., Iva, M., Olteanu, R. (1975)** Eocène moyen-supérieur et Oligocène inférieur des environs de Cluj. In *"Guide micropal. Mésoz. Tert. des Carpates Roum."*, *Inst. Géol. Géophys.*, p. 163–174, Bucureşti.

Dragoş, V. (1966) Paleogenul din regiunea Călăţele-Huedin. *An. Com. Stat Geol.*, XXXV, p. 347–395, Bucureşti.

Gheţa, N. (1984) The Eocene of NW Transylvania. A New Biochronostratigraphic Zonation Based on Calcareous Nannoplankton. *D. S. Inst. Geol. Geofiz.*, LXIX/3, p. 95–106, Bucureşti.

Hauer, Fr., Stache, G. (1863) Geologie Siebenbürgens. 636 p., Ed. W. Braumüller, Wien.

Hofmann, K. (1879) Bericht über die im östlichen Theile des Szilágyer Comitates während der Sommercampagne 1878 vollführten geologischen Specialaufnahmen. *Földt. Közl.*, IX, 5–6, p. 231–283, Budapest.

Joja, T. (1956) Observaţii de ordin stratigrafic în regiunea din jurul oraşului Jibou. *An. Com. Geol.*, XXIX, p. 309–323, Bucureşti.

Koch, A. (1894) Die Tertiärbildungen des Beckens der Siebenbürgischen Landestheile. I. Paläogene Abtheilung. *Mitt. a. d. Jahrb. d. kgl. ung. geol. Anst.*, X, p. 177–399, Budapest.

— **, Hofmann, K. (1887)** Harta geologică, scara 1:75.000. Umgebungen von Bánffy-Huniad, Budapest.

Mantea, G., Ştefan, A., Rusu, A., Dimitrescu, R. (1987) Harta geologică a României, scara 1:50.000, Foaia Răchiţele. Inst. Geol. Geofiz., Bucureşti.

Martini, E. (1971) Standard Tertiary and Quaternary calcareous nannoplankton zonation. *Proc. II Plank. Conf. Roma, 1970*, p. 739–785, Roma.

Mateescu, Şt. (1926) Observaţii morfologice şi geologice asupra depresiunii Huedinului. *An. Inst. Geol. Rom.*, XI, p. 349–388, Bucureşti.

Maxim, I. Al. (1959) Câteva consideraţiuni asupra valorii actuale a lucrării lui Anton Koch: "Formaţiunile terţiare ale Bazinului Transilvaniei. I. Paleogenul, 1894; II. Neogenul, 1900." *Studia Univ. "Babeş-Bolyai", Geol., Geogr.*, II, 1, p. 7–22, Cluj.

Mészáros, N. (1960) Studiul litofacial şi paleogeografic al depozitelor marine eocene medii de la vest şi sud-vest de Cluj. *Stud. Univ. "Babeş-Bolyai"*, II, 1, p. 87–104, Cluj.

— **(1992)** Nannofossil zones in the Paleogene and Miocene deposits of the Transylvanian Basin. *Knihovnička ZPN*, 14 b, 2, p. 87–92, Hodonin.

— **, Moisescu, V. (1991)** Bref aperçu des unités lithostratigraphiques du Paléogène dans le Nord-Ouest de la Transylvanie (région de Cluj-Huedin), Roumanie. *Bull. Inf. Géol. Bass. Paris*, 28, 2, p. 31–39, Paris.

— **, Moga, V., Ianoliu, C. (1987)** Studying the various groups of fossil organisms of Leghia - Leghia Băi. In *"The Eocene from the Transylvanian Basin"*, p. 143–150, Cluj-Napoca.

Niţă Pion, P., Vasilescu, G., Vasilescu, M. (1966) Contribuţii la stratigrafia depozitelor eocene din regiunea Huedin-Călăţele-Păniceni (regiunea Cluj). *D. S. Inst. Geol.*, LII/1, p. 301–316, Bucureşti.

Odin, G. S. (1978) Isotopic Dates for a Paleogene Time Scale. *AAPG, Studies in Geology*, p. 247–257, New York.

Pávay, A. (1873) Die geologischen Verhältnisse der Umgebung von Klausenburg. *Mitt. a. d. Jahrb. d. kgl. ung. geol. Anst.*, I/3, p. 351–441, Pest.

Popescu, B. (1978) On the lithostratigraphic nomenclature of the NW Transylvania Eocene. *Rev. roum. géol., géophys., géogr., (Géol.)*, 22, p. 99–107, Bucureşti.

— **(1984)** Lithostratigraphy of cyclic continental to marine Eocene deposits in the NW Transylvania, Roumania. Arch. Sc. Genève 37, 1, p. 37–73, Geneva.

— **, Bombiţă, G., Rusu, A., Iva, M., Gheţa, N. Olteanu, R., Popescu, D., Tăutu, E. (1978)** The Eocene of the Cluj-Huedin Area. *D. S. Inst. Geol. Geofiz.*, LXIV/4, p. 295–357, Bucureşti.

Răileanu, Gr., Mészáros, N. (1966) Studiul geologic al complexului vărgat inferior din NW Transilvaniei. *St. cerc. geol., geofiz., geogr., (Geol.)*, 11, 2, p. 451–462, Bucureşti.

— **, Saulea, E. (1956)** Paleogenul din regiunea Cluj şi Jibou (NW bazinului Transilvaniei). *An. Com. Geol.*, XXIX, p. 271–308, Bucureşti.

Rusu, A. (1970) Corelarea faciesurilor Oligocenului din regiunea Treznea-Bizuşa (nord-vestul bazinului Transilvaniei). *St. cerc. geol., geofiz., geogr., (Geol.)*, 15, 2, p. 513–525, Bucureşti.

— **(1987)** Ostreina Biohorizons in the Eocene of the North-West Transylvania (Romania). In *"The Eocene from the Transylvanian Basin"*, p. 175–182, Cluj-Napoca.

— **(1994)** Paleoclimatic Meaning of Paleogene Mollusca in NW Transylvania (Romania). *Rom. J. Paleont.*, 76, p. 47–52, Bucureşti

Szádeczky-Kardoss, E. (1930) Contribuţii la geologia Ardealului de NW. *D. S. Inst. Geol. Rom.*, XIV, p. 39–58, Bucureşti.

Vlaicu-Tătărîm, N. (1963) Stratigrafia Eocenului din regiunea de la sud-vest de Cluj. Ed. Acad. R. P. R., 204 p., Bucureşti.

— *** * * (1979)** Guide stratigraphique international, classification, terminologie et règles de procédures. Doin Editeurs, 233 p., Paris.

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Rom. J. Tect. & Reg. Geol., 1995, 76, p. 73–82, Bucureşti

TECTONICS OF THE SĂVÂRŞIN GRANITE (DROCEA MOUNTAINS)

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Key words: Petrography. Structural elements. Magma flow direction. Granitic body form. Granite tectonics. Drocea Mountains. South Apuseni Mountains.

Abstract: The Săvârşin granitic body represents the southern intrusion of the composite pluton - the Late Kimmerian granitoid massif of Săvârşin. This massif is situated in the Mureş Valley, in the south of the Drocea Mts, the Mureş zone, respectively. In order to determine the tectonics of the Săvârşin Granite measurements have been effectuated as for the position (strike, dip, pitch) of the orthoclase-albite megacrystals with a zonal structure, of the melanocrate and leucocrate autoliths, joints (Q, L, S), aplites and microgranitic porphyry dykes, mineralized fissures, as well as two fracture systems which cross the granite. The plotting of the joints on the structural diagrams and the other structural elements measured showed that the magma intruded from NE to SW the Săvârşin granitic body, an asymmetrical laccolith in shape. The paper has also demonstrated that the Săvârşin granitoid massiv is formed of two intrusions: a northern one, which penetrated from SW to NE, trending N46^{0}E, and a southern one, of the Săvârşin Granite, which penetrated from NE to SW, trending S75^{0}W. The magma of the two intrusions started from the same place in the depth as shown by the gravity anomaly in the Săvârşin area.

Introduction

The Săvârşin Granite belongs to the Late Kimmerian granitoid massif with the same name, situated in the southern part of the Drocea Mts, in the Mureş Valley, nearby the southern margin of the Mureş zone. This massif was first studied in petrographic and petrochemical respect by Szentpétery (1928). During the Second World War the researches began for sulphide mineralizations, including molybdenite, a mineral rendered evident in the northern part of the massif on this occasion (Socolescu, 1944; Papiu, 1945). In 1953 Savu (unpublished report) showed that the Săvârşin granitoid massif is not constituted of a single intrusive body differentiated *in situ*, as previously presumed (see Giuşcă, 1950), but it represents a composite pluton, formed of two successive intrusions: the northern intrusion consisting of the Temeşeşti granodiorite and diorites, and the southern one, formed of the Săvârşin Granite (described in this paper), a name given by Savu and Vasiliu (1966). The petrology and geochemistry of the granitoid massif, the Săvârşin Granite inclusive, were studied by Savu et al. (1967 a, 1967 b) who also elaborated a geological sketch-map of the massif with the two intrusions. Until recently the granitoid massif was assigned to the Laramian magmatites (banatites). In 1986 Savu et al. established that the radiometric dating situates the massif on the isochrone of 128 Ma and, therefore, it belongs to the Late Kimmerian intrusions in this area (Savu et al., unpubl. rep., 1984), being linked to the southern island arc of the Mureş Zone. Ştefan (1986) considered this massif of Eocretaceous age.

The structural elements concerning the northern granodiorite-diorite intrusion of Temeşeşti have been presented in a previous paper (Savu, Mândroiu, 1980).

In the present paper the structural elements of the Săvârşin Granite, the position of this intrusion within the Late Kimmerian granitoid massif and its tectonics were described.

Constitution of the Granitic Body and Petrographic Aspects

The Săvârşin Granite lies in the Mureş Valley, along the national highway no 7 (NH 7), between the localities of Cuiaş, Săvârşin and Hălăliş on about 7.2 km.

It extends up to the Contrava Valley to the north and up to the locality of Valea Mare to the south on about 4.2 km. It has been concluded that the granitic intrusion represents an intrusive body elongated toward N75^{0}E (Pl. I). This granitic body was mostly eroded by the Mureş River and its main tributaries – the Troaş and Vineşti valleys – and covered by alluvial deposits. In the Mureş flood plain, south of Săvârşin, in the area lying between the two main tributaries of the river, there is a remnant of the granitic body, represented by the Cetăţuia Hill or the "sugar loaf hill" (Pl. II, Fig. 1).

The rocks constituting the Săvârşin granitic body display a massive texture which indicates isotropic – postorogenic – granites. Within the structure of the body three characteristic facies can be distinguished: the marginal facies, the common granite facies, and the porphyritic granite facies (Pl. I).

The marginal facies is developed on a larger area in the north-easternmost part of the body, and extends as a narrow strip on a certain distance along its northern contact, being thus separated from the twin intrusion of Temeşeşti (see the map annexed to the paper elaborated by Savu and Mândroiu (1980). This facies is constituted of a fine-grained porphyritic granite which displays potash feldspar and albite-oligoclase phenocrysts included in a groundmass formed of quartz, potash feldspar, albite-oligoclase, biotite and rarely green hornblende.

The common granite facies forms a strip with a maximum width of 400 m (Pl. I). This facies lies between the marginal and the porphyritic facies, in the north-eastern area of the granitic body; at Hălăliş it is exposed on the western margin of the intrusion. It is constituted of a common grained (uniform) granite formed of xenomorphic quartz, potash feldspar, albite-oligoclase, biotite, green hornblende and accessory minerals. South-west of Cuiaş at the eastern contact of the body the common granite includes wedges and xenoliths of Liassic ophiolites (dolerites), poorly affected by the contact metamorphism.

The porphyritic granite facies (Pl. II, Figs. 2 and 3) occupies almost the whole Săvârşin granite body (Pl. I). It is by this granite that the granitic body differs from other acid intrusions (Savu, Vasiliu, 1966) and its denomination is based on this reason. The porphyritic granite consists of a medium-grained granitic ground-mass formed of quartz, potash feldspar, oligoclase, biotite, common green hornblende, and accessory minerals represented, as in the previous facies, by apatite, titanite, zircon and opaque minerals. The plagioclase from the groundmass displays locally a zonal structure. The granitic groundmass includes huge phenocrysts (megacrystals) of orthoclase-albite feldspar with a zonal structure and an angle -2V of 50–60^0 (Pl. II, Figs. 2 and 3). The zones which succeed alternatively consist of orthoclase and albite, as shown in Figure 1, established by means of a French electronic microprobe of CAMECA type. Savu and Vasiliu (1966) showed that the megacrystals of orthoclase-albite feldspar with a zonal structure crystallized rhythmically from the granitic magma at 660^{0}C. The zonal structure of the megacrystals belongs to the Rapakivi structure, whose formation was explained in different ways. Yoder et al. (1957) established that the zonal megacrystals from the Scandinavian shield granites were formed at a temperature of 720^{0}C and a pressure of 5,000 bars from a melt with an equal content of Or and Ab and a low content of An.

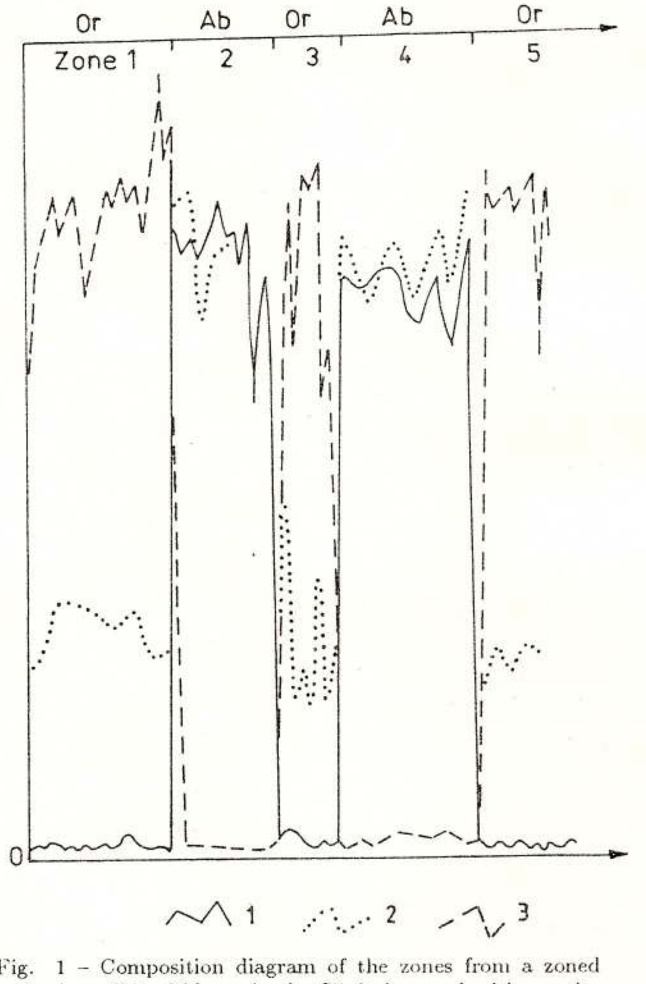

Fig. 1 – Composition diagram of the zones from a zoned orthoclase-albite feldspar in the Săvârşin porphyritic granite (Hălăliş). 1, Ca variation; 2, Na variation; 3, K variation.

The granitic body contains numerous melanocrate and leucocrate autoliths (Balk, 1937; Didier, 1964) and it is penetrated by acid rocks and epidosite veinlets. Due to the weathering of the granitic rocks change into a typical white-pinkish grit. In some places, for instance west of Vineşti, the alteration of the porphyritic granite has been radially (Fig. 2), similar to what the French geologists called "alteration en boule", which is active on the main joints of the granitic intrusion.

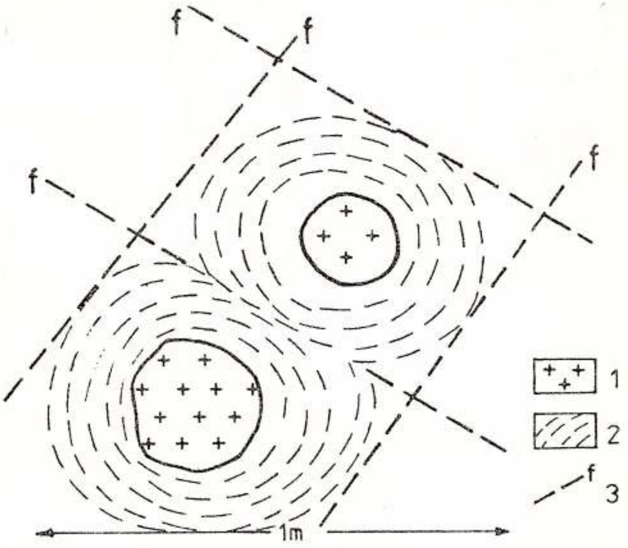

Fig. 2 - Radial alteration process ("alteration en boule") of the Săvârşin porphyritic granite. 1, granite; 2, grit; 3, joints.

Structural Elements

Considering the remarks of Cloos (1923), Balk (1937), Martin (1953), Möbus (1959) etc. concerning the granite tectonics, several structural elements have been measured by us in order to establish the Săvârşin Granite tectonics: the phenocrysts of albite-orthoclase feldspar, autoliths (Fig. 3), joints and mineralized fissures. The strike and dip of the aplitic and porphyry dykes and the fractures which affect the granitic body have been measured, too.

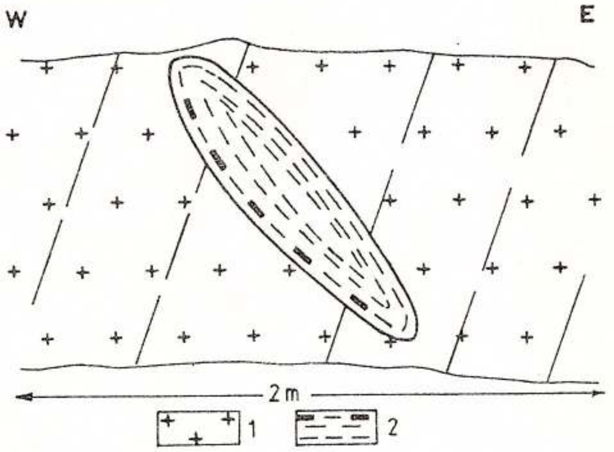

Fig. 3 - Melanocrate autolith (1) with megacrystals of albitic-orthoclase feldspar (2) in the Săvârşin granite crossed by Q joints, approximately perpendicular on the autolith elongation, Vineşti on the NH7 highway.

Albite-orthoclase feldspar phenocrysts. The pitch angle of the more widely developed megacrystals from the porphyritic granite facies in the west of the intrusive body has been measured. The results of the measurements are plotted on the diagram in Figure 4a (lower hemisphere). This diagram shows that in the Hălăliş-Vineşti area the megacrystals are trending N38^{0}E and are pitching 34^{0}NE, a position which points to an ascending direction of the magma flow in this area of the granitic body.

Melanocrate and leucocrate autoliths. Although frequently found in the granitic body, the autoliths are more numerous toward the lower part of the intrusion, in the porphyritic granite facies and more rare in the marginal facies. The melanocrate autoliths are made up mostly of biotite and green hornblende (Pl. II, Fig. 4), whereas the leucocrate autoliths contain crystals of plagioclase and potash feldspar. The potash feldspar occurs as megacrystals even in the melanocrate autoliths (Fig. 3). These autoliths are either fusiform or disc-shaped and are oriented in the flow plane of the magmas. The direction and the pitch of the autoliths, like those of the megacrystals, are indicated on the annexed structural map (Pl. I) which shows that they are more abundant in the Hălăliş-Vineşti area, where granite exposures are good for researches, therefore toward the bottom of the intrusion.

Like the albite-orthoclase feldspar phenocrysts, the autoliths are trending ENE-WSW, in the direction of the granitic magma flow, and they have pitch angles oriented toward ENE of 15–35^0, rarely above these values. (Pl. I).

Joints. After the cooling of the granitic body, numerous joints occurred which belong to three major systems marked with the letters Q, L and S, according to Cloos (1923) as shown in Figures 4, 5 and 6. Diagonal joints are more rarely found. Q joints (cross joints), which occurred first in the granitic body (Balk, 1937), are perpendicular to the elongation trend of the albite-orthoclase megacrystals and of the melanocrate and leucocrate autoliths, therefore perpendicular to the direction of the magma flow or the magma intrusion trend. L joints (longitudinal joints) are parallel to the direction of the magma flow and perpendicular to its plane. S joints (flat-lying joints) are parallel to the magma flow plane and, therefore, perpendicular to the plane of the Q and L joints.

Maxima for each of the three joint systems resulted on the diagrams in Figures 4, 5 and 6 built up in the lower hemisphere. For the most characteristic diagrams the direction of the intruding magma flow has also been determined, which is rendered by a line linking the center of the diagram with the crossing point of the L and S joints, which is theoretically situated

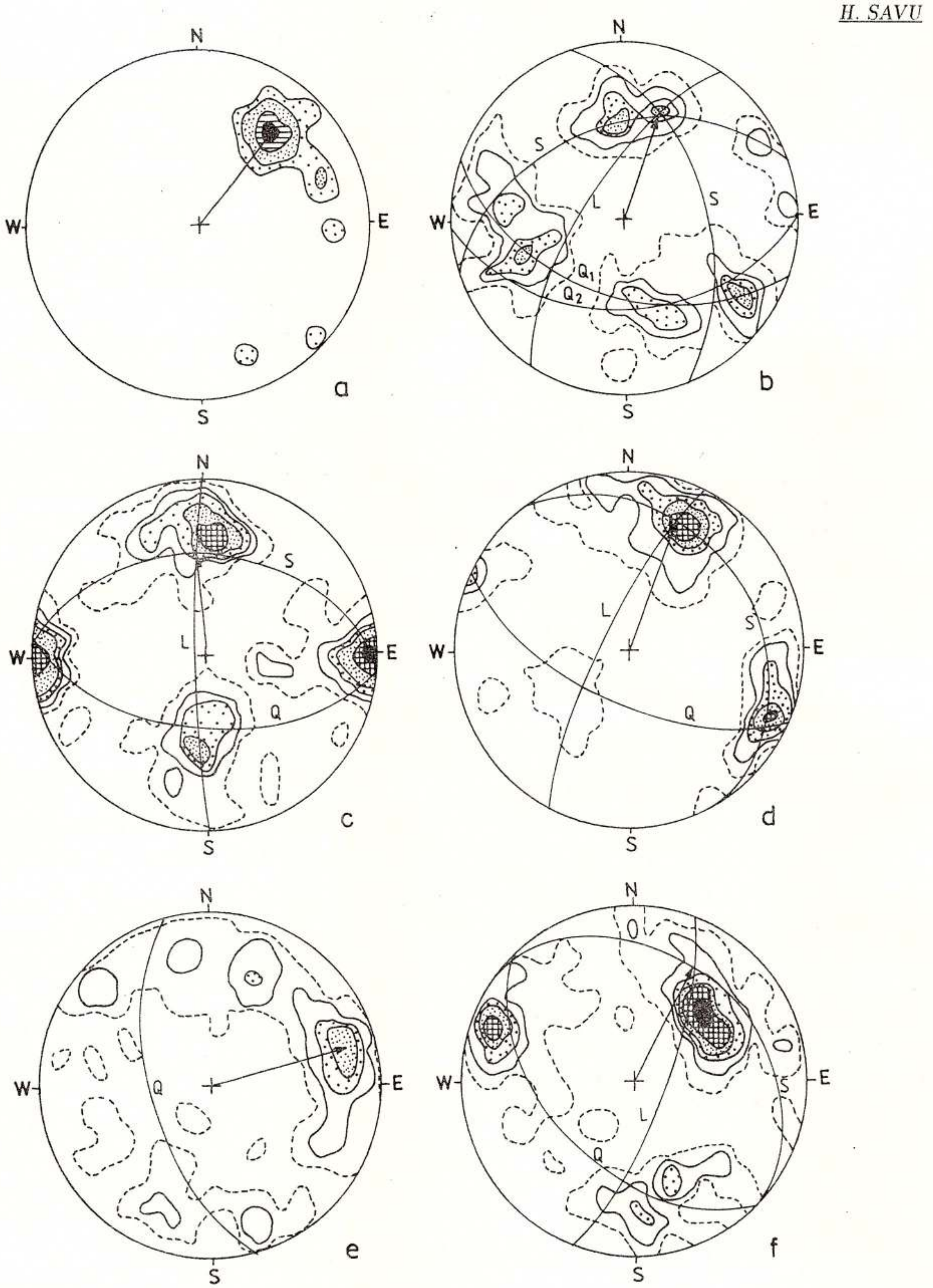

Fig. 4 – Structural diagrams. a, pitch angle of the orthoclase-albite feldspar megacrystals between Hălăliş and Vineşti (40 megacystals), isolines: 1, 2, 5, 7 %; b, diagram of the joints on the Contrava Valley (146 joints), isolines: 1, 3, 5, 7, 9 %; c, diagram of the joints on the Mutu Brook, tributary of the Contrava Valley (150 joints), isolines: 1, 3, 5, 7, 16, 24 %; d, diagram of the joints on the Nucu Brook (226 joints), isolines: 1, 3, 5, 7, 9 %; e, diagram of the joints in the Troaş Valley-Laliţa Brook area (243 joints), isolines: 1, 3, 5, 7 %; f, diagram of the joints between the Banieşu and the Troaş valleys (146 joints), isolines: 1, 3, 5, 7, 9, 14 %.

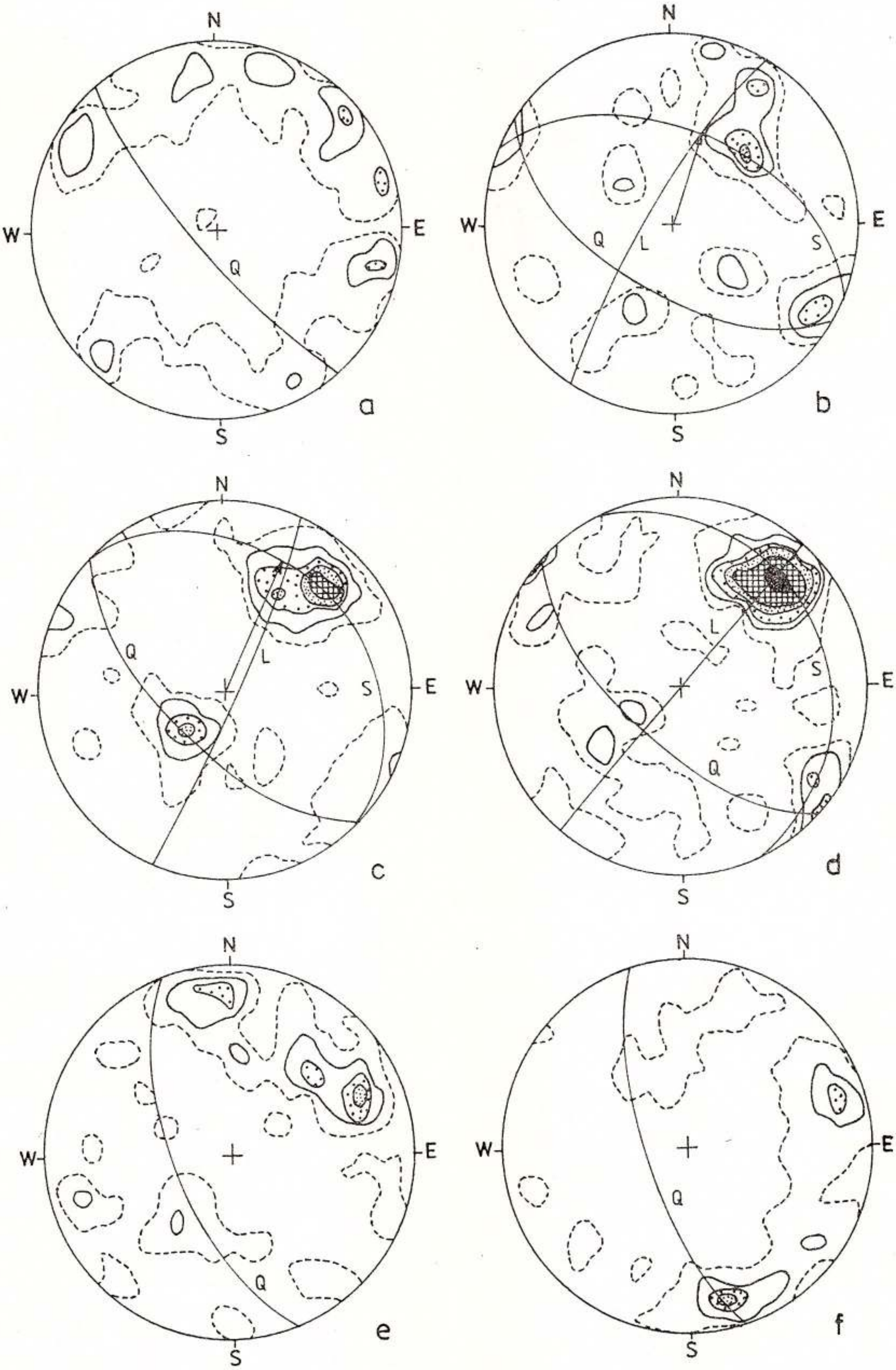

Fig. 5 – Structural diagrams. a, diagram of the joints in the Cetăţuia Hill (225 joints), isolines: 1, 3, 5 %; b, diagram of the joints in the Porlata Brook (Vineşti) (80 joints), isolines: 1, 3, 5, 7, 9 %; c, diagram of the joints between Vineşti and Hălăliş (200 joints), isolines: 1, 3, 5, 7, 9 %; d, diagram of the joints in the Hălăliş Valley (130 joints), isolines: 1, 3, 5, 7, 9, 14 %; e, diagram of the joints on the NH7, between Vineşti and Hălăliş (100 joints), isolines: 1, 3, 5, 7 %; f, diagram of the joints in the Valea Mare quarry (175 joints), isolines: 1, 3, 5, 7 %.

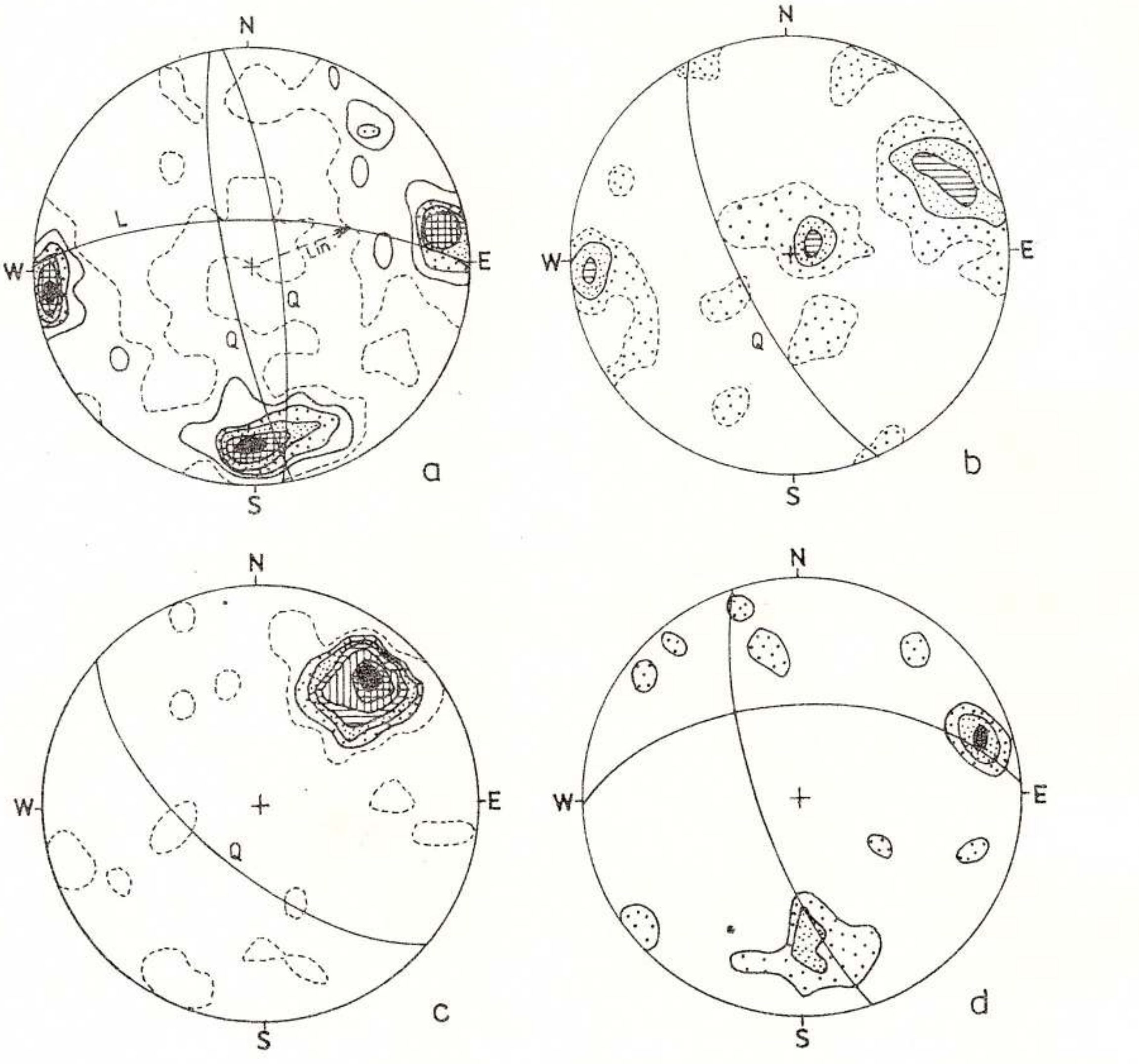

Fig. 6 – Structural diagrams. a, diagram of the joints at Valea Mare (150 joints), isolines: 1, 2, 3, 5, 7, 9, 16 %; b, diagram of the aplite dykes (80 dykes), isolines: 1, 3, 5, 7 %; c, diagram of the joints with kaolinizations and pyrite depositions (85 joints), isolines: 1, 3, 5, 7, 9, 15, 19 %; d, diagram of the fractures (50 fractures), isolines: 1, 3, 5 %.

in the maximum formed by the poles of the Q joints (see Fig. 4c) and the plotting maximum of the linear elements (megacrystals and autoliths) determined by their pitch angles. However, certain small discrepancies occur when the crossing of the two joint systems does not fall exactly in the mentioned maxima (Fig. 5b and 5c). These discrepancies probably depended on the last movements of the magma, which has become more and more viscous before cooling. It is also possible that some of the discrepancies might have been determined by later tectonic deformations (Fig. 6a). The line which indicates the direction of the magma flow is perpendicular to the plane of the Q joints which crosses at the angle of 90^0 this flow trend.

As shown in the two diagrams rendered for the north-easternmost part of the granitic body (Figs. 4b and 4c), the pitch angle of the magma flow in this part of the body varies from $N4^0W/44^0N$ to $N18^0E/32^0N$ depending on the position of the measured zone versus the feeder channel of the Săvârşin granitic body. On the diagram in Figure 4c L joints are almost vertical. Farther south-eastwards in the area of the Nucu Brook (Fig. 4d) the pitch angle of the magma flow is $N21^0E/22^0N$, and farther westwards, between the Laliţa Brook and the Troaş Valley it is approximately $N75^0E/20^0N$ (Fig. 4e). Farther westwards, in the zone north of the Mureş, in the area lying between the Troaş Valley and the Pârneşti Valley the direction of the magma flow is $N28^0E/28^0N$ (Fig. 4f). South of this area, in the Cetăţuia Hill, the Q joints are strongly dipping, almost vertical (Fig. 5a), but they keep the trend of the above-mentioned Q joints. This situation shows that here the pitch angle of the magma flow was less dipping as in Figure 4e, but the direction was the same (N52°E/12°N).

West of Vineşti, in the Porlata Valley basin the direction of the magma flow is N20°E/49°N (Fig. 5b) and farther westwards nearby the locality of Hălăliş it is N28°E/24°N (Fig. 5c), a value close to the pitch angle in the same direction of the megacrystals and autoliths in this area (Pl. I). It is to note that the diagram in Figure 5c was first presented in the Guidebook to Excursion 48 AC (Savu in Giuşcă et al., 1968). At Hălăliş the direction of the magma flow is N40°E/22°N (Fig. 5d). In this area the L joints are vertical, and the S joints are very little dipping. The pitch angle of the megacrystals and autoliths in this area is a bit higher (Pl. I).

On the national road (NH 7), between the localities of Vineşti and Hălăliş, an area close to the axial zone of the granitic body the Q joints are also strongly dipping, the line perpendicular to them (direction of the magma flow) trending N70°E, and pitching of 24°N (Fig. 5e), as in the Cetăţuia Hill area which is also situated close to the axial zone of the granitic body.

South of the Mureş, in the quarry north-east of the locality of Valea Mare, below the Măgura Hill, the situation is almost similar to that in the Cetăţuia Hill and also that on NH7, between Vineşti and Hălăliş, because the direction of the magma flow is approximately N73°E/18°N. At Valea Mare, probably due to some fractures (Fig. 11) concealed by the Mureş River and its alluvia, the granitic mass has been disturbed so that the Q joints, which are almost vertical, doubled; the aplite veins occurring on these joints have the same position. The direction of the magma flow changes either to N80°E/10°N or S80°W/10°S (Fig. 6a). The pitch angle of the melanocrate autoliths in this area is, however, N70°–80°E/52°N, probably due to the southern contact of the granitic body.

Vein rocks. The numerous dykes and veins which are crossing the granitic body are formed of aplites (Pl. II, Fig. 4) and, more rarely, rocks with a porphyritic microgranite structure. They are fine-grained and of a white-pinkish colour. The porphyritic rocks consist of an applite- or microganite-like groundmass within which phenocrysts of quartz, orthoclase and locally acid plagioclase can be observed. These rocks are located especially on the Q and S joints and more rarely on other joints (Fig. 6b, Fig. 7). In some places they occur in association with epidosite veinlets (Fig. 8) or contain in their median zone quartz veins and epidote nests (Figs. 9 and 10).

Unlike the granodiorite-diorite body of Temeşeşti in the north, no sulphide mineralizations occur in the granitic body. Depositions of fine pyrite and quartz crystals situated on Q joints, accompanied by granite kaolinization on the joints walls, are rarely noticed.

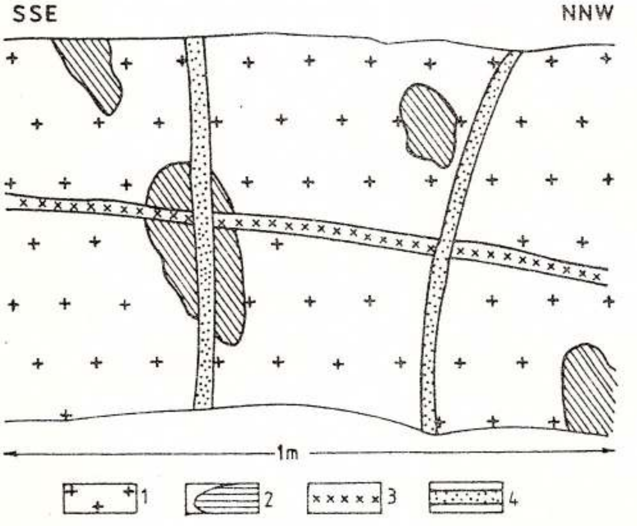

Fig. 7 – Aplite veins in the porphyritic granite on Valea Satului (Valea Mare). 1, porphyritic granite; 2, melanocrate autolith; 3, aplite; Qap, aplites situated on Q joints; Sap, aplites situated on S joints.

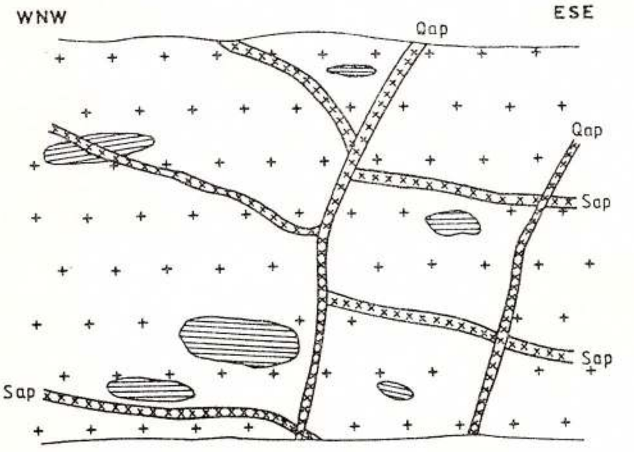

Fig. 8 – Porphyritic granite (1) with melanocrate autoliths (2) on the Valea Satului (Valea Mare), penetrated by aplites (3) and epidosites (4).

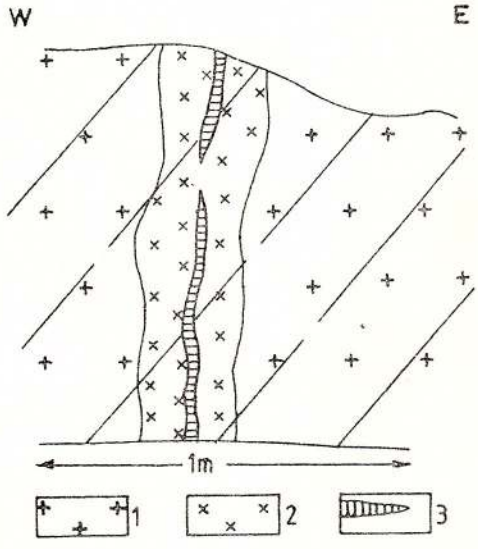

Fig. 9 – Porphyritic granite with Q (joints) (1), penetrated by an aplite dyke (2), with quartz depositions in the median zone (3). Vineşti.

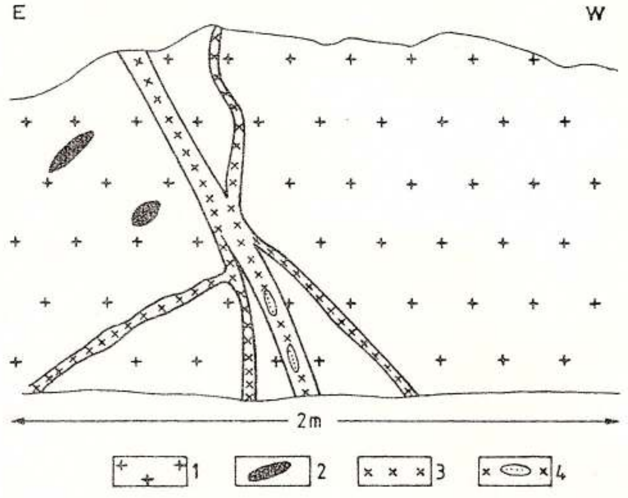

Fig. 10 – Porphyritic granite (1) with melanocrate autoliths (2) penetrated by an aplite dyke (3), with epidosite depositions in the median zone (4). NH7 (Hălăliş).

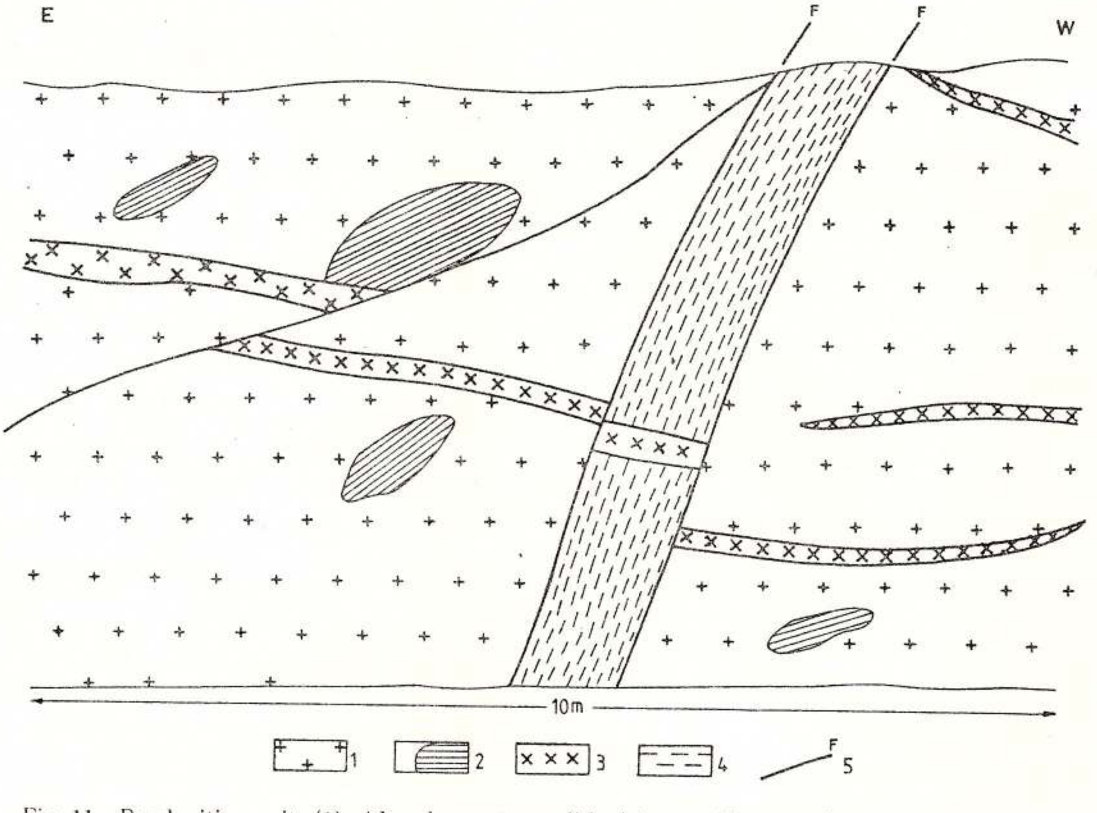

Fig. 11 – Porphyritic granite (1) with melanocrate autoliths (2) crossed by aplite (3) that was affected by a fault system (5) along which it was deformed (4). Mureş Valley.

The granitic body has been affected by two fault systems, which locally determine the granite shearing and a poor shifting of the blocks separated by them, as one can observe at Valea Mare (Fig. 11). One of the fault system is trending N18^{0}W and is deeping 75^{0}SSW (Fig. 6d). This system is the oldest and it correlates with the Q joints. It corresponds to a system of regional fractures affecting the Drocea Mts, which has almost the same direction. The second fault system is trending approximately E–W (N88^{0}E) with strong dippings (65^0) northwards (Fig. 6d). This system seems to be more recent. It correlates with the system of fractures along which the Lăpugiu Basin, south of the Săvârşin granitoid massif, has sunk.

Conclusions

The field evidence has pointed out that in the axial zone of the granitic body the direction of the magma flow has been quite close to the axis of the granitic body and the Q joints are approximately perpendicular to it (Pl. I). In the marginal zones of the granitic body although the strike and dip of the three joint systems and the direction of the magma flow are similar, as concerns the general direction, to those in the axial zone, several modifications occur as regards their strike and dip, probably determined both by the influence of the walls of the magmatic body during the magma intrusion and by the later fractures.

The direction of the magma flow (the intrusion trend) correlates well with the structural elements represented on the map (Pl. I).

In the Săvârşin area a gravity anomaly has been rendered evident by Andrei (1964) (Fig. 12). It shows two minima (gravity lows): one in the Contrava Valley and another south of the former, in the Găunoasa Hill.

A comparison of the strike and dip of the joints and the direction of the magma flow in different sectors of the granitic body with the axis of the southern part of the gravity anomaly stressed out:

The axial direction of the intrusion of the magma which has generated the granitic body (as shown on the map – Pl. I – and the gravity anomaly – Fig. 12) shows that the starting place in the depth of this magma is the same with that of the magma that has generated the northern granodiorite-diorite intrusion of the massif – the Temeşeşti intrusion – (Savu, Mândroiu, 1980; Savu et al., 1986). This area lies between the Contrava Valley and the Găunoasa Hill in which the enrootments of the two intrusions are marked by the two gravity lows of the gravity anomaly (Fig. 12). Nearby the surface magma has changed its upward direction, approximately penetrating the plane between the sheeted-dyke and the basaltic complexes of the Liassic ophiolitic megaslab obducted from the crust of the Mureş Ocean which hosts the Late Kimmerian granitoid massif of Săvârşin.

From the zone south of the Contrava Valley the granitic magma penetrated first towards south, then its direction was west-southwest and then it changed gradually to south-west (Fig. 12).

The result is a laccolith-like granitic body, elongated and excentrical versus the feeder channel, situated in the northeasternmost part of the granitic body.

The Săvârşin granitic body differs thus from the Temeşeşti intrusion of the granitoid massif not only

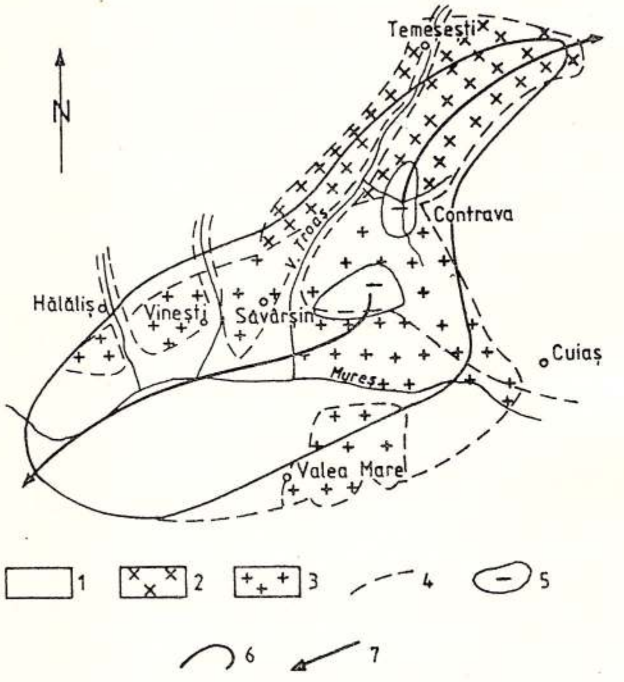

Fig. 12 – Sketch of the minimum gravity anomaly in the Săvârşin area (according to Andrei, 1964, unpubl. data). 1, alluvia; 2, granodiorite-diorite intrusion of Temeşeşti; 3, granite intrusion of Săvârşin; 4, granitoid massif contact; 5, gravity low; 6, boundary of the gravity anomaly; 7, direction of the granitic magma flow (intrusion).

by its petrographic constitution but also by the morphology and its strike and dip because the northern body is a dyke (sphenolith) trending N46^{0}E and dipping south-eastwards.

The mineralizations associated to the granitic body are rare and they occur as pyrite veinlets related to kaolinization phenomena situated on the Q joints.

The granitic body was affected by two fault systems: the former trending NW-SE and the latter trending approximately E-W.

As a conclusion we may state that this paper has established the tectonics of the Săvârşin granite and clearly demonstrated that the Late Kimmerian granitoid massif of Săvârşin consists of two intrusions: a northern one which penetrated from SW to NE, trending N46^{0}E, and a southern one, of the Săvârşin Granite, which penetrated from NE to SW, trending S75^0 W (Fig. 12).

References

Andrei, J. (1964) Repport. *Arch. I. G. R.*, Bucureşti.

Balk, R. (1937) Structural behaviour of igneous rocks. *Geol. Soc. Am. Mem.*, 5, p. 1–177, Washington.

Cloos, H. (1923) Das Batholithenproblem.. *Fortschr. Geologie u. Palaeontologie*, Ht. 1, Berlin.

Didier, J. (1964) Étude pétrographique des enclaves de quelques granites du Massif Central Français. *Ann. Fac. Sci. Univ. de Clermont*, 23, p. 1–245, Clermont Ferrand.

Giuşcă, D. (1950) Le massif éruptif de Vlădeasa. *Ann. Inst. Géol.*, XXIII, p. 200–252, Bucureşti.

— , **Bleahu, M., Borcoş, M., Dimitrescu, R., Kräutner, H., Savu, H. (1968)** Neogene volcanism and ore deposits in the Apuseni Mts. *Intern. Geol. Congr., XXIII Ses., Prague, Guide to Excursion, 48AC*, Romania.

Martin, N. R. (1953) The structure of the granite massif of Flamanville, Manche, North-West France. *Quart. Journ. Geol. Soc.*, CVIII, 4, p. 311–341, London.

Möbus, G. (1959) Zür Tektonik der Ganggesteine im Lausitzer Granitmassiv. *Z. Geologie*, p. 601–611, Berlin.

Papiu, V. C. (1945) Repport. *Arch. I. G. R.*, Bucureşti.

Savu, H. (1953) Repport. *Arch. I. G. R.*, Bucureşti.

— , **Vasiliu, C. (1966)** Temperatura de formare a granitului de Săvârşin (Munţii Drocea). *D. S. Inst. Geol.*, LII, 1, p. 141–157, Bucureşti.

— , **Mândroiu, V. (1980)** Studiul metalogenetic al masivului banatitic de la Săvârşin (Munţii Drocea). *D. S. Inst. Geol. Geofiz.*, LXIV, 2, p. 133–151, Bucureşti.

— , **Vasiliu, C., Udrescu, C. (1967a)** Contribuţii la studiul geochimic al rocilor banatitice de la Săvârşin (Munţii Drocea). *D. S. Com. Geol.*, LII/2, p. 359–382, Bucureşti.

— , **Vasiliu, C., Udrescu, C. (1967b)** Géochimie du massif banatitique de Săvârşin. *Assoc. Géol. Carp.-Balk.*, VIII^{eme} Congr., Belgrade, p. 265–270, Belgrade.

— , **Vâjdea, E., Romanescu, O. (1986)** The radiometric age (K/Ar) and the origin of the Săvârşin granitoid massif and of other Late Kimmerian intrusions from the Mureş Zone. *D. S. Inst. Geol. Geofiz.*, 70–71/1 (1983; 1984), p. 419–429, Bucureşti.

Socolescu, M. (1944) Les affleurements de minerais de la région Vaţa - Şoimuş Buceva - Săvârşin - Zam. *C. R. Inst. Géol. Roum.*, XXVIII, p. 93–125, Bucureşti.

Szentpétery, S. (1928) Petrologie des südlichen Teiles des Drocea Gebirges. *A. m. kir. Földt. Int. Evk.*, XXVII, p. 1–128, Budapest.

Yoder, H. S., Stewart, D. S., Smith, J. R. (1957) In Report of the director of the Geophysical Laboratory, Carnegie Inst. Washington Year Book, 56, p. 206–214, Washington.

Ştefan, A. (1986) Eocretaceous granitoids from the South Apuseni. *D. S. Inst. Geol. Geofiz.*, 70–71/1 (1983; 1984), p. 229–241, Bucureşti.

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Plate I

Geological Map of the Late Kimmerian Săvârşin Granitic Body

1, alluvia and terraces; 2, microgranitic porphyries; 3, granite in marginal facies; 4, common granite facies; 5, porphyritic granite facies; 6, contact of the granitic body; 7, boundary between the granitic facies and the microgranitic porphyries; 8, aplite dykes; 9, strike and dip of autoliths; 10, strike and dips of the contact between granites and ophiolites; 11, pitch angle of the orthoclase-albite megacrystals and of the autoliths; 12, fracture; 13, location of the structural diagrams in Figures 4, 5 and 6.

Plate II

Fig. 1 – Cetăţuia Hill constituted of porphyritic granite, an erosion remnant in the Mureş flood plain.
Fig. 2 – Aspects of the porphyritic granite at Hălăliş. Life size.
Fig. 3 – Porphyritic granite on NH7, west of Vineşti. Life size.
Fig. 4 – Porphyritic granite with melanocrate autoliths, crossed by aplite dykes. Valea Mare.

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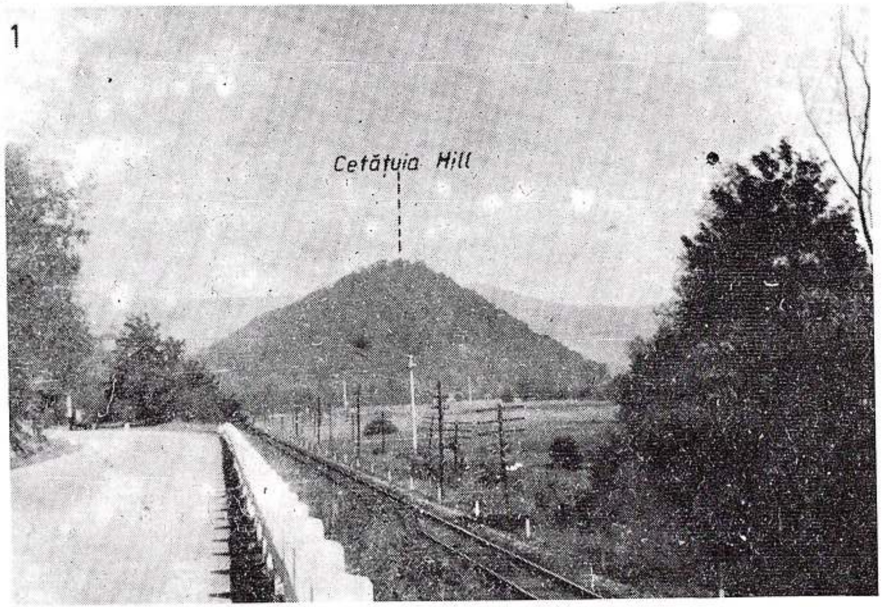

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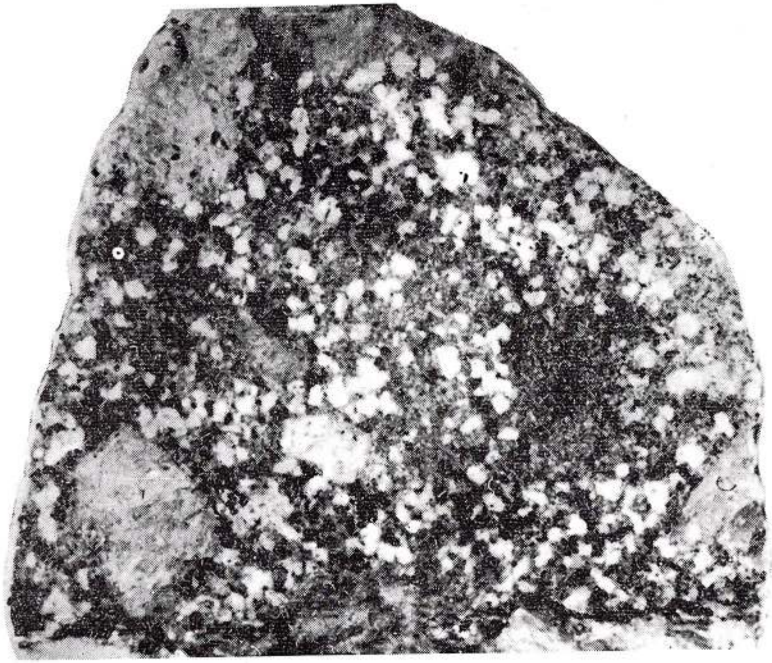

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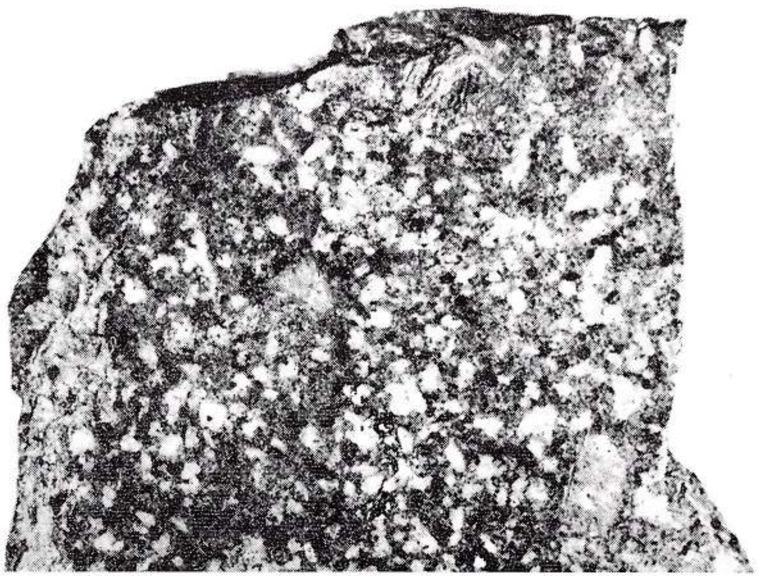

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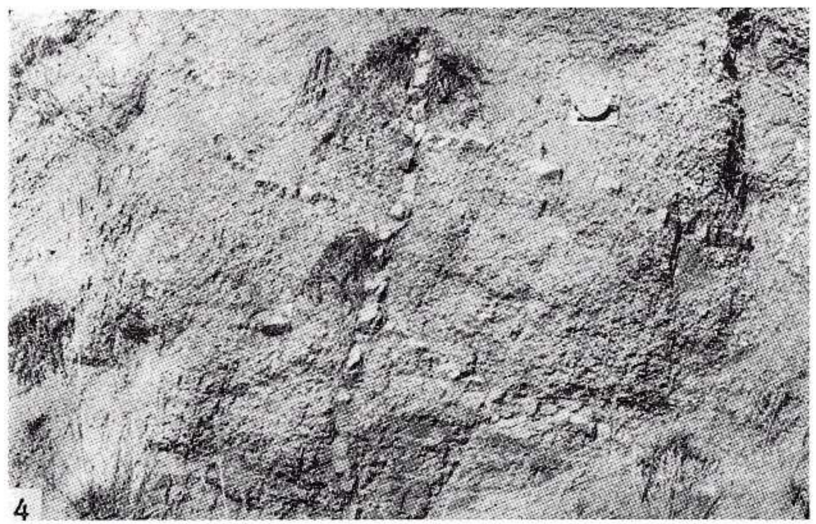

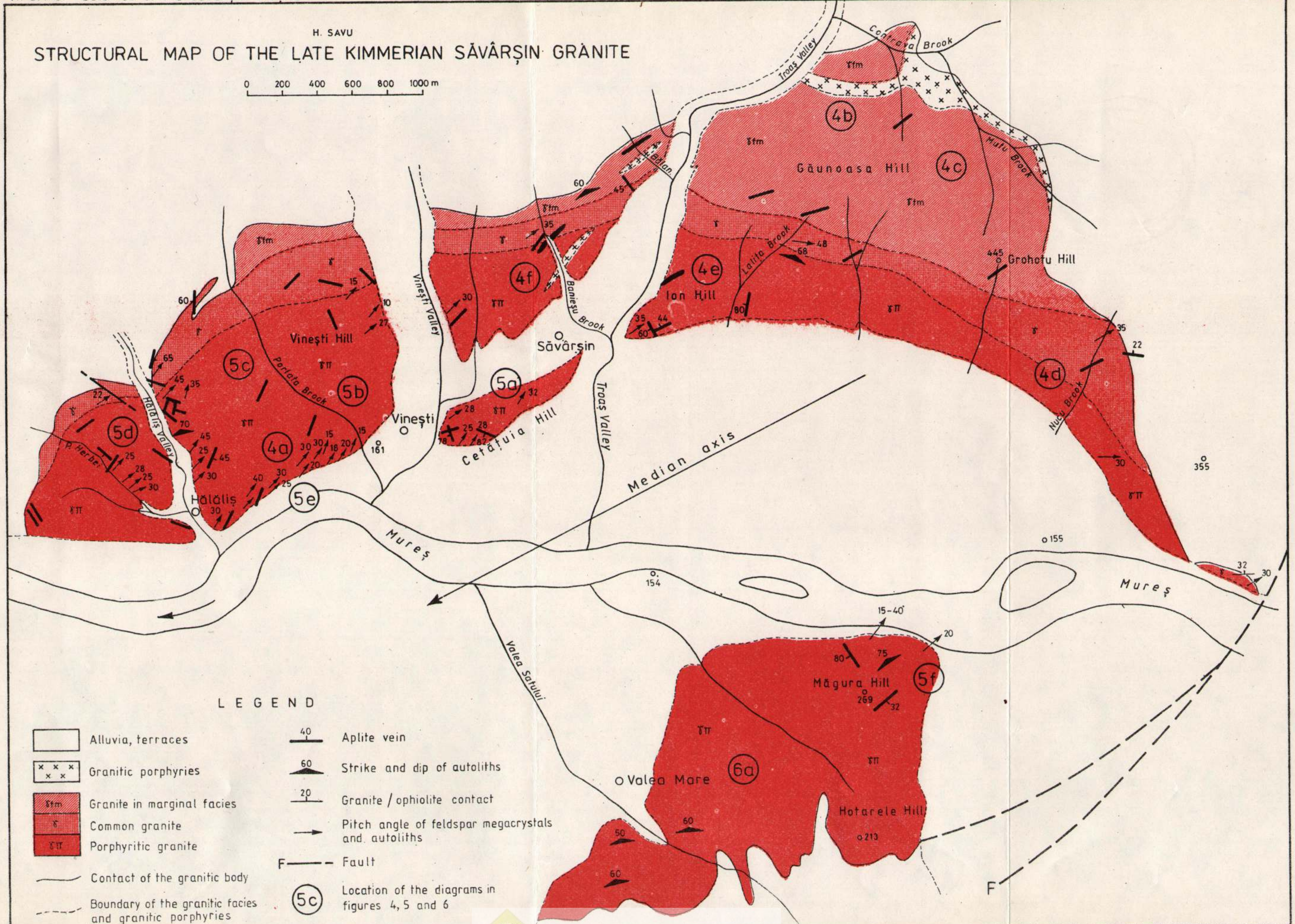

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TECTOGENETIC TRANSFORMATIONS IN THE PRE-ALPINE GEOLOGIC BASEMENT OF THE ORŞOVA-TOPLEŢ-BĂILE HERCULANE REGION (BANAT - SOUTH CARPATHIANS)

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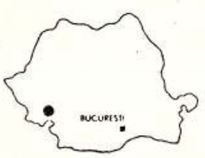

Key words: Caledonian tectogenesis. Hercynian tectogenesis. Retromorphism. Granitoid crystalline schists. Volcano-sedimentary formation. Rhyodacites. Keratophyres.

Abstract: The pre-Alpine basement has been transformed by a Caledonian tectogenesis and a Hercynian tectogenesis. The Caledonian tectogenesis affected the Precambrian-Cambrian crystalline schists of the Neamţu Group. This took place by the dynamic metamorphism as an effect of the reverse faults or/and of the incipient overthrusts. The tectogenetic zones show mylonitic structures and are strongly retromorphosed from the amphibolitic to the greenschist facies. The Hercynian tectogenesis involved the Neamţu Group, the Permian volcano-sedimentary formation and the Sesemin Granitoid. The Neamţu Group underwent a static retromorphism due to the numerous disjunctive dislocations at the end of the Permian. The water necessary to retromorphism has been taken from the overlying volcano-sedimentary formation strongly compressed by the Hercynian movements. Granitoids were cataclased, and the plagioclase from rhyodacites within the volcano-sedimentary formation albitized in places; in this way some rhyodacites became keratophyres.

Introduction

The region consists of a Precambrian-ante-Carboniferous crystalline-granitic basement crossed by vein-like rocks and syenitic rocks. All these rocks are conformably overlain by three sedimentary covers: the oldest, Carboniferous-Permian in age, the intermediate one, Jurassic-Cretaceous in age, and the youngest one of Badenian-Sarmatian age. The Permian sedimentary formation is associated with rhyodacites with keratophyre aspects in places.

Tectonically most of the region belongs to the Danubian Domain (Danubian Autochthon). The Getic Domain (Getic Nappe) crops out over a small area of about 4–5 km^2. Both the Getic Domain and the Danubian Domain from this area were assigned to the Carpian Supergroup (Upper Precambrian A) by Kräutner (1981).

The crystalline basement of the Danubian Domain is made up of the Neamţu Group with which the Ogradena and the Sesemin granitoids as well as the syenites are associated.

Significant tectogenetic transformations took place in the Neamţu Group as a result of the Caledonian movements.

The Hercynian movements with a smaller intensity on the rocks have affected the Neamţu Group, the Sesemin Granitoid, and the Permian volcano-sedimentary formation.

Caledonian Tectogenesis

The Caledonian tectogenesis involves only the Neamţu Group.

The Neamţu Group consists mainly of amphibolites, amphibolic gneisses, plagioclase gneisses, more rarely micaschists. At various levels quartzite, quartzite gneisses, sericite-chlorite schists, calcareous shales, limestone and crystalline dolomites, manganese silicate rocks and metagabbrodiorite interbeds occur subordinately. These rocks do not form continuous levels but lenses greatly varying in size (1–20 m thick, 10–2,000 m long).

The prograde regional metamorphism of the Neamţu Group took place in the almandine amphibolite facies during the Dalslandian orogenesis (Bercia, Bercia, 1975; Savu et al., 1978; Kräutner, 1980).

There are certain intensely tectonized lineaments within the Neamţu Group, trending NNE or N-S that made up as a consequence of the dynamic movements subsequent to the prograde regional metamorphism. These zones are connected by reverse faults or/and overthrust planes.

The most important strip lies in the eastern part of the Cherbelezu and Sfârdin Granitoids. It is 30 km long and up to 1,5 km wide. This zone was described as a distinct lithological entity as the Corbu Series (Codarcea, 1937, 1940; Bercia, Bercia, 1975, 1980; Kräutner, 1980).

Other strips reaching only 3–4 km in length and up to 200 m in thickness, also trending NNE-SSW, can be identified within the Neamţu Group on the Valea Satului Brook, on the Vodna Valley near the Ogradena granitoids. Some crystalline rock septa from the Ogradena Granitoid also underwent a retrograde dynamic metamorphism before the intrusion of the magmatic body. These crystalline schists, like the Corbu Series, were first described as a separate lithological entity, with the name of the Vodna Series (Codarcea, 1937, 1940).

The retromorphic character of the Corbu and Vodna Zones was pointed out by Gunnesch and Gunnesch in 1978. According to these authors, retromorphism took place during the Hercynian orogenesis at the expense of the Neamţu Group.

Stan et al. (1981, 1982, unpubl. data), Mărunţiu and Seghedi, 1983 a, b), Stan (1984, 1985) come to the conclusion that the Corbu and Vodna Series are the result of the dynamic retromorphism manifested during the Caledonian movements at the expense of the Neamţu Group. It is also possible that the Corbu-Vodna tectogenetic zones include a Cambrian younger series, too, as suggested by the palynological data (Visarion, in Savu et al., 1978), transgressive over the Neamţu Group basement. But this series together with the Neamţu Group were subsequently strongly mylonitized so that the boundaries between these lithological entities cannot be marked.

Another three intensely tectonized zones about 10 km long were outlined in the Topleţ-Mehadia sector, and an 8 km long, tectonized one was pointed out in the Bolvaşniţa sector.

The tectogenetic zones consist of sericite-chlorite schists, actinolite schists, quartzite sericite-chlorite schists ± partly chloritized garnet, quartzites, limestones, dolomitic limestones, calc-schists, quartzite schists with albite porphyroblasts, chlorite-epidote schists, micaschists bearing sericitized muscovite and chloritized biotite, very rarely graphite schists and serpentinites. The texture of these rocks is mylonitic and retromorphic. They show the greenschist facies. In places these crystalline schists contain andalusite, staurolite, biotite and garnet porphyroblasts (Stan, 1981, 1982, unpubl. data, 1984, 1985; Mărunţiu and Seghedi, 1983; Mercus et al., 1985, unpubl. data). Dinică (1989) quotes also sillimanite. The nontectonized crystalline schists of the Neamţu Group gradually pass to the intensely tectonized zones. No obvious geological boundary can be observed. In thin sections the gradual transition from the less tectonized zones, where the minerals are little altered and deformed, to the zones with ever more altered and crushed minerals can be also observed. Thus for instance the amphibolite gneisses change into actinolite-sericite gneisses, the amphibolic schists into epidote chlorite schists, the gneisses into sericite schists or sericite quartzite schists. If tectogenesis is very intense the rocks show strongly schistose, in places phyllitic structures. These rocks are associated with "hydrothermal" quartz veins, calcite, ankerite, pyrite, chalcopyrite, blende and galena, locally even molybdenite. Under these stress conditions acid hydrothermal solutions form, their temperature not exceeding 200–300^{0}C (Helgeson, Garrels, 1968). They can metasomatically carry elements from the crushed crystal network and form stable crystallographic edifices under the new temperature and pressure conditions. Thus in this case the main primary source of carbonates may consist in the basic plagioclases within rocks; the protore for iron and magnesium within ankerite consists of amphiboles. The femic minerals provide the metallic minerals for the pyrite, chalcopyrite, blende and galena mineralizations. The genesis of the calc-schists, of the ankerites and even that of some limestone lenses could be also explained through this process of "squeezing" by stress. The small size of the limestone lenses (10–50 cm thick) is probably not accidental; they cannot be hunted after on the field as they do not exceed 10–20 m. Quartzites from the tectogenetic zones, also lens-shaped and difficult to follow along strike are partly the product of the SiO_2 residual accumulation resulted through the squashing of the quartzitic gneisses. So, these tectonodynamic-metasomatic processes, manifested at the expense of the Neamţu Group, give rise to the "Corbu and Vodna Series". They could be named Corbu-Vodna-Neamţu tectogenetic zones (C.V.N.).

Because the Ogradena granites are of an old Hercynian age, Devonian-Carboniferous (345 m.a., acc. to Mânzatu et al., 1974), and pierce both the Neamţu Group (Upper Precambrian A) and the C.V.N. tectogenetic zones, too, the level latter very likely formed during the Caledonian movements within the Cambrian-ante-Devonian time-span.

Hercynian Tectogenesis

The Hercynian movements at the end of the Permian period affected the volcano-sedimentary formation. However, these movements manifested also in the crystalline-granitic basement (Neamţu Group and Sesemin granitoid).

In the Mehadia-Bolvaşniţa sector numerous rhyodacitic veins trending N-S cross both Permian conglomerates and the basement of the Neamţu Group. The abundance of these veins points out the intensity of the Permian disjunctive dislocations which affected also the crystalline basement. Consequently the Neamţu Group is marked to a great extent by a static regional retromorphism pointed out by the extension of the chloritization and sericitization phenomena. The Neamţu Group can be seldom recognized by its initial features found south of Topleţ, in the Bolvaşniţa sector. The intensity of the static retromorphism coincides with the maximal development of the Permian volcano-sedimentary formation. This observation suggests that during the Hercynian movements the volcano-sedimentary formations released a certain amount of water that penetrated on fractures and fissures, regionally retromorphosing the Neamţu Group. In the zone of Topleţ, where the volcano-sedimentary formations are lacking, the static retromorphism occurs only sporadically.

The gneissic-cataclastic structure of the Sesemin granitoid is also due to the tectonic movements at the end of the Permian. In this sense the argument is suggested by the Permian volcano-sedimentary formation from the Mişlep-Iablaniţa sector, where the respective beds, almost vertical, are strongly compressed and cataclassed.

The feldspar albitization within some rhyodacites (keratophyres from the Permian volcano-sedimentary formation) was interpreted also as an effect of the coercive tectonic movements manifested during the Hercynian (Stan, 1987).

Conclusions

The Caledonian and Hercynian tectogenesis greatly influenced the pre-Alpine infrastructure of the region.

The Caledonian tectogenesis determined the formation of the C.V.N. zones; it has a dynamic character that manifested lineary, on north-south directions by incipient faults and/or overthrusts. The rocks of the Neamţu Group, involved in the Caledonian tectogenetic processes, show obvious mylonitic structures both macro- and microscopically. The transition from the nontectonized crystalline schists of the Neamţu Group to the strongly tectonized ones (Corbu, Vodna Series) takes place gradually.

The Hercynian tectogenesis affected the Permian volcano-sedimentary formation penetrating also the subjacent formations. It manifested by disjunctive dislocations and compressions, locally raising vertically the beds of the volcano-sedimentary formations. The Hercynian tectogenesis brought about a static retromorphism within the Neamţu Group, which superposed in places over the Caledonian dynamic retromorphism. The static retromorphism is well represented in the Neamţu Group, where the Permian volcano-sedimentary formations are also present. The water released by these formations during the tectonic movements at the end of the Permian probably played an important role in this process. In this sense it is also likely that the Permian volcano-sedimentary formations underwent an incipient regional metamorphism that should be demonstrated. The Hercynian tectogenesis also determined the gneissic structure and the more or less increased cataclasis of the Sesemin granitoid as well as the albitization of the plagioclase feldspars from the rhyodacites of the Permian volcano-sedimentary formation.

It is also possible that the alpine tectogenesis should have more or less affected the pre-Mesozoic infrastructure of the region.

References

Bercia, I., Bercia, E. (1975) Formaţiunile cristaline din sectorul românesc al Dunării (Banat-Carpaţii Meridionali). *An. Inst. Geol. Geofiz.*, XLIII, p. 5–73, Bucureşti.

— **(1980)** The crystalline of the Danube domain from the Banat (România). *Rev. roum. géol., géophys., géogr.*, 24, 1, p. 3–13, Bucureşti.

Codarcea, A. (1937) Note sur la structure géologique et pétrographique de la région Ogradina-Sviniţa (department de Severin). *C. R. Inst. Géol. Roum.*, XXI, p. 179–198, Bucureşti.

Codarcea, A. (1940) Vues nouvelles sur la tectonique du Banat Méridional et du Plateau de Mehedinţi. *An. Inst. Géol. Roum.*, XX, p. 1–75, Bucureşti.

Dinică, I. (1989) Contributions to the geological evolution of the crystalline fromations in the south Banat (Danubian Realm). *D. S. Inst. Geol. Geofiz.*, 74/1, p. 191–207, Bucureşti.

Gunnesch, K., Gunnesch, M. (1981) Formaţiunile cristalofiliene din sud-estul Munţilor Almajului (Banat). *St. cerc. geol., geofiz., geogr., (Geol.)*, 23/1, p. 23–32, Bucureşti.

Helgeson, H. C., Garrels, R. M. (1968) Hydrothermal transport and deposition of Gold. *Ec. Geol., Bul. Soc. Ec. Geol.*, 63 (6), p. 622–635,

Kräutner, H. G. (1980) Lithostratigraphic Correlation of the Precambrian in the Romanian Carpathians. *An. Inst. Geol. Geofiz.*, LVII, p. 229–296, Bucureşti.

Mărunţiu, M., Seghedi, A. (1983 a) New data concerning the metamorphic rocks and metamorphic processes in the Eastern Almaj Mountains. *Rev. roum. géol., géophys., géogr., (Géol.)*, 27, p. 29–36, Bucureşti.

— **(1983 b)** Mylonites in the Almaj Mountains. *An. Univ. Bucureşti*, XXXII, Bucureşti.

Mânzatu, S., Lemne, M., Tiepac, I., Colios, E. (1974) Données géochronologiques concernant les massifs granitoides de l'Autochtone danubien situé au nord du Danube (Roumanie). *Rev. roum. géol., géophys., géogr., (Géol.)*, 18, p. 9–17, Bucureşti.

Savu, H., Maier, O., Bercia, I., Berza, T. (1978) Assyntic metamorphosed formations in the Southern Carpathians. *Rev. roum. géol., géophys., géogr., (Géol.)*, 22, p. 19–29, Bucureşti.

Stan, N. (1984) Polimetamorfismul şisturilor cristaline situate în partea de est a masivului granitoid de Cherbelezu (M. Almaj). *D. S. Inst. Geol. Geofiz.*, LXVII/1, p. 293–300, Bucureşti.

— **(1985)** Cherbelezu and Sfârdin granitoides. Their relationship with the adjacent rocks. *D. S. Inst. Geol. Geofiz.*, LXIX/1, p. 167–180, Bucureşti.

— **(1987)** Upper Carboniferous and Permian volcanics from Romania. In Flügel, H. W., Sassi, F. P., Grecula, P.: Pre-Variscan and Variscan Events in the Alpine-Mediteranean Mountain Belts. p. 445–456, Bratislava.

— **, Colios, E., Bratosin, I. (1986)** Formations permiennes de Topleţ-Mehadia-Bolvaşniţa (Banat). *D. S. Inst. Geol. Geofiz.*, 70–71/1, p. 217–228, Bucureşti.

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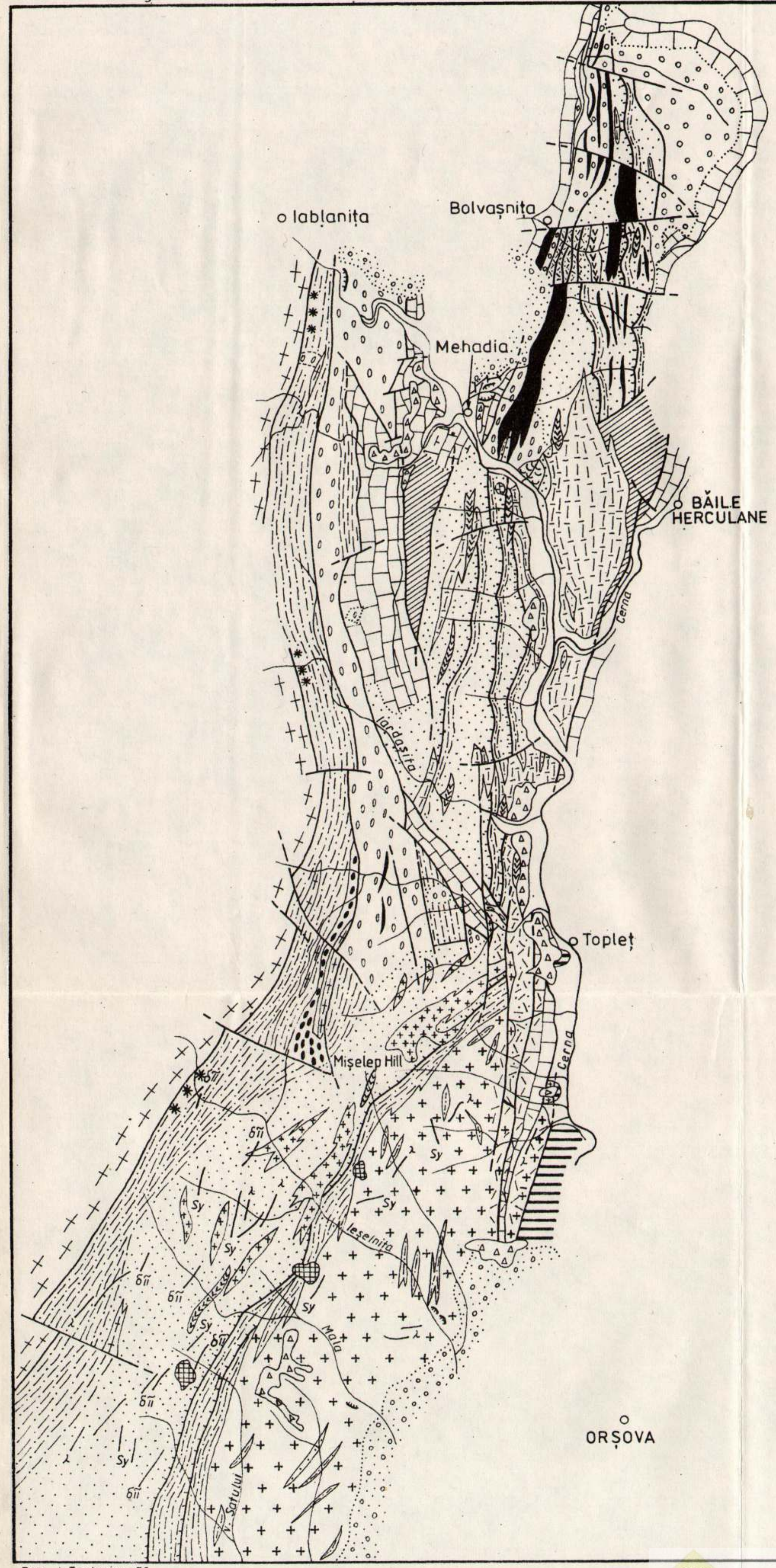
N. STAN

GEOLOGICAL MAP OF THE ORȘOVA-TOPLEȚ-BĂILE HERCULANE (BANAT)

0 1 2 3 4km



LEGEND

SEDIMENTARY ROCKS

QUATERNARY — Alluvia (a), terraces (b), slidings (c)

NEOGENE — Conglomerates, marls, clays, sands, gravels, limestones, argillaceous shales

JURASSIC-LOWER CRETACEOUS — Quartz conglomerates, microconglomerates, sandstones, limestones, argillaceous shales.

PERMIAN — Polygenetic conglomerates (a), red sandstones, siltites (b)

CARBONIFEROUS — Conglomerates, microconglomerates, sandstones, marly clays.

IGNEOUS ROCKS

MESOZOIC — Alkaline foidic syenites; bodies (a) porphyric microsyenites (b); Sy

PERMIAN — Rhyodacites, dacites with quartz keratophire aspect; dyke (a); vein (b)

LOWER PALEOZOIC
- (micro) granodiorite-porphyries, (micro) diorite porphyries ± quartz (δπ), lamprophyres (λ).
- Sesemin granitoids: granites, subordinately granodiorites; gneissic structure.
- Ogradena granitoids: granites, subordinately granodiorites; massive, sometime slightly oriented (a), aplitic microgranular facies (b).
- Cherbelezu and Sfârdin granitoids: granites, subordinately granodiorites.

CRYSTALLINE ROCKS

DANUBIAN DOMAIN

Corbu - Vodna - Neamțu tectogenetic-retromorphic zone (C.V.N.); greenschist facies.

UPPER PRECAMBRIAN-CAMBRIAN — Sericite-chlorite schists, micaschists, paragneisses ± micaceous, graphite schists, calc-schists, quartzites, crystalline limestones and dolomites; serpentinized metaperidotites accidentaly

Neamțu Series (group); Barrovian metamorphism in the almandine amphibolite facies; Hercynian regional retromorphism up to the greenschist facies - chlorite zone

UPPER PRECAMBRIAN
- Amphibole gneisses ± biotite ± garnet, amphibolites, micaceous gneisses ± garnet, biotite micaschists, migmatic gneisses, quartzites, cristalline limestones.
- Metagabbrodiorites ± quartz (hornblende orthogneisses)
- Plagioclase gneisses, micaceous gneisses.

GETIC DOMAIN

Lotru-Sebeș Series: Barrovian metamorphism in the almandine amphibolite facies; low pressure, intermediary metamorphism.

- Gneisses, amphibolites, garnet micaschists, staurolite ± kyanite ± andalusite, ± cordierite ± sillimanite.

SYMBOLS

- Thermal contact metamorphism: andalusite schists ± staurolite ± garnet ± biotite ± muscovite
- Brecciation zones
- Quarry
- Fault
- Reverse fault

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