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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

STRUCTURE AND PETROGRAPHY OF THE REGION WEST OF ŢARCU MASSIF (EASTERN BANAT)[1]

by

NICOLAE GHERASI[2], HORST PETER HANN[3]

Getic Nappe. Danubian Units. Metamorphic rocks. Volcanoclastics. Volcanic rocks. Major elements. Sedimentary cover. Anticlinoria. South Carpathians — Crystalline Danubian Realm — Ţarcu Mountains — Muntele Mic; Sedimentary Danubian Domain — Feneş zone.

Abstract

The formations making up this region belong to two main units: the Getic Nappe and the Upper Danubian. Within the Danubian units the nappe pile consists, from top to bottom, of the Arjana, Măru, Muntele Mic and Presacina Units. The Arjana Unit consists of the Lias-Dogger anchimetamorphic volcano-sedimentary formation and of a sedimentary cover represented by deposits assigned to the Malm and the Upper Cretaceous respectively. The volcanic rock complex has a clear alkaline character and consists mainly of trachytes, keratophyres and pyroclastics. The Măru Unit consists of the Măru Amphibolite Series. The Muntele Mic Unit consists of the Muntele Mic granitoid which underlies a Jurassic sedimentary cover exhibiting peculiar facies features. The Presacina Unit consists of a weakly metamorphosed Paleozoic basement underlying a Jurassic cover in which rare alkaline igneous rock dykes are also found. The whole structure presented was post-overthrust folded in an antiform which would represent the southward continuation of the Rof anticlinal structure.

Résumé

Structure et pétrographie de la région à l'ouest du mont Ţarcu (Banat de l'Est). Les formations de cette région appartiennent à deux unités principales : nappe gétique et danubien supérieur. La pile de nappes des unités danubiennes est considérée par nous être constituée, de haut en bas, par les unités d'Arjana, Măru, Muntele Mic et Presacina. L'unité d'Ar-

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jana est formée d'une couverture sédimentaire représentée par des dépôts appartenant au Malm, respectivement au Crétacé supérieur et par la formation volcano-sédimentaire anchimétamorphique d'âge liasique-dogger. Le complexe de roches volcaniques présente un caractère nettement alcalin et comporte en particulier des trachytes, des kératophyres accompagnés de pyroclastites. L'unité de Măru comprend le granitoïde de Muntele Mic qui supporte une couverture sédimentaire jurassique aux caractères faciaux particuliers. L'unité de Presacina est constituée d'un soubassement paléozoïque légèrement métamorphisé (ordovicien-silurien) surmonté d'une couverture jurassique où se développent aussi de rares dykes de roches éruptives alcalines. Toute la structure a été plissée postcharriage en une antiforme qui représenterait le prolongement vers le sud de la structure anticlinale de Rof.

This paper is based on the investigations started by N. Gherasi in 1960 and continued in the period 1962–1963 east of the Ilova and Armeniş localities, dealing mainly with the keratophyre volcano-sedimentary complex. These investigations were subsequently extended to the basins of the Rîul Lung, Rîul Alb and Sebeşul Mare Valleys by a mapping on scale 1 : 10 000, being carried out together with E. Matsch and H. P. Hann. The presence of keratophyres was pointed out since 1958 by the microscopic determination (of an agglomerate) after an investigation made by N. Gherasi and I. Rădulescu on the Sebeşul Mare Valley. It is also worth noting that in 1929 A. Streckeisen, traversing the Ţarcu-Pleaşa crest, pointed out to N. Gherasi that the eruptive rock complex which appears round the Pleaşa ridge would not represent Verrucano, as F. Schafarzik had established in 1898, but Liassic formations.

1. Historical Background

The first data on the geology of this region are mentioned by Stur (1889), Inkey (1891) and Schafarzik (1898 a) who, in a first paper, assigns to Verrucano the reddish arkosian sandstones which occur from Dragota to the Deavoia Valley and are overlain by the conglomeratic and argillaceous Liassic crossed by diaclases. In 1898 (b) he described felsites and felsite-porphyries, removed in conglomeratic rocks at Pleaşa and Dosul Ilovei, which he considered to be Verrucano based also on the presence of some argillaceous shales fragments, arbitrarily assigned to Culm on the Bolvaşniţei Valley. Codarcea (1940) establishes the presence in the Danubian of some nappe structures, among which also the Arjana Nappe consisting of Liassic tuffs and basic effusive rocks, Upper Jurassic reef limestones as well as of Upper Cretaceous Arjana flysch. The tectonic map drawn by him is bounded northwards by that presented by us (Pl. I). Rădulescu and Rădulescu (1957, 1958) described keratophyres, porphyries, diabases and syenites in the Sebeşul Mare Valley basin, east of Borlova. Boldur (in Răileanu et al., 1958, respectively in Năstăseanu et al., 1958, 1959) investigated the area extending between Feneş and the Ţarcu Massif. They established the presence of both a large anticline formed of Paleozoic deposits in the axial zone and the Arjana Nappe west of Ţarcu and in the Pleaşa ridge, which would be overlain by the Upper Cretaceous flysch nappe. Hurduzeu (1962) prospected the area situated between Feneş and Armeniş for iron ore deposits. Năstăseanu (1967) points out that the volcano-sedimentary series extends up to the Rîul Lung Valley, underlying wildflysch deposits. In 1968 he carried out a systematic study of the Paleozoic and Mesozoic deposits from the Ideg Valley to the Rîul Lung Valley. The Jurassic limestones are considered olistoliths in a wildflysch type Upper Cretaceous. Năstăseanu (1979) contributed important specifications regarding the lithology and stratigraphy of the area situated between the Cerna and the Hideg Valleys. Concerning the Jurassic, he distinguishes between a Presacina facies and a Feneş facies, comprising the gritty conglomeratic Lower Lias, the prevailingly clay Middle Jurassic containing *Entolium liasicum* Nyst., the argillaceous Toarcian bearing *Harpoceras* which supports a volcano-sedimentary complex assigned to the Middle Jurassic-Upper Jurassic pro parte.

Researches were carried out by Gherasi and Savu (1969) and Savu et al. (1973) on the Muntele Mic Granitoid and the crystalline schists delimiting it.

In recent years the researches on the geology of the Ţarcu-Godeanu region have developed due to the mappings for the elaboration of the Geological Map of Romania, scale 1 : 50.000 Godeanu sheet (Conovici et al., 1986, 1987). These investigations cover an area which is delimited by or, in a restricted zone northeastwards, overlaps the area presented in our map. It is worth mentioning the studies made by S. Năstăseanu, D. Russo-Săndulescu, T. Berza, V. Iancu, A. Seghedi, E. Negrea which will be further discussed in this paper. We also mention the recent communications regarding this area, presented by Năstăseanu and Negrea (1986, unpublished) and Iancu et al. (1990).

2. Geologic Setting

The formations making up the presented region belong to two principal units : the Getic Nappe and the Upper Danubian. The latter comprises several recently established units (Berza et al., 1983). Within it we distinguished the following units : the Arjana Nappe with the Upper Cretaceous flysch, the Măru Unit, the Muntele Mic Unit and the Presacina Nappe, which is in lower position. The partly anchimetamorphic Arjana Nappe consisting of a Jurassic volcano-sedimentary formation with keratophyres, trachytes, included in a tuff and tuffite mass, thrusts over both the Liassic of the Presacina, respectively the Feneş Units (sensu Năstăseanu, Negrea, 1986) and the Muntele Mic Granitoid. The Măru Unit, consisting especially of amphibolites which contain also gneiss interbeds, represents the southern termination of a zone which reaches the maximum thickness east of Bistra Mărului, enclosed between the Getic Nappe and the Muntele Mic Unit which it overthrusts. The latter unit comprises the granitoid massif of the same name as well as Jurassic deposits developed over small areas. North of the Craiu Valley it is noticed that the overthrust of the Muntele Mic Granitoid and of the neighbouring crystalline schists (Gherasi, Savu, 1969 ; Kräutner et al., 1981) was in its turn resumed by the overthrust plane of the Arjana Nappe. Regarding the latter nappe it is worth mentioning that Codarcea (1940) established the presence of two facies of the Liassic and Dogger: the Presacina and the Arjana facies. The latter consists of basic effusive rocks, tuffs and reef limestones bearing corals and brachyopods. The Arjana Unit thrusts over the Lower Jurassic and Cretaceous in Presacina facies within which the volcanic products are rare. The Arjana Nappe comprises also Upper Cretaceous flysch which is underlain by wildflysch.

3. Getic Nappe

The medium intensity metamorphics of the Sebeș-Lotru Group develop between Vîrciorova to the north and Ilova to the south, therefore in the north-western extremity of the investigated region, making up the Getic Nappe or, more exactly, the Turnu Ruieni Scale. The presence of this scale was established further north, in the Slatina Valley basin by Hann (in Savu et al., 1981) and the succession of lithological sequences characterizing it has been presented also by Hann (1987). A horizon consisting of white, quartz-feldspathic gneisses associated with numerous amphibolite levels is characteristic. The respective sequence, assigned by Kräutner (1980) to the leptyno-amphibolic formation, is found also in the region south of Vîrciorova. The white quartz-feldspathic gneisses are found here on the Plaiul Mare Crest in the Albinele Valley and on the Gîrdea ridge constituting a horizon. We also note the large development of the amphibolites. In addition, there are alignments marked by mylonites and ultramylonites striking approximately N—S, along some reverse faults, parallel to the Getic overthrust, which represents a continuation of those pointed out also in the Turnu Ruieni-Slatina Valley zone. On these fracture lines, around which the effects of a dynamic metamorphism are visible, mineralizations are also located (pyrite, pyrrhotite, chalcopyrite), e.g. in the right bank of the Bolvașniței Valley, at Piatra Popii. These mineralizations are identical with those encountered further north in the Turnu Ruieni Scale, described by Savu and Hann (1981) who show that they are part of the Turnu Ruieni-Teregova-Lăpușnicel metallogenetic district, situated on the third alignment of Laramian eruptions from the banatite metallogenetic province.

On the Bolvașniței Valley sillimanite paragneisses, migmatites and rare pegmatite veins were also encountered.

4. Upper Danubian Units

The map presented (Pl. I, II) comprises the south-western end of a zone exhibiting these structural features, which borders further north the Hațeg Basin in the east-west direction and then bends southwards.

4.1. *Arjana Nappe*

In the investigated region this tectonic unit consists of the volcano-sedimentary formation belonging to the Lias, Dogger and possibly the Lower Malm as well as of a cover represented by deposits belonging to the Malm and the Upper Cretaceous respectively. To the north-east of the investigated region this unit was named the Muroniu Nappe, comprising also Paleozoic deposits and at the same time showing a scale structure (Năstăseanu, Negrea, unpublished).

4.1.1. *Upper Cretaceous and Malm Sedimentary Deposits*

In the western part of the investigated area the Upper Cretaceous deposits constitute an important zone extending from the western part of the Sadova Nouă towards the south and bends in the eastern part towards the north up to below the Țarcu Massif. South of the Deavoia Valley an Upper Cretaceous patch extends over the Dosul Deavoiei ridge up to the Hideg Valley. These sediments overlie transgressively the volcano-sedimentary deposits as well as the Jurassic limestones. Năstăseanu (1968) established their transgressive character. Codarcea (1940) considered these deposits as belonging to the Arjana flysch. The deposits consist of sandstones, conglomeratic sandstones, conglomerates. In their basal part there are also conglomeratic sandstones and conglomerates including rounded elements formed of prevailingly white quartz, limestones, but also Liassic black shales and igneous rocks. Conglomerate levels containing especially quartz elements were encountered at Poiana Înaltă. In the Rîul Alb Valley the sandstones constitute diaclased massive banks with gritty shale interbeds and subordinately thin (5—10 cm), slightly micaceous black argillaceous shale interbeds. Within the succession conglomeratic sandstone levels occur, suggesting a rhythmical sedimentation.

The Upper Jurassic limestones crop out at the top of the volcano-sedimentary formation and under the Upper Cretaceous flysch deposits. Fine, hard, red marly, often schistosed limestones are noticed in the Deavoia Valley, which grade to whitish limestones showing red "flames" and spathic limestones. Owing to some Saccocmidae remnants ("pelagic crinoids) these limestones were assigned to the Kimmeridgian. In places the spathic limestones contain keratophyre fragments.

4.1.2. *Volcano-Sedimentary Formation*

Starting with the Deavoia basin this formation develops northwards, reaching the maximum thickness (over 5 km) between Sadova Nouă and the Pleașa crest, and ending in the left bank of the Craiu Valley, under the Muntele Mic Granitoid. The volcanic products prevail, and the Liassic sediments appear as discontinuous intercalations, greatly varying in thickness (Pl. I). Psephitic-psammitic deposits are found in the base, such as those on the Armeniș Valley, downstream the granite window, where coarse sandstones occur. Conglomeratic sandstones and massive quartzitic sandstones underlying pyroclastics were noticed on the Boghii Valley, a tributary of the Armeniș Valley, both downstream and upstream the granitoids. These sandstones reaching a few meters in thickness differ from those on the Rîul Lung Valley which belong to the Liassic and Presacina facies and reach 400 m in thickness (Năstăseanu, 1979). Further north of the spring of the Sadovița Valley, at the height of 950 m (under Poieni) quartzitic sandstones are found. Grey-quartz sandstones are distinguished on the Morii Valley, beyond the second granitoid body. Conglomeratic and feldspathic sandstones are found also on the crest east of the Porcului Brook, tributary of the Bolvaşniţei Valley, towards Dosul Ilovei, at the altitude of 750 – 780 m. The Liassic deposits consist mostly of black argillaceous, siltic shales forming often thin, metric interbeds that cannot be mapped.

The volcanic rock complex exhibits a clearly alkaline character and consists of trachytes, keratophyres which prevail, accompanied by pyroclastics, bostonites, rarely oligophyres, dacites and augite andesites as well as small syenite bodies. Pyroclastics are represented by microconglomerates and lapilly tuffs, while agglomerates occur only in a few outcrops. Tuffs and tuffites are widespread in the Armeniş Valley basin.

Trachytes are rocks that underwent the least albitizations; they occur as dykes or lava flows, the most widespread of which was mapped in Mount Pleaşa . They look like felsites, are greenish or violaceous, respectively dark grey south of the peak, showing sometimes a chlorite network. South of this peak a pyroclastic lava including trachyte fragments exhibiting a vitrophyric texture, surrounded by a microgranular lava resulted through devitrification (Pl. III, Fig. 1). Karlsbad twinned orthoclase is noticed in the same section, in a felsitic lava. Porphyric, rarely glomeroporphyric structures occur, while phenocrystals may lack in places. The groundmass is as a rule trachytic, bearing parallel feldspar microlites or a fluidal structure appears (Pl. III, Fig. 2). The groundmass is frequently a devitrified glass determining an intersertal type structure with feldspar microlites randomly arranged in a cryptocrystalline mass (Pl. III, Fig. 3). The orthoclase often shows concealed spots that indicate an incipient exsolution stage, while three universal stage values are of $-2V = 60 - 65°$.

The tectonic movements resulted in the rock laminations preceded by cataclasations. The dynamic metamorphism generated "S" planes that generally adapted to the flow directions; along them sericites formed, which are disposed in thin, parallel bands, along which the fragmented feldspars are aligned (Pl. III, Fig. 4). Sometimes the sericite mass includes small relict orthoclase crystals.

South of the investigated region, in the Deavoia Valley and at Poiana Bradului, trachytes show a K-feldspar which is inhomogeneous at extinction and exsolution cryptoperthites lending a spotted aspect to the phenocrysts. Plagioclase is an acid oligoclase. The matrix shows a clear trachytic structure with parallel feldspar microlites.

Keratophyres are the prevailing rocks, being associated with pyroclastics. The latter are crossed by keratophyre dykes, more rarely by bostonites. The rocks are whitish-yellowish or greyish, in places dark grey. Due to the laminations sericite and chlorite stripes form over the respective surfaces. The feldspar phenocrysts are usually distinguished with the naked eye and their presence enabled us in most cases to distinguish between keratophyres and massive tuffs on the field. The structures visible under the microscope are generally porphyric, while the groundmass is trachytic, more rarely microgranular. The feldspar microlites are most often divergently disposed (Pl. IV, Fig. 2) or parallel and undulated, when the texture is fluidal. The main mineral is the usually albitized orthoclase resulting in the schachbrettalbite. Chlorite bearing biotite and augite relicts were sporadically noticed. The orthoclase crystals, which show universal stage values of $-2V = 60-64°$, are often cataclased and rotated. The presence of anorthose was mentioned by Rădulescu and Rădulescu (1959). Anyway the percentage of mafic minerals was reduced within the rock.

The keratophyric pyroclastics have been classified from the granulometric point of view as follows : tuff 1—4 mm in diameter, lapilly tuffs 4—32 mm, microagglomerates 32—50 mm (Wentworth, Williams, 1932). The cartographic separation of these products was not possible, only the agglomerate outcrops could be distinguished. The latter are found over a restricted area, between the upper course of the Irighiș Valley and Mount Pleașa . Pyroclastite may contain also argillaceous shales fragments that may reach 25 cm in diameter, where, east of the southern peak, the agglomerates contain enrolled trachyte fragments exhibiting a vitrophyric texture (Pl. IV, Fig. 3).

Microconglomerates consist of whitish fragments included in a dense mass which is brown-greenish, sometimes red-violaceous in colour due to the iron hydroxides. This coloration made Schafarzik (1898 a) consider them Verrucano. Laminations determine in places ocular aspects, the rocks becoming schistous. Microconglomerates contain keratophyre angular or rounded fragments (Pl. IV, Fig. 1) displaying a trachytic or divergent arborescent texture as well as isolated albite, more rarely orthoclase fragments. The basic microconglomerates can be recognized by their green sometimes violaceous colour ; they contain a lot of chlorite and femic mineral relicts. They are associated with ferruginous types, in their turn associated with tuffites, with an argillaceous component and a Fe content ranging between 6 and 12% (Gherasi et al., 1971). These rocks are found in the Deavoia Valley basin, in the Pietroasa Valley and in the crest situated on the right side as well as in the Călina Mare and Larga Mare Valleys. Microconglomerate and ferruginous tuffite levels crop out over more restricted areas in the Armeniș Valley basin and in the Rîul Lung Valley.

The lapilly tuffs enable the best observation of pyroclastics in thin sections, showing rather varied aspects. In addition to keratophyre fragments exhibiting a trachytic or microgranular, more rarely of intersertal type structure, hyalopilitic textures with dark brown glass also occur (Pl. V, Fig. 1). These tuffs were affected by a quite evident anchimetamorphism which starts with a cataclasation and the sericite formation (Pl. V, Fig. 2) ; a foliation is noticed among the schachbrettalbite crystals, while a microfolding occurred in a more advanced stage (Pl. V, Fig. 3). Regarding the distribution, it is found that the lapilly tuffs prevail over the microconglomerates. A basic lapilly tuff from the Sadovița Valley — Coama Fîntînele — Valea Morii area contains opaque fragments of basic lavas with microlites, trachyte fragments, feldspar and angular quartz crystals (Pl. VI, Fig. 1).

Crystalloclastic tuffs were encountered at Dosul Ilovei and in the Armeniș Valley. Those at Dosul Ilovei contain brown glass fragments and colourless fragments of microgranular feldspar aggregates, the vitreous groundmass being crypto-microcrystalline. The presence of some unalbitized K-feldspar crystals indicates a trachytic tuff. In the Armeniş Valley, the tuffs and tuffites are widely distributed, having at their bottom Lower Liassic coarse sandstones or microconglomerates. In the right bank of the valley a tuff is found, formed of a cineritic mass bearing feldspar microlites, which is in fact a devitrified glass of very fine granulation and small oxidized pyrite crystals. The cement includes feldspar crystals, quartz, muscovite, sericite, biotite grains.

Bostonites form grains crossing the keratophyre pyroclastics in the Craiu Valley, the Irighiş Valley and in the Crucişoara Valley or in the Deavoia Valley and in the Poiana Bradului ridge. The feldspar developed divergently in small and equal prisms, but randomly disposed, is characteristic of the bostonitic structure. In places a porphyric structure is found (Irighiş Valley), with small unalbitized K-feldspar phenocrysts, which is orthoclase ($-2V = 65-70°$). It forms sometimes elongated prisms (Pl. VI, Fig. 2), replaced by schachbrettalbite. The rocks present flow textures and are affected by cataclasations and laminations.

Dacites are found on the Faţa Rugilor crest in the form of a dyke piercing the pyroclastics. Small quartz crystals can be distinguished also megascopically. Under the microscope one can notice that small poikilitic melanocrate mineral phenocrysts are completely carbonated (pyroxene) and oxidated (hornblende).

Plagioclases (oligoclase) are elongated along the "c" axis, indicating a viscous magma (Giuşcă, 1974). Quartz shows rounded phenocrysts which are corroded on the margin, while the groundmass is microgranular and consists only of plagioclases.

Syenites generally form small bodies both in the north and south of the investigated region. Rădulescu and Rădulescu (1959) described them for the first time. The syenite bodies are here crossed by keratophyres, therefore are older. Under the microscope the K-feldspar is spotted, being at present an exsolution of patch-perthite and is replaced by schachbrettalbite, which penetrates within it. Chlorite and carbonates are also noticed. On the Crucişoara Valley two small bodies are encountered, and close to the one situated in the vicinity of the confluence with the Sebeş Valley a silicate marble formed through infiltration metasomatosis is present. The marble contains forsterite and small green spinel crystals. The presence of this rock is difficult to explain as the pyroclastics do not contain also limestones at the level where the intrusion took place. Microsyenites consisting of K-feldspar, albite and very small amounts (below 1%) of hornblende and biotite are found in the Deavoia Valley.

The dark grey *oligophyres* crop out on the Irighiş Valley, in the Strîmbu Valley, on the Vărateca Peak as veins piercing black argillaceous shales, tuffs and agglomerates. They contain plagioclase phenocrysts in a microgranular groundmass. Plagioclase (An 12–16) is often twinned. Hornblende is highly oxidated, biotite is chloritized. The groundmass is microgranular, but also vitreous. The rocks are often intensely laminated, even mylonitized.

Andesites are rocks which were previously mentioned or described as diabases, diabase-porphyrites (Schafarzik, 1898 a), respectively uralitic porphyrites and diabase porphyrites (Gherasi et al., 1971). According to the present classifications, these rocks are pyroxene andesites. Andesites form dykes crossing the pyroclastics from west of the Vărateca crest, the Cuntu Valley and north-east of the Pleașa Peak, then in the vicinity of the granite windows, south of the Bolvașniței Valley, in the Morii Valley, the Sadovița and Armeniș Valleys. There are two veins on the last two valleys, one of which containing pyroxene crystals reaching 1.5 cm in size. On the Rîul Alb Valley they appear in the vicinity of the confluence with the Pietroasa Valley, and at Poiana Bradului they pierce a keratophyric dyke, being therefore subsequent to the latter. Andesites are dark green rocks, in which pyroxene crystals are noticed, developed in short prisms and faintly pink-brownish coloured. The anomalous birefringence colours as well as an hourglass zoning indicate the presence of a titaniferous augite in which ng : c = 44 — 48°. The pyroxene uralitization may lead to their complete transformation, an acicular actinolite being formed. Hornblende frequently occurs chloritized. Plagioclases occur as idiomorphic, zoned or polysynthetically twinned crystals. The microlite groundmass is hardly distinguishable, being filled with opaque minerals (Pl. VI, Fig. 3).

The Jurassic sediments, represented by black argillaceous shales are intercalated in the pyroclastic mass. On the Vărateca crest greyish sericite anchimetamorphic argillaceous shales are found, marking the foliation, being succeeded by fine white sandstones at a greater altitude (1.040 m height). Argillaceous shales crop out also at the confluence of the Sebeș Valley with the Crucișoara Valley. White micaceous sandstones alternate with argillaceous shales, microconglomerates and schistose tuffites forming steep rocks on the Răchita Valley, tributary of the Cuntu Valley.

The Alpine, mostly dynamic metamorphism affected obviously but unevenly the rocks of the volcano-sedimentary formation. This is why the structure of the eruptive rocks was not uniformly modified. When lamination planes get formed, they contain sericite, constituting bands that mark the foliation. Pyroclastics were affected to the same extent. Within them "S" planes were generated, the clearest anchimetamorphic aspects being noticed in the lapilly tuffs.

4.2. *Măru Unit*

(Kräutner et al., 1981 ; Berza et al., 1983)

In the area presented in this paper only the southern part of this unit can be followed between Ilova in the south to the Sopot Valley towards the north, constituting a narrow zone which is better exposed on the Bolvașniței Valley. The Măru Unit is overthrust by the Getic Nappe, while the overthrust plane covers calcareous sedimentary deposits (intensely laminated, probably Mesozoic, white-greenish microcrystalline limestones which overlie transgressively the Măru amphibolites. It is worth mentioning that the overthrust plane of the Getic Nappe was previously figured (Gherasi et al., 1970) farther west inside the metamorphic pile, belonging to the Sebeș-Lotru Group. The Măru amphibolites (the Măru Series respectively, Savu et al., 1984) are represented by an alternation of amphibolites more or less migmatized amphibolic gneisses with quartz-feldspathic gneisses and sometimes almandine and kyanite paragneisses.

4.3. *Muntele Mic Unit*
(Kräutner et al., 1981 ; Berza et al., 1983)

In the north-western part of the investigated region, this unit is overthrust by the Măru Series amphibolites, therefore by the Măru Unit (Pl. I). Towards the east the Muntele Mic Unit underlies also tectonically the volcano-sedimentary formation of the Arjana Nappe. The unit consists of the Muntele Mic plutonic massif which preserves north-eastwards also a crystalline schist cover—the Bărniţa Series (Savu et al., 1981) ; the schists are arteritically migmatized or not, in places forming enclaves in the granite mass. Also here the Muntele Mic Unit overthrusts the Presacina Unit which is thus in lower position with respect to the other units.

The Muntele Mic Granitoid underlies transgressively and unconformably the sedimentary cover of the tectonic unit represented by Jurassic deposits.

4.3.1. *Sedimentary Cover*

The Muntele Mic Granitoid is overlain transgressively by quartzitic arkosian sandstones, light grey, slightly micaceous, fine sandstones altered by limonite, as can be seen, for instance, on the Bobului Valley, right tributary of the Bolvaşniţei Valley. White massive arkoses are found on the hill north-east of the Runcu Peak. Based on lithofacial similarities, it is assumed that these rocks represent the Lower Lias. Round the Runcu Peak conglomeratic sandstones bearing black quartz curvicortical coaly sandstones and fine sandstones are also encountered. Feldspathic sandstones as well as coarse quartz sandstones are well exposed in the Satului Valley. On the right bank of the Ilovița Valley, south of the Runcu Peak and quite close to the boundary with the granitoid, magnetite blocks containing hematite (82.20% Fe_2O_3 ; 57.33% Fe) are found in some greyish-reddish brittle sandstones. The round quartz grains within sandstones, observed under the microscope, indicate their clastic origin, namely their connection with the Liassic sandstones. South of Piatra Ilovei, in the base of the Upper Jurassic limestones, spathic yellowish limestones and bedded greyish limestones, limestones presenting hematite veins or, farther north, limy sandstones with echinid remains (ossicles), grey-blackish quartzitic calcarenites are found, while on the path under the hill spathic dark grey limestones and yellowish limestones are present.

These rocks were assigned to the Dogger. South-east of Piatra Ilovei, the granites are overlain by limy-gritty shales with greyish-silvery phyllite interbeds that can be seen in a ravine, tributary of the Morii Valley. These deposits probably belong to the Lias.

Massive grey and white limestones follow in continuity of sedimentation, which would represent the Malm. They are microcrystalline, do not contain organic traces, but contain siderites instead ; they are schistose in the base, microfolds being sometimes noticed. These limestones were assigned by Schafarzik (1898 a) to the Malm, showing similarities to the Stănuleţ Limestones (Piatra Iorgovanului), which belong partly to the Malm based on a *Nerinea* sp. form.

4.3.2. *Granitoid Massif*

The granitoids make up a plutonic massif, reaching the maximum thickness in the north of the presented map (5 km) and continues uninterruptedly southwards up to Sadova Nouă. South-east of the latter locality the granitoid outcrops represent windows of the unit to which they belong, which underlie the volcano-sedimentary formation of the Arjana Nappe. The granitoid is often quite laminated due to the dynamic metamorphism, which may determine even sericite schist aspects (Valea Boghii-Sadovița Valley).

The main eruptive body consists of granodiorites and granites. Subordinately biotite quartz diorites are found. As regards the petrography of these rocks, we mention that the respective data are found in the papers written by Gherasi, Savu (1969) and Savu et al. (1973).

4.4. *Presacina Unit*

The rocks belonging to this unit, which are in a lower position with respect to the other units, belong to a Paleozoic basement and a Jurassic cover, sporadically crossed by eruptive rocks dykes (bostonites, keratophyres and oligophyres) and rare tuffite interbeds.

4.4.1. *Eruptive Rocks, Agglomerates, Tuffs, Tuffites*

On the right bank of the Rîul Lung Valley, east and south-east of the Pleașa crest, the only zone where the eruptive rock dykes are somewhat more widespread is to be found. These dykes occur only sporadically in the rest of the area covered by the Jurassic formations.

Bostonites form dykes, under the microscope show a trachytic structure with parallel or divergent feldspars, contain schachbrettalbite with typical twins, accompanied by about 15% chlorite.

Oligophyres form also dykes and contain plagioclases (An 12) partly calcitized . The mafic minerals are chloritized. The brown groundmass exhibits a trachytic texture with feldspar microlites containing glass relicts in various devitrification stages.

Agglomerates occur only east of the Miuțon locality in the basin of the Cuntu Valley, covering restricted areas. The tuffs and tuffites are somewhat more widespread as compared to the agglomerates. They are found, for instance, on the Seroni ridge, in the north-eastern part of the investigated region. Tuffites show a marked foliation ; under the microscope small oligoclase crystals are noticed as elongated laths, aligned along the laminated planes. The rock mass is a siltite consisting of millimetric lamellae, containing angular quartz associated with calcite. This tuffite is underlain by Lower Liassic conglomeratic sandstones. North of the Seroni Peak greenish tuffs presenting an evident foliation crop out along with a bostonite (Pl. VI, Fig. 2). The tuff shows a microlenticular structure due to the laminations and contains rare oligoclase crystals. The glass, devitrified to a large extent, contains fine-grained feldspar associated with sericite and opaque minerals.

We point out the anchimetamorphic processes which affected especially the tuffs and tuffites.

One should also mention that in the Craiu Valley basin a few serpentinite bodies are found, which have been recently considered (Savu et al., 1986) as olistoliths.

4.4.2. *Jurassic Sedimentary Deposits*

Since the Lias sediments of an age which includes also the Dogger (Gherasi et al., 1970, 1971) deposited transgressively over a Paleozoic basement. In the Cuntu zone, in the upper basin of this valley as well as on that of the Craiu Valley, sandstones prevail; they are whitish feldspathic bearing white quartz veins, or coarse feldspathic, almost conglomeratic. On the Strîmbu Valley slightly metamorphosed sericite feldspathic sandstones were encountered. Gritty shales and phyllites, dark greyish limestones with echinid remains (ossicles), were also found, being intercalated in black argillaceous shales also in this region. The sandstones are generally overlain by the black argillaceous shales. According to Boldur (in Năstăseanu, Boldur, 1959), in the basin of the Rîul Lung and Rîul Alb Valleys the Lower Liassic is represented by conglomerates, microconglomerates, sandstones and arkoses. The Middle Lias consists of black shales, the age of which was established based on the *Frondicularia* cf. *pulchra* Terquem, found by Boldur (in Năstăseanu, Boldur, 1958) in these rocks, on the Rîul Alb Valley.

The recent investigations indicate that the whole pile, which, as shown by Gherasi et al. (1971), contains also sandstone and argillaceous black shale interbeds, in which hard fine greyish-blackish limestone layers are found, and which is crossed by the above-described dykes, includes also the Dogger. On the other hand, Năstăseanu and Negrea (unpubl.) assign the zone situated in the north-east of the map (Seroni Peak), considered by us of Liassic age, to the Paleozoic, and at the same time place this sector in the Muroniu Unit.

4.4.3. *Paleozoic Formation (Rîul Alb Formation*, Năstăseanu, 1975)

The Upper Ordovician-Lower Silurian age of this formation, also named the Rîul Alb Formation, was established by Boldur and Visarion (1972) based on palynologic and stratigraphic data. The rocks of this formation were encountered over larger areas in the basins of the Rîul Lung and Rîul Alb Valleys (Pl. I). The outcrops from the banks of these valleys constitute large rocky areas as the rocks are hard, compact, with a slightly marked, though visible greyish-greenish or greenish schistosity. The quartz grains, visible also macroscopically, indicate a detrital character. As shown by Gherasi et al. (1971), chlorite greenish pelitic rocks, more rarely subgraywackes, dark greyish argillaceous shales and siltite rocks, feldspathic sandstones bearing chlorite and microconglomerates (especially on the Rîul Lung Valley) are also encountered. Siltites contain small subrounded quartz and plagioclase fragments within a fine argillaceous, sericite ± chlorite mass. The pelitic schists show a fine sericite aggregate. Small muscovite lamellae form through the sericite recrystallization. Calcite appears on diaclases in places. The chlorite feldspathic sandstones are marked by the fact that the main minerals have the long axes parallel with the foliation.

The Rîul Alb Formation contains also some intermediate eruptive rock dykes which cannot be described so far due to the lack of information.

5. Geochemical Data

The chemical analyses presented in Table 1 concern exclusively eruptive rocks of the volcano-sedimentary formation. Two of them (samples nos. 296 and 287), however, come from a region situated outside the presented map, namely north-east of Mount Căleanu, close to the north-eastern crest (287) and on the Plaiul Mare Crest of the Vîrciorova crest above the springs of the Suculeț Brook (296).

TABLE 1

No.	Sample no.	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	PO_5	CO_2	S	Fe(S)	H_2O	TOTAL
1	296	47 .35	14 .57	5 .86	6 .64	0 .23	6 .75	5 .81	2 .50	3 .10	2 ·62	.79	.20	0 .00	0 .00	3 .15	99 .57
2	58	62 .70	14 .70	1 .10	4 .16	0 .90	2 .34	1 .35	6 .40	1 .06	.62	0 .00	1 .20	0 .82	0 .71	0 .12	98 .18
3	50	64 .00	17 .18	3 .09	á .89	0 .00	2 .08	.51	6 .20	3 .30	.30	0 .00	0 .00	0 .00	0 .00	0 .10	98 .71
4	61	64 .36	15 .53	5 .02	1 .60	0 .50	1 .40	0 .70	7 .70	1 .01	.00	0 .04	0 .00	0 .10	0 .05	0 .07	98 .52
5	118	65 .14	16 .59	.74	3 .01	0 .09	.65	1 .54	4 .70	5 .18	.54	0 .05	0 .88	0 .24	0 .21	0 .33	98 .89
6	287	66 .68	18 .58	1 .68	.76	0 .06	1 .16	.98	3 .77	3 .50	.31	0 .07	0 .00	0 .00	0 .00	1 .99	99 .25
7	54	73 .30	16 .25	1 .68	0 .13	0 .00	2 .38	0 .85	2 .00	2 .56	.00	0 .03	0 .00	0 .08	0 .07	0 .05	99 .38

296 Andesite,Căleanu-Obirșia Hidegului, 2,050 m altitude ; 58 Trachyte, Bolvașniței Valley ; 50 Keratophyre, Tîlva ridge, 1,030 m altitude ; 61 Keratophyre, Irighiș Valley, 895 m altitude ; 118, Keratophyre, Fața Rugilor ; 287 Dacite, Brustur Formation, the Suculeț Spring, 1,900 m altitude ; 54 Trachyte, Tilva ridge, 1,190 m altitude.

The analyses indicate a high alkali content in most cases. The Ritt man norm (Tab. 2) and the plotting in the QAP triangle (Fig. 1) confirm this tendency in the case of sample no. 1, which plots in the latite field, and of samples nos. 2, 3, 5 that plot in the alkaline quartz trachyte field. Sample no. 4 contains a relatively low K_2O, but very high Na_2O amount. The normative mineralogical composition (Tab. 2) contains a great albite amount, while sample no. 2, which shows values of the chemical analysis very similar to those of sample no. 4, contains anorthoclase as normative mineral. If we assigned the albite of the sample to the "p" parameter, it would lead us to an eroneous plotting in the quartz andesite field. It would be difficult to consider this rock as andesite, taking into account the extremely low CaO amount (0.70% — Tab. 1). But considering the albite of this rock as primary, we can assign it to the "A" parameter of the QAP diagram and thus the respective rock will plot also in the alkaline quartz trachytes field, which seems natural.

Samples nos. 6 and 7, by their chemism, are plotted in the field characteristic of the rhyodacites (Fig. 1).

TABLE 2

Rittman Norm

Minerals \ Sample no.	296	58	50	61	118	287	54
Quartz	—	13.04	7.03	8.49	13.61	23.69	40.34
Sanidine	27.68	—	—	—	72.46	23.11	17.87
Anorthoclase	—	64.46	78.17	—	—	—	—
Albite	—	—	—	79.39	—	—	—
Oligoclase	—	—	—	—	—	27.80	—
Andesine	—	—	—	—	—	—	12.48
Labradorite	40.93	—	—	—	—	—	—
Muscovite	—	—	—	8.39	—	10.34	6.97
Biotite	—	—	—	—	3.20	—	—
Diopside	2.65	—	—	—	—	—	—
Hypersthene	12,90	6.68	5.66	—	—	—	—
Olivine	8.58	—	—	—	—	-	—
Cordierite	—	9.71	7.89	1.91	7.13	14.24	21.92
Sillimanite	—	—	—	—	0.01	—	—
Magnetite	1.89	1.02	0.93	1.54	0.66	0.35	0.20
Apatite	1.75	—	—	0.08	0.09	0.14	0.06
Ilmenite	3.09	0.06	0.31	—	0.42	0.33	—
Calcite	0.52	2.43	—	—	1.94	—	—
Pyrite	—	1.65	—	0.21	0.46	—	0.16
Q	0	16.83	8.25	9.66	15.81	31.76	57.07
A	40.35	83.17	91.75	90.34	84.19	30.97	25.27
P	59.65	0	0	0	0	37.27	17.86
$= \frac{Al_2O_3-Na_2O}{TiO_2}$	4.61	13.39	36.60	--	22.02	47.81	—

In Niggli's alk-al diagram (Fig. 2) samples nos. 2, 3, 4, 5 plot in the alkaline rock field, samples nos. 1 and 6 in the intermediate rock field, while sample no. 7 in that of the poor alkali rocks.

In the Peccerillo, Taylor diagram (1976) (Fig. 3) sample no. 1 plots in the shoshonite series field (under the absarokite field), samples nos. 3

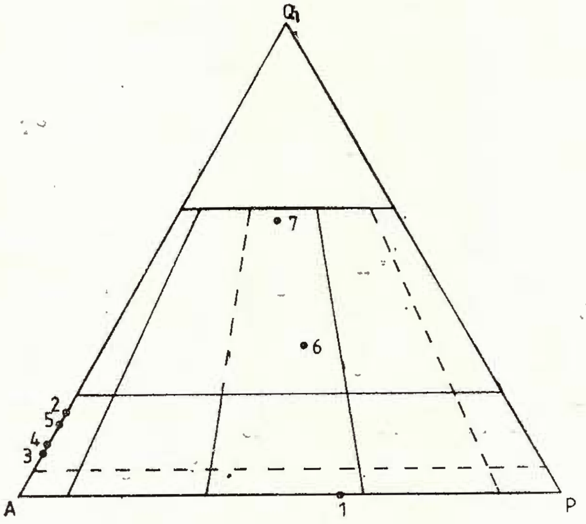

Fig. 1 — QAP diagram (Streckeisen, 1979).

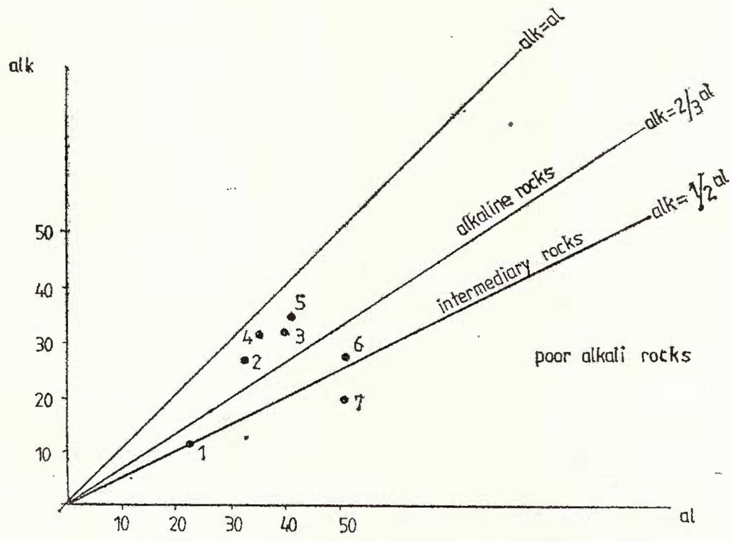

Fig. 2 — Niggli diagram (alk-al).

and 6 in that of the calc-alkaline series rich in K, while samples nos. 2, 4 and 7 in the calc-alkaline series field, a fact explained through the relatively low K_2O content ; sample no. 5 plots ouside the diagram, but likely in the prolongation of the shoshonitic series field.

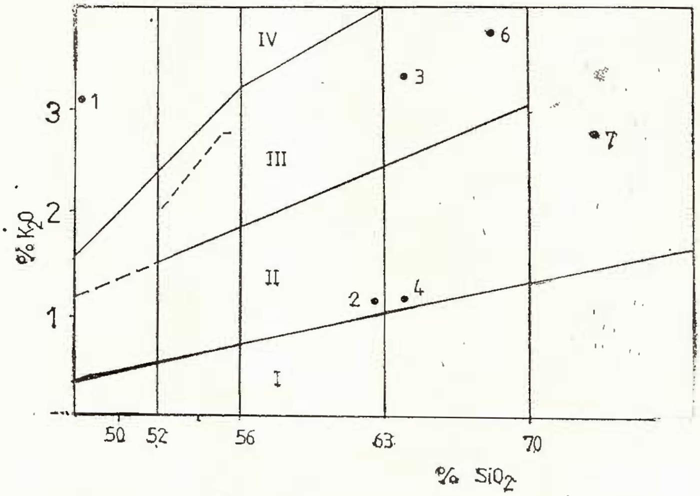

Fig. 3 — K_2O/SIO_2 diagram (Pecerillo, Taylor, 1976). I, field of arc tholeiite series; II, field of calc-alkaline series; III, field of calc-alkaline series rich in K; IV, field of shoshonitic series.

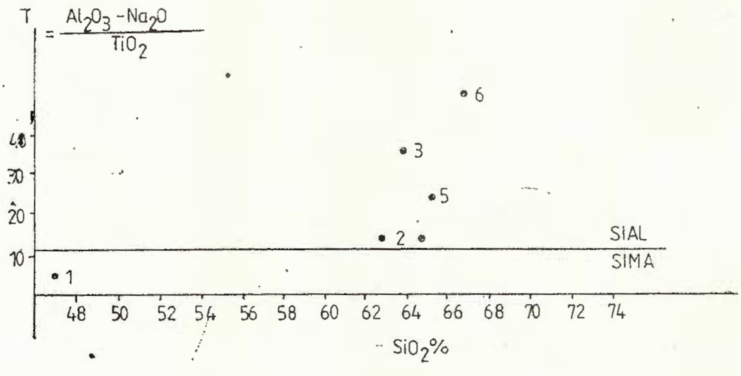

Fig. 4 — τ/SiO_2 diagram (Gottini, 1969).

In the τ/SiO_2 diagram (Gottini, 1969) only 5 samples plot, samples nos. 4 and 7, which do not contain TiO_2, have a τ index that tends towards the infinite. Of these only sample no. 1 (latite) has a smaller index, determining the plotting of the sample in the domain of the rocks of simatic origin, while the other samples plot above the boundary that separates this domain from that of the rocks of sialic origin. This latite, by the high Fe_2O_3 total, MgO and very low SiO_2 content indicates a basic character, while the great K_2O content shows that it is the equivalent of a potassic trachybasalt.

In conclusion, the first five samples have a clear alkaline character, sample no. 1, a latite, being the equivalent of an absarokite or of a potassic trachybasalt, while samples nos. 2–5 are alkaline quartz trachytes. The last two samples (nos. 6 and 7) are calc-alkaline rhyodacites.

6. Structure of the Region

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The region situated west of the Țarcu Mount, figured on the map, represents the intermediate segment of a complicated structural ensemble which is not yet sufficiently known and can be followed towards the north and north-east as well as southwards. The data presented attempt to contribute to a better understanding of the difficult problems connected with the tectonics of this region.

On the presented map we distinguished the Getic Nappe and the Upper Danubian Units. The pile of nappes from the Danubian Units consists from top to bottom of the Arjana, Măru, Muntele Mic and Presacina Units.

The Getic Nappe is found in the north-west and south-east of the region, it occupies the highest position and overthrusts directly, in the north-western part, the Măru Unit which belongs to the Upper Danubian. The Getic Nappe consisting of the medium intensity metamorphics of the Sebeș-Lotru Group, is represented in this zone by the Turnu Ruieni Scale and overthrusts Mesozoic calcareous deposits which overlie transgressively the metamorphic formations of the Măru Series. In the south-eastern part of the region the Getic Nappe thrusts over the Upper Cretaceous flysch of the Arjana Nappe.

In our opinion, the Arjana Nappe is in the highest position among the Danubian Units (Pl I, Pl. V). This unit thrusts both over the Muntele Mic Unit in the western part and over the Presacina Unit. As regards the contact between the Muntele Mic Unit and the Arjana Unit, it should be mentioned that some authors (Kräutner et al., 1981) had previously considered that the Muntele Mic Unit thrusts over the Arjana Unit. Following the boundaries according to our mappings on scale 1 : 10,000, it is obvious that the volcano-sedimentary formation of the Arjana Nappe is in higher position throughout its extent, a fact also proved by the numerous tectonic windows, usually situated along river valleys, in which the Muntele Mic Granitoid occurs frequently intensely laminated, surrounded by the pyroclastics or the Liassic schists of the volcano-sedimentary formation. In the few places where tectonic windows are encountered also on crests, this is explained by the fact that the overthrust plane of the Arjana Nappe shows sinuous aspects due to the post-overthrust tectonics (Pl. I). The Arjana Nappe consisting of the Jurassic (Lias-Dogger) volcano-sedimentary formation is bounded by Kimmeridgian limestones at the top and underlies the Upper Cretaceous flysch deposits.

Năstăseanu and Negrea have recently (1986) included in the north-eastern part of the map, in what they have separated as the Arjana Unit, also deposits considered by us to be of Liassic age and belonging to the Presacina Unit. But the above-mentioned authors assign these deposits (Seroni Zone) to the Paleozoic and consider them to belong, together with the volcano-sedimentary formation, to a recent unit, named by them the Muroniu Nappe, characterized also by the presence of some scales.

In the western part of the investigated region, the Măru Unit crops out from under the Getic Nappe, consisting of the Precambrian mesometamorphic metamorphics of the Măru Series and a thin Mesozoic limestone cover. This different interpretation is due to the local absence of the Arjana Nappe which crops out farther south as well as in the western flank of the antiform constituted by the nappe pile from the Ţarcu Mountains.

The Măru Unit overlies the Muntele Mic Unit consisting of granitoids bearing the same name, associated northwards also with the Băuţa Series metamorphics which occur both as septa within granitoids or borders them north-eastwards; the granitoids underlie a Jurassic sedimentary cover of particular facial characters. This unit overthrusts the Presacina Unit, which thus becomes the lowermost one of the Upper Danubian units distinguished by us.

The Presacina Unit consists of a Paleozoic (slightly metamorphosed —Ordovician-Silurian) basement that underlies a Jurassic (Lias-Dogger) cover. The Rîul Alb Paleozoic formation occurs as windows situated along the Rîul Lung and Rîul Alb Valleys, where they crop out over restricted areas. The Jurassic sedimentary deposits overlying this formation show a transgressive character. The Presacina Unit was named after the "Presacina Zone", established by Codarcea (1940, Codarcea et al., 1961); immediately south of the presented region it crops out over large areas, presenting the same lithological and stratigraphic characteristics. It is also worth mentioning that Năstăseanu and Negrea (1986) name the unit under discussion the Rîul Rece Nappe, excluding the Seroni Zone, which they included in the upper Muroniu Unit, therefore it would be partly equivalent to the Arjana Unit.

As shown on Plate I, the whole presented structure was in its turn post-overthrust folded, becoming an antiform which would represent the northward continuation of the Rof major anticlinal structure (Morariu, 1982). An important fault striking north-northeast-south-southeast, having the lowered eastern compartment, affects the axial zone of the anticline, exposing the Paleozoic formations in the uplifted compartment.

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STRUCTURA ŞI PETROGRAFIA REGIUNII DE LA VEST DE MUNTELE ŢARCU (BANATUL DE EST)

(Rezumat)

Formaţiunile care alcătuiesc regiunea aparţin la două unităţi principale : pînza getică şi danubianul superior. În cadrul unităţilor danubiene, stiva de pînze o considerăm ca fiind alcătuită, privind de sus în jos, din unităţile de Arjana, Măru, Muntele Mic şi Presăcina. Pînza getică, reprezentată prin solzul Turnu Ruieni, este constituită din metamorfitele grupului Sebeş-Lotru şi prinde sub planul de şariaj calcare mezozoice (?) ce stau transgresiv peste şisturile seriei de Măru.

Unitatea de Arjana este constituită din formaţiunea vulcano-sedimentară, anchimetamorfică de vîrstă Lias-Dogger, poate şi Malm inferior şi dintr-o cuvertură sedimentară reprezentată prin depozite calcaroase aparţinînd Kimmeridgianului, respectiv prin flişul cretacic superior. Complexul de roci vulcanice are un caracter net alcalin şi constă în special din trahite, keratofire care predomină, însoţite de piroclastite. Au mai fost întîlnite dyke-uri de bostonite, oligofire, dacite, andezite şi mici corpuri de sienite. Piroclastitele sînt reprezentate prin microconglomerate şi tufuri lapilice, subordonat aglomerate. Chimismul rocilor eruptive a evidenţiat caracterul alcalin clar al acestor roci. Sedimentele jurasice care formează intercalaţii în masa piroclastitelor sînt reprezentate prin şisturi argiloase negre, uneori în alternanţă cu gresii micacee, microconglomerate. Metamorfismul alpin, preponderent dinamic, a afectat evident, dar inegal rocile acestei formaţiuni. Foliaţia de laminare este marcată de benzi constituite din sericit.

Unitatea de Măru este constituită din seria de Măru şi încalecă spre est peste granitoidele de Muntele Mic.

Unitatea de Muntele Mic, constituită din granitoidul cu acelaşi nume, asociat spre nord şi cu metamorfitele seriei de Bărniţa, suportă o cuvertură sedimentară jurasică cu caractere faciale particulare. Această unitate şariază peste unitatea de Presacina, aşa cum se observă în versantul drept al văii Craiu, dar este încălecată de pînza de Arjana.

Unitatea de Presacina este constituită dintr-un fundament paleozoic slab metamorfozat (ordovician-silurian), ce suportă o cuvertură jurasică în care se întîlnesc şi rare dyke-uri de roci eruptive alcaline şi cîteva intercalaţii de tufuri sau tufite.

Întreaga structură prezentată a fost la rîndul ei cutată post-şariaj într-o antiformă care ar reprezenta continuarea spre sud a structurii anticlinale majore Rof. O falie importantă, orientată nord-nord-est — sud-sud-vest, cu compartimentul estic coborît, afectează zona axială a anticlinalului şi aduce la zi, în compartimentul vestic ridicat, formaţiunile paleozoice.

QUESTIONS

I. Drăguşin: Could the authors insist on the iron mineralization occurring in the region with special regard to their economic prospects?

Answer: The iron mineralizations in the region are of three types: 1) pyroclastics and limonitic tuffite-argillites connected with the basic volcanism that manifested at the top of the rock pile constituting the Arjana Nappe; 2) hematite lenses situated in the base of the limestones belonging to the Malm from Piatra Ilovei, which might represent oxidations of some siderite-ankeritic limestones. A research gallery of about 10 m was dug before 1914; 3) magnetite blocks associated with Lower Liassic quartzitic sandstones which are encountered on a plateau round the Runcu Summit, north-east of Ilova. Owing to the lack of outcrops, trenches were dug before 1940. The last two types are found in the sedimentary cover of the Muntele Mic Unit.

The mineralizations at Piatra Ilovei, namely from the Runcu Hill, represent important iron concentrations, but so far only small areas are known. The mineralizatios within pyroclastics and tuffites cover large areas (the zones in question have been described in the text, therefore they are quantitatively important, but the iron contents are low — up to 12% FeO).

M. Săndulescu: 1. Why do you use the denomination of Presacina Unit in the investigated area, since it is known that the lithofacies of the Mesozoic series does not show the characteristics of the Presacina zone? On the contrary, it belongs to the Feneş zone (according to S. Năstăseanu).

2. To what structural context does the tectonic window on the Rîul Lung Valley belong, in which Jurassic deposits crop out from under Paleozoic formations? The mentioned window has been pointed out also in recent oral communications by S. Năstăseanu. Why is it not figured on the map you presented?

3. Which is the vergence of the structures connected with the deformations contemporaneous with the Alpine dynamic metamorphism from the Arjana Nappe?

Answers: 1. We used the denomination of the Presacina Unit, extending the Mesozoic series of the same name, established by Codarcea (1940) to the north of the Hideg Valley.

Thus the black argillaceous schists with gritty interbeds from the basins of the Rîul Alb and Rîul Lung Valleys resemble the schists of the same type from Mehadia and Cornereva, assigned by Codarcea (1940) to the median part of the Lias in the Presacina zone. South of the zone investigated by us, Middle and Upper Jurassic calcareous deposits are found indeed, which also belong to the Presacina zone, but are not encountered north of the Hideg Valley. This fact may be explained by the overthrust of the Arjana Nappe as a result of which the respective deposits were covered in this region.

We did not use the denomination of the Feneş Unit, established by Năstăseanu and Negrea (unpubl.) as these authors include in this unit also some of the pyroclastics, formations which we include exclusively in the Arjana Nappe.

2. In a paper (Năstăseanu S., Russo-Săndulescu D., Iancu V., 1988), the presence of a half-window situated on the Rîul Alb Valley is mentioned only in the text, without the specification of its location, where "black fossiliferous clays of the type of those defining the Pliensbachian from the Sirinia Unit", lying "under the volcano-sedimentary deposits which are similar to those on the Rîul Rece Valley" are described. We assign the latter deposits to the Arjana Nappe. Therefore, Paleozoic formations are not mentioned in this context. We also mention that the map presented by the quoted authors does not include either the basin of the Rîul Alb Valley or that of the Rîul Lung Valley, being limited northwards by the right bank of the Rîul Rece Valley (Hideg).

3. The eastern vergences generally prevail; this can be better noticed in the basins of the Irighiş, Bolvaşniţa and Armeniş Valleys.

DISCUSSIONS

S. Năstăseanu: Regarding the tectonic units in the internal part of the Danubian Domain presented by the authors, the following specifications are made:

— The Arjana Nappe west of the Ţarcu Mount was first mentioned by Boldur and Boldur (1959) in the sense established by Codarcea (1940) only in the zone south of the Rîul Rece Valley.

— The volcano-sedimentary formation from the Feneş region does not differ either as regards its lithostratigraphy or its petrography from that of the Ţarcu Mount.

— The Muroniu Nappe (redefined as the Arjana Nappe) is underlain by the deposits of the Feneş zone which belong to the Rîul Rece Nappe; the latter overlies the deposits of the Presacina zone, as results from the succession of the nappes from the Rîul Rece-Rîul Lung antiform (sensu Năstăseanu, Negrea, 1987).

EXPLANATION OF PLATES

Plate III

Fig. 1 — Trachyte, Pleaşa Crest, saddle south of the altitude of 1.413 m. Microgranular lava with a trachyte fragment exhibiting vitrophyric structure (black), Karlsbad twinned orthoclase in felsitic trachyte. N +, 35 ×.

Fig. 2 — Trachyte. Ridge south of the Pleaşa crest, towards the Armeniş Valley, altitude of 1,050 m. Trachytic paste of clear fluidal flow structure. N II, 35 ×.

Fig. 3 — Trachyte. Ridge south of the Pleaşa Peak, altitude of 1,245 m. Elongated orthoclase phenocrystal in an interstitial groundmass, the glass containing feldspar microlites. N +, 60 ×.

Fig. 4 — Trachyte, ridge from the Golieş Peak towards the south-west, at the altitude of 990 m. Narrow sericite band bordering idiomorphic (black) feldspar crystals ; in the opposite part fragmented orthoclase, widespread somewhat farther from the sericite lamination zone. N +, 35 ×.

Plate IV

Fig. 1 — Agglomerate outcrop on the Irighiş Valley.

Fig. 2 — Keratophyre. Răchiţei Valley, right tributary of the Cuntu Valley. Divergent trachytic structure. Twinned albitic feldspars, schachbrettalbite, top left. N +, 80 ×.

Fig. 3 — Agglomerate. Mount Pleaşa — eastern zone. Two rounded trachyte fragments exhibiting a vitrophyric structure, fragment exhibiting an intersertal structure and albites showing a macrogranular structure. N II, 35 ×.

Plate V

Fig. 1 — Crystallolithoclastic lapilly tuff. Sadoviţa Valley, between the andesite dykes. Dark brown volcanic glass fragment and glass fragment exhibiting an obvious devitrification, with semiangular, rarely rounded feldspars ; sericite among feldspar crystals. N +, 30 ×.

Fig. 2 — Laminated lapilly tuff. Right tributary of the Armeniş Valley, downstream the granitoid window. Sericite among schachbrettalbite phenocrysts, the one in the middle being plastically deformed. N +, 80 ×.

Fig. 3 — Lapilly tuff, Tilva ridge, east of the peak. Feldspar lens surrounded by sericite, a microfold being noticed. Two (black) vitrophyric fragments, top right. N +, 60 ×.

Fig. 4 — Strongly laminated lapilly tuff. Irighiş Valley, upstream the granitoid. Band consisting of feldspar and sericite grains, microfold (centre) bordered by feldspar fragments (black) in a microgranular paste. N +, 80 ×.

Plate VI

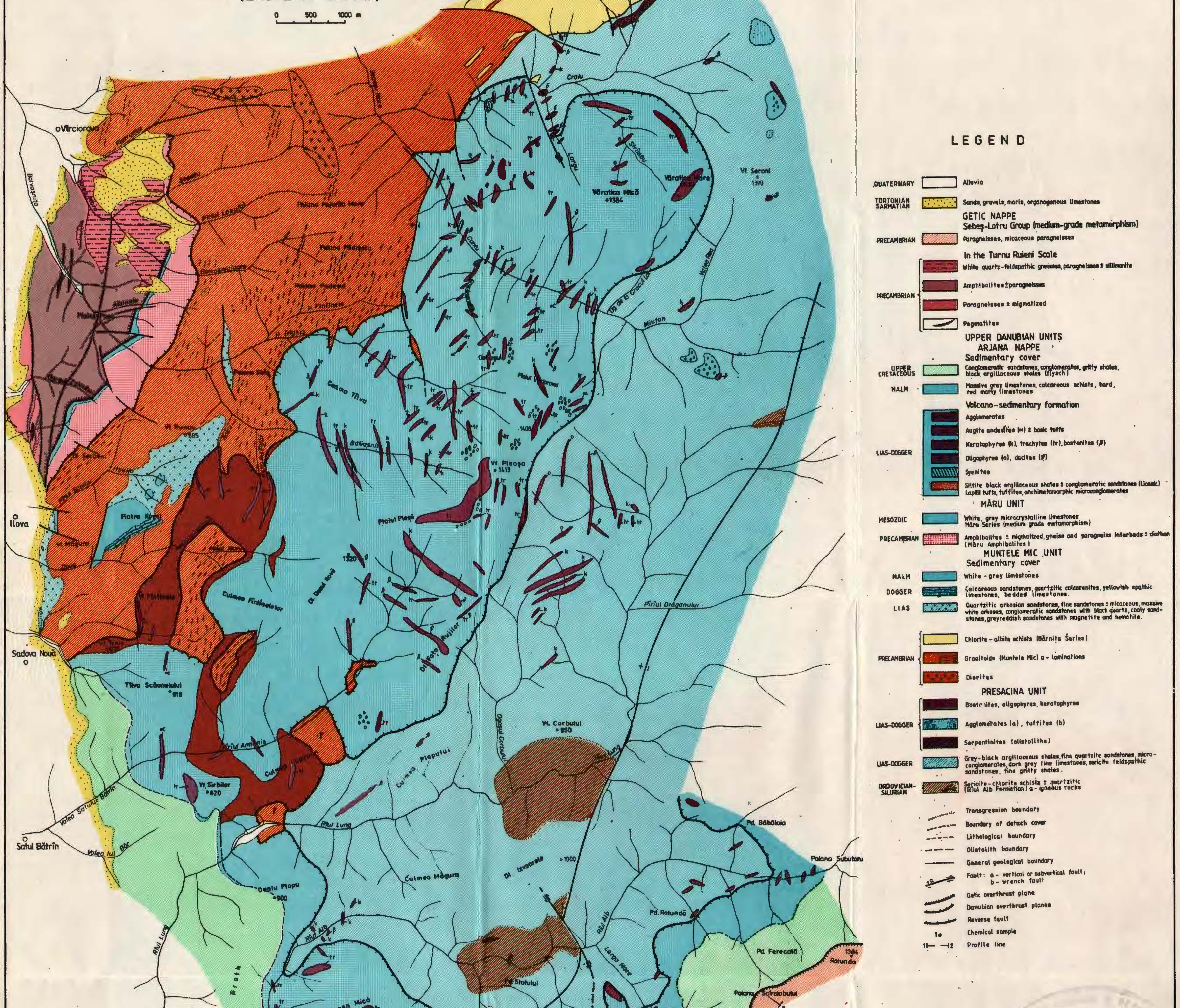
Fig. 1 — Lapilly basic tuff. Sîrbilor ridge, the peak at the altitude of 780 m. Trachyte fragments with a lot of phenocrystals, two andesite fragments with vitrophyric paste, angular feldspars in isotropic ash. N +, 40 ×.

Fig. 2 — Bostonite. Seroni Summit. Divergent schachbrettalbite, chlorite in interstices. N +, 80 ×.

Fig. 3 — Andesite. Sadoviţa Valley. Twinned titaniferous augite with basal cleavage and crystal (001) in the microlithic paste with opaque minerals. N +, 60 ×.

N. GHERASI, H. P. HANN

GEOLOGICAL MAP OF THE REGION WEST OF THE ȚARCU MOUNTAIN (EASTERN BANAT)



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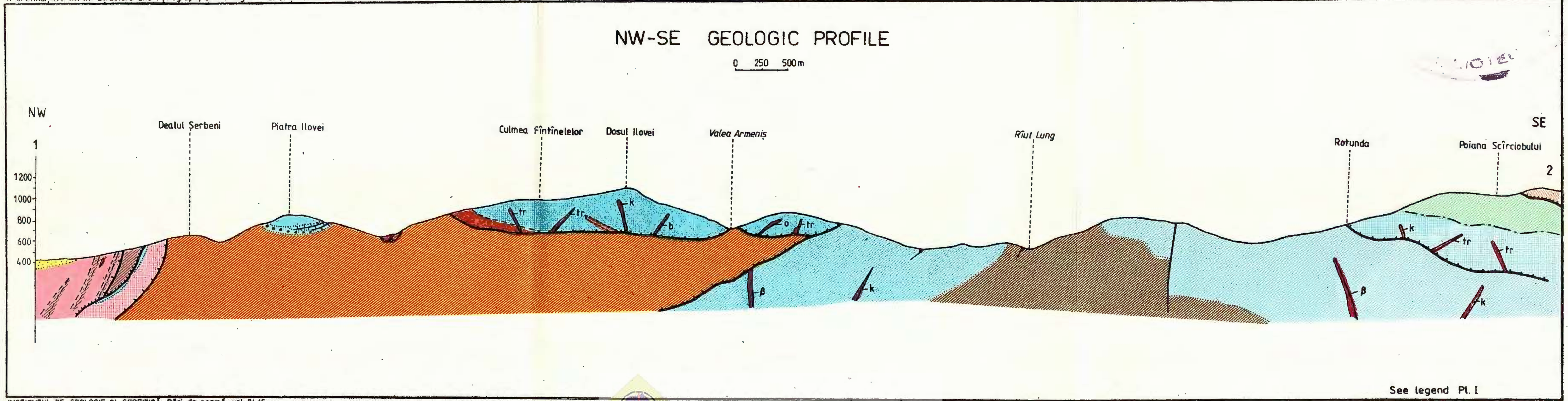

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. D. S. Inst. Geol. Geofiz., vol. 74/5 (1987), 1990, pag. 29—40
5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

NOTĂ ASUPRA PREZENŢEI FORMAŢIUNII DE VALEA LUPULUI ÎN FĂGĂRAŞUL CENTRAL, VERSANTUL NORDIC[1]

de

TEOFIL GRIDAN[2], GEORGE DUMITRAŞCU[2]

Tectonic units. Schists. Paragneiss. Migmatites. Petrology. Major elements. Structural analysis. Faults. Thrust fault. South Carpathians — Crystalline Getic and Supragetic Realms — Făgăraş Mountains.

Abstract

Note on the Presence of the Valea Lupului Formation in the Central Făgăraş, Northern Slope. The Valea Lupului Formation is an independent Mesocretaceous tectonic unit due to its Cumpăna type lithological constitution (a prevailingly gneiss-micaceous background showing phaneroblastic aspects and ophtalmitic migmatite levels), the structural and textural characteristics, the lack of the graphite rocks and of the crystalline limestone-amphibolite association. It represents a fragment of the Făgăraş Supragetic Nappe (Kräutner, 1983; Hann, Szász, 1984) or an outlier of the Argeş Nappe (Balintoni et al., 1986).

Résumé

Sur la présence de la formation de Valea Lupului dans le mont Făgăraş central, versant septentrional. La formation de Valea Lupului se présente comme une unité tectonique méso-crétacée indépendante par sa composition lithologique de type Cumpăna (à prédominance gneissique-micacée avec des aspects phanéroblastiques et des niveaux de migmatites opthalmitiques), par les aspects structuraux et texturaux, par l'absence des roches graphiteuses et de l'association calcaires cristallins-amphibolites. Elle représente un fragment de la nappe surgétique de Făgăraş (Kräutner, 1983; Hann, Szász, 1984) ou un lambeau de la nappe d'Argeş (Balintoni et al., 1986).

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Introducere

Ca urmare a cartărilor și studiilor efectuate pentru Harta geologică, foaia Suru, sc. 1 : 50000, au rezultat, pe lîngă o mai bună sistematizare și interpretare a cunoștințelor anterioare, și date noi de cunoaștere.

Obiectul acestei note îl constituie prezentarea unei noi entități litostratigrafice în partea nordică a Făgărașului central, situată în bazinul mediu și superior al Văii Lupului și parțial în bazinul superior al văii Mîrșa, bazine ce se află la sud de localitatea Mîrșa.

Geologia regională

Evoluția geologică a arealului Valea Lupului nu poate fi desprinsă de contextul regional. Acest areal, aparținînd cristalinului de Făgăraș, se încadrează în ceea ce Ghica-Budești (1940) și Dimitrescu (1963) numeau seria de Poiana Neamțului, Arion (1970), zona cu granat și biotit, Balintoni et al. (1984), formațiunea de Suru — partea retromorfă, iar Chivu (1985), formațiunea de Poiana Neamțului-Bîlea.

În accepțiunea Balintoni et al. (1985) — privită ca o adaptare a codului Hedberg la problematica metamorfitelor — cristalinul de Făgăraș ar reprezenta un subgrup, ce face parte din grupul Negoi.

Indiferent de încadrarea sa — după criterii tectonice în pînze (pînza de Făgăraș, după Kräutner, 1983 ; Hann, Szász, 1984 ; pînza de Moldoveanu, după Balintoni et al., 1985) sau după criterii litostratigrafice — cristalinul de pe clina nordică a Făgărașului a fost privit de către cercetătorii anteriori (Ghica-Budești, 1940, Dimitrescu 1963, 1978 ; Arion 1969, 1970 ; Chivu 1970, 1980, 1985 ; Giușcă et al. 1977 ; Balintoni et al. 1984, 1985) ca fiind caracterizat prin abundența rocilor carbonatice și a amfibolitelor pe un fond de micașisturi, paragnaise și cuarțite ce admit și secvențe de roci grafitoase, precum și prin absența migmatitelor chiar și în ariile de apariție a sillimanitului. Face excepție de la această opinie, îndeobște acceptată, Pană și Ricman (date nepublicate), care consideră posibilă prezența unor gnaise oculare și în formațiunea de Suru, retromorfă.

Geologia locală

Constituția petrografică a arealului de la sud de Mîrșa este net deosebită de rocile întîlnite pe clina nordică a Făgărașului central. Fondul acestui areal este dat de micașisturi faneroblastice și paragnaise micacee cu granulație medie, oricum, în general, mai mare decît la paragnaisele formațiunii de Suru. Acest fond admite secvențe cu grosimi metrice pînă la cîțiva zeci de metri de amfibolite și gnaise amfibolice. Nota caracteristică a formațiunii de Valea Lupului o constituie însă prezența a trei nivele de migmatite oftalmitice, cu treceri marginale sau pe direcție la migmatite lineare, precum și prezența extrem de rară a calcarelor și dolomitelor cristaline.

Micașisturile faneroblastice apar sub formă de fîșii cu grosimi de zeci sau sute de metri, în alternanță cu paragnaisele micacee, împreună cu care asigură fondul acestei formațiuni. Au șistozitate marcantă, conformă cu stratificaţia, granulaţie uniformă şi sînt constituite dintr-o alternanţă de pături milimetrice micacee cu pături cuarţo-feldspatice. Biotitul este mica predominantă şi apare ca şi muscovitul în lamele mari ce asigură o şistozitate pronunţată. De obicei, biotitul este brun-roşcat sau brun-gălbui. Cuarţul dă granule milimetrice cu contururi neregulate

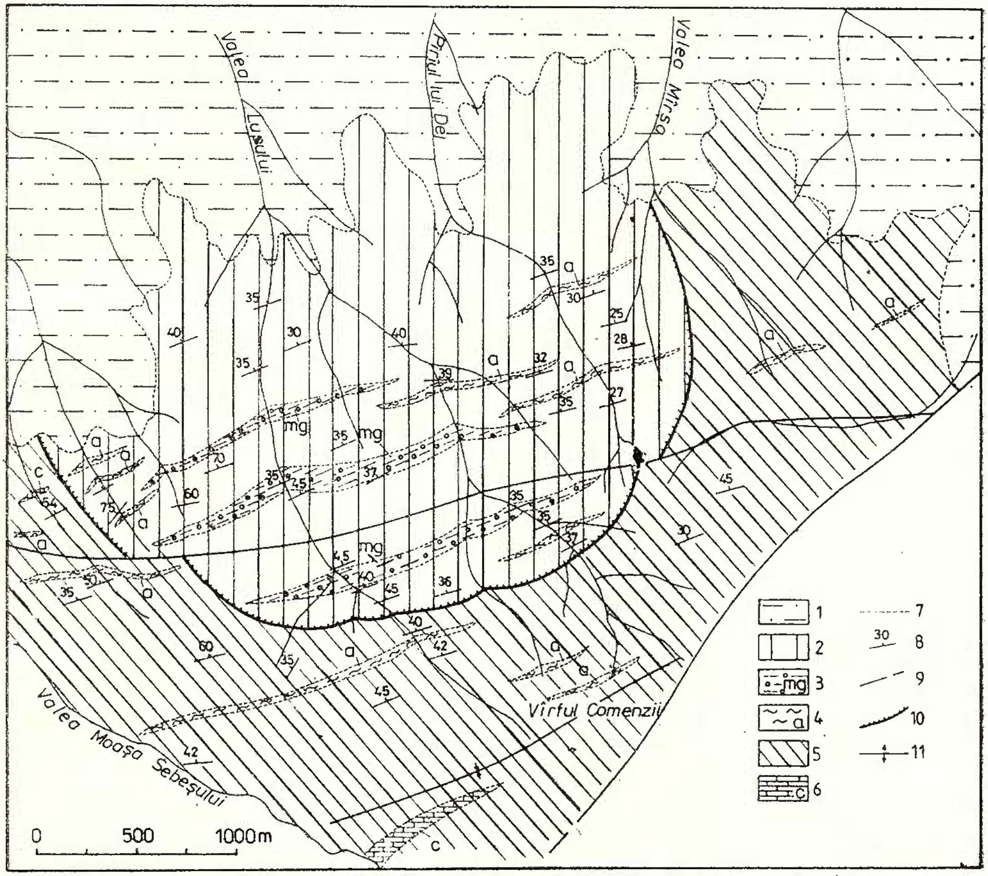

Fig. — Schiţa geologică a perimetrului Valea Lupului.

1, depozite neogene; formaţiunea de Valea Lupului: 2, paragnaise şi micaşisturi faneroblastice; 3, migmatite; 4, amfibolite; formaţiunea de Suru — retromorfă (după Balintoni et al., 1986): 5, paragnaise, micaşisturi cu granaţi ± staurolit ± disten ± sillimanit; 6, calcare; şi 4, amfibolite; 7, limită geologică; 8, poziţia foliaţiei; 9, fractură; 10, şariaj; 11, anticlinal.

Geological sketch of the Valea Lupului area.

1, Neogene deposits; Valea Lupului Formation; 2, phaneroblastic paragneisses and micaschists; 3, migmatites; 4, amphibolites; Suru retromorphic Formation (according to Balintoni et al., 1986): 5, paragneisses, garnet ± staurolite ± kyanite ± sillimanite micaschists; 6, limestones; and 4, amphibolites; 7, geologic boundary; 8, foliation position; 9, fracture; 10, overthrust; 11, anticline.

şi cu extincţie ondulatorie, uneori accentuată. Plagioclazul este un oligoclaz cu 15 — 30% An şi are o participare procentuală mai redusă decît a cuarţului. Uneori, în apropierea migmatitelor, atît la micaşisturi, cît şi la paragnaise se observă o tendinţă de creştere porfiroblastică a plagioclazului, care în acest caz include şi granulele din jur. Micaşisturile mai conţin în cantităţi cu totul neînsemnate : granaţi, staurolit, disten, zircon, turmalină, iar cînd sînt în apropierea amfibolitelor şi hornblendă. Este interesant modul de prezentare al granaţilor, care frecvent sînt sparţi (cu spărturi adesea paralele cu planul de şistozitate), rotiţi (cu „S" caracteristice) şi cu incluziuni de cuarţ, feldspat, mai rar de zircon.

Paragnaisele micacee au o structură, textură şi compoziţie mineralogică asemănătoare cu aceea a micaşisturilor, distingîndu-se de acestea printr-o participare mai mare a feldspatului şi cuarţului, ceea ce imprimă şi rocii o granulaţie medie, dar uniformă. Uneori — mai ales în apropierea amfibolitelor, dar şi în masa paragnaiselor micacee — apar benzi de paragnaise fin biotitice, la care granulaţia este măruntă, datorită faptului că biotitul se prezintă sub formă de mici lamele dispuse cu axul lung în planul de şistozitate. Au şistozitatea marcantă, conformă cu stratificaţia şi uneori sînt rubanate datorită alternanţei de strătuleţe mai bogate în biotit, cu strătuleţe bogate în feldspat şi cuarţ. Biotitul şi hornblenda, care apar în vecinătatea amfibolitelor, şi mai ales granatul prezintă evidenţe sigure de cataclazare. La biotit se constată, uneori, şi o deferizare pe planele de clivaj. Considerăm responsabile de aceste aspecte mişcările tectonice care au provocat apariţia unei fracturi ce secţionează paragnaisele biotitice.

Amfibolitele şi gnaisele amfibolitice apar sub forma unor benzi subţiri ce se menţin pe direcţie pe cîteva sute de metri, pînă la cel mult 1 km. Sînt constituite în principal din hornblendă verde şi epidot, clinozoizit, alături de care, într-o proporţie redusă, apar : plagioclaz (mai frecvent un andezin cu 30—35% An), biotit şi uneori granat. Este interesant că unele benzi de amfibolite prezintă o granulaţie medie spre mare, pe cînd altele sînt mărunt cristalizate, dar în ambele situaţii se remarcă o uniformitate a granulaţiei (nu se întîlnesc porfiroblaste).

Migmatitele oftalmitice constituie trei benzi cu grosimi aparente de la zeci de metri la 200 m şi o dezvoltare pe direcţie pînă la 2 km. Cum, în general, megablastele de microclin au diametrul cuprins între 0,5 şi 5 cm, ni se pare mai potrivită folosirea denumirii de gnaise oculare sau migmatite oftalmitice, chiar dacă pe direcţie există o variaţie a dimensiunilor megablastelor de microclin. Textura acestor roci este evident şistoasă, iar structura este lepidogranoblastică cu aspecte porfiroblastice. Constituţia mineralogică este următoarea : cuarţ, plagioclaz, microclin, biotit, muscovit, granat, apatit, minerale opace, minerale secundare (clorit şi oxizi de fier). Alături de cuarţ, micele sînt cele mai abundente şi ele mulează, de regulă, ocelii de microclin. La toate mineralele componente ale migmatitelor se constată o mare variabilitate a dimensiunilor pentru acelaşi mineral, mult mai evidentă fiind în cazul microclinului şi uneori al plagioclazului. De asemenea, sînt evidente aspecte de segregare pe specii mineralogice.

Cuarţul este un constituent principal al rocii ce apare fie în aglomerări monominerale (cu indivizi de dimensiuni, în general, mai mari), fie inclus în microclin. Prezintă o extincţie ondulatorie slabă şi rareori are incluziuni din celelalte specii minerale.

Micele — cu predominanţă netă a biotitului, muscovitul putînd uneori să lipsească — apar în agregate cu indivizi dispuşi haotic sau cu o oarecare orientare planară, agregate care, prin dispunerea lor mai mult sau mai puţin în acelaşi plan, dau şistozitatea rocii, vizibilă în special la scară macroscopică. Lamelele de biotit sînt adesea deformate (îndoite, fragmentate, cu margini franjurate etc.) pe cînd muscovitul (atunci cînd apare) prezintă contururi rectilinii slab deformate. În situaţiile cînd muscovitul este inclus în cristalele de plagioclaz are o dispoziţie întîmplătoare.

Microclinul se prezintă sub formă de megablaste heterogene, care atată o blasteză sindeformaţională în sensul lui Reinhard (1906) şi Balintoni (1975). Destul de des, megablastele sînt rupte în bucăţi, spaţiile dintre ele fiind umplute cu cristale tot de microclin, dar de dimensiuni reduse. Maclele în grătar sînt evidente şi adesea curbate, iar unii indivizi apar cu structuri pertitice. La megablaste se observă o sericitizare, care nu este însă prezentă la cristalele de microclin din masa rocii. Megablastele de microclin includ poikilitic plagioclaz, cuarţ, zircon etc., astfel încît zone întregi din cadrul lor au aspect de mozaic. Interesant ni se pare şi cazul unor incluziuni în microclin de plagioclaz, care la rîndul lui include muscovit. Acest fapt ne sugerează formarea pe cale migmatică a microclinului şi deci existenţa unor procese metasomatice. Balintoni (1975) consideră că microclinul migmatitelor din Făgăraş şi-a refăcut reţeaua în mod repetat după sfărîmări aproape complete, aspectele sale structurale excluzînd ipoteza cristalizării dintr-o topitură sau preexistenţa în metasedimente.

Plagioclazul (oligoclaz) se prezintă fie sub formă de porfiroblaste cu dimensiunile asemănătoare granulelor de cuarţ, fie sub formă de aglomerări de indivizi mărunţi cu contururi cvasipoligonale. Porfiroblastele de plagioclaz includ adesea plagioclaz mărunt, cuarţ şi muscovit, iar marginal pot fi sericitizate.

Granatul, atunci cînd apare, prezintă deformări şi cataclazări. Cloritul apare pe seama biotitului şi granatului.

După Balintoni (1975), în formarea migmatitelor din Făgăraş, succesiunea fenomenelor în raport cu deformarea este următoarea: într-o primă etapă s-au distrus structurile preexistente şi s-au segregat constituenţii, martor rămînînd granatul; în etapa următoare a avut loc un aport de substanţă care a determinat blasteza microclinului, sericitizarea plagioclazului şi recristalizarea biotitului; într-o ultimă etapă, post-deformaţională, cuarţul recristalizează total şi pot să apară structuri mirmekitice.

Migmatitele lineare apar, cu totul subordonat, în unele porţiuni periferice ale corpurilor de migmatite oculare. Ele nu alcătuiesc corpuri de sine stătătoare, comportîndu-se mai degrabă ca nişte particularităţi structurale ale migmatitelor la care microclinul nu a reuşit să formeze oceli.

Problematica generală a metamorfismului şi apartenenţa formaţiunii de Valea Lupului la cristalinul grupului de Sebeş-Lotru (Cumpăna)

Prin modul de prezentare a rocilor predominant micacee, prin aspectele structurale şi texturale şi mai ales prin prezenţa migmatitelor şi lipsa rocilor carbonatice şi a celor grafitoase, formaţiunea de Valea Lupului aparţine cristalinului multigrupului de Sebeş-Lotru, grupul de Cumpăna, în sensul Hann, Balintoni (1988). Condiţiile metamorfice de temperatură şi presiunea sînt cel puţin ale izogradului cu disten, dată fiind prezenţa migmatitelor. Cum însă mineralele tipomorfe identificate pînă în prezent sînt doar almandinul, staurolitul, distenul şi feldspatul potasic, nu putem să facem mai multe precizări. Menţionăm că în metamorfitele grupului Cumpăna (vezi Harta geologică a R. S. România, 1 : 50 000, foaia Cumpăna) sillimanitul apare foarte rar, dar este totuşi prezent (Dimitrescu et al., 1985).

Studiul secţiunilor subţiri, ca şi observaţiile de teren duc la recunoaşterea şi aici, ca de altfel în toate terenurile metamorfice proterozoice carpatice, aşa cum precizează Balintoni et al., 1985, a trei evenimente metamorfice prealpine. Primul eveniment are caracter prograd şi poate fi considerat cu certitudine proterozoic şi se înscrie în condiţiile faciesului amfibolitelor cu almandin (cu presiuni indicate deocamdată doar de prezenţa staurolitului şi a distenului), iar ultimul are caracter retrograd (în formaţiunea de Valea Lupului, fiind mai puţin evident, doar, uneori, slabe cloritizări ale biotitului, hornblendei şi granatului). Despre al doilea eveniment, care a fost legat de orogeneza hercinică, posibil şi alpină, se pot afirma mai puţine lucruri, Balintoni (1975) legînd migmatizarea de acest al doilea eveniment, fără însă a-i stabili poziţia în timp.

Probleme de petrogeneză. Din punct de vedere petrogenetic constatăm deosebiri importante între cristalinul de Făgăraş (formaţiunea de Suru) şi formaţiunea de Valea Lupului. Astfel, pentru cristalinul de Făgăraş (formaţiunea de Suru) materialul premetamorfic are caracter predominant terigen cu sedimente mature, bine diferenţiate, sedimentate în condiţii de platformă stabilă şi cu o mare răspîndire a depozitelor carbonatice, pe cînd la cristalinul formaţiunii de Valea Lupului, rocile predominant micacee şi gnaisice indică un material premetamorf mai slab diferenţiat, depus în condiţii de agitaţie tectonică, iar migmatizarea pe arii importante indică o mobilizare şi migrare de substanţă în condiţii sinmetamorfe. Cum această formaţiune este acoperită în partea ei nordică de sedimentele neogene ale bazinului Transilvaniei şi în ansamblu nu reprezintă, în accepţiunea noastră, decît un „rest" dintr-o unitate tectonică cu o extindere mult mai mare, nu excludem posibilitatea de a fi identificate eclogite şi metaultrabazite sau martori ai acestor tipuri de roci ce constituie o particularitate a cristalinului tip Cumpăna.

Formaţiunea de Valea Lupului se individualizează şi în ceea ce priveşte retromorfismul, care este aici mult mai slab decît la cristalinul de Făgăraş, unde ajunge uneori să fie foarte pronunţat.

Elemente structurale. Ca elemente structurale în formaţiunea de Valea Lupului se recunosc, cu dificultate, trei plane S. Planele S_1 reprezintă amprentele unei stratificaţii iniţiale, ele se observă foarte greu şi numai privite în secţiuni transversale ac, cînd se evidenţiază prin diferenţe de culoare datorate compoziţiei petrografice diferite a microstratelor. Situaţia pe teren arată clar că şistozitatea iniţială de stratificaţie se observă destul de rar, ea fiind ştearsă aproape complet de şistozitatea de clivaj S_2, care o întretaie sub un unghi mic. Cele cîteva situaţii în care a fost identificată arată pentru S_1 o direcţie E-V cu înclinări nordice mici, sub 30°. Şistozitatea de clivaj S_2, bine exprimată, este orientată ENE—VSV, cu căderi nordice sub 45°. Sincron cu şistozitatea S_2 se formează o serie de cute vizibile pe teren datorită prezenţei unor „litoane" anterioare şistozităţii, care au fost cutate „sinşistos" sau budinate între planele de şistozitate. Au fost întîlnite şi cîteva deformări plicative tardive, care cutează şistozitatea S_2 şi care, în general, sînt cute cu rază de curbură mare şi cu alunecare concentrică.

În secţiuni subţiri se observă uneori şi o foliaţie S_3, foarte slab exprimată pe teren, pe cînd în cristalinul de Făgăraş din vecinătate S_3 este mult mai bine exprimat.

Tectonica disjunctivă se face remarcată prin prezenţa unei importante falii cvasidirecţionale, cît şi a fisuraţiilor, mai bine exprimată fiind fisuraţia ac.

Falia cvasidirecţională normală antitectică taie atît formaţiunea de Valea Lupului, cît şi cristalinul de Făgăraş, are înclinări sudice în jur de 70° şi determină ridicarea compartimentului sudic. În formaţiunea de Valea Lupului falia este marcată de prezenţa unei argile şi brecii de falie cu o grosime în jur de 2 m şi o zonă de milonitizare între 30 şi 50 m, pe cînd în cristalinul de Făgăraş efectele ei sînt mai restrînse.

Dacă avem în vedere raporturile de încălecare între formaţiunea de Valea Lupului şi cristalinul de Făgăraş (formaţiunea de Suru) şi faptul că falia secţionează ambele unităţi litostratigrafice, rezultă că vîrsta ei este postşariaj.

Raporturile între formaţiunea de Valea Lupului şi cristalinul de Făgăraş sînt de încălecare, formaţiunea de Valea Lupului fiind şariată peste cristalinul de Făgăraş, şi anume peste ceea ce Ghica-Budeşti (1940) numea seria de Poiana Neamţului, Balintoni et al (1984), formaţiunea de Suru retromorfă, iar Chivu (1985), seria de Poiana Neamţului-Bîlea.

Prin constituţia sa litologică de tip Cumpăna, formaţiunea de Valea Lupului — ca unitate tectonică mezocretacică bine individualizată — ar putea aparţine, dacă privim în general, fie pînzei de Făgăraş în sensul Kräutner (1983), Hann şi Szász (1984), fie pînzei de Argeş, în sensul Balintoni et al. (1985). Pînza de Argeş, începînd de la cotul Dîmboviţei spre vest, ar sta peste pînza de Moldoveanu (constituită din formaţiuni ale grupului de Făgăraş). Din cadrul grupului Cumpăna, formaţiunea de Măgura Cîinenilor prezintă cele mai multe similitudini litologice, structurale şi texturale cu formaţiunea de Valea Lupului.

Deoarece apare doar un fragment al unităţii tectonice Valea Lupului de sub depozitele neogene ale bazinului Transilvaniei, iar relaţiile tectonice nu pot fi urmărite pe direcţie, este dificil de a urmări cu mai multă precizie şi a stabili raporturile ei cu pînza de Moldoveanu. Totuşi se constată prezenţa unui segment, sub Vîrful Comenzii, în zona de izvoare a Văii Lupului, în lungul căruia raporturile de încălecare sînt evidente :

TABEL

Compoziția chimică a rocilor formațiunii de Valea Lupului

	Formațiunea de Valea Lupului					Formațiunea de Suru		Formațiunea de Măgura Clinenilor			Media în litosferă (Lucițki, 1949)
	micașisturi		paragnaise	migmatite		după Gridan, Dumitrașcu, 1985					
	nr. probă 843	nr. probă 1133	nr. probă 840	nr. probă 875	nr. probă 1143	paragnaise 35 analize	micașisturi 28 analize	paragnaise 11 analize	micașisturi 7 analize	migmatite 4 analize	gnaise migmatice
SiO_2	61,90	45,80	71,82	71,78	70,62	63,49	60,61	66,49	60,11	75,01	72,44
TiO_2	0,82	2,00	0,36	0,36	0,53	0,98	1,11	0,63	1,05	—	—
Al_2O_3	19,25	16,96	15,60	13,73	15,06	15,10	17,52	16,20	19,76	12,38	12,31
Fe_2O_3	2,65	6,00	1,07	1,73	1,74	3,75	3,21	0,35	0,63	0,30	3,83
FeO	5,04	8,93	3,45	2,07	1,59	3,25	3,55	3,93	5,88	1,500	0,81
MnO	0,05	0,19	0,11	0,06	0,04	0,14	0,15	0,11	0,16	0,04	—
MgO	2,35	3,70	1,00	0,90	1,90	2,76	2,93	2,29	2,67	0,85	0,27
CaO	0,81	7,71	1,21	1,01	1,37	1,20	1,14	2,51	1,70	1,56	1,59
K_2O	2,91	3,05	0,37	4,37	3,72	3,56	3,54	3,19	3,66	4,46	4,31
Na_2O	1,11	0,52	4,57	2,85	2,75	2,32	1,57	2,95	1,97	3,43	3,63
P_2O_5	0,09	0,37	0,06	0,10	0,14	0,18	0,15	0,15	0,12	0,03	
H_2O^+	2,36	0,79	0,00	0,03	1,06	2,34	3,04	0,75	1,96	0,31	—
H_2O^-	0,15	—	0,16	0,15	—	—	—	—	—	0,22	
CO_2	0,18	3,62	0,15	0,66	0,28	—	—	—			
S	0,00	0,16	0,04	0,00	0,04					0,04	
Total	99,94	99,95	99,97	99,80	99,97	99,85	98,52	99,80	99,67	100,03	99,19

	Elemente minore în ppm										
Pb	32,00	2,5	24,00	26,00	8,5	16,90	14,16	18,41	14,29	26,5	
Cu	28,00	30,00	9,00	15,00	7,5	31,58	30,74	21,05	33,29	11,00	
Zn	46,00	160,00	80,00	< 30,00	40,00	82,18	91,54	50,73	93,29	39,75	
Sn	2,5	2,00	4,00	7,5	4,00	3,49	3,74	4,09	4,58	4,75	
Ga	26,00	18,00	26,00	23,00	17,00	21,92	24,62	18,46	23,43	17,50	
Ni	46,00	60,00	4,50	10,00	15,00	47,66	53,83	28,73	44,15	4,88	
Co	23,00	30,00	3,00	6,00	5,00	17,95	18,53	11,00	17,72	2,25	
Cr	100,00	115,00	< 10,00	23,00	24,00	83,89	103,37	51,96	100,86	4,38	
V	150,00	220,00	8,0	70,00	40,00	112,15	130,19	84,73	128,58	14,25	
Sc	23,00	22,00	7,00	12,00	10,00	17,00	19,04	12,78	19,29	4,13	
Nb	21,00	< 10,00	160,00	22,00	18,00	12,20	12,86	12,82	13,86	15,35	
Mo	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	< 2,00	
Zr	230,00	200,00	< 1000,00	220,00	200,00	344,29	280,54	223,19	222,15	146,25	
Be	4,00	1,80	8,00	3,80	1,80	3,26	3,55	2,96	3,65	2,70	
B	150,00	< 30,00	< 30,00	< 30,00	< 30,00	130,40	127,08	38,19	51,72	< 30	
Vb	2,10	1,80	14,00	2,80	2,10	3,08	3,10	2,58	2,50	4,53	
V	46,00	30,00	145,00	33,00	25,90	33,52	32,08	32,28	34,00	39,00	
La	36,00	30,00	135,00	< 30,00	< 30,00	45,63	44,61	40,46	55,72	35,25	
Sr	170,00	260,00	18,00	130,00	62,00	145,46	128,43	157,91	180,72	62,00	
Ba	590,00	165,00	115,00	750,00	620,00	743,29	726,79	772,73	921,34	385,00	

Analiști: Dănescu A., Șerbănescu A.

formațiunea de Valea Lupului stă peste pînza de Moldoveanu, respectiv peste seria de Poiana Neamțului (Ghica-Budești, 1940), seria de Poiana Neamțului-Bîlea (Chivu, 1985), formațiunea de Suru retromorfă (Balintoni et al., 1984, 1985).

Date chimice. În tabel prezentăm trei analize chimice pe probe recoltate din formațiunea de Valea Lupului: un paragnais, un micașist și un migmatit, precum și, pentru comparație, medii ale analizelor efectuate pe aceleași tipuri de roci din formațiunea de Suru (care nu are migmatite) și formațiunea de Măgura Cîinenilor (care are migmatite). Inserăm în tabel și media în litosferă pentru gnaise migmatice, după Lucițki (1949).

Din compararea chimismului major, pe tipuri de roci, rezultă similitudini clare între formațiunile de Valea Lupului și de Măgura Cîinenilor (ambele fiind alcătuite din metamorfite caracteristice grupului de Cumpăna) și unele diferențieri între acestea și formațiunea de Suru (alcătuite din metamorfite ale cristalinului de tip Făgăraș).

Astfel, luînd în considerație silicea și într-o oarecare măsură și ceilalți oxizi principali, Al_2O_3, FeO, MnO etc., se constată o participare procentuală a lor sensibil egală la formațiunile de Măgura Cîinenilor și Valea Lupului și mai mare la aceste două formațiuni față de formațiunea de Suru, pe tipuri de roci comparabile (paragnaise, micașisturi). Deosebirile sînt mai puțin semnificative în ceea ce privește comportamentul oxizilor MnO, MgO și CaO, dar destul de nete dacă avem în vedere creșterea accentuată a Fe_2O_3 la rocile formațiunii de Suru (creștere pusă pe seama retromorfismului). O situație anormală o prezintă conținutul de K_2O mai scăzut dintr-o analiză pe o probă de gnais a formațiunii de Valea Lupului. O posibilă explicație ar fi mobilizarea K_2O în cadrul proceselor metasomatice pentru a forma ocelii de microclin ai migmatitelor din vecinătate.

Formațiunea de Suru se individualizează de celelalte două prin lipsa migmatitelor (roci cu un regim geochimic deosebit), prin creșterea netă a Fe_2O_3, H_2O și a gradului de oxidare.

Compararea chimismului migmatitelor formațiunii de Măgura Cîinenilor cu cel al migmatitelor formațiunii de Valea Lupului relevă asemănări evidente. Totodată, conținuturile oxizilor sînt apropiate de media gnaiselor migmatice din litosferă dată de Lucițki (1949). Conținutul în SiO_2 ceva mai ridicat la migmatitele formațiunii de Măgura Cîinenilor, corelat cu o oarecare creștere a K_2O, denotă un grad de migmatizare mai avansat pentru zonele din care au fost recoltate probele.

Considerații finale

Formațiunea de Valea Lupului se constituie ca o unitate tectonică de sine stătătoare prin componența sa litologică de tip Cumpăna (un fond predominant gnaisic-micaceu, cu aspecte faneroblastice și cu nivele de migmatite oftalmitice), prin aspectele structurale și texturale, prin lipsa rocilor grafitoase și a asociației calcare cristaline-amfibolitite. Cea mai importantă caracteristică o constituie însă prezența migmatitelor, inexistente în tot ceea ce Dimitrescu (1978) numea seria de Făgăraş, Chivu (1985), seria de Poiana Neamţului-Bîlea, iar Balintoni et al. (1985), grupul Negoi — subgrupul de Făgăraş.

De asemenea, formaţiunea de Valea Lupului se caracterizează printr-un chimism mai degrabă comparabil cu al formaţiunii de Măgura Cîinenilor din grupul Cumpăna.

Din punct de vedere tectonic, formaţiunea de Valea Lupului ar putea reprezenta un fragment nordic al pînzei de Argeş (ce stă aici peste pînza de Moldoveanu) în sensul Balintoni et al. (1986), sau un fragment al pînzei supragetice de Făgăraş în sensul Kräutner (1983), Hann şi Szasz (1984).

A fost avansată de către H. Hann (informaţie verbală) şi ideea unei antiforme majore, orientată aproximativ est-vest, în cadrul Hărţii geologice a R.S.R., scara 1 : 50000, foaia Suru, antiformă în care limita dintre formaţiunea de Şerbota şi formaţiunea de Suru este considerată normală. Conform acestei idei, preluată şi dezvoltată de D. Pană (informaţie verbală) formaţiunea de Şerbota s-ar afla în axul antiformei, pe flancuri fiind formaţiunea de Suru, peste care ar sta pe flancul sudic, cu o poziţie incertă (şariaj sau nu?), formaţiunea de Măgura Cîinenilor. Dacă se va accepta această antiformă — în baza unei argumentaţii convingătoare — atunci formaţiunea de Valea Lupului ar fi echivalentul, pe flancul nordic, al formaţiuni de Măgura Cîinenilor.

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NOTE ON THE PRESENCE OF THE VALEA LUPULUI FORMATION IN THE CENTRAL FĂGĂRAŞ, NORTHERN SLOPE

(Summary)

The Valea Lupului Formation consists prevailingly of micaceous paragneisses and phaneroblastic micaschists; ophtalmitic migmatites, amphibolic gneisses and amphibolites are also found in smaller amounts.

Due to its lithological constitution, especially the presence of migmatites, the structural and textural aspects, the major chemism and the metamorphism grade (kyanite isograde), the Valea Lupului Formation is comparable to the Măgura Cîinenilor Formation, thus belonging to the Sebeş-Lotru multigroup, Cumpăna group, in the sense of Hann, Balintoni (1987). It differs from the Negoi group crystalline (Suru Formation) by the lack of the graphite rocks and of the crystalline-limestone-amphibolite association, the structural and textural aspects of the rocks and a much weaker retromorphism.

The Valea Lupului Formation respresents a Mesocretaceous tectonic unit, an outlier that might belong either to the Făgăraş Supragetic Nappe in the sense of Kräutner (1983) and Hann and Szász (1984) or to the Argeş Nappe, in the sense of Balintoni et al. (1985). It overlies the Moldoveanu Nappe, the Poiana Neamţului Series respectively (in the sense of Ghika-Budeşti, 1940), the Poiana Neamţului-Bîlea Series (in the sense of Chivu, 1985), the Suru retromorphic formation (in the sense of Balintoni et al., 1984, 1985).

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

COZIA GNEISS (SOUTH CARPATHIANS): PETROGRAPHY, STRUCTURE, GENESIS[1]

by

HORST PETER HANN[2]

Metamorphic rocks. Gneisses. Paragneiss. Augen gneiss. Tectonic breccia. Metasomatism. K-feldspar. Overthrust faults. Dip faults. Tectogenesis. South Carpathians — Crystalline Getic and Supragetic Domains — Cozia Peak.

Abstract

The petrographic constitution of the massif is formed of more or less micaceous paragneisses (plagiogneisses), showing gradual passings to microblastic gneisses of blastomylonitic origin. Migmatization, the formation of the augen gneisses took place through metasomatic processes, mainly at the expense of blastomylonites. The characteristics and evolution of the main mineral constituents of paragneisses indicate the retromorphic character of both blastomylonitization and metasomatosis, but all the processes take place within the limits of the almandine amphibolite facies. The augen gneisses comprise rocks of heterogeneous characters and aspects. The formation of the feldspar within phenoblasts represents a polyphasic metasomatic process that took place during a long time interval. The description and classification of the shape, outline and position of the K-feldspar augen as well as the aspects related to the associations among the K-feldspar augen enable the understanding of the petrographic-tectonic convergences involved in the formation process of the augen gneisses. The breccias consisting of crystalline elements belong to different genetic types, most of them being subaerial or tectonic formations. The gneisses of the whole massif belong to the Cumpăna Group. The most obvious structural element is the Cozia Fault which delimits the massif northwards along a distance of over 50 km and raises the southern compartment (in the Olt Valley Zone) by over 1,000 m. The crystalline massif belongs to the Făgăraş Supragetic Unit, to the Laramian Boia Nappe respectively, and overthrusts in its western extremity, in the right slope of the Stan Valley, the Santonian-Coniacian deposits along an overthrust plane. Part of the system of transversal and directional faults from the interior of the massif is prior both to the Cozia Fault and to the Paleogene, respectively Upper Cretaceous deposits.

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Résumé

Massif de Cozia-Frunţii-Ghiţu, Carpathes Méridionales — données pétrographiques et structurales. Le fond pétrographique du massif est constitué de paragneiss (plagiogneiss) plus ou moins micacés, à passages graduels vers des aspects blastomilonitiques. La migmatisation, la formation des gneiss oeillés s'est produite par des processus métasomatiques surtout au profit des blastomilonites. Les caractéristiques et les évolutions des principaux minéraux constituants des paragneiss indiquent le caractère rétromorphe de la blastomilonitisation ainsi que de la métasomatose, tous les processus se déroulant toutefois en des limites du faciès des amphibolites à almandin. Les gneiss oeillés comprennent des roches à caractères et aspects hétérogènes, mais liés au point de vue descriptif. La formation du feldspath des phénoblastes représente un processus métasomatique polyphasique, déroulé pendant un intervalle de temps large. La description et la classification de la forme, du contour et de la position des yeux de feldspath potassique ainsi que les aspects portant sur des associations d'entre les yeux de feldspath potassique permettent de comprendre les convergences pétrographique-tectoniques du processus de formation des gneiss oeillés. Les brèches constituées d'élements de cristallin appartiennent aux types génétiques différents, la plupart étant des dépôts continentaux subaériens ou bien d'origine tectonique. Les gneiss de tout le massif appartiennent au groupe de Cumpăna. L'élément structural principal est la faille de Cozia qui délimite le massif vers le nord le long de plus de 50 km et soulève le compartiment meridional (dans la zone de la vallée de l'Olt) plus de 1000 m. Le massif cristallin appartient à l'unité surgétique de Făgăraş, respectivement à la nappe laramique de Boia, et chevauche l'extrémité ouest des dépôts santoniens-coniaciens dans le versant droit de la vallée Stan, tout au long du plan de charriage. Une partie du système de failles transversales et directionnelles de l'intérieur du massif est antérieure autant à la faille de Cozia qu'aux dépôts sédimentaires paléogènes, respectivement crétacés supérieurs.

The Cozia-Frunţii-Ghiţu crystalline massif is known especially for its augen gneisses covering large areas. So far these gneisses have been often described or only mentioned.

The entire massif was first mapped by Max Reinhard (1906) and later on by Şt. Ghika-Budeşti (1958). But, after Reinhard's investigations, there existed no unitary presentation of any petrographic and structural researches concerning the whole massif.

The investigations and mappings which I carried out in view of drawing up the Geologic Map of Romania, scale 1 : 50,000, sheets Vînturariţa, Mălaia, Călimăneşti, Titeşti and Cumpăna made possible a detailed mapping of the region (Pl. I). At the same time they allowed me to tackle problems connected with the structure and petrography of the crystalline schists making up the massif, laying stress on the augen gneisses.

1. Previous Researches and Conceptions

The name of "Cozia augen gneisses" was first used by Primics (1884) who, nevertheless, did not separate it on the presented map. Sabba Ştefănescu (1884, 1886) described the augen gneisses as "porphyroids". Mrazec (1897, 1904) delimited the Cozia gneiss on the map and considered it as belonging to the crystalline schists of high crystallinity (group I). He was the first to put forward a hypothesis on its origin, according to which, it would be a protoclastic consolidated granite, its primary parallel structure being the result of the foldings that were simultaneous with the metamorphism. Max Reinhard's thesis of doctor's degree contains a detailed mineralogical and petrographical description of the Cozia gneisses which is, on the whole, valid even today, as well as petrochemical researches. The author points to the various aspects shown by the augen gneisses. He noticed the different shapes of the augen consisting of the often perthitic microcline, and the biotite, muscovite, plagioclase inclusions within it, which he considered to have formed at the end of the petrogenetic process. Reinhard also described hornfels-like "psammitic gneisses" or gneisses exhibiting cataclastic structures and shear zones. The genesis of the augen gneisses is connected to the supposed existence of a laccolith situated between the Olt and the Dîmbovița Valleys, consolidated under stress, while the surrounding schists were subsequently not assimilated by the granitic magma. Later on (1910) Reinhard stated that among the augen gneisses, conglomeratic shales, arkoses and acid lavas, subjected to thermal metamorphism, could be distinguished. Schmidt (1930) considers that the Cozia gneisses are rocks formed as a result of an "injection metamorphism" which was followed by a regional, catazonal metamorphism, therefore initially also of magmatic origin.

Streckeisen (1934) also admits that the Cozia gneiss represents an important injection zone, marked by a pneumatolythic influence, constituting an entity together with the Cumpăna gneisses, from which it was later on tectonically separated.

Ghika-Budești (1940, 1942) introduced the name of "Cumpăna Series", within which he separated also the Cozia Zone. The genesis of the Cozia gneisses is explained as a consequence of the conditions imposed by the ultrametamorphism, palyngenesis, with the participation of some metasomatic processes, the whole phenomenon taking place in a strong stress zone. Thus he assigns the augen gneisses to mylonites; he also notices paragneisses exhibiting hornfels structure. In 1958 Ghika-Budești presented a map of the entire massif and the Loviștea Depression on scale 1 : 100, 000. Besides augen and banded gneisses, he separates essexitic paragneisses, mesocatazonal paragneisses, amphibolites, biotite paragneisses and micaschists. He also describes various aspects concerning the breccias from the Cozia massif, considering them deposits of continental origin and not tectonically formed.

Dessila-Codarcea and Dragomir (1960) mapped an area extending between the Olt River and Valea lui Stan on scale 1 : 10,000, pointing out the presence of the breccias south of the Năruțiu Peak. They also noticed the grading of the initial paragneisses to augen gneisses as a result of some injections of quartz-feldspathic material. Pitulea et al. (1960) drew up a map on the same scale, covering an area situated between the Olt Valley to the west and the Argeș Valley to the east. They distinguished a paraschist complex and the migmatite complex, the genesis of the latter complex being explained through the injection of a quartz-feldspathic material originating from an anatectic zone. Also in 1960 Dimitrescu presented a map of the Argeș Valley basin, separating among other rocks, the augen gneisses and the "clastic dikes" — veins filled with breccias from crystalline elements. He admits the contribution of feldspathic material in the formation of the augen gneisses, originating through metasomatosis within the paragneiss mass.

Pitulea and Arion (1966) prospected the Cozia gneiss, mapping the area between the Ghițu Peak and the Cheia Valley, in the Vîlsan Valley basin, on scale 1 : 10,000, while Arion et al. (1966) mapped the area extending between the Argeș and Topolog Valleys. The map on scale 1 : 200,000, Pitești Sheet, which comprises also the Cozia massif, appears in 1968, the augen gneisses being represented as metablastic migmatites. Dimofte (1966) presented new data on the petrography and areal extension of the Cozia augen gneisses from a zone situated in the eastern vicinity of the Olt Valley. Arion and Ignat (1970) discussed in detail the formation of the migmatites from the southern slope of the Făgăraș Mountains. As regards the augen gneisses, they pointed out two main aspects, namely the synkinematic augen, conformable with the schistosity, and the postkinematic ones, in oblique position; all the aspects formed metasomatically, by the contribution of alkali and silica rich solutions, generated by complex metamorphism and ultrametamorphism processes. They also underlined the fact that the formation of the augen gneisses was preceded by cataclasation, lamination and even retromorphism processes (at the level of the chlorite isograde). Quoting Lemne, Pitulea (1970), Arion and Ignat (1970) point to the high radiometric values due to the The concentrations in the augen gneisses as compared to the paragneisses, inferring the formation of augen gneisses with the contribution of metasomatic feldspathic material. In his thesis of doctor's degree maintained in 1970 and published in 1975, which refers also to the migmatites from the Cozia Mountains, Balintoni explained the formation of the augen gneisses, showing that the megablastesis took place syntectonically in the intense shearing zone, in solid state, without the impact of a silicate melting; the material supply was metasomatic, achieved through water. He demonstrates the retromorphic character of the rocks (tectonites) from the migmatized spaces with respect to the parageneses of the host metamorphics. Migmatites are therefore located in areas of maximal synmetamorphic deformation, the inter- and intragranular cohesions within the respective rocks being destroyed and thus enabled the access of the substance. The sheets 1 : 50,000 Călimănești (Hann, Schuster, in Popescu et al., 1977), Mălaia (Savu et al., 1977) and Vînturarița (Olănești), published in 1978 (Hann, in Lupu et al., 1978) include the map of the massif between Valea lui Stan and the Cozia Peak. Tran (1979) presents a detailed and thorough study of the Cozia gneisses in the area delimited by the Olt River and Valea lui Stan, which will be farther discussed in connection with the elaboration of the thesis of doctor's degree. The paper contains precise mineralogical data along with roentgenographic studies or by means of infrared absorbtion spectra, petrographical and petrochemical studies that demonstrate the metasomatic origin of the augen gneisses. The author thinks that the formation of the augen gneisses is preceded by a dynamic metamorphism which gave rise to retromorphic blastomylonites, while the paragenetic assemblages indicate the thermodynamic conditions under which the metasomatosis took place, namely those corresponding to the epidote-amphibolite facies, therefore the metasomatosis has also a retromorphic character. The description, classification and genetic interpretation of the breccias from the Cozia massif were presented in the paper published by Hann and Szász (1981). The mapping of the massif was achieved in the sheets 1 : 50, 000 Titeşti (Ştefănescu et al., 1982) and Cumpăna (Dimitrescu et al., 1985). In recent years prospection works on scale 1 : 5,000, which also provided new data, have been carried out by Panaite and Panaite (1981) between the Topolog and Baiaşu Valleys, by Pitulea et al. (1984) in the Vîlsan Gorges, Pitulea et al. (1985) between the Argeş an Vîlsan Valleys, Badea et al. (1985) in the Perişani area and Stoica et al. (1986) between the Argeş, the Topolog Valley and Perişani.

A recent study made by Dimofte (1984) demostrates, based on geochemical data (distribution of minor elements), the metasomatic origin of the microcline porphyroblasts. Taking into account the minor elements, the author also observes the intensity of deformations that accompanied metamorphism during the formation of the augen gneisses.

2. Petrographic Constitution

2.1. *Petrographic Background*

The petrographic background of the massif consists of biotite or biotite and muscovite paragneisses (plagiogneisses) which are in places more micaceous, rarely micaschists, within which various migmatites develop, the ophtalmitic ones, therefore the augen gneisses, prevailing. The mass of these rocks often shows gradings to sequences exhibiting a microgranoblastic texture, which lends a hornfels-like aspect to the rocks. In fact these rocks show a blastomylonitic character, within them developing prevailingly the K-feldspar phenoblasts.

We have also found amphibolites, in places with garnets (Argeş Gorges), very rarely white quartz-feldspathic gneisses (north of the Ghiţu Peak) and in a single place (Stan Valley, right tributary of the Argeş River) crystalline limestones which are rather tectonized, but contain diopside and zoisite.

It is worth mentioning from the very beginning that the unmigmatized paragneisses do not actually contain K-feldspar. But the characteristics and evolution of the plagioclases, of the quartz and micas from these rocks have various significances. Tran (1979) made important contributions to these studies. It is found out that plagioclase belongs both to a first generation (oligoclase, 24% An) which prevails in the unmigmatized initial paragneisses, and to a second generation exhibiting a more acid chemism (oligoclase-albite, 5–12 % An) which is recrystallized in the blastomylonite groundmass, partially replacing the oligoclase porphyroblasts depending on the intensity of the retromorphic processes. During the blastomylonitization the broken plagioclases, the peripheric parts of porphyroblasts recrystallize into an oligoclase-albite when the potassium metasomatosis occurs, both plagioclases are replaced or form inclusions within the K-feldspar mass.

Quartz also shows an interesting evolution. It either forms in a first generation elongated xenoblasts intergrown with the other minerals or appears as inclusions within garnet and staurolite. The ground and recrystallized quartz within blastomylonites belongs to a later generation.

In addition to the above-mentioned plagioclase with small biotite crystals, it forms the microgranoblastic texture. Veinlets consisting of monomineral aggregates piercing the blastomylonites indicate the presence of a subsequent phase in the quartz formation. The quartz within Alpine mylonites has recently formed, being usually accompanied also by chlorite.

Biotite forms in a first generation large crystals, sometimes bearing apatite and zircon inclusions, exhibiting an intense dark brown-reddish pleochroism, such as the titanium-rich biotites. The refraction indices (determined by M. Bălan — oral communication) conform the optical data as they range between 1. 657—1.662. Biotite within blastomylonites forms small, recrystallized brown-yellowish or brown-greenish lamellae, associated with sphene and epidote. When the recrystallization persists, biotite of the second generation may form more largely crystallized lamellae (refraction indices 1.639—1.668 — M. Bălan — oral communication).

Muscovite is present in the unmigmatized paragneisses, forming lamellae usually developed along with biotite and contains apatite and zircon inclusions. Muscovite is found within blastomylonites also largely crystallized and associated with secondary biotite. Tran (1979) shows that it is in fact a ferriphengite. It has surely formed at the expense of biotite, as proved by the frequent iron oxide inclusions.

All the data presented indicate that most of the paragneisses were subjected to a blastomylonitization process of different intensities, that is to almost simultaneous cataclasations and blasteses, brought about by a dynamic metamorphism under lower thermodynamic conditions, as compared to those that controlled the formation of the initial paragneisses; it is therefore of retromorphic character, the recrystallization being isochemical; but all the processes took place within the limits of the almandine amphibolite facies, even if they occurred near the upper limits of the latter facies. On this more or less blastomylonitic background the formation of the augen gneisses took place through metasomatic processes and, at the same time, through the change in the chemical composition of the whole system. The hypothesis of the formation of the gneisses through metasomatosis and tectonization at the expense of some pararocks was supported since 1923 by Angel for the Gleinalm massif from the Eastern Alps. Tran (1979) describes several types of blastomylonites depending on the deformation intensity: (1) blastomylonites with mortar formed through the breaking and recrystallization of the marginal parts of crystals; (2) porphyroblastic blastomylonites, the rock mass being characterized by a cataclasation and intense recrystallization, in which uncataclased plagioclase porphyroblasts are still preserved; (3) blastomylonites with lens-like petroclasts, representing a still more advanced deformational stage, characterized by microlepidoblastic structures; plagioclase porphyroblasts are changed into lense along with the other relicts of the primary rock; hornfels-like ultramylonites (thermal contact metamorphism determining identical aspects) characterized by a uniform microgranoblastic structure.

2.2. *Augen Gneisses*

The augen gneisses are found under various aspects, the present paper attempting to explain this situation in the field and the phenomena which generated it. Thus I achieved some comprehensive descriptions and a systematization necessary for eliminating the multitude of intermediate aspects which would have hindered us from distinguishing between different phenomena or processes and which do not help us by their mere existence. The name of augen gneiss designates rocks of heterogeneous aspects and characteristics which are united from the descriptive-terminological point of view. The conclusion that the petrogenetic-tectonic processes determine a certain convergence of the petrographic aspects was reached after consulting the relevant literature (Heinisch, Schmidt, 1982).

The feldspar augen are usually at most 10–12 cm in diameter. There is, however, an exception, represented by an augen which is 40 cm in diameter. In this case one can macroscopically observe what can be only microscopically seen in other cases, a reason why we figured it (Fig. 1). First of all we noticed the substitution, irregular, metasomatic textures and the fine biotitic-blastomylonitic gneiss enclaves in the K-feldspar megablastic aggregate. This mineral is prevailingly white in the left half of the drawing and pink in the right one.

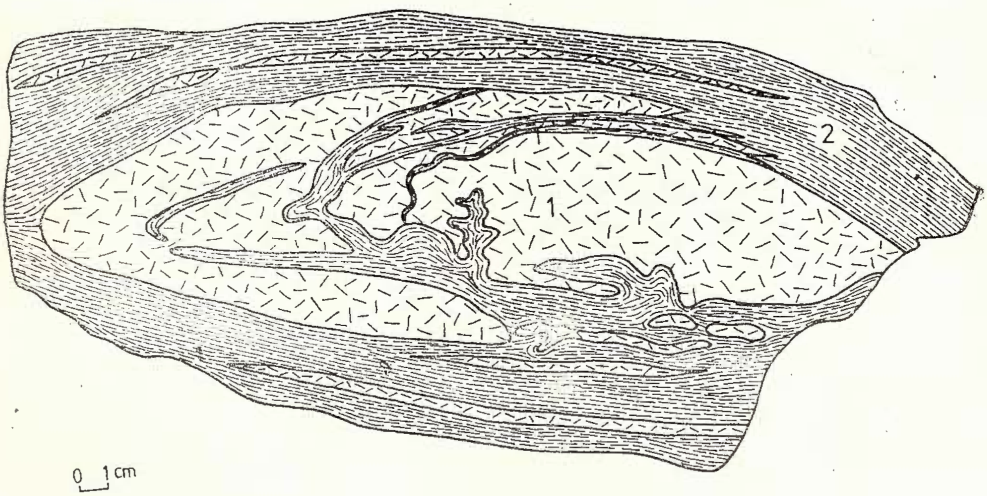

Fig. 1 — K-feldspar augen (1) uncommonly large (40 cm in diameter), prevailingly white on the left and pink on the right, hosted by biotitic schists (2) microblastic schists (blastomylonites). Metasomatic structures and inclusions of blastomylonites within the K-feldspar megablast.

The microscopic study shows in fact that the white variety corresponds to an irregular monoclinic K-feldspar (orthoclase), while the pink one is a maximal microcline. Tran (1979) presents a detailed study of the K-feldspars which I could not complete but only confirm for the whole area of augen gneisses. Orthoclase is always untwinned, with the angle of the optical axes 2V (—) = 65–70°, without exception, smaller than that of the microcline (80–90°) which constitute a distinguishing criterium between them, when the microcline is not twinned. Orthoclase is closely associated with microcline, constituting a residual monoclinal modification in the triclinical one, formed by polymorphic inversion, there existing no limit between them. Microcline often shows albite-pericline twins, sometimes with asymmetrical texture. K-feldspars often contain perthite, formed from a fresh albite or include partly replaced plagioclase, usually having an altered aspect, marginal muscovitized biotite corroded lamellae and quartz grains along the cleavages. The K-feldspar phenoblasts and porphyroblasts include also blastomylonite fragments of microgranoblastic texture during their metasomatic growth, being therefore formed prior to feldspathisation. Microcline is often associated with plagioclase crystals bearing mirmekitic quartz, a phenomenon always accompanying the potassium metasomatosis processes.

It is inferred from the above-presented facts that the feldspar formation represents a polyphasic metasomatic process taking place during a long time interval. The irregular monoclinal orthoclase formed initially then changed partly into regular and twinned triclinic microcline through polymorphic inversion. A first perthite generation forms through disintegration in the inversion phase. Afterwards the untwinned microcline formed directly as irregular, stable, triclinic modification. The substitution perthites as well as some albite veinlets formed as a result of a subsequent sodic metasomatosis of low intensity and unimportant as compared to the potassic one.

The development of the strictly mineralogical processes of the K-feldspar was parallel to that of the petrographic-structural ones, leading to the formation of the augen gneisses. In order to better understand these phenomena, some descriptive aspects as well as their genetic interpretation are presented.

2.2.1. *Shape, Outline and Position of the K-Feldspar Augen*

Feldspar augen exhibiting normal (augen) lenticular shapes (Fig. 2) or with sharp endings of the lenses (Fig. 2 b) were observed. In some other cases the pheno- and respectively megablasts may be rectangular, being idiomorphic (Fig. 2 c) or have the small sides of the square rounded (Fig. 2 d). Various other combinations are found, such as that in Figure 2e. Round augen exhibiting a more or less spherical shape (Fig. 2 f) are noticed sometimes. Augen showing compound, sinuous (Fig. 2 g) or irregular, atypical shapes, such as those in Figure 2 h, are also noticed.

It should be mentioned that not only augen of combined aspects, but also combinations among augen of various aspects may be noticed concomitantly in one and the same outcrop.

The usual lenticular shape is the most frequent and reflects the deformation type, the action of which prevailed during blastesis. The phenoblasts presenting sharp endings indicate certain modifications of the field in which the stress manifests and which determine the elongation of the spaces that constitute the "pressure shadows". The idiomorphic shape of megablasts suggests the formation of the respective crystals in the absence or under low stress, their position being frequently unconformable with respect to the foliation. Therefore crystallization took place in a phase subsequent to the formation of the other aspects. Thus the augen of combined aspects reflect the transition among the various types and directions of deformation, respectively the moments situated between the prevalence and persistence of certain types of deformational phenomena. The augen exhibiting compound, sinuous shapes formed due to the resumption of the shearings and boudinages that acted also on the already formed augen, which indicates a long process and the recurrence of the tectonic phenomena that controlled the genesis of these rocks. The irregular atypical aspects support this interpretation in the sense of practically unlimited possibilities of combinations between the undisturbed blastesis on the one hand and that manifested under stress on the other hand, taking into account at the same time also the modifications of the already existing aspects due to the resumptions characteristic of the deformational processes.

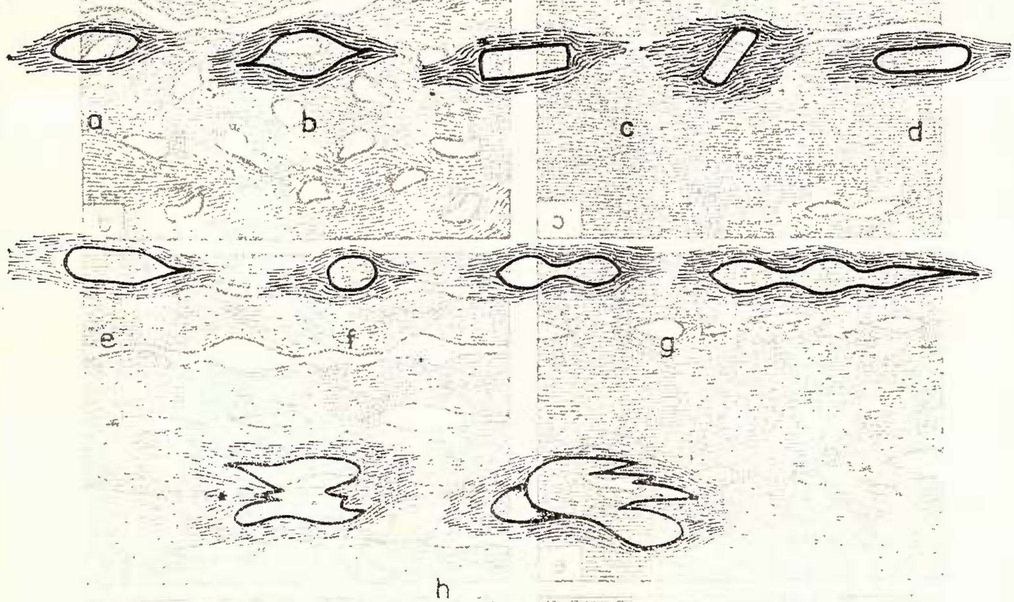

Fig. 2 — Shape, outline and position of the K-feldspar augen : a, normal lens-shaped augen ; b, the lens endings are sharp ; c, idiomorphic, sometimes unconformably grown megablast ; d, idiomorphic megablast with the small sides of the rectangle rounded ; e, combinations between various aspects ; f, round augen ; g, augen showing compound, sinuous shapes ; h, augen showing atypical, irregular shapes.

As regards the position of the K-feldspar augen, it is noticed that they may be conformable, unconformable, which means in many cases that they are postkinematic or in various rotation stages in respect of the two basic positions.

The contact between phenoblasts and the surrounding rocks is in most cases sharp. But sometimes it is diffuse, probably when the intensity of the feldspathisation was great, as it is only found where the rock s packed with augen and feldspathic bands.
i

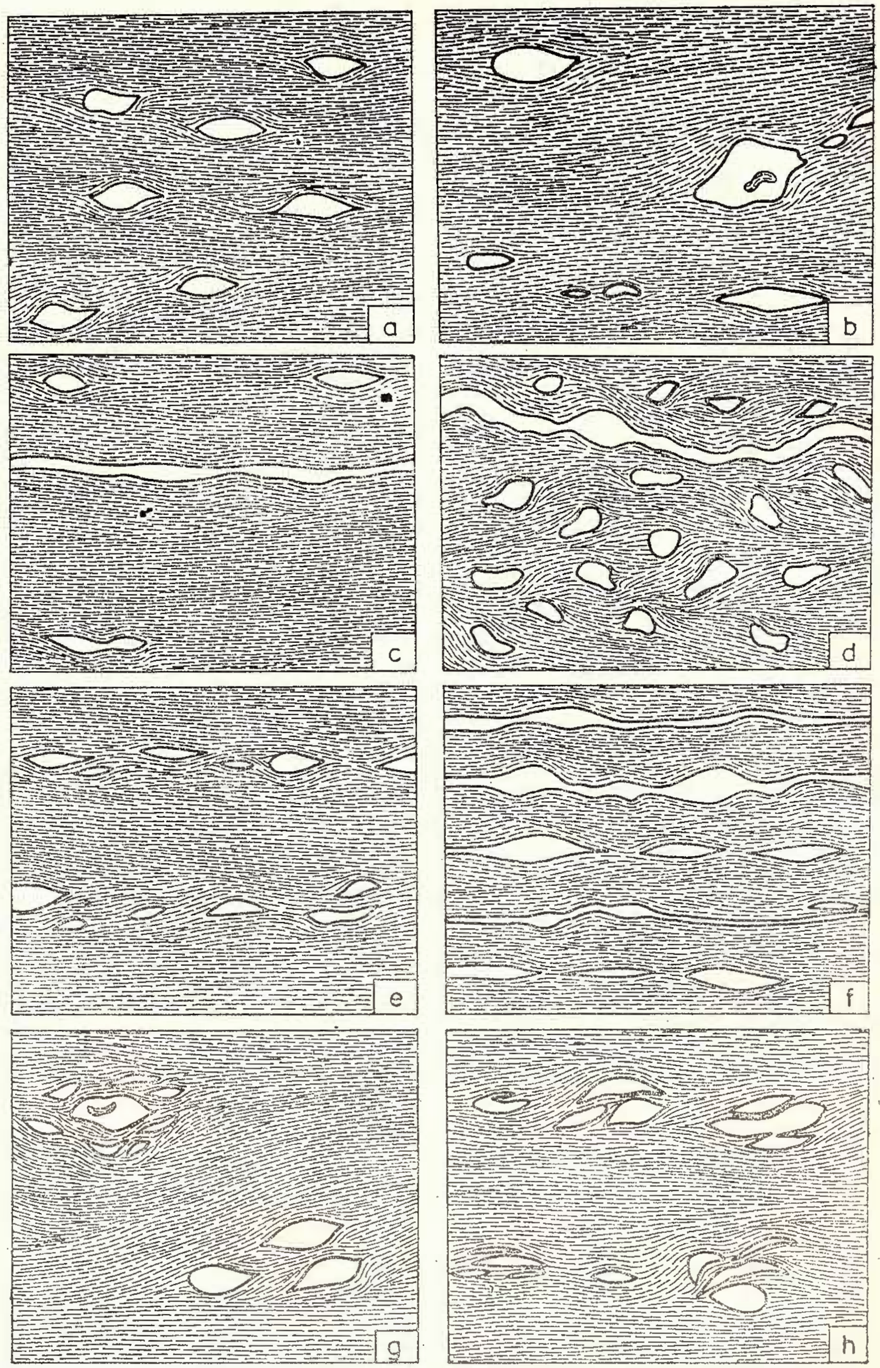

Fig. 3

2.2.2. *Aspects Related to the Associations among the K-Feldspar Augen*

a) The feldspar augen may be distributed equally and uniformly, unequally and ununiformly or may be isolated and rare in the rock mass (Fig. 3 a, b, c), function of the metasomatosis intensity, while the deformations remained within the usual limits. When an increase of the amount of movement takes place, the augen appear rotated (Fig. 3 d).

b) The augen gneisses are represented quite often in fact by the occurrence of some conformable levels, separated by unmigmatized sequences, along which there are feldspar augen (Fig. 3 e) which often detach also from feldspathic bands (Fig. 3 f). These aspects can be explained on the one hand by the way in which the infiltration of the feldspathic material and the crystallization of phenoblasts took place along some more accessible levels, of low pressure gradients, and on the other hand it is obvious that the same augen aspects result from the mechanical segmentation of some previously formed bands.

c) A third characteristic aspect is represented by the agglomeration of phenoblasts (Fig. 3 g, h). Their formation may be explained both by the concentration of some blastesis centres in the conditions of a greater material supply in areas which are favourable also in respect of the distribution of low pressure gradients zones and by the tectonic fragmentation of some large-sized, previously formed augen.

It follows from the above statements that the petrogenetic process of the augen gneiss formation reveals convergence phenomena and takes place in a long time interval involving resumptions, overlappings and changes in the physico-chemical and thermo-dynamical conditions. Blastesis or deformation prevailed in certain periods, one of these two processes being always present. At the same time alkaline metasomatosis manifested in an open system (otherwise pegmatitic processes occur in this way), water being the main component of the solutions that brought about the feldspathisation.

Recently Mehnert and Busch (1985) have studied by electron microprobe analysis the distribution of the Ba content in K-feldspar megacrysts in granites, migmatites and augen-gneisses from various European crystalline massifs. They noticed significant similarities between the K-feldspar megacrysts in granites and those in augen-gneisses in respect of the zonal distribution of the Ba content in the centre of the crystals. The marginal parts reflect deformations and recrystallizations caused by some metamorphic events. These authors state that these augen gneisses may originate from granites. The following question arises : if we made such analyses also on the K-feldspars from the Cozia augen gneiss and obtained the same result, could we interpret it in the same way?

Fig. 3 — Aspects related to the associations between the K-feldspar augen : feldspar augen approximately equally and uniformly distributed. (a) unequally and ununiformly (b), isolated and rare (c) or rotated (d). K-feldspar augen located along some levels separated by unmigmatized sequences (e) or detached from feldspathic bands (f). Agglomerations of K-feldspar augen (g, h).

Based on a mainly geochemical study (major elements), Heinisch and Schmidt (1982) conclude that a large part of the augen gneisses developed in the old crystalline of the Eastern Alps represent metavolcanics and did not likely form through the metamorphosis of some plutonic rocks. The participation of the metasomatic processes is therefore excluded even in an area (Gleinalpe) where Angel (1923) made a classical description of these phenomena.

The above-mentioned rocks are based on detailed studies, analyses with a deep, but narrow investigation spectrum, and less on the understanding of the relationships among minerals (which are more difficult to analyse either chemically or by electron microprobe analysis); at the same time, they do not take into account the analysis of all the petrographic and structural data, the great variety of aspects which is a metasomatosis characteristic.

However, we should bear in mind that the validity of the hypotheses discussed is relative as all the phenomena and processes explained by these hypotheses cannot always directly represent the geological reality. The hypotheses and theories refer only to what we think we know about this reality, thus reflecting only our present possibilities of understanding.

2.3. *Breccias and Mylonites*

Breccias consisting of crystalline elements are often found in the interior of the massif or at its periphery, consisting of crystalline elements, the origin of which has been variously interpreted along time. Studying this problem, Hann and Szász (1981) demonstrate that there exist four distinct genetic types: (1) fault breccias formed as a result of the friction between two rock packets (e.g. along the Cozia Fault); (2) breccias formed by the filling of some fissures or cracks (breccia veinlets) with fragments formed through the disintegration of gneisses due to the physical agents; (3) breccias representing subaerial continental sedimentary deposits, formed during some strong uplifting movements through the disintegration of the crystalline basement and the accumulation of the paleorelief at the base of the slopes; (4) breccias of tectonic nature (Năruţiu breccia) which show large and significant developments at the southern margin of the massif and can be followed intermittently from the Lotrişor Valley (right tributary of the Olt River) to the Argeş Valley. The last two categories of breccias are surely pre-Senonian in age, being transgressively overlain by the Senonian sedimentary deposits.

The rocks presenting fluxion structure (sensu Berza et al., 1984) are found throughout the massif, accompanying the various transverse or strike faults. Protomylonites, mylonites and ultramylonites, in some cases phylonites, are noticed. Typical mylonites develop along the Vîlsan Gorges, in the Limpedea Valley and in the Stan Valley (Argeş tributary). They are quite visible also south and south-west of the Perişani and Băiaşu localities.

2.4. *Succession of Lithological Sequences*

Based on the petrographic features observed in the field and completed by the microscopic study, the gneisses of the whole massif were assigned to the Cumpăna Group (Cumpăna Series, Ghika-Budești, 1940; Dimitrescu, 1978; Aluta Group; Cumpăna Series — Kräutner, 1980; Sebeș-Lotru Multigroup, Cumpăna Group, Balintoni et al., 1986).

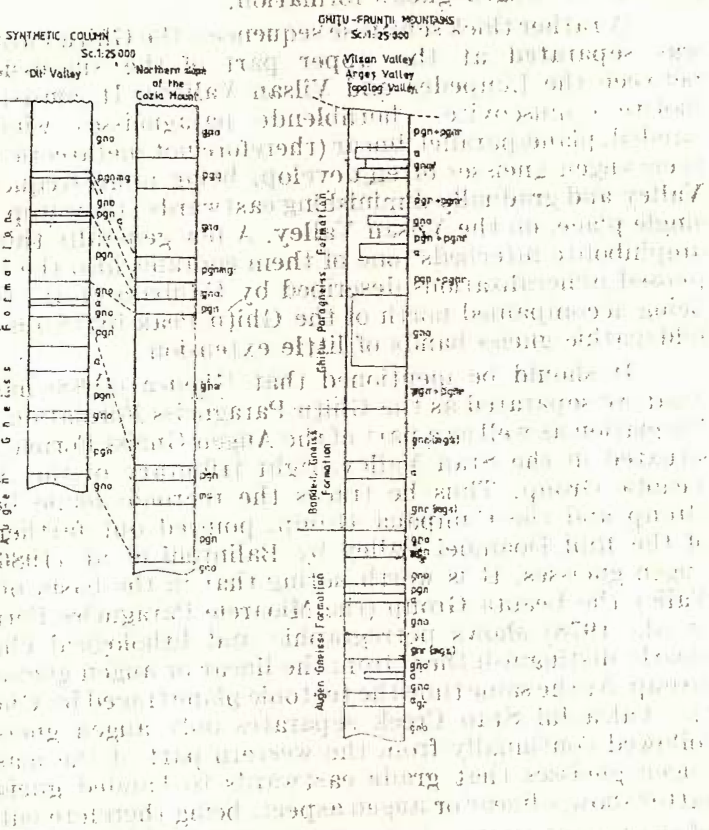

Fig. 4 — Succession of lithological sequences within the massif.

The described crystalline schists are in fact tectonites, a reason why a lithostratigraphic column representing the real thickness and position of the sequences cannot be achieved. Thus the lithologic columns in Figure 4 are built especially for indicating the position and geometric distribution of the lithological content. It follows that the "Cozia Augen Gneiss Formation" develops in the base of the pile, covering the whole outcrop area of the massif between the Olt and Argeș Rivers. Paragneiss sequences (in places micaschists, amphibolites and stromatitic migmatites-banded gneisses) develop within it. It is overlain by the "Albina Banded Gneiss Formation" consisting of a WSW-ENE trending strip which forms the southern half of the Argeş Gorges and extends eastwards to the Limpedea Valley, a left tributary of the Argeş River. Stromatitic migmatites, which have sometimes linear aspects, are characteristic, presenting also augen aspects. Amphibolite interbeds are also found. Generally thin biotite paragneisses ± muscovite, in places with grey-bluish quartz, were also noticed, constituting the petrographic background which is similar to that of the augen gneiss formation.

A rather thick schistose sequence—the Ghiţu Paragneiss Formation — was separated at the upper part of the succession, which develops between the Limpedea and Vîlsan Valleys. It consists mainly of usually biotite ± muscovite ± hornblende paragneisses which are often fine-banded, plane-parallel, linear (therefore not metatectic migmatites); ithin them augen gneisses often develop, being more frequent in the Limpedea Valley and gradually diminishing eastwards; they crop out also farther in a single place, on the Vîlsan Valley. A few generally thin and discontinuous amphibolite interbeds (one of them contains also the Cu—Ni unmetamorphosed mineralizations described by Udubaşa et al., 1987) are also found, being accompanied north of the Ghiţu Peak by two narrow white quartz-feldspathic gneiss bands of little extension.

It should be mentioned that Gheuca (1988) has recently assigned what we separated as the Ghiţu Paragneiss Formation, the Banded Gneiss Formation as well as a part of the Augen Gneiss Formation (up to a locality situated in the Stan Valley, right tributary of the Argeş River) to the Leaota Group. Thus he traces the tectonic plane between the Leaota Group and the Cumpăna Group, pointed out farther east in the basin of the Rîul Doamnei Valley by Balintoni et al. (1986) within the Cozia augen gneisses. It is worth noting that in the basin of the Rîul Doamnei Valley the Leaota Group (the Mioarele Paragneiss Formation, Dimitrescu et al., 1978) shows petrographic and lithological characteristics which clearly distinguish them from the linear or augen gneisses of the Cumpăna Group. At the same time the tectonic plane traced by Gheuca (1988) through the Valea lui Stan Creek separates only augen gneisses, which can be followed continually from the western part of the massif, from the same augen gneisses that grade eastwards to banded gneisses; in places the latter show a linear or augen aspect, being therefore entirely characteristic of the Cumpăna Group. On the other hand, where a certain lithologic and petrographic change is noticed (Limpedea Valley basin), such a tectonic plane cannot be pointed out. Here the augen gneisses show interbeds and interfingerings that gradually disappear eastwards. The amphibolite or quartz-feldspathic gneiss levels are too discontinuous and reduced in thickness and extension to constitute lithologic markers which could be followed at a distance or to be reliable for some lithostratigraphic correlations. However, there are similarities between the rocks making up the petrographic background of the Ghiţu Formation from the Vîlsan Valley and those from the basin of the Rîul Doamnei Valley. The comparative study of the amphibolites in both zones might contribute to the solving of this problem which, for lack of sufficient data, is still open.

3. Structure of the Massif

The most obvious structural element is represented by the Cozia Fault trending east-westwards, with the upper compartment dipped by over 1,000 m in the Olt Valley basin, and delimiting northwards the massif from the Brezoi-Titești sedimentary basin along a distance exceeding 50 km. The Cozia Fault extends westwards along the Lotru Valley (Lotru Fault, Savu, 1967).

In its eastern extremity, in the Vîlsan Valley basin, north and east of the Ghițu Peak, a first impression suggests its transgressive overlying by the Lower Miocene sediments, indicating that this fault functioned to the Upper Paleogene inclusive. It is sure that the fault steep slope decreases to a great extent from the Argeș Valley eastwards. Ștefănescu (1985) shows, however, that the Cozia Fault gradually disappears due to some more recent faults trending NW-SE which, by the gradual dipping of the north-eastern compartments cause disappearance of the crystalline massif outcrop eastwards.

It is also worth noting that this tectonic accident was recognized for the first time by Reinhard (1910). Later on the fault was figured also on the maps drawn by Popescu-Voitești (1918), Streckeisen (1934), Ghika-Budești (1940, 1958) on sheets 1 : 50,000, mentioned above, and by Săndulescu (1984). Hann and Szász (1984) show that the fault functioned in two phases : first it brought about a strike slip by over 2 km of the southern compartment westwards, while subsequently the northern compartment sank greatly ; this effect was emphasized also by the Olt Fault trending N—S, with the lowered eastern compartment, and simultaneous with the second phase of the Cozia Fault.

In the western extremity, the Cozia gneisses overthrust the Santonian-Coniacian, Upper Cretaceous deposits along an overthrust plane situated in the eastern slope of the Stan Valley (Hann, Szász, 1984). Balintoni et al. (1986) show that this plane is transgressively overlain by Campanian-Lower Maastrichtian deposits, being thus of Laramian age. But the first to point out this overthrust plane was Schmidt (1930). Later on Ghika-Budești (1940) admitted here only a common tectonic contact, turning the Cozia Fault southwards.

The Cozia massif was considered by Streckeisen (1934) to belong to the "superior nappes from the Getic Nappe". Ghika-Budești (1940) did not recognize the existence of these nappes and consequently included the Cozia gneiss in the Cumpăna Series along with the "Făgăraș Series" within the Getic Group (Nappe). Codarcea et al. (1967) includes the Cozia gneiss in the Austrian "Supragetic Unit". Hann and Szász (1984) show that the Cozia massif belongs to the Făgăraș Supragetic Unit, at the same time distinguishing the Omul Scale, situated east of the Cozia Peak, and the Valea lui Stan Scale. According to Balintoni et al. (1986), the Cozia gneiss belongs to the Argeș Nappe, which is one of the Austrian units forming the Laramian Boia Nappe and overthrusting the Călinești Mesocretaceous Unit along the Stan Valley. The latter unit makes up together with other Austrian units the other Laramian Nappe, the Lotru Nappe. Recently Hann and Balintoni (1987) have shown that the Călinești and

Cîineni Nappes may represent two subunits of the same larger tectonic unit.

The crystalline schists of the Cozia-Frunții-Ghițu massif generally trend east-westwards. The monoclinal structure prevails, and only in the Olt Valley, where the outcropping pile reaches its maximum thickness, folded structures are found, represented by an anticline already observed by Ghika-Budești (1940); the southern slope of the anticline contains secondary folds, most of them being involved in a system of directional faults.

The structural image of the massif is marked by the presence of some transversal faults, the most important being that crossing the Olt Valley in the vicinity of the Turnu Monastery, which strikes the Upper Cretaceous deposits, and the one situated west of the Băiașu Brook, with the lowered eastern compartment, enabling the contact between the Upper Cretaceous and the Eocene ones. A system of transversal and longitudinal faults of local importance is present inside the massif. It is interesting to note that part of this system formed prior not only to the Cozia Fault, but also to the Paleogene and respectively Upper Cretaceous deposits.

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GNAISUL DE COZIA (CARPAȚII MERIDIONALI): DATE PETROGRAFICE, STRUCTURALE, GENEZĂ

(Rezumat)

Alături de paragnaise (plagiognaise) mai mult sau mai puțin micacee, la fondul petrografic participă și intercalații de amfibolite, rar gnaise cuarțo-feldspatice și într-un singur caz, calcare cristaline. Paragnaisele trec uneori gradat în roci cu structură microgranoblastică, care sînt de fapt blastomilonite. Paragnaisele nemigmatizate nu conțin feldspat potasic. Migmatizarea, formarea gnaiselor oculare s-a produs prin procese metasomatice cu precădere pe seama blastomilonitelor care pot constitui uneori mici fragmente înglobate de fenoblastele de feldspat potasic. Caracteristicile și evoluțiile principalelor minerale constituente ale paragnaiselor indică caracterul retromorf atît al blastomilonitizării, cît și al metasomatozei, toate procesele desfășurîndu-se în limitele faciesului amfibolitelor cu almandin. Formarea gnaiselor oculare s-a produs pe un fond blastomilonitic prin schimbarea compoziției chimice a întregului sistem. Gnaisele oculare cuprind roci cu caractere și aspecte heterogene, dar unite din punct de vedere descriptiv. Feldspatul potasic este reprezentat

atît prin ortoză, cît şi prin microclin, mineralele fiind strîns asociate, fără limită între ele. Formarea feldspatului din profiroblaste reprezintă un proces metasomatic polifazic, ce are loc într-un interval de timp larg. Iniţial s-a format ortoza monoclinică neordonată, ce se transformă apoi parţial în microclin triclinic ordonat şi maclat prin inversiune polimorfă. Microclinul nemaclat se formează direct în modificaţia stabilă triclinică ordonată. Se observă prezenţa a două generaţii de pertite. Descrierea şi clasificarea formei, conturului şi poziţiei ochilor de feldspat potasic, cît şi a aspectelor legate de asociaţiile dintre ochii de feldspat potasic permite înţelegerea convergenţelor petrografico-tectonice incluse în procesul de formare a gnaiselor oculare.

Breciile constituite din elemente de cristalin aparţin la tipuri genetice diferite, majoritatea fiind depozite continentale subaeriene sau de origine tectonică (brecia de Năruţiu), ambele de vîrstă presenoniană. Milonite caracteristice se dezvoltă la sud şi sud-vest de localităţile Băiaşu şi Perişani, în Valea Limpedea, Valea lui Stan — afluentul Argeşului şi în Valea Vîlsanului, însoţind fracturi şi falii transversale sau longitudinale.

Elementul structural cel mai evident este falia Coziei, care delimitează masivul cristalin spre nord pe o distanţă de peste 50 km şi ridică compartimentul sudic (în zona Văii Oltului) cu peste 1000 m.

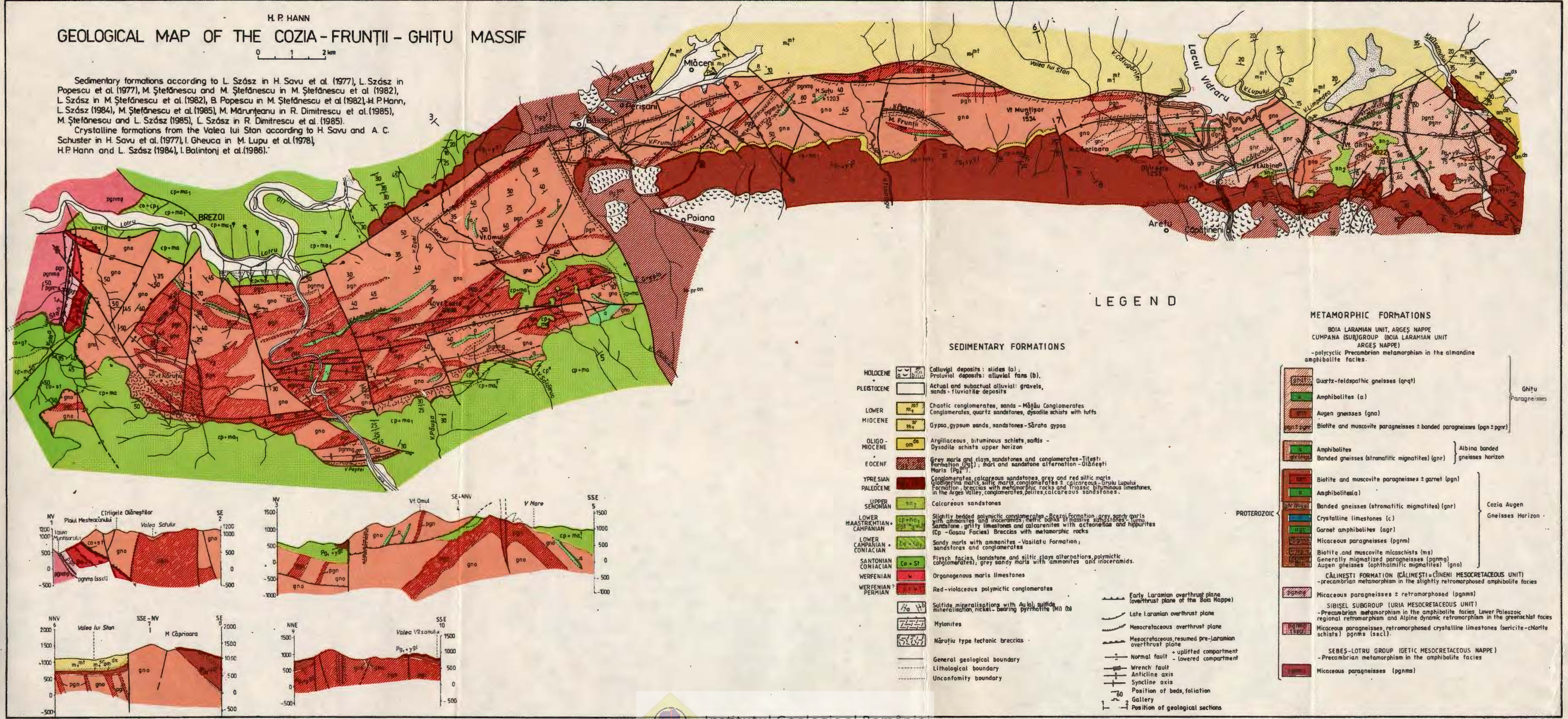
Masivul cristalin aparţine unităţii supragetice de Făgăraş, respectiv pînzei laramice de Boia şi încalecă în extremitatea lui vestică, în versantul drept al Văii lui Stan, în lungul unui plan de şariaj, depozitele santonian-coniacene. În zona Văii Oltului se remarcă o structură anticlinală cu flancul sudic cutat. O parte a sistemului de falii transversale şi direcţionale din interiorul masivului este anterior atît faliei Coziei, cît şi depozitelor sedimentare paleogene, respectiv cretacice superioare.

atît prin ortoză, cît și prin microclin, mineralele fiind strîns asociate fără limită între ele. Formarea feldspatului din porfiroblaste reprezintă un proces metasomatic polifazic, ce are loc într-un interval de timp larg, în care s-a format ortoza monoclinică neordonată, ce se transformă apoi parțial în microclin triclinic ordonat și, mereu, prin intermediul polimorfilor. Microclinul neomorf se formează direct în modificația stabilă triclinică ordonată. Se observă prezența a două generații de pertite. Descrierea și clasificarea formei conturului și poziției ochilor de feldspat potasic [illegible] relațiilor de asociație dintre cristalele de feldspat potasic permite înțelegerea convergențelor petrografo-tectonice implicate în procesul de formare a gnaiselor oculare.

Breciile constituite din elemente de cristalin apar la tipuri genetice diferite, majoritatea fiind depozite continentale subaeriene sau de origine tectonică (breccia de Nărățiu), ambele de vîrstă presenoniană. Milonite cataclastice se dezvoltă la sud și sud-vest de localitățile Brăoși și Perișani, în Valea Lîmpedea, Valea lui Stan, [illegible] Argeșului și în Valea [illegible], însoțind fracturi transversale sau longitudinale.

Elementul structural cel mai evident este falia Cozia, care delimitează cristalinul spre nord pe o distanță de peste 30 km și ridică compartimentul sudic (în zona Văii Oltului) cu peste 1000 m.

Masivul cristalin aparține unității supragetice de Făgăraș, respectiv pînzei [illegible] și încalecă în extremitatea lui vestică, în versantul drept al Văii lui Stan, în lungul unui plan de șariaj, depozitele senoniene. În zona Văii Oltului se remarcă o structură anticlinală cu flancul sudic cutat, iar partea sistemului de falii transversale și diagonale din interiorul masivului este anterioară faliei Cozia, cît și depozitelor sedimentare paleogene, respectiv cretacice superioare.



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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

THE STRUCTURAL BACKGROUND OF THE BRUSTUR FORMATION IN THE INNER DANUBIAN NAPPES[1]

by

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Tectonic units. Danubian Nappes. Brustur Formation. Conglomerate. Sandstones. Low-grade metamorphism. Tectogenesis. Structural analysis. Sedimentary structures. South Carpathians—Sedimentary Danubian Realm—Feneș zone, Svinița zone-Svinecea.

Abstract

The Brustur Formation from the Inner Danubian Realm is a megasequence consisting of: Baicu Conglomerates (with gabbro and peridotite lithoclasts originating from the Caledonian Tișovița-Iuți ophiolitic complex); green sandstones bearing amphibole clasts; black pelites; limestones; white or grey sandstones and microconglomerates reworking quartz-feldspathic material. The whole sequence, which also contains rhyolite bodies, is folded and affected by a low grade metamorphism. Structurally it belongs to a Danubian basement unit—the Baicu Nappe. The Baicu Nappe is overridden by the Arjana Nappe (redefined) consisting of two digitations: Căleanu and Feneș. The Arjana Nappe (Codarcea, 1940) consists of Mesozoic formations associated with alkaline magmatites (Russo-Săndulescu, unpublished data), originating from the "Arjana-Severin Realm". It overlaps the basement nappes of the Inner Danubian Realm and it is overridden by the Getic Nappe. The Brustur Formation represents the northern extension, within the pre-Alpine basement, of the Caledonian paleosuture cropping out in Southern Banat.

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Résumé

Formation de Brustur du domaine danubien interne dans le contexte structural d'ensemble. La formation de Brustur du domaine danubien interne est une mégaséquence contenant de : conglomérats de Baicu (à éléments de gabbros et à peridotites provenant du complexe ophyolitique calédonien de Tişoviţa-Iuţi), grès verts à clastes d'amphiboles, pélites noires, calcaires, grès et microconglomérats blancs ou gris, à matériel quartzo-feldspathique. Toute la séquence, avec aussi des corps de rhyolites, est plissée et affectée par un métamorphisme de faible intensité. Au point de vue structural, elle fait partie d'une unité de socle danubien : nappe de Baicu. En position géométrique supérieure elle supporte la nappe d'Arjana (redéfinie), où se développent les digitations de Căleanu et de Feneş. La nappe d'Arjana (Codarcea, 1940) est constituée des formations mésozoïques associées aux magmatites alcalines (Russo-Săndulescu, données inédites), provenant du „domaine d'Arjana-Severin". Elle dépasse les nappes de socle du domaine danubien interne et supporte la nappe gétique. La formation de Brustur marque, dans le cadre du soubassement préalpin, l'extention vers le nord de la paléosuture calédonienne qui affleure dans le Banat du Sud.

1. Introduction

North-west of the Ţarcu Mountain, Gherasi (1937) distinguished a sequence of "green tuffogeneous rocks", subsequently named "the Brusturi Formation" (Gherasi et al., 1974). Formerly this sequence was considered Liassic, Triassic or older (Gherasi, 1937) and later ascribed to the Ordovician-Silurian (Gherasi et al., 1974) based on palynological evidence of A. Visarion (chitinozoans and microphytoplanktonic remains typical of this time span).

Ghera i (1937) has separated the conglomerates with gabbro clasts as the "Şucu-Poiana Nedeii Conglomerates and breccias". Gherasi et al. (1974) used the name "Baicu Conglomerates" for conglomerates with gabbro clasts, as opposed to the "Şucu Conglomerates" containing flassergabbro clasts. Their age was considered Lower Carboniferous (Schafarzik, 1899), Mesozoic and possibly also Paleozoic (Gherasi et al., 1974) Ordovician (Morariu, 1976).

This paper yields evidence that the Baicu (Şucu) Conglomerates and the grey or green Brustur Sandstones are members of a lithofacies association within a unitary and continuous lithostratigraphical sequence. For the entire sequence (well exposed in the Olteana Creek-Culmea Brusturului area) we propose the name "the Brustur Formation" with the Baicu Conglomerates making up the dominant facies in the basal part.

The mineralogical-petrographical composition of the coarse facies and vertical facies associations strongly point to a common source area. Vertical facies changes reflect changes in sedimentary environment, while mineralogical changes reflect variations in the source area composition due to the erosion of a basement with complex structure and petrographic constitution.

The Brustur Formation redefined crops out on the right tributaries of the Şucu Creek (Fig. 1) : Şuculeţ, Mîrşeşti, Olteana, as well as on the divides between them (Culmea Nedeii—Bistricioara Peak—Baicu Mountain; Fărcăşeşti, Culmea Brusturului). Isolated patches of the Brustur Formation (previously separated as Baicu Conglomerates (Gherasi et al., 1937) or as rocks of the "Scorila Formation" (Gherasi et al., 1974) occur on the left bank of the Şucu Creek, south-east of the Muntele Mic area.

2. Regional Structural Background

Information concerning the structural relationships of the Brustur Formation with its Proterozoic basement and with the Mesozoic deposits of the area shown in Figure 1 has been integrated into a general structural

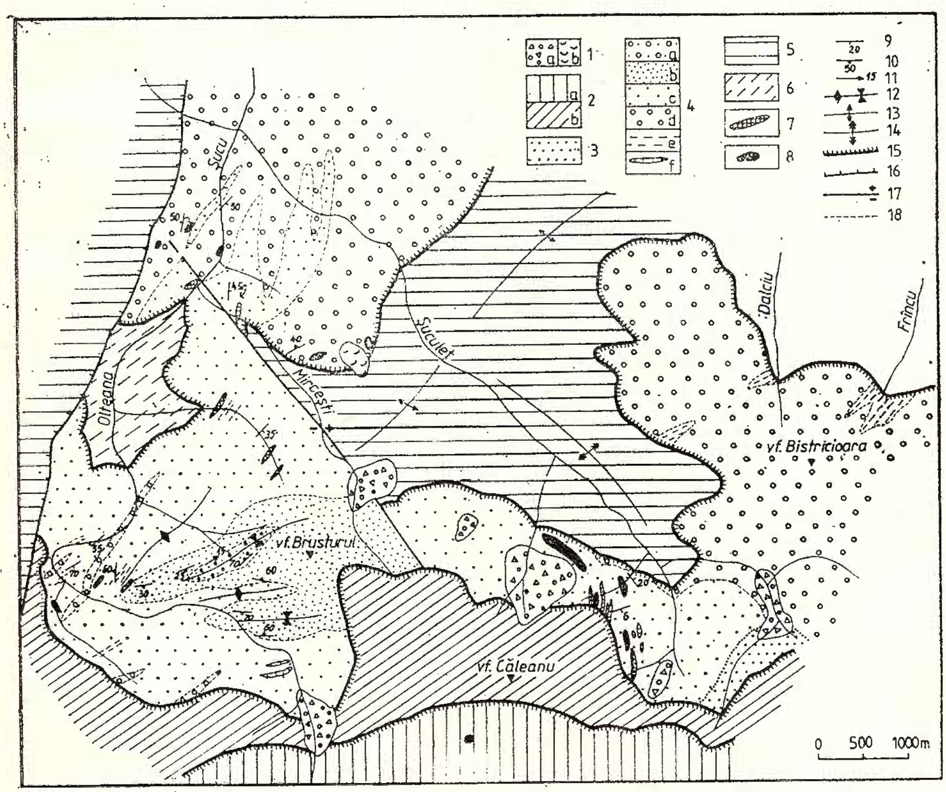

Fig. 1 — Geological map of the Şucu area.

1, Quaternary: a, glacial deposits; b, debris fans; 2, Arjana Nappe: a, Feneş thrust; b, Căleanu thrust; 3, Wildflysch formations (Upper Cretaceous); 4, Brustur Formation: a, grey microconglomerates; b, grey sandstones, black slates; c, amphibolic sandstones and siltstones; d, Baicu Conglomerates (with gabbro clasts); e, black siltstones and mudstones; f, limestones and carbonate rocks; 5, Poiana Mărului Unit; 6, Olteana Window. Magmatic rocks associated with the Brustur Formation: 7, rhyolites; 8, serpentinites; 9, bedding (S_0); 10, cleavage (foliation) (S_1); 11, B_1 fold hinges; 12, B_1 folds; a, anticline; b, syncline; 13, 14, post-nappe anticlines; 15, overthrust; 16, reverse fault; 17, dip slip fault; 18, geologic boundary.

scheme (Pl. I), useful in understanding the position within the pre-Alpine units of rocks which contributed the clasts for the Brustur Formation.

The area illustrated in Figure 1 lies in the north-eastern part of the Cornereva-Rîul Alb-Rîul Lung zone (Pl. I) where the geological maps (scale 1 : 200 000) (Hunedoara and Baia de Aramă sheets) show Mesozoic deposits (Jurassic and Cretaceous) unconformably overlying the metamorphic basement (Paleozoic-Precambrian) of the Inner Danubian.

Recent overviews (Kräutner et al., 1981; Berza et al., 1983) revealed that the pre-Alpine basement is involved in several upper Danubian nappes; the Mesozoic deposits of the "Arjana zone" (Codarcea, 1940) representing the normal cover of a Danubian basement nappe.

Our researches carried out during several years in the Danubian Realm (Jiu, Cerna, Lăpuşnic-Rîul Mare basins, Cornereva sheet, as well as the correlation sections on the area of Armeniş sheet) enabled us to separate the following units:

— Getic units, with typical deformed Mesozoic cover;

— Infragetic units originating from the Severin-Arjana sedimentary realm (Codarcea, 1940; Săndulescu, 1984), made up by Mesozoic deposits which underwent differential deformations, locally accompanied by low grade matamorphism;

— Danubian units, with specific Mesozoic covers, deformed together with their basement.

Superposed deformations (folds and nappes) resulted in a complex Alpine structure in this area of the Inner Danubian Realm. The Alpine nappes, involving Proterozoic, Paleozoic and Mesozoic formations have been produced during the dacidic period (Săndulescu, 1984) in the Austrian and Laramian events. The Austrian event created lower amplitude nappes within each realm, in contrast with the Laramian event, when Getic and Infragetic nappes overthrust the Danubian units. The whole structural assemblage was subjected to post-nappe folding (with different trends) and faulting (regional reverse faults) (Pl. I, Fig. 1).

The Brustur Formation lies in the axial zone of a large, post-nappe (post-Laramian) anticlinorium, trending NE—SW, which we designate as the Rîul Lung-Şuculeţ anticlinorium; in respect of this, the Getic Nappe outliers lie in synclinorium axes. The younger folds trending WNW—ESE, complicate the post-nappe structures by axial depressions and culminations.

In the areas of axial culmination (basins of Rîul Lung-Rîul Alb Valleys) the nappes with upper geometric position crop out, made up by the Mesozoic formations of the Arjana Realm. The lowest nappes, dominated by the pre-Alpine basement are exposed in the areas of highest axial culmination, situated to the NE (Şucu and Bistra Mărului Valleys) and SE (Rîul Rece, Bela Reca to the Danube).

In the area detailed in Figure 1 we could separate the following structural units, situated under the Getic Nappe (from top to bottom): the Arjana Nappe, with two subdivisions: the Feneş and Căleanu Nappes or digitations; the Baicu Nappe; the Poiana Mărului Nappe; Olteana Window.

2.1. *The Arjana Nappe*

Under the Godeanu outlier of the Getic Nappe (Murgoci, 1905; Codarcea 1940), on the Cuntu-Ţarcu slope and in the upper reaches of the Olteana Creek, the Mesozoic deposits of the Arjana Nappe occur (separated south of the Ţarcu Mountain as Arjana Nappe — Codarcea, 1940); this nappe partly coincides with the Feneş-Arjana Nappe, separated by Năstăseanu et al. (1988) west of the Godeanu outlier.

2.1.1. *The Feneş digitation.* W and NW of Ţarcu Peak, the Jurassic sequences in Feneş facies (Năstăseanu, 1979) make up a pre-Laramian nappe (digitation) tectonically overlying (along a mylonitic contact) the Jurassic sequences in Gresten facies of the Căleanu zone. Both as position as well as in lithology this nappe coincides with the "Feneş-Arjana Nappe" separated in the Cornereva area (Năstăseanu et al., 1988) or in the Armeniş area (Boldur, Boldur, in Răileanu et al, 1958).

The Feneş digitation, which is a component of the Arjana Nappe, is made up of Jurassic-Lower Cretaceous deposits, with the typical presence of the Mesojurassic-Neojurassic volcano-sedimentary formation, containing alkaline magmatites, typical of intracontinental rift areas (Russo-Săndulescu, in Năstăseanu et al, 1988; in Conovici et al., 1987). The volcano-sedimentary formation underlies an Upper Cretaceous terrigenous sequence showing flysch features (Boldur, Boldur, in Răileanu et al, 1958) — the "Bodia flysch" of Morariu, Morariu (1982).

The post-Austrian Upper Cretaceous facies, dominated by sandstones, unconformably overlie the folded associations which make up the Feneş and Căleanu Nappes (digitations).

The base of the Feneş Nappe is marked by mylonites accompanied by low grade recrystallisation (blastomylonites), noticed at the springs of the Rîul Rece, on the Cuntu-Ţarcu slope as well as in the tectonic windows mapped by Boldur, Boldur (in Răileanu et al., 1958).

2.1.2. *The Căleanu Nappe (digitation).* This unit, separated in this paper, underlies the Feneş Nappe (Pl. I; Fig. 1). It consists of terrigenous deposits; white or grey coloured conglomerates and sandstones, sometimes showing cross bedding, interbedded with siltstones and mudstones, which may have a high coal content typical of the Lower Jurassic associations in the Gresten facies (Năstăseanu et al., 1974). These deposits have been ascribed to the Lower Liassic (Gherasi, 1937) or to the Lower and Middle Liassic (Năstăseanu, in Conovici et al., 1986). The presence of alkali magmatic bodies (Russo-Săndulescu, in Conovici et al., 1987) suggests that this lithofacies association was initially situated in the "Arjana zone", similar to the Feneş Formation.

The facies described are tightly folded, with folds trending NNE—SSW, obviously discordant with the internal structure of the underlying Brustur Formation and of the Feneş Formation. The mudstones and siltstones show a penetrative slaty cleavage which parallels the axial planes of the mesoscopic folds. The coarse, competent sequences are dominated by concentric folds, the fine grained interbeds show microcrenulations and cleavage, with incipient metamorphic neoformation.

The contact of these formations with the Brustur Formation and the Upper Cretaceous deposits of the underlying unit (at the springs of the Suculeț Creek) is a low angle thrust plane with irregular pattern, which intersects the internal structures and is accompanied by mylonites.

2.2. *Baicu Nappe*

This nappe was separated here for the first time in respect of the Proterozoic basement and the formations of the Arjana Nappe. In the area investigated by us, it consists of the Brustur Formation, as defined in this paper. Considering the probably Paleozoic age of the Brustur Formation (which underlies Liassic conglomerates in the Cornereva area — Mărunțiu, Seghedi, 1983 ; Năstăseanu et al., 1988) and its common folding with the Precambrian basement, a pre-Alpine age for this contact is possible.

Although the Baicu Nappe includes the Baicu Conglomerates (separated by Kräutner et al., 1981, and Berza et al., 1983, as the "Şucu Unit") it has a completely different significance than the latter. The tectonic contact of the Brustur Formation with the Precambrian basement is marked by ductile deformations (shear cleavages and microfolds), sometimes accompanied by calcite and quartz blastesis. The mylonites resulted on rocks belonging both to Baicu and Poiana Mărului Units were subjected to subsequent deformations (crenulation, microfolds and cleavages, but also major folds).

2.3. *The Poiana Mărului Nappe*

This unit crops out on the Şuculeț Creek, where it consists of Proterozoic formations (amphibole gneisses, amphibolites, mica plagiogneisses) affected by pre-Alpine regional retrograde metamorphism (Caledonian and possibly also Variscan).

Kräutner et al. (1981) and Berza et al. (1983) used the name "Poiana Mărului Unit" for the tectonic unit separating the Inner Danubian (in the basins of the Rîu Şes, Bistra Mărului and Bistra Bucovei Valleys) from the units of the Outer Danubian, also called the "Mehedinți-Retezat Unit" (Stănoiu, 1973).

Both basement units (Baicu and Poiana Mărului) correlate in spatial position and componence to the Nijudimu Nappe in the Cornereva zone (Iancu, Visarion, 1988) and to the Sirinia Nappe (Năstăseanu et al., 1988).

2.4. *The Olteana Window*

The Olteana Window has a very restricted outcrop area on the Olteana Creek, where it represents the lowermost tectonic unit. This unit, revealed by hydrogeological borings (Durdun, 1982) is marked by the overthrust of the Brustur formation (folded together with rocks of the Proterozoic basement exposed on the left slope of the Olteana Creek) over Liassic sandstones and conglomerates with mudstone-coal layers, which possibly belong to the Cornereva Unit (Năstăseanu et al., 1988). To the west, this window is limited by a post-nappe reverse fault with regional extension; on its western, upthrown side, the Muntele Mic Granites, retrogressed Proterozoic formations and rocks of the Brustur Formation are exposed.

3. The Brustur Formation

3.1. *Lithology*

The Brustur Formation is an upward fining sequence, consisting of several dominant lithofacies:

— Baicu Conglomerates (containing gabbro clasts as a typical feature); green sandstones (with amphibole clasts) or grey sandstones (arkoses); light grey microconglomerates and conglomerates (with quartz-feldspar lithoclasts). Grey or black siltstones and mudstones, limestones and carbonate sandstones subordinately occur associated with the main facies. The entire sequence is associated with serpentinitic rocks as large blocks or lens-shaped bodies (reaching meters or tens of meters in size), acid volcanic bodies (rhyolites-keratophyres — Gherasi, 1937) and basic rocks (porphyritic microgabbros), affected by deformations and low grade metamorphism.

The lithofacies of the Brustur Formation occur in three main lithofacies associations (Fig. 1):

— Baicu Conglomerates; green sandstones and siltstones;

— green sandstones and siltstones with subordinate Baicu Conglomerate bodies; black mudstones; white arkosic sandstones; limestones and carbonate rocks;

— polymictic microconglomerates and conglomerates, associated with white-grey sandstones.

3.1.1. *Green conglomerates and sandstones*. The green coloured rocks — Baicu (Şucu) Conglomerates and sandstones-siltstones — are closely associated and rework similar clasts, suggesting a common source area. The coarse, conglomeratic facies dominates the lower part of the formation. The conglomerates grade upward to and interfinger with green sandstones which become the dominant facies in the middle part of the formation.

The Baicu Conglomerates, well exposed on the Şucu Creek, Culmea Nedeii, Baicu and Bistricioara Peaks, occur as massive beds, 3—6 m in thickness, which lack internal structure and organisation. They contain rounded or subangular clasts mainly of green rocks (gabbros, amphibolites), but white coloured rocks may also occur (quartz-feldspar gneisses; plagioclase granites, paragneisses). Serpentinite clasts may be seen in places. The facies lacks size sorting. In places conglomerates interfinger with coarse, microconglomeratic sandstones, or the structureless beds show a faint grading at the bed tops to massive microconglomerates or to coarse grained sandstones displaying horizontal laminations. In the outcrops on the Olteana, Şucu and Şuculeţ Creeks, the conglomerates are interbedded with thin continuous beds or lens-shaped bodies of green or black sandstones. The thickness of sandstone bodies ranges between centime-

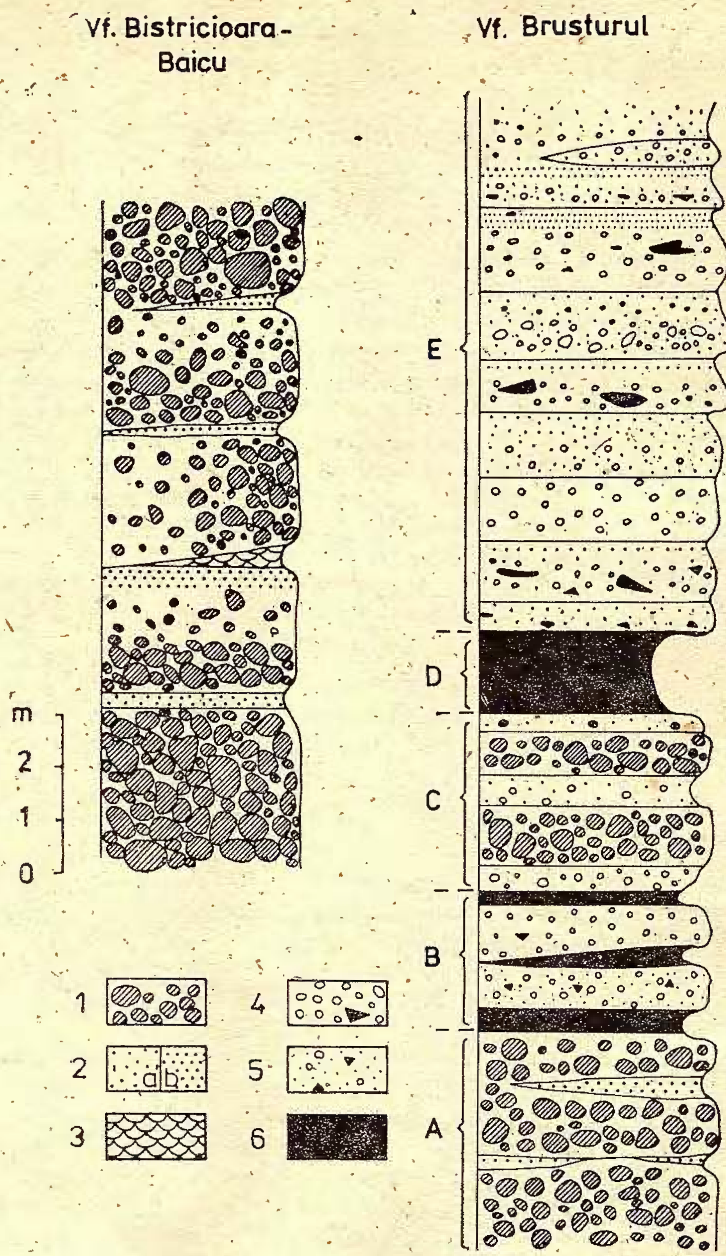

Fig. 2 — Lithological columns in the Brustur Formation.

1, Baicu Conglomerates ; 2, massive (a) and horizontally laminated (b) amphibolic sandstones ; 3, ripple cross-laminated amphibolic sandstones ; 4, grey conglomerates, microconglomerates ; 5, grey sandstones ; 6, black slates.

ters and decimeters. Sandstones are massive but frequently show horizontal or ripple cross laminations (Fig. 2, col. 1). The coarse sandstones, with well preserved hornblende clasts, have erosional base and sharp contacts.

In some areas (the upper reaches of Olteana and Şuculeţ Creeks) the green sandstones make up thick sequences, interlayered with conglomerates, coarse sandstones with large amphibole clasts, grey sandstones, detrital limestones, black slates.

Coarseness, absence of sorting and internal organisation of conglomerates suggest high discharge and high sediment load, typical of upper flow regime conditions. The conglomerates are probably proximal fan deposits.

Horizontally bedded sandstones are deposited in upper flow regime, from high velocity flows (Mc. Kee et al., 1967). Cross laminations may result at lower velocities, at low water stages. Horizontal laminations may result in lower flow regime, in sands coarser than 0.6 mm (Harms et al., 1975).

The association of conglomerates and green sandstones in dominantly coarse sequences (in the Şuculeţ reaches and Baicu and Bistricioara Peaks) reflect alternation between high stage and low stage deposits.

3.1.2. *White-grey sandstones, microconglomerates and conglomerates.* This facies consists of well sorted microconglomerates and conglomerates, with rounded clasts, attaining 1–2 cm in diameter and consisting mainly of white-coloured rocks. Microconglomerates occur as lenses attaining 2 m in thickness, associated with coarse or medium-grained sandstone beds ranging in thickness from centimeters to decimeters. Conglomerates are massive, with clearly erosional beds and irregular contacts with the underlying beds. Various sized intraclasts of black shales frequently occur in sandstones. Some conglomerates grade upward to coarse or medium-grained sandstones, possibly showing a faint lamination. The conglomerates are bimodal, with clasts supported by a sandstone matrix.

The white-grey coloured sandstones are lens-shaped or show a laterally persistent bedding. They have irregular or planar, erosional base with frequent mudstone intraclasts concentrated toward the base or unevenly dispersed throughout the bed. They may be massive or may show graded bedding. In the Olteana basin and on the lower reaches of the Mîrşeşti (Mirceşti) Creek, sandstones and black mudstones occur in upward fining sequences, attaining several meters in thickness and interbedded with coarse-grained green rocks. On the Brusturul Peak, sandstones and microconglomerates are the dominant facies (Fig. 2, col. 2).

The features of this facies — erosional bases, good sorting, presence of bedding and of internal organisation — suggest deposits from low viscosity flows in upper flow regime conditions possibly representing distal fan facies and eventually channelized flows.

3.1.3. *Massive black mudstones.* This facies occurs as thick interbeds (meters or decimeters in thickness) in the previously described facies and often occurs as intraclasts in the overlying coarser beds. The pelitic beds have sharp bases and a frequently irregular, eroded top. The facies suggests deposition from suspensions in pools of standing water.

Limestones represent a minor facies, usually situated in the sequence intermediate between conglomerates (locally with carbonate matrix) and the green sandstones (Şuculeţ springs, Plaiul Borlovei). The limestones occur as metric or decimetric, grey coloured and massive layers containing crinoidal ossicles.

Several carbonate rocks associated with the green sandstones are in fact sandstones with metamorphic neoformation of calcite. Carbonate rocks associated with the serpentinites have resulted by secondary alteration of ultramafic rocks.

3.2. *Petrography and Mineralogy*

The Baicu Conglomerates are orthoconglomerates (Morariu, 1976) consisting mainly of dark lithoclasts of basic and ultramafic rocks, amphibolites and amphibolitic gneisses. Clast lithotype analysis revealed : Iuti type gabbros (locally altered) ; Plavişeviţa type blastomylonitic gabbros (showing oriented fabric and neomineralisation with actinolite + epidote + chlorite ± fucsite) ; pegmatoid gabbros ; clinopyroxenites ; serpentinized dunites (showing relict cumulus type structures) ; listwenitized peridotites etc. ; all these rocks derive from the petrological association of the Caledonian ophiolitic complex (Mărunţiu, 1984) occurring within the pre-Alpine structural units from the basement of the Inner Danubian. The conglomerates contain minor amounts of leucocratic rocks — granitoids and paragneisses — usually associated with the amphibolitic rocks of the pre-Alpine Danubian units. Conglomerates are usually clast supported, but in places matrix supported conglomerates with disseminated clasts occur. Rocks are bimodal, the conglomerate matrix consisting of a very coarse-grained sandstone with plagioclase, hornblende, epidote, chlorite and actinolite clasts.

The sandstones associated with the Baicu Conglomerates are immature to mature containing various amounts of matrix and subangular clasts. Clasts are mainly represented by primary minerals of gabbros and amphibolites (hornblende and plagioclase) and various secondary alteration products of these minerals ; lithoclasts are subordinate, consisting of gabbros, stripped amphibolites and plagioclase gneisses. The green hornblende is usually fresh, sometimes with pyroxene cores ; large clasts of uralite with pyroxene cores or ghosts also occur. The hornblende shows various stages of alteration to actinolite or chlorite, the latter occurring as rims or pseudomorphoses on the former. In places, the hornblende may alter to calcite.

The plagioclase feldspar is fresh or it may show various stages of sericitisation or albitisation (accompanied by neoformation of sericite, albite and epidote). The sphene and opaque minerals occur in minor amounts. According to the classification of Folk (1968), these rocks are litharenites and feldspathic litharenites and may be named amphibolic sandstones.

The grey or white coloured microconglomerates and conglomerates consist mainly of light-coloured lithoclasts — granites, plagioclase gneisses, quartzites, rhyolites and frequently gabbros, with amphiboles altered to calcite and plagioclases, to sericite and partly calcite. The light-coloured sandstones associated with these microconglomerates consist of angular and subangular clasts of quartz, plagioclase and alkali feldspar (altered to schachbrettalbite-calcite), chlorite and opaque minerals pseudomorphing amphiboles, detrital muscovites, oxides and minor amounts of lithoclasts — plagioclase gneisses, rhyolites and highly calcitised gabbros. The fine grained sandstones may show cross laminations marked by heavy mineral concentration. From the mineralogical point of view, these rocks are plagioclase arkoses.

All these rocks show an argillaceous matrix, now recrystallised due to the very low grade metamorphism which affected the Brustur Formation. The absence of cement is also typical.

The black pelites consist mainly of clay minerals, with scarce clasts of quartz and detrital muscovite.

The compositional and textural immaturity of the Baicu Conglomerates and the associated green sandstones suggest a rapid transport and/or a short distance of transport (Morariu, 1976). The presence of clasts of amphibolic rocks as well as of fresh amphiboles indicates a quick deposition and burial, conditions required by the chemical instability of the basic material. The presence of large amounts of amphiboles suggests a source area with an excessive climate either cold or dry, where mechanical weathering was the dominant process, and not the chemical alteration, as it is known that minerals of the basic rocks are partly hydrolised in water (Pettijohn et al., 1972).

The petrographical and mineralogical features of the grey microconglomerates and sandstones show a similar textural and compositional immaturity suggesting quick transport and burial. The better sorting and the advanced alteration of the amphiboles and the basic rock clasts point to a longer distance of transport than those which preserve the primary nature of the clasts, thus to distal deposits.

4. Structures and Metamorphism

The folding of the Brustur Formation took place under very low grade metamorphic conditions. Large scale folding is accompanied by development of mesoscopic folds in incompetent sequences (sandstones, siltstones, mudstones) and of the partly penetrative cleavages, accompanied by differential and inhomogeneous metamorphic neomineralisation. Sedimentary rocks as well as associated magmatites show both sedimentary and metamorphic fabric and mineral assemblages.

The sequence of the Brustur Formation, dominated in its lower part by green conglomerate facies, occurs in the core of a large synclinorium with the basal part cut by the thrust plane at the base of the Baicu Nappe.

While the limbs of this main structure are represented by the thick sequences of conglomerates with competent deformations, the axial zone of the structure is involved in minor folds trending mainly SW—NE (Fig. 1), well exposed in the Olteana basin. The Baicu Conglomerates in the base of the sequence make up thick, apparently homogeneous beds, showing irregularly distributed S_1 shear cleavages : strongly sheared zones, with disrupted clasts, alternate with massive "unsheared" zones.

The middle and upper part of the sequence, consisting of alternations of rocks with different competence (multilayered sequence), is affected by buckle or flexural slip folds as well as by shear folds showing concentric or similar morphology. Buckle folds, more frequent in the thick homogeneous sandstones, show deformations of bedding, with the cleavage (S_1) parallel to S_0. Clast deformation (shearing, subgranulation) is accompanied by reorganisation and recrystallisation of the matrix constituents along the S_1 planes. Meanwhile, a mimetic, apparently static crystallisation of the neoformation minerals, possibly occurs either within the matrix or at the expense of the clasts (by overgrowing or substitution).

The multilayered sequences show metric folds well exposed in outcrops; folds are open to tight, symmetrical or asymmetrical, sometimes overturned to NW. These folds are accompanied by axial-plane cleavages intersecting bedding (S_0) along constant, parallel planes (Fig. 3a, c) or along variously inclined planes, due to cleavage refraction (Fig. 3d).

In several places mechanical transpositions are visible, marked by distortion of lithology (bedding) and slip along the cleavage planes (Fig. 3c). Mainly in the grey sandstones (with carbonates) cleavages are accompanied by intense neoformation (of calcite, quartz, epidote) creating an inhomogeneous aspect of rocks : layers showing foliation and metamorphic neomineralisation alternate with layers with well preserved sedimentary structures (Fig. 3b). The fine-grained facies (siltstones, mudstones) show highly penetrative cleavages and bedding (S_0) is deformed by rootless intrafolial microfolds (tight and extremely sheared).

The same features are visible under the microscope (Pl. II — VI) : preservation of pre-metamorphic (sedimentary or magmatic) fabrics; differential development of cleavage, due to the relative competence (petrographical-mineralogical composition, thickness, position in the fold etc.).

Like deformations, metamorphic recrystallisation is inhomogeneous differentiated mineralogically and quantitatively as well as in intensity. In most coarse-grained rocks, the newly formed minerals occur in subordinate amounts in respect of the pre-existing minerals, showing static, random growth or a certain orientation in the cleavage plane (S_1). The random growth is explained either by the weak stress within the entire rock volume (which correlates to the low and inhomogeneous density of the cleavages) or by blastokinematic development of neoformation minerals in the slip zones, inhomogeneously distributed within the general recrystallisation background under the influence of lithostatic P and T.

The following mineral parageneses are dominant, depending on the previous mineralogical-petrographical constitution :

— chlorite, albite, sericite, carbonate (calcite, siderite, magnesite), epidote, stilpnomelane, in green rocks;

— muscovite (phengite), chlorite, quartz, albite — in quartz feld-

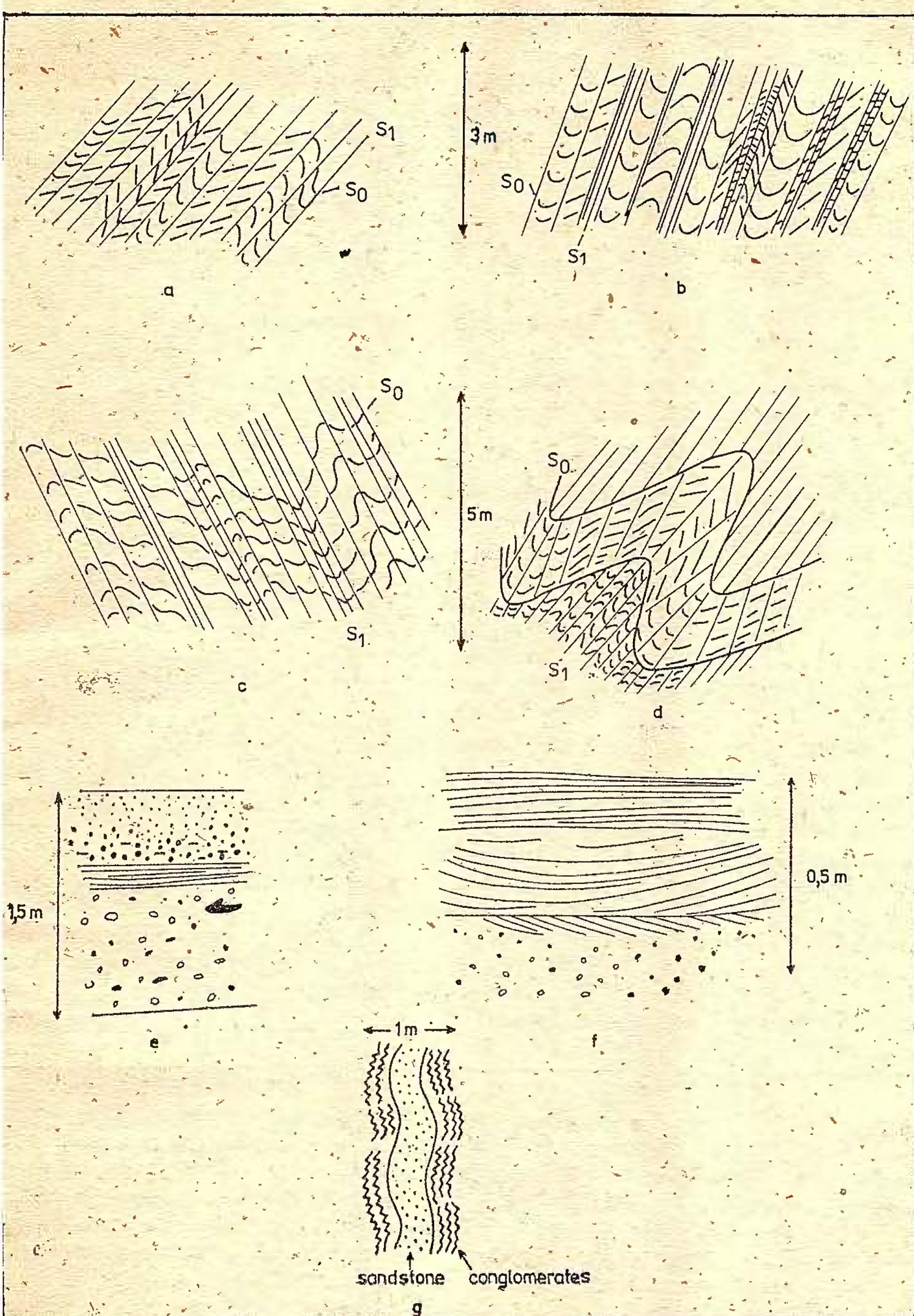
S1
S0
a
3 m
S0
S1
b
S0
S1
c
5 m
S0
S1
d
1,5 m
e
0,5 m
f
1 m
sandstone
conglomerates
g

Fig. 3

spar and micaceous rocks (sandstones, siltstones, microconglomerates rhyolites).

This assemblage, as well as the large scale preservation of the pre-existing minerals and structures, suggest a very low grade metamorphism of rocks. The S_1 cleavages, accompanied by recrystallisation, are marked either by shearing, similar to the mylonitic cleavages (the rocks showing a blastomylonitic appearance), by phyllosilicates (phyllosilicate foliation or by strips or lenses of neoformation minerals (quartz-calcite) marking an incipient metamorphic differentiation (Pl. II—VI).

5. Paleotectonic Significance

The description of the lithofacial features and of the lateral and vertical facies association in the Brustur Formation suggests several conclusions.

The Brustur Formation is an upward fining sequence, dominated in its lower part by proximal fan associations and by distal fan association in its upper part. The green conglomerates (Baicu Conglomerates) associated with green (amphibolic) sandstones show upward and distal facies changes to grey microconglomerates and sandstones interbedded in black mudstones. These deposits have resulted by the erosion of a source area consisting of mafic and ultramafic rocks similar to the Tisovița-Iuți ophiolitic complex and to rocks resembling metamorphics from the crustal units. They were brought in contact during the tectogenetic phases of the Caledonian and Variscan cycles.

The material eroded underwent a quick transport and deposition close to the source area. Deposition of a thick sequence of mainly conglomeratic deposits suggests an intense tectonic activity in the source area which was probably folded and uplifted on active faults. This phase of high uplifting (Fig. 2, col. 2 A) marked by the accumulation of the thick sequences of green conglomerates and sandstones is interrupted from time to time when distal deposits (mudstones and white coloured sandstones prograde over the conglomerates (Fig. 2, col. 2 B). After a period of tectonic rejuvenation, marked by the progradation of the green conglomerates over mudstones (C), a relative tectonic quiescence follows (D), and afterwards the source area is reactivated again (E). As a consequence

Fig. 3 — Outcrop sketches in the Brustur Formation.

a, B_1 shear folds and associated spaced cleavages (S_1) affecting bedding (S_0); b, bedding (S_0) and cleavages (S_1) accompanied by mechanical transposition and metamorphic neomineralization; c, distorsion of bedding (S_0) along S_1 cleavages and production of incipient metamorphic layering; d, buckle fold in a sandstone bed with related S_1 cleavage. (Figures a-d — Olteana reaches); e, sequence of grey conglomerates capped by laminated siltstones, underlying a graded, coarse-grained sandstone with sharp, erosional base; siltstone intraclasts in sandstone and conglomerates (Culmea Brusturului); f, green, amphibolic, very coarse grained, massive sandstones underlying ripple cross-laminated fine grained green sandstones (Bistricioara Valley); g, open concentric folds of sandstone bed in ultramylonitic Baicu Conglomerates (Șucu Creek) a thick sequence of sandstones and microconglomerates is deposited, showing the same features over great thicknesses, but the erosion affects other lithological levels (richer in leucocratic rock) in the source area, which probably has a lower relief or has migrated.

The presence of proximal detrital deposits north of Țarcu Peak (Șucu-Poiana Mărului area) resulted by erosion of an ophiolitic complex similar to that outcropping in Southern Banat, provides valuable evidence concerning the continuation over 70 km to NNE of this Paleozoic ophiolitic paleosuture, separating two important crustal domains (Iancu, 1983) brought in contact during the Caledonian orogeny.

Within the structural framework making up the pre-Alpine basement of the Alpine upper Danubian units, in Southern Banat several "zones" and petrographic complexes are evidenced, separated by important structural breaks. These complexes are :

— polymetamorphic Proterozoic formations belonging to the Mraconia-Ielova Group (Poiana Mraconia and Ielova Series — Bercia, Bercia, 1975). These rocks crop out within the Mraconia Variscan Nappe, unconformably overlain by the Variscan molasse deposits ;

— mafic-ultramafic igneous rocks of the Tisovița-Iuți-Plavișevița Caledonian ophiolitic complex (Mărunțiu, 1984) partly affected by dynamic metamorphism and recrystallisation ;

— Proterozoic polymetamorphic formations of the Neamțu Group lying east of the paleosuture and separated from it by a blastomylonitic "zone" (Mărunțiu, Seghedi, 1983) named "Corbu zone" by Codarcea (1940).

North of Mehadia, in Cornereva area, a sequence of Lower Paleozoic low grade metamorphic rocks (Nijudimu Formation — Iancu, Visarion, 1988) is tectonically interposed between the two sequences ;

— Caledonian granitoids intruding the previously mentioned complexes (Stan, 1985 ; Stan et al., 1986).

North of the Mraconia Creek, the rocks of the ophiolite complex become narrower, occurring as thin tectonic slices and northward, the paleosuture becomes cryptic.

The Poiana Mraconia Series, unretrogressed, tectonically overlies the metamorphic rocks west of the paleosuture and partly the ophiolitic complex. To the north, in the Cornereva-Ideg area, metamorphic rocks are dominant with the pre-Alpine basement cropping out as scattered inliers. In this area, the metamorphic rocks of the Variscan Mraconia Nappe are no more exposed, but only formations belonging to the underlying structural unit occur, showing a strong Lower Paleozoic retrogression (the Padina amphibolite Formation — Iancu, Visarion, 1988).

On the divide between Ohaba and Bela Reca Creeks, terrigenous deposits occur which rework rocks belonging to the Lower Paleozoic ophiolitic complex and adjacent metamorphic rocks : Baicu Conglomerates and green sandstones, remarked by Mărunțiu, Seghedi (1983) and separated by Năstăseanu et al. (1988) as "Culmea Românului Formation". Thus, in the Nijudimu Unit (Iancu, Visarion, 1988) rocks belonging to the Brustur Formation occur with amphibolites and Baicu Conglomerates, rocks which north of the Țarcu Peak, occur in the Baicu and the Poiana

Mărului Units respectively. In these nappes, the rocks of the ophiolitic complex are reworked in the conglomerates and sandstones of the Brustur Formation, which crop out to the west, next to the Muntele Mic granite and to the north, next to the Poiana Mărului area.

The data presented here have important bearing or the correlation of the Proterozoic formations from the basement of the Inner Danubian, which have been enclosed to the Zeicani Group in the latest syntheses (Kräutner et al., 1981; Berza et al., 1981).

In order to avoid the assignment to a single group of entities lying in distinct Alpine and pre-Alpine nappes, we think it more proper to use the names of formations as they were used by Dimitrescu (1986). For the retrogressed amphibolites in the Șucu Creek, which constitute the "basement" of the Brustur Formation, we used the name "the Șuculeț Formation", which can be correlated to the Padina Formation from Cornereva area, on the basis of structural and paleotectonic position.

The separation of pre-Alpine structural units within the Poiana Mărului Alpine Nappe (which may be correlated to the Sirinia Nappe from the Mehadia basin — Năstăseanu et al., 1988) (Pl. I) as well as the separation of lithostratigraphical units have to be made in the future.

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FORMAŢIUNEA DE BRUSTUR DIN DOMENIUL DANUBIAN INTERN ÎN CONTEXTUL STRUCTURAL DE ANSAMBLU

(Rezumat)

Lucrarea de faţă aduce argumente de natură litologică, mineralogică şi structurală, ce permit redefinirea „formaţiunii de Brustur" (Gherasi, 1937; Gherasi et al., 1974) prin includerea în această formaţiune a „conglomeratelor de Baicu (Şucu)" (Gherasi, 1937). Formaţiunea de Brustur astfel redefinită reprezintă o megasecvenţă ce conţine, ca faciesuri dominante: conglomeratele de Baicu (cu elemente de gabbrouri şi blocuri de peridotite), gresii verzi cu amfiboli clastici, gresii şi microconglomerate albe sau cenuşii (cu elemente de roci cuarţo-feldspatice).

Deşi formaţiunea este afectată de cutare strînsă, însoţită de apariţia clivajelor şi de neoformare metamorfică incipientă, prezervarea conţinutului mineral premetamorfic şi a structurilor moştenite permit reconstituirea succesiunii litostratigrafice. Aceasta începe cu conglomeratele de Baicu, continuă cu o secvenţă alcătuită predominant din gresii verzi cu claste de amfiboli, urmată de o secvenţă constituită din gresii cenuşii, micacee, cu nivele de microconglomerate.

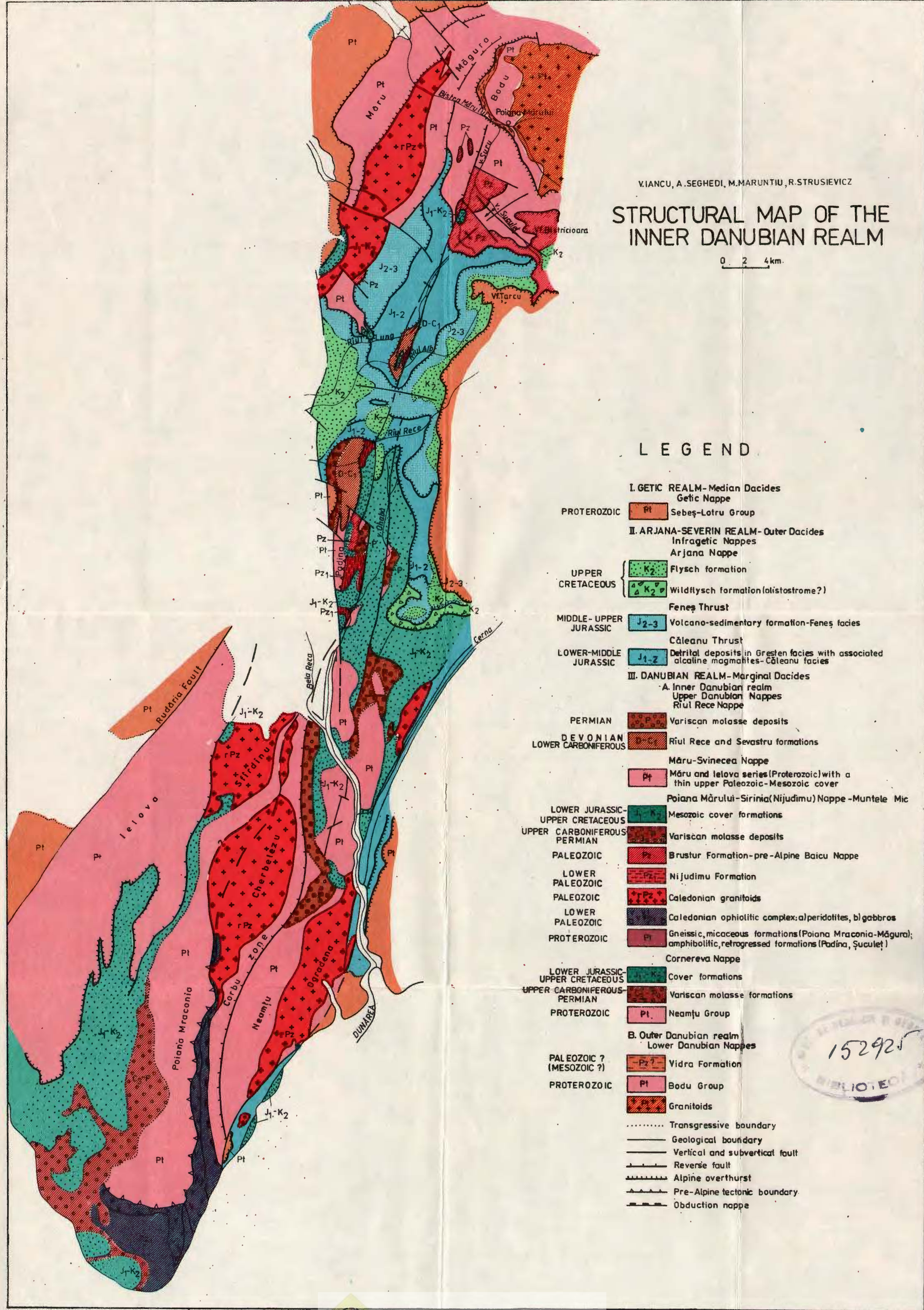
Prima parte a lucrării prezintă structura în pînze de şariaj a zonei Şucu-Ţarcu (fig. 1), fiind evidenţiată poziţia allohtonă (în pînze de cuvertură) a formaţiunilor mezozoice din zona Cuntu-Ţarcu, pe care le considerăm ca provenind în ansamblu din domeniul Arjana (Codarcea, 1940). Majoritatea autorilor le-au considerat drept cuvertură în loc a cristalinului din zona Cornereva-Poiana Mărului.

În aria detaliată de noi au putut fi delimitate următoarele unităţi structurale, situate geometric sub pînza getică: pînza de Arjana (cu digitaţiile Feneş şi Căleanu), pînza de Baicu (ce cuprinde formaţiunea de Brustur), pînza de Poiana Mărului şi fereastra Olteană. Este posibilă vîrsta prealpină a contactului tectonic dintre pînza de Baicu, constituită din formaţiunea paleozoică de Brustur şi formaţiunile proterozoice ale pînzei de Poiana Mărului. Planşa prezentată în lucrare reprezintă o încercare de corelare structurală a pînzelor de cuvertură ale domeniului Arjana şi a pînzelor de soclu, cu cuvertură paleozoic superior-mezozoică ale domeniului danubian intern.

În partea a doua a lucrării se prezintă descrierea formaţiunii de Brustur. Se aduc argumente privind asocierea litofacială a celor două faciesuri dominante: conglomerate de Baicu (Şucu) şi gresii verzi sau cenuşii de Brustur, într-o succesiune litostratigrafică unitară şi continuă.

În cadrul formaţiunii de Brustur au fost descrise următoarele asociaţii litofaciale: conglomerate de Baicu, asociate cu gresii şi siltite verzi; gresii şi siltite verzi în care apar subordonat conglomerate de Baicu, pelite negre, gresii albe (cuarţo-feldspatice), calcare şi roci carbonatice; microconglomerate şi conglomerate polimictice asociate cu gresii alb-cenuşii.

Formaţiunea de Brustur este afectată de cutare la scară regională, cutele mezoscopice fiind caracteristice secvenţelor mai fin granulare din zona axială a sinclinoriului pe care îl formează (gresii fine, siltite,



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pelite). Atît la scară mezoscopică, cît şi la scară microscopică, se constată: conservarea structurilor premetamorfice, sedimentare sau magmatice; comportarea diferenţiată a clivajului, în funcţie de competenţa relativă a rocilor; blasteza metamorfică neomogenă şi diferenţiată mineralogic şi cantitativ.

Asociaţiile paragenetice de neoformaţie sînt clorit, albit, sericit, carbonaţi (calcit, siderit, magnezit), epidot, stilpnomelan (în rocile verzi); muscovit (fengit), clorit, cuarţ, albit (în roci cuarţo-feldspatice şi micacee).

O serie de date mineralogice şi structurale permit sublinierea unor caractere particulare ale formaţiunii de Brustur. Astfel, imaturitatea compoziţională şi texturală a conglomeratelor de Baicu şi a gresiilor verzi, sugerează un transport rapid şi/sau de mică distanţă.

Prezenţa clastelor de roci amfibolice şi a amfibolilor proaspeţi indică o depunere şi îngropare rapidă, condiţii necesare datorită naturii instabile a materialului bazic. Aceeaşi imaturitate texturală şi compoziţională se constată şi la microconglomeratele şi gresiile alb-cenuşii, însă sortarea mai bună şi alterarea avansată a amfibolilor şi a clastelor de roci bazice permite caracterizarea lor ca depozite distale.

Toate aceste caracteristici, ca şi prezenţa blocurilor mari (de la metrice la zeci de metri cubi) de gabbrouri şi serpentinite (amplasate neregulat fie în conglomerate, fie în gresii sau siltite) pledează pentru depunerea formaţiunii într-un regim tectonic activ, cu intervale de timp în care a avut loc progradarea depozitelor distale.

Sursa materialului component o constituie fundamentul antepermian al domeniului danubian intern, fiind recunoscute atît roci polimetamorfe ale fundamentului proterozoic (amfibolite, gnaise amfibolice sau micacee, granitoide), cît şi roci aparţinînd complexului ofiolitic Tişoviţa-Iuţi (peridotite serpentinizate, gabbrouri şi flassergabbrouri).

EXPLANATION OF PLATES

Plate II

Fig. 1 — Irregular, erosive base of sandstone overlying siltstone (S_0); S_1 cleavage occurs in the siltstone layer. N//, 36 ×.

Fig. 2 — Rough, discontinuous cleavage (S_1) intersecting sedimentary laminations (S_0) in sandstone. N//, 36 ×.

Plate III

Fig. 1 — Erosive base of green sandstone (with amphibole clasts) (S_0) overlying siltstone. N +, 18 ×.

Fig. 2 — Local metamorphic crystallization in siltstones: quartz and carbonate lenses. N +, 36 ×.

Plate IV

Fig. 1 — Differential crystallization of metamorphic phyllosilicates in siltstones and sandstones. N +, 36 ×.

Fig. 2 — Intense metamorphic neomineralization and incipient layering in sandstones and siltstones due to quartz and carbonate segregation. N +, 36 ×.

Plate V

Fig. 1 — Crenulation cleavages (S_2) affecting S_1 // S_2 in siltstones. N //, 36 ×.

Fig. 2 — Tight intrafolial microfolds (B_1) in black slates. N //, 36 ×.

Plate VI

Fig. 1 — Quartz and plagioclase feldspar phenocrysts in the slightly recrystallised groundmass of a metarhyolite. N +, 36 ×.

Fig. 2 — S_1 cleavages in metarhyolite, marked by phyllosilicates (muscovite, chlorite) crossing the relict phenocrysts. N +, 36 ×.

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

CENTRAL AND NORTHERN FĂGĂRAŞ — LITHOLOGICAL SEQUENCES AND STRUCTURE[1]

by

DINU PANĂ[2]

Tectonic units. Metamorphic rocks. Retrograde metamorphism. Key beds. Shear zone. Mylonites. Antiform folds. Tectogenesis. South Carpathians — Crystalline Getic and Supragetic Realms — Făgăraş Mountains.

Abstract

A series of lithological units in the Făgăraş polymetamorphic Group are here separated, based on pre-retromorphic distinctive features. Some discontinuous lithological markers show a constant location inside the synmetamorphic built up sequences. A remarkable along strike continuity of the lithon sequences is noted. Their lithological structure and metamorphic significance are revealed and a comparison with the previously described divisions is presented. A large mylonitic-blastomylonitic series is pointed out. Some major tectonic unconformities and a subsequent folding, controlling the present location of the lithon sequences are emphasized.

Résumé

Făgăraş central et septentrional — successions lithologiques et structure. Des unités lithologiques ont été séparées dans le groupe polymétamorphique de Făgăraş à partir des caractères prérétromorphiques distincts. Quelques repères lithologiques discontinus montrent une localisation constante dans le cadre des séquences synmétamorphiques. Une continuité directionnelle importante des séquences de lithon a été remarquée. Leurs structure lithologique et signification métamorphique ont été relevées et une comparaison des divisions décrites antérieurement a été présentée. Une série milonitique-blastomilonitique a été observée. Quelques discordances tectoniques majeures et un plissement postérieur en contrôlant la présente localisation des séquences de lithon ont été mis en évidence.

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The present paper is an attempt to approach, based on "lithostratigraphic" criteria, the complex problems related to the metamorphics from the central and northern zone of the Făgăraş massif, which are assigned to the "Făgăraş Group" in the modern classifications (Kräutner et al., 1988; Balintoni et al., 1984, 1986). It is based on personal mapping data obtained in recent years; some lithological details, important for the here presented correlations, have been provided by the previous prospection works (Arion et al., 1965, 1966; Arion, 1969, 1970; Chivu, 1966—1982; Chivu et al., 1983—1985; Lazu et al., 1985; Mînecan et al., 1985; Pitulea, 1961; Pop et al., 1982; Stroică et al., 1984).

Only certain observations on the metamorphism and deformation which were considered significant for the problems discussed will be presented.

We adopted the lithostratigraphic terminology, but considering the "lithostratigraphic units" to be distinct lithological assemblages within which the present lithological sequences and thickness are tectonically controlled (anyway, they are not the initially sedimentary ones). Thus, for simplification, we considered that the polymetamorphism resulted only in foliations and lithons transpositions within the various initial lithological assemblages, without involving, however, the tectonic mixing between the adjacent formations. In this way the separated lithons sequences may be referred to as "stratigraphic intervals", which make possible some "lithostratigraphic correlations" on the scale of the whole massif as well as some structural considerations.

Historical Background

According to the modern classifications, based on lithostratigraphic criteria, the Făgăraş Group represents an upper rank unit exhibiting very characteristic features within the Făgăraş massif metamorphics, this entity being separated for the first time by Ghika-Budeşti (1940) as the "Făgăraş Series".

Several correlation attempts were made in the Central and Eastern Făgăraş after its first delimitation and subdivision. In a first stage, these correlations were made based on the concepts of "isograde", "mineralogical zone", "metamorphic facies". The limits of the divisions and the sense in which the old denominations coming from Ghika-Budeşti are used, differ from one author to another and even by the same author in successive works. This stage is represented by the contributions of: Manea (1962), Dimitrescu (1962, 1963, 1964, 1967 a, b), Dimofte (1967), Dessila-Codarcea, Stancu (1968), Pitulea (1961, 1967—1970), Dimitrescu et al. (1974), Dimitrescu (1976), Chivu (1985).

As the retromorphism concept developed and remarkable results were obtained in some other regions through correlations based on lithostratigraphic criteria, a new stage of knowledge starts also in Făgăraş. Dimofte (1967) and Dimitrescu (1967, 1978) mention the existence of retromorphism phenomena in the upper part of the "Şerbota complex". Giuşcă et al. (1977) emphasized the impossibility of tracing some isogrades in the Făgăraş crystalline and the importance of the retromorphism imprint from the crest zone towards the north.

The lithostratigraphic terminology is introduced by Kräutner (1978, 1980), Kräutner et al. (1988), and the lithostratigraphic approach of the problems related to the metamorphics of the Făgăraş massif begins with the studies made by Balintoni et al. (1984, 1986) and Gheuca (1988).

It is obvious that the stratigraphic classifications based on some features do not necessarily coincide with those based on some other features. The case of the Făgăraş Group is one of the most significant examples in this sense. There are often great differences between the classifications proposed in the two stages, the first of them being based on the criterion of the (apparent) intensity of metamorphism, while the second one is based on lithostratigraphic criteria, taking into account also the polymetamorphic evolution of the pile.

We shall present in detail these differences when describing each lithostratigraphic unit of lower rank separated according to the recommendations of the international stratigraphic guide — Hedberg et al. (1976) — within the Făgăraş Group, based on the distinctive pre-retromorphic lithological characteristics. On the northern slope of the Făgăraş massif the following recommendation of the I S G has been taken into account: "the inclusion within a single lithostratigraphic unit of the adjacent beds, separated by a regional unconformity, should be avoided even if lithological differences justifying the separation cannot be found". The rank established for each unit depends on the detail degree required by necessities of drawing up the map.

The boundaries of the lithostratigraphic units representing formations are of tectonic nature, constituting therefore also tectonic units, the relative shifting of which are difficult to estimate.

1. Pîrîul Moaşa Formation

This unit is now separated for the first time in the north-western part of the Făgăraş massif, between the Ionelu Valley to the west and the Laita Valley to the east, based on some distinct lithostratigraphic characteristics. Its name derives from the Moaşa Brook which provides a complete transverse. It comprises metamorphics assigned by Ghika-Budeşti (1940) to the "Şerbota garnet paragneiss zone" and to a part of the Poiana Neamţului chlorite-schist zone". Balintoni et al. (1984, 1985) assign this entity to the "Şerbota Formation", which would be intensely retromorphosed northwards — at the top (?)

Based on the new observations we could separate sequences of various lithologies within this unit, considering them members. In this case, the use of the Şerbota denomination in connection with some other entities exhibiting different lithological features from those characteristic of the "Şerbota garnet paragneiss zone" — Ghika-Budeşti, 1940 — is no longer justified. It was necessary to denominate the separated subdivisions and to re-denominate the unit including them

The boundaries between the members are more or less arbitrary, depending on the alteration degree of the previous medium-grade particular aspects through the subsequent retromorphic imprint. Within the Pîrîul Moaşa Formation the following members will be described: the

Moldele Member in lower position, cropping out in the axis of an antiform trending approximately E—W; the Şerbota Member s.l. (the Bîrcaciu horizon + the Şerbota Member s.str.) that develops along the entire southern flank and on the northern flank up to the Lupului Valley meridian; farther eastwards the Poiana Neamţului Member crops out both on the northern flank and in the zone of the Făgăraş main ridge, where it is involved in several folds. Its relations with the other units of the Pîrîul Moaşei Formation are subject to several interpretations due to the lack of the lithological markers and of the conclusive structural elements. The Poiana Neamţului Member might represent either (a) a lateral facies of the Şerbota Member or (b) a unit transgressively overlying the other units of the formation, the stratigraphic unconformity boundary being subsequently transposed.

The Pîrîul Moaşei Formation crops out in the tectonic window beneath the Suru Nappe, being the only entity assigned (so far) to the Făgăraş Group which does not present migmatization aspects.

1.1. *Moldele Member*

Its name derives from the only toponym appearing on maps in the outcrop region of this lithostratigraphic unit. The introduction of a new unit is justified by the lithological contrast in respect to the neighbouring units. This denomination is used for the first time, and represents a part of the unit initially separated by Ghika-Budeşti (1940) as the "Poiana Neamţului chlorite-schist zone". It partly corresponds to the "Poiana Neamţului-Piscul Moaşei subzone", separated by Dimofte (1967). Its distinct lithological character is marked by the prevalence of some quartz-chlorite-muscovite schists (retromorphics on phaneroblastic micaschists) with rare garnet micaschist/paragneiss levels, slightly micaceous massive quartzites, entirely subordinated amphibolites and some relatively thick crystalline limestone lenses, without continuity (but quite probably the same lithostratigraphic interval—Pîrîul Mînii Limestone).

The sequence is characterized by the generally silky aspect due to the dismembered muscovite and sericite, the frequency of the relict muscovite megablasts and a very characteristic crenullation of the micaceous laminae, especially in the central part of the member. The texture is very characteristic plane-schistose due to the alternation of quartz and micas, after which the rock splits in thin leaflets. It crops out in the axis of an antiform trending approximately E—W, which explains its crenullated aspect.

It represents a terrigenous sequence that initially underwent a medium-grade metamorphism. It is entirely affected by retromorphism.

The northern boundary with the suprajacent Poiana Neamţului Member is apparently continuous. It could be followed up to the Fătu Valley westwards and to the right bank of the Porumbacu Mare Valley eastwards.

Southwards, westwards and partly northwards the gradual passing to the Şerbota s.str. suprajacent Member can be noticed. The paragneisses prevail over the micaschists, decimetric to metric amphibolite lithons occur, the relict muscovite megablasts gradually disappear. These aspects suggest transition pre-retromorphic lithological features between the Moldele and Şerbota s. str. Members. At the same time the retromorphism intensity decreases from the Moldele Member towards the Şerbota s. str. Member: sericite gradually disappears, biotite prevails over chlorite, garnet occurs frequently, while staurolite only sporadically. As the retromorphism does not reach the lithological boundary of the Şerbota Member s. str. — which manifests through the massive occurrence of staurolite kyanite paragneisses and phaneroblastic micaschists with frequent amphibolic lenses — we are entitled to introduce an independent lithostratigraphic unit showing transition pre-retromorphic characters — the Bîrcaciu horizon, which is part of the Şerbota Member s.l.

1.2. *Şerbota Member s.str.*

Its name derives from the Şerbota Mount, where this unit crops out. The geographic component of the denomination was used first by Ghika-Budeşti (1940) to designate a metamorphic zone: "the Şerbota garnet paragneiss zone" included in the Făgăraş Series. Dimitrescu (1962) uses the name of "Şerbota zone" in the Eastern Făgăraş for a sequence which is in upper position with respect to the "Măgura Cîineni zone". Dimofte (1967) uses the geographic part of the name in defining the "Şerbota-Piciorul Surului zone", which he assigns to the "Făgăraş Series". In 1974 Dimitrescu defines a "Şerbota complex" in the Eastern Făgăraş above the Măgura Cîineni complex and below the Suru horizon; in 1976 the Şerbota complex comprises also the Suru upper horizon, while in 1978 the "Şerbota complex" comprises a lower Suru horizon and a metapelitic upper one (quartzite-muscovite schists). Savu et al. (1978) use the term of "Şerbota complex" (Măgura Cîineni) as a subdivision of the Cumpăna Series. Chivu (1985) uses the name of "Şerbota Formation" for the upper part of the Cumpăna Series. Balintoni et al. (1984, 1985) refer to the "Şerbota Formation" as a lower part of the Făgăraş Group.

In the present paper the name is used to designate a well-defined sequence generally trending NW—SE and reaching a thickness of about 700 m. It consists of an alternation of garnet, staurolite, kyanite phaneroblastic micaschists/paragneisses as well as numerous thin decimetric-metric amphibolite ± garnet lenses, the continuity of which along strike is difficult to prove. Kyanite occurs especially in the eastern (Şerbota glacial caldera) and western (Tătaru Valley, Fătu Valley) extremities of the members — sometimes as decimetric kyanitite lenses.

The member extends on the southern flank of the antiform with the Moldele Member in its axis. It appears also on the northern flank of the Tătarului and Surului edges. Eastwards, in the Şerbota Mount, it seems to be involved in several folds. This unit is characterized by the very frequent presence of the staurolite megablasts, kyanite, the isolated occurrence of the kyanitite lenses, the almost entire lack of retromorphism phenomena.

It represents an aluminous pelite sequence affected by a medium-grade metamorphism. The boundaries with the Moldele subjacent Member are formed by a gradual passage, through the Bîrcaciu horizon; eastwards it continues with the Poiana Neamţului Member (the relation discussed in chapter 1); the boundary with the Şuru suprajacent Formation is tectonic, being represented by a continuous graphite mylonites level.

1.3. *Poiana Neamţului Member*

Its name derives from an artificial geographical element (Poiana Neamţului chalet) in the vicinity of which Ghika-Budeşti (1940) initially separated a low grade metamorphism zone: "the Poiana Neamţului chlorite-schist zone", part of the Răşinari Series. Later on Dimitrescu extends this name to the metamorphics from the Central Făgăraş (1963), placing it in the Făgăraş Series and subsequently (1967) in the Eastern Făgăraş, considering it to be equivalent to the Făgăraş Series. Pitulea adopts this denomination in the reports between 1967—1970 in connection with a narrow sequence at the boundary with the sedimentary rocks of the Transylvanian Basin. In all the quoted papers the denomination was assigned to a northernmost sequence at the boundary with the sedimentary deposits which underwent a regional chlorite zone prograde metamorphism. Dimofte (1967) and Arion (1969, 1970) use this name making some useful specifications: "the Poiana Neamţului zone" does not show the phyllitic character initially assigned to it by Ghika-Budeşti, and develops between the more eastern formations representing a retromorphosed mesozone. Savu et al. (1978) use the geographic component of this denomination to designate a sequence of much smaller extent at the boundary with the sedimentary rocks, but consider it to be of a higher rank: the "Poiana Neamţului Series". Chivu (1985) comprises under the name of the "Poiana Neamţului-Bîlea Series" the whole pile showing "epizonal" characters east of the Moaşei Brook on the northern slope of the Făgăraş massif. Balintoni et al. (1985), Pană and Ricman (1986) consider that the denomination was applied to the retromorphic part of the "Şerbota Formation".

In this paper the denomination is assigned to a lithostratigraphic unit which crops out in two areas: one north and the other south of the Moldele Member, and which continues westwards in both areas with the Şerbota Member s.l. (the relationships presented in chapter 1).

The Poiana Neamţului Member consists of an alternation of quartz paragneisses/micaschist ± garnet as well as limestone lenses, entirely subordinated amphibolites, and is characterized by: the small size of the micas, a fine lamellar, shiny, often fresh biotite, the frequency of the slightly micaceous quartzites appearing as small plates or as fragments of garnet pencils (or only garnet molds); the rock mass does not show a consistent lamination foliation, but only largely spaced reactivation without the neoformation of the metamorphic foliation; a fine limonitic pigment can be often noticed.

This lithostratigraphic unit represents a terrigenous sequence, initially metamorphosed under medium-grade conditions; subsequently it was discontinuously affected by retromorphism, which lead to the alternation of quartz paragneisses ± garnet and biotite micaschists, totally unaffected, with intensely retromorphic rocks resembling phyllites, sericite quartz schists, chlorite-sericite schists, in which relict minerals are often recognized either as "quer" micas (deferrized biotite, partly chloritized biotite, chlorite pseudomorphoses upon biotite, dismembered muscovite) or partly or totally chloritized or only fragmented garnet.

In the northern outcrop area, in the vicinity of the Poiana Neamtului chalet, the sequence generally extends on the northern flank of the antiform with the Moldele Member in its axis, the boundary seeming to be continuous. The northern boundary with the Suru suprajacent Formation is of tectonic nature. The Poiana Neamtului Member could be followed eastwards up to the Laita Valley and westwards to the Comenzii Hill, where it continues with the Serbota Member s.l.

In the southern outcrop area of the Făgăraş range, the Poiana Neamtului Member can be followed from the Sărătii sharp crest eastwards, continuing the Serbota s.l. Member, being involved in several folds. It is tectonically overlain by the Suru Formation to the south, east and partly north, the boundary being represented by a continuous graphite mylonitic rock level — appearing as a peri-antiform sinking eastwards. In the axial zone of this structure a few intensely tectonized limestone boudines occur, which probably constituted a unique lithological level — the "Laita Peak Limestone". The sequence separated here is geophysically clearly represented; it is overlapped by aeroradiometric maximal anomalies for Th, U, K (Scurtu, in Gheuca et al., 1987).

This entity assigned herein to the Poiana Neamtului Member would correspond to a part of the "Vemeşoaia zone" (chlorite zone) — Dimitrescu (1963, 1976, 1978) and to a part of the "greenschist and crystalline limestone complex" — Giuşcă et al. (1977).

The presence of kyanite in the vicinity of the T.C.H. chalet on the Transfăgărasan road (Pl. III, Fig. 1) indicates that the initial metamorphism of this sequence reached at its climax the medium-grade conditions.

2. Suru Formation

Its name comes from the Suru Mount in the vicinity of which Ghika-Budeşti (1940) separated for the first time a sequence displaying its own characteristics, "the Suru crystalline limestone and para-amphibolite zone", within the Făgăraş Series.

After I separated it based on lithostratigraphic criteria and followed it along strike, I found that it extends at least from the Olt Valley (to the west) to the eastern Făgăraş and shows sometimes important lithological along strike variations.

Dimitrescu (1963) includes the Suru zone (sensu Ghika-Budeşti, 1940) in the Vemeşoaia zone — redefined — from the Făgăraş Series. Dimitrescu (in Patrulius et al., 1967) assign a limestone and amphibolite sequence from the Eastern Făgăraş to the Cumpăna Series, while a limestone and amphibolite sequence from the Western Făgăraş is assigned by Dessila-Codarcea and Stancu (1968) to the Făgăraş Series. Manea (1962) and Pitulea (1967—1970) state that the limestone and amphibolite sequence in the Central Făgăraş represents the upper part of the Măgura Cîinenilor zone (Cumpăna Series). Dimofte (1967) assigns the "Suru-Boului Peak zone" from the Western Făgăraş to the Făgăraş Series.

In 1974 Dimitrescu assigns the "Suru horizon" to the Cumpăna Series, in 1976 specifying that this horizon constitutes the upper part of the Şerbota complex (from the Cumpăna Series).

Schuster (1977) includes the limestone and amphibolite sequence in the Şerbota Formation which he assigns to the Făgăraş Series. According to Giuşcă et al. (1977), this sequence would also partly represent the "greenschist and crystalline limestone complex".

In 1978 Dimitrescu considers the Suru horizon as a lower part of the Şerbota complex from the Cumpăna Series.

Savu et al. (1978) assign the sequence under discussion to the Şerbota complex of the Cumpăna Series. Kräutner (1980) calls it the "carbonatic formation". According to Chivu (1985), part of this sequence represents the Bîlea Formation of the Poiana Neamţului-Bîlea Series, while that separated by Ghika-Budeşti remains in an upper tectonic unit, the Cumpăna Series. According to Kräutner et al. (1988), the Suru Formation represents the base of the Făgăraş Group.

Balintoni (1984) describes the calcareous-amphibolitic sequence in the Suru Formation and in the Valea Rea Formation within the Făgăraş Group; based on the lithological and metamorphism similarities Pană and Ricman (1989) consider that the Suru Formation includes also the "Valea Satului-Porceşti micaschist zone" — Ghika-Budeşti (1940).

The Suru Formation is clearly represented within the Făgăraş metamorphics by massive dolomitic limestone and amphibolite lithons against a micaceous metapelite background, frequently containing graphite. These lithons may reach several hundred metres in thickness and present a remarkable continuity along strike. The pelite background presents sometimes very great variations along strike; under these conditions the continuity of the dolomitic-limestone and amphibolite lithons surely indicate the continuity of the formation (or of the various lower rank stratigraphic intervals). That is the case in the glacial caldera of the valleys on the northern slope in the central part of the massif, where from the Ucea Valley westwards a very characteristic pile of phaneroblastic micaschists bearing a great amount of garnet (followed between the Dejani and Ucea Valleys) laterally grades to fine paragneisses/micaschists.

If from the Olt to the basins of the Capra and Bud a lleys the limestones and amphibolites generally reach great thickness (the amphibolites on the Rîiosu and Lespezi Peaks, the limestones on the Rîiosu Peak), their thickness gradually decreases eastwards, the limestones occur much more rarely, while the pelitic-micaceous background often shows phaneroblastic aspects, the distinctive features of the formation disappearing. This entity situated east of the Moldoveanu Peak meridian was separated and named by Balintoni et al. (1984, 1986) "the Valea Rea Formation", in upper position with respect to the Suru Formation. The continuity of a limestone level accompanied by an amphibolite one between the Moldoveanu Spring and the Dejani Valley basin, close to the boundary with the Măgura Cîinenilor Formation clearly indicates, however, that it is the same stratigraphic interval.

Particular aspects of the Suru Formation were pointed out in the north-western part of the Făgăraş massif (Arion, 1970; Pană, Ricman, 1989). Gridan, Dumitraşcu (1989) assign a restricted sequence of this region to an independent tectonic unit. Limestones occur discontinuously as ever smaller lenses towards the east. In the Lupului Valley basin a few metric augen gneiss levels conformably occur between paragneisses/micaschists and biotite amphibolites extending over some hundred metres. The sequence crops out over a smaller area eastwards, being overlain by sedimentary deposits, but it reappears on the right bank of the Rîul Mare-Avrig Valley and could be followed to the Porumbacul Mare Valley. Here the augen gneisses occur also conformably, without continuity along strike, reaching almost the same thickness as those in the Lupului Valley and forming the same lithological assemblage. Although the microcline augen gneisses occur completely isolated, the sequence contains ophtalmitic rocks that often show aspects suggesting metasomatism phenomena.

There are some elements in common with the Suru Formation which are arguments for preserving the sequence from the north-western extremity of the massif within this formation such as: dolomitic limestones identical amphibolites (either with biotite, or carbonatic laminae), micaschists, paragneisses with small tourmaline and staurolite prisms, the ophtalmitic rocks exhibiting mirmekitic structures (Lupului Valley, Moaşei Valley, Păului Peak).

The metamorphism of the Suru Formation reached at its climax the medium grade conditions (kyanite and staurolite zone). Part of the pile was downgraded to chlorite zone. During these phenomena the "metamorphosed syngenetic-stratiform" sulfide mineralization was entirely remobilized and transposed along some new foliation planes. In some other cases it reconcentrated on fissures showing epigenetic aspects.

On the northern slope of the massif from the Pojorta Valley eastwards the retromorphism affecting the boundary of the Suru Formation is more pronounced. The occurrence of some microcline gneiss levels, do also occur, both aspects being characteristic of a suprajacent lithostratigraphic unit. The single delimiting criterium we could use was the continuity of the mylonitic level followed from the western side of the massif.

The Suru Formation is bounded by the metamorphics of the Cumpăna Group (Măgura Cîinenilor Formation) to the south, by the sedimentary formations of the Transylvanian Basin (in the Western Făgăraş) to the north, and by the metamorphics of the Albota Formation (eastwards). In the north-western and central parts of the massif the boundary between the Suru Formation and the subjacent formation (the Moaşei Brook Formation) also occurs. The boundaries with the Măgura Cîinenilor and Albota Formations are represented by (major?) tectonic unconformities; the one with the Pîrîul Moaşei Formation could be considered an overthrust plane (see chapter 5).

The boundary between the Suru Formation and the Măgura Cîinenilor Formations has lately given rise to much controversy. Kräutner et al. (1988) consider it normal, Balintoni et al. (1986) consider it of tectonic nature, while Gheuca (1988) finds it also normal.

Starting from the above presented opinions and by detailing the mapping in the basins of the Capra and Buda Valleys, we placed this boundary on a graphite mylonitic rock level, based on the following arguments :

a) The revaluation of the lithologic content of the Măgura Cîinenilor Formation. On the Florea Valley, right tributary of the Capra Valley, crystalline dolomitic limestones were encountered in a lithological assemblage which undoubtedly belongs to the Măgura Cîinenilor Formation. In this way one of the former criteria for establishing the boundary — the first limestone occurrence, corresponding to the passage to a carbonatic character of sedimentation, typical of the "Făgăraș Group" — is no longer valid. Thin and discontinuous carbonatic schist levels in a lithological assemblage of Măgura Cîinenilor type also occur on the Mizgavului Spring, the Florea Mount, the Lespezi, Capra and Buda Valleys. Thus we might briefly define the Măgura Cîinenilor Formation as an alternation of metric amphibolite micaschists/plagiogneisses along with isolated limestone/carbonatic schists lenses, while southwards staurolite, kyanite micaschists and discontinuous migmatic gneiss levels occur ever more frequently.

b) Direct indices : mylonitic rocks which would constitute the shearing plane in the base of the Făgăraș Group (part of the Moldoveanu Nappe — sensu Balintoni et al. — 1984, 1986). In the basins of the Capra, Mușeteica, Izvoru Rîiosul, Buda, Doamna Valleys a single important level of graphite and sulfide mylonitic rocks ranging between metres and tens of metres in thickness could be followed. From the Florea Mount westwards along strike this tectonic plane becomes a tectonic zone of over one kilometer in width, represented by numerous decimetric conformable ultramylonite levels. At least from the Capra Valley basin eastwards the mentioned level may be accepted as a boundary marker either in a lithostratigraphic sense (graphite rock levels) or in a tectonic sense (mylonitic rocks), the more so as it separates two sequences showing their own characteristics : on the one hand the Suru Formation with its characteristic dolomitic-limestone and amphibolite assemblage ranging between tens-hundreds of metres in thickness in a generally medium-grade, unhomogeneously downgraded pelitic background ; on the other hand the Măgura Cîinenilor Formation bearing phaneroblastic micaschists and numerous metric amphibolite lithons.

c) Geophysical information : the petrophysical characters of amphibolites differ clearly on either side of the mylonitic level (Romanescu, in Balintoni et al., 1984).

3. Albota Formation

It is named after the Albota Peak situated in the maximal apparent thickness zone of this unit, separated and named now for the first time. It would represent part of the "chlorite-biotite series", described by Manea (1962), part of the "Făgăraș Series", sensu Dimitrescu (1963), part of the "Breaza Valley zone", Pitulea (1967—1970) ; it probably constitutes the upper metapelitic horizon (quartz-muscovite schists) of the Şerbota complex (from the Cumpăna Series, Dimitrescu, 1978) and, according to Kräutner et al. (1988), the lower horizon of the Cirţişoara Formation from the Făgăraş Group. Chivu et al. (1983, 1984, 1985) assign this sequence to the upper part of the Bîlea Formation.

The unit consists of a monotonous sequence of paragneisses/micaschists with or without garnet, numerous separations of quartzite-muscovite schists and micaceous quartzites. It is marked by its more quartzitic character as compared to the neighbouring lithostratigraphic units. It differs from the Suru Formation by the more reduced dimensions of the micas, and from the Sîmbăta Formation by the subordinated gneiss participation and the lack of lamination phenomena. The K-feldspar (microcline) was found only in an isolated level, in the vicinity of the Bîlea Cascadă chalet — without, however, showing augen aspects.

In the central part of this sequence isolated limestone lenses occur, varying in thickness (from a few centimetres — Colţii Brezei — to several hundred metres — Piatra Caprei Caţiavei) and accompanied or replaced sometimes by a thin metric amphibolite or amphibolic gneiss level. Although these limestone and/or amphibolite lenses occur discontinuously, they lie at the same "stratigraphic interval" within the formation. It was named the "Piatra Dracului horizon", after the only toponym appearing in its vicinity on maps.

The horizon was followed along strike for 22 km. The constancy of this marker proves the continuity along strike of at least this stratigraphic interval while : (a) the pelitic background of the Albota Formation shows some variations, and (b) the boundaries of the formation are of tectonic nature.

The understanding of the structural position of this unit is possible only on a regional scale. Between the Bîlii and Sîmbăta Valleys, the foliations are generally east-west with northern dippings, suggesting the superposition of the Albota Formation with respect to the Suru Formation (sections 2, 3, 4, Pl. II) ; at the northern boundary of all the glacial calderas between the two mentioned valleys reverse limb minor folds with slight eastern sinking can be measured ; this would imply the acceptance of a position of antiform nucleus reversed southwards of the Albota Formation. Eastwards, between the Drăguş and Trăsnita edges, the Albota Formation (and its boundaries) dips southwards (sections 5 and 6, Pl. II), therefore it would be situated over the Sîmbăta Formation and below the Suru Formation, and farther eastwards it shows northern dippings (section 7, Pl. II). Taking into consideration the general structural context, in my opinion, the superposition of the Albota Formation with respect to the Suru Formation can be accepted. It is certainly in lower position in respect of the Sîmbăta Formation.

The boundary between the Albota Formation and the Suru Formation is represented by a tectonic alignment that can be continuously followed between the Porumbacu Valley and the Hermeneasa edge — consisting of mylonites and ultramylonites with or without graphite and sulfides. Mylonites are microfolded (Drăguş edge, Caţiavei edge, Pl. III, Fig. 2) and show various dips — which demonstrates that they were involved in foldings after their formation.

The northern boundary with the Sîmbăta Formation is represented by a lamination zone that will be described together with the presentation of the Sîmbăta Formation. This unit has so far been followed eastwards to the Dejani Valley and westwards to the Porumbăcel Valley.

4. Sîmbăta Formation

The name derives from the Sîmbăta Valley where one of the most characteristic profiles in this unit crops out.

The introduction of the new unit is justified by the lithological contrast with the neighbouring units, and the rank assigned to the formation meets the necessities of the map and the achieved degree of detail.

The name is used for the first time and would correspond to : the northern part of the "chlorite-biotite series" — Manea (1962); to the northern extremity of the "Făgăraş Series" — Dimitrescu (1963, 1964, 1967, 1974, 1976, 1978); partly to the Poiana-Neamţului zone" — Pitulea (1967—1970); to the northern extremity of the "Vemeşoaia Formation" — Schuster (1977); partly to the "phyllitized micaschist complex" — Giuşcă et al. (1977) and to the "Făgăraş Series" — Savu et al. (1978), to the "micaschist-quartzite formation" — Kräutner (1980), to the Tunsu Formation—Chivu (1985); to the upper part of the "Cîrţişoara Formation" — Kräutner et al. (1988), partly to the "Cîrţişoara Formation" — Dimitrescu (1988).

The unit is represented by an alternation of micaschists, plagiogneisses, sometimes with garnet, in which pearly gneiss levels (acid oligoclase) or augen gneisses occur (in which the K-feldspar is partly replaced by albite). The frequency of the gneiss levels increases towards the upper part of the pile; thin and discontinuous amphibolite and black banded quartzite levels occur, the whole sequence being affected by regional retromorphism (partial or total substitutions of garnet, biotite, amphibolite by pennine and of K-feldspar by Na-feldspar) and by a discontinuous dynamic metamorphism which is more evident in the upper (northern) part, which affects preferentially the micaceous ± graphite levels. Macroscopically the rocks show a massive texture due to the homogeneous mylonitic aggregate. Under the microscope the oriented texture with tectonically rolled monomineral or composite clasts can be distinguished. The mineral relicts make possible the reconstruction of the following polymetamorphic evolution : medium-grade metamorphism, migmatization, regional retromorphism, mylonitization (Pl. III, IV, V). Frequent incipient neoformation of blastomylonitic type are to be noticed (microlithons consisting of a recrystallized subgrained quartz aggregate and acicular chlorite neoformation (Pl. VI). These aspects determine the mylonitic-blastomylonitic "shear belt" character of the Sîmbăta Formation.

In its lower part (southwards) two lithologic markers were identified, which are of major importance for demonstrating the continuity of the formation on the northern slope of the Făgăraş massif : Piatra Albă limestone horizon and Boldanu quartzite level.

They were pointed out, named and followed in a restricted area by Dimitrescu, Cocîrţă (1980) and Dimitrescu (1988) respectively.

Although discontinuous, these markers steadily occur at the same level within the pile, the distance among them being generally constant.

The Boldanu quartzite is a massive glassy quartzite of high purity degree; so far it has been followed for a distance of 20 km. It crops out continuously from the right bank of the Arpaşu Mare Valley to the Viştea Mare Valley, then it disappears and crops out again in the Sîmbăta Valley; in the source area of the Apa Rîului Valley its development is remarkable, while on the Colţilor edge it is hardly found; the last identified eastward occurrence of interest is in the right bank of the Dejani Valley, immediately downstream the confluence with the Budiu Brook. Except the latter occurrence (in a structural closure position), this level is in east-west position dipping northwards.

The Piatra Albă limestone horizon was followed over a distance of 30 km between the right bank of the Porumbăcel Valley to the west and the left bank of the Pojorta Valley to the east: a) either as a single level ranging in thickness from a few tens of metres (Arpaşu Mare Valley, the right bank of the Apa Rîului Valley) to several tens of centimetres in the Colţilor edge or completely disappearing (the Drăguş edge; Zănoaga edge etc.) or b) as a consistent level accompanied by one or several levels of smaller thicknesses and without continuity (Cîrţişoara, Lupoaia and Viştea Mare Valleys).

The Sîmbăta Formation disappears westwards at the meridian of the Porumbăcel Valley, under the sedimentary deposits, while eastwards it seems to be much more developed.

The southern boundary with the Albota Formation (subjacent) corresponds to a lamination zone which is oblique with respect to the lithological markers: eastwards it is superimposed on a biotite, in places, garnet and/or graphite micacshist packet; westwards it overlies a graphite schist level (the Viştea, Gîrdomanu, Tărîţa, Albota edges), farther micaschists (Cîrţişoara Valley) and graphite quartzite-chlorite schists (Tunsu Valley, Seaca Valley — right tributary of the Porumbacu Valley).

The lamination zone shows quite varied thicknesses, dips and morphological aspects. Thus on the Cîrţişoara Valley, the Tărîţa, Gîrdomanu, Viştea Mare edges, the Vistişoara Valley it is well exposed, on the Zănoaga, Drăguş and Trăsnita edges there are poor outcrops, and, farther, at the springs of the Apa Rîului Valley, on the Hermeneasa edge as well as on the Dejani Valley, it is well exposed again. It is marked by characteristic geomorphological aspects: either pronounced saddles (Tărîţa, Buteanu, Hermeneasa edges) or abrupt changes of slope (Tolfia Hill, Gîrdomanu, Viştea Mare, Apa Rîului edges).

Poor sulfide accumulations locally occur along this alignment (Vistişoara Valley — abandoned gallery, Cîrţişoara and Seaca Valleys) which are very likely of tectogenetic nature.

To the north the Sîmbăta Formation is transgressively overlain by the Badenian deposits of the Transylvanian Basin.

5. Structure

The cartographic extension of the described lithological sequences as well as their spatial relationships lead to some preliminary structural considerations as the "lithostratigraphic" correlation along the whole massif are not yet finished and the metamorphic evolution of the sequences separated based on lithological criteria is not sufficiently known.

Immediately south of the main ridge of the Făgăraş massif, the axis of an E—W trending, longitudinally faulted, antiform can be distinguished. This structure was identified and mapped based on the divergent foliations by Reinhard (1911), Pitulea (1961), Arion et al. (1965, 1966), Dimitrescu (1963, 1978). In the presented map the faulted eastward sinking antiform of the ridge appears well marked also based on lithology; in the Transfăgărăşan road zone the Suru Unit covers the nucleus of the antiform represented by the Poiana Neamtului Member. The fault of the crest was followed between the Negoiul Peak and the Ţiganului saddle, being marked by spectacular narrow saddles or changes on an east-westward direction of some valleys (Urlea, Langa, Sebeş Valleys).

Another antiform outlined based both on the foliations and lithology that resumes on opposite flanks was mapped in the north-western part of the Făgăraş massif. Beside these well outlined major folds, based on the divergent foliations numerous secondary structural axes can be traced on the northern slope, especially west of the Transfăgărăşan road meridian.

It is worth noting that on the flanks of the Făgăraş major antiform, beyond the boundaries of the Suru Formation, lithological entities crop out, that cannot be equated. The initial metamorphism under medium-grade conditions is comparable in all here separated entities; within a several kilometers thick sequence the isogrades could not be identified. Garnet is omnipresent and aluminium silicate minerals occur only in places. Migmatites are in upper geometric positions on both flanks; the retromorphic transformations in the chlorite zone affect a compozite sequence (Făgăraş Group) of several kilometers apparent thickness. All these aspects are difficult to understand without accepting the existence of the tectonic unconformities.

The here first noticed tectonic planes (Pl. I) represent practically continuous mylonitic rock levels; the extent and sense of the shifting on these planes is still difficult to estimate.

a) The mylonitic alignments mapped on the northern slope separate sequences with a similar lithological background, difficult to distinguish on many profiles. Under these conditions their importance will be revealed only by their relationships with the planes pointed out in the Eastern Făgăraş massif. The Sîmbăta Formation seems to correspond to the gneiss sequence described in the "Strîmba Unit", the plane of which is marked by sedimentary deposits; the Albota Formation likely corresponds to the sequence described in the "Bîrsa Fierului Unit" (Nedelcu, Anton, 1984). Savu et al. (1978) mention a tectonic plane situated approximately at the level of the plane in the base of the Sîmbăta Formation as a boundary between the "Făgăraş Series" and the "Cumpăna Series", while Săndulescu (1984) prolongs approximately along the same line the plane of the Bîrsa Fierului Nappe. Dimitrescu (1988), describing the Boldanu quartzite, suggests among other assumptions that the base of the latter might represent a pre-Alpine overthrust with southern vergence as the prolongation of one of the planes pointed out in the Eastern Făgăraş. Balintoni et al. (1986) prolong the plane of the Bîrsa Fierului Unit to the Viştea Valley (cross-cutting the lithological markers described for the first time in the present paper). The Sîmbăta Formation on the northern slope represents the deep structural level of a shearing that affected a lithological sequence showing pre-retromorphic characters similar to those of the Măgura Cîinenilor Formation. The tectonic zone followed on the southern slope between the Măgura Cîinenilor and the Suru Formations is supposed here to represent the upper structural level of the same major shear zone.

b) The mylonitic level at the base of the Suru Formation suggests by its cartographic outline a several kilometers shifting. The fact that it separates in lower position a formation (Moaşa Brook) totally devoid of migmatization from one in which the metasomatic aspects are frequent and even isolated augen gneisses are found (Suru), may indicate a metamorphic unconformity. At least in the regions marked by structural closing an angular unconformity can be also noticed. All these would constitute arguments supporting the overthrust rank of this tectonic discontinuity.

Although the configuration of this plane indicates that it is prior to the planes mapped on the northern slope, no strong arguments have been found in this sense so far. The lack of some significant dynamic neoformations in the rocks making up this plane testifies its formation in the upper crustal domain.

The structural closures in which this plane is involved are suggested by mesoscopic structural elements both in the retromorphic sequences and in the completely unretromorphosed ones. The coaxial deformations during the successive pre-Alpine events is the traditional Romanian explanation. The syn-collisional Alpine shearing was favoured by pre-existing structural directions, and the post-nappe folding only reoriented older synmetamorphic structural elements.

Perceiving the development of stratigraphic and structural knowledge since Ghika-Budeşti's original work, this paper is an attempt of synthesis of the Făgăraş metamorphics structure based on lithologic pre-retromorphic features taking into account the effects of the Alpine mobilization of the Supragetic crust. With a keen awareness that future work will undoubtedly modify, if not refute, the conclusions presented herein, we consider the time is ripe for a new framework on which future discussions can be based.

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FĂGĂRAŞUL CENTRAL ŞI NORDIC — SUCCESIUNI LITOLOGICE ŞI STRUCTURĂ

(Rezumat)

În cadrul polimetamorfitelor „grupului Făgăraş" (sensu Kräutner et al., 1983; Balintoni et al., 1986) sînt separate mai multe unităţi litologice. Este adoptată terminologia stratigrafică cu precizarea: „unităţile litostratigrafice" reprezintă asociaţii litologice distincte în cadrul cărora succesiunile litologice şi grosimile actuale sînt rezultate prin transpoziţia avansată sinmetamorfică.

Unităţile sînt separate conform recomandărilor Ghidului stratigrafic internaţional (Hedberg et al. 1976), avînd în vedere caracterele litologice distinctive preretromorfe. Pe baza reperelor litologice este demonstrată continuitatea pe direcţie a unor unităţi litostratigrafice de-a lungul întregului masiv Făgăraş. Dispoziţia lor spaţială actuală este

efectul deplasării pe plane de discontinuitate tectonică majoră şi al cutărilor ulterioare. Rangul acordat unităţilor litostratigrafice separate este dat de gradul de detaliere impus de necesităţile întocmirii hărţii. Sînt prezentate pentru fiecare unitate sinonimiile cu diviziuni separate anterior în cadrul metamorfitelor din Făgăraş. Unităţile litostratigrafice sînt descrise de la inferior la superior, după poziţia lor geometrică.

Formaţiunea de Pîrîul Moaşa aflorează în nord-vestul şi centrul masivului Făgăraş ca fereastră tectonică sub pînza de Suru. În cadrul ei au fost separate următoarele subunităţi: subformaţiunea de Moldele, care ocupă o poziţie de nucleu de antiformă; prin intermediul unui orizont cu caractere de tranziţie (orizontul de Bîrcaciu) subformaţiunea de Moldele trece gradat la subformaţiunea de Şerbota s.str. care îmbracă antiforma amintită la sud, la vest şi parţial la nord. Subformaţiunea de Şerbota s.str. şi orizontul de Bîrcaciu trec lateral la subformaţiunea de Poiana Neamţului.

Formaţiunea de Suru acoperă tectonic formaţiunea de Pîrîul Moaşei; a fost urmărită de la Olt pînă în Făgăraşul de est. Suferă pe parcurs atît variaţii litologice, precum şi unele complicaţii structurale. Limita sudică cu formaţiunea de Măgura Cîineni este reprezentată printr-o discontinuitate tectonică majoră. La nord-vest este acoperită transgresiv de depozite sedimentare ale bazinului Transilvaniei; începînd din valea Porumbacul spre est, formaţiunea de Suru suportă tectonic formaţiunea de Albota. În cadrul formaţiunii de Albota a fost remarcat „orizontul calcarelor de Piatra Dracului", cu dezvoltare foarte capricioasă.

La nord, formaţiunea de Albota este acoperită tectonic de formaţiunea de Sîmbăta — gnaisică, milonitizată în masă, cu recristalizări dinamice locale la nivelul cloritului. În cadrul formaţiunii de Sîmbăta sînt descrise două repere litologice discontinue, dar cu poziţie constantă în stivă: nivelul cuarţitelor de Boldanu şi orizontul calcarelor de Piatra Albă.

Formaţiunile descrise şi planele tectonice care le separă sînt implicate în cutări ulterioare. Importanţa planelor tectonice prezentate va fi relevată de relaţiile cu discontinuităţile tectonice majore evidenţiate în Făgăraşul de est.

EXPLANATION OF PLATES

Plate III

Fig. 1 — Relict kyanite (ky) in a quartz-sericite aggregate under the Poiana Neamţului Member, the Transfăgăraş road, N +, 100 ×.

Fig. 2 — Disharmonic folds in the mylonites at the base of the Albota Unit, Drăguş edge.

Fig. 3 — Garnet (gr) partly replaced by chlorite (cl) and corroded by undeformed quartz (qz), the Sîmbăta Formation, Ţăriţa edge, N //, 25 ×.

Fig. 4 — Partly chloritized biotite (bi), transverse with respect to the protomylonitic foliation, the Sîmbăta Formation, Ţăriţa edge. N +, 25 ×.

Plate IV

Fig. 1 — Armoured oligoclase (olig.) inclusion in K-feldspar (F_K, partly replaced by albite (Ab) — the Sîmbăta Formation, Sîmbăta Valley, N +, 100 ×.

Fig. 2 — K-feldspar (F_K) peripherically replaced by albite (Ab). M — Sîmbăta Formation, Drăguş edge, N +.

Fig. 3 — Oligoclase clast (olig.) enrolled in the protomylonitic foliation — Sîmbăta Formation, Colţilor edge, N +, 25 ×.

Fig. 4 — Muscovite micaceous septum (mu) incompletely transposed on the protomylonitic foliation — Sîmbăta Formation, Cîrţişoara Valley, N + 25 ×.

Plate V

Fig. 1 — Composite clast — biotite (Bi) + quartz (qz) in the mylonitic foliation — Sîmbăta Formation, Hermeneasa edge, N +, 25 ×.

Fig. 2 — Fish-like composite clast-quartz (qz) + albite (Ab) in the mylonitic foliation — Sîmbăta Formation, Drăguş edge, N +, 25 ×.

Fig. 3 — Protomylonitic foliation specific for the Sîmbăta Formation — Tolfia Hill, N + 25 ×.

Fig. 4 — Intergranular quartz recrystallizations — Sîmbăta Formation, Ucea Valley, N+ 25 ×.

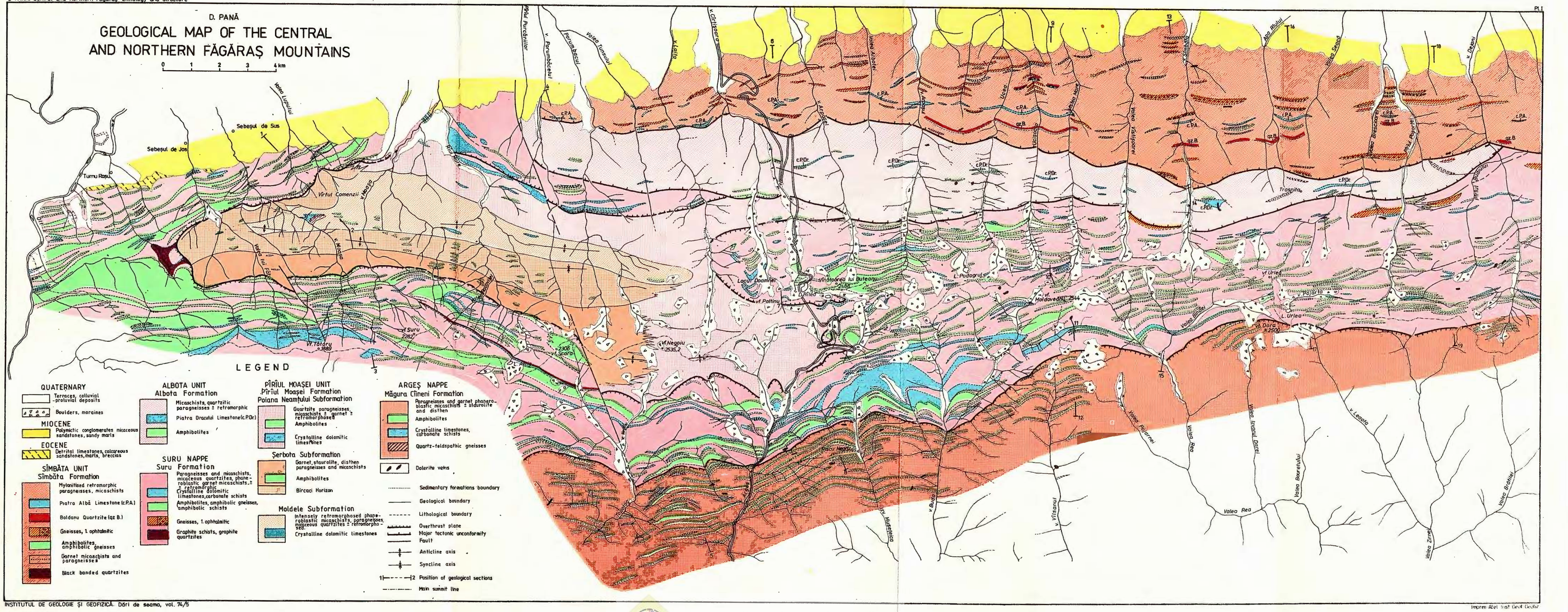
Plate VI

Fig. 1 — Blastomylonitic microlithon — Sîmbăta Formation, Cîrţişoara Valley, N + 25 ×.

Fig. 2 — Detail — the subgrained quartz mylonitic aggregate and very fine acicular chlorite neoformation, N + 100 ×.

Fig. 3 — The same detail — the acicular chlorite neoformation, N // 100 ×.

Fig. 4 — Late quartz recrystallizations — the mylonites at the base of the Suru Formation — Capra Valley, N + 100 ×.



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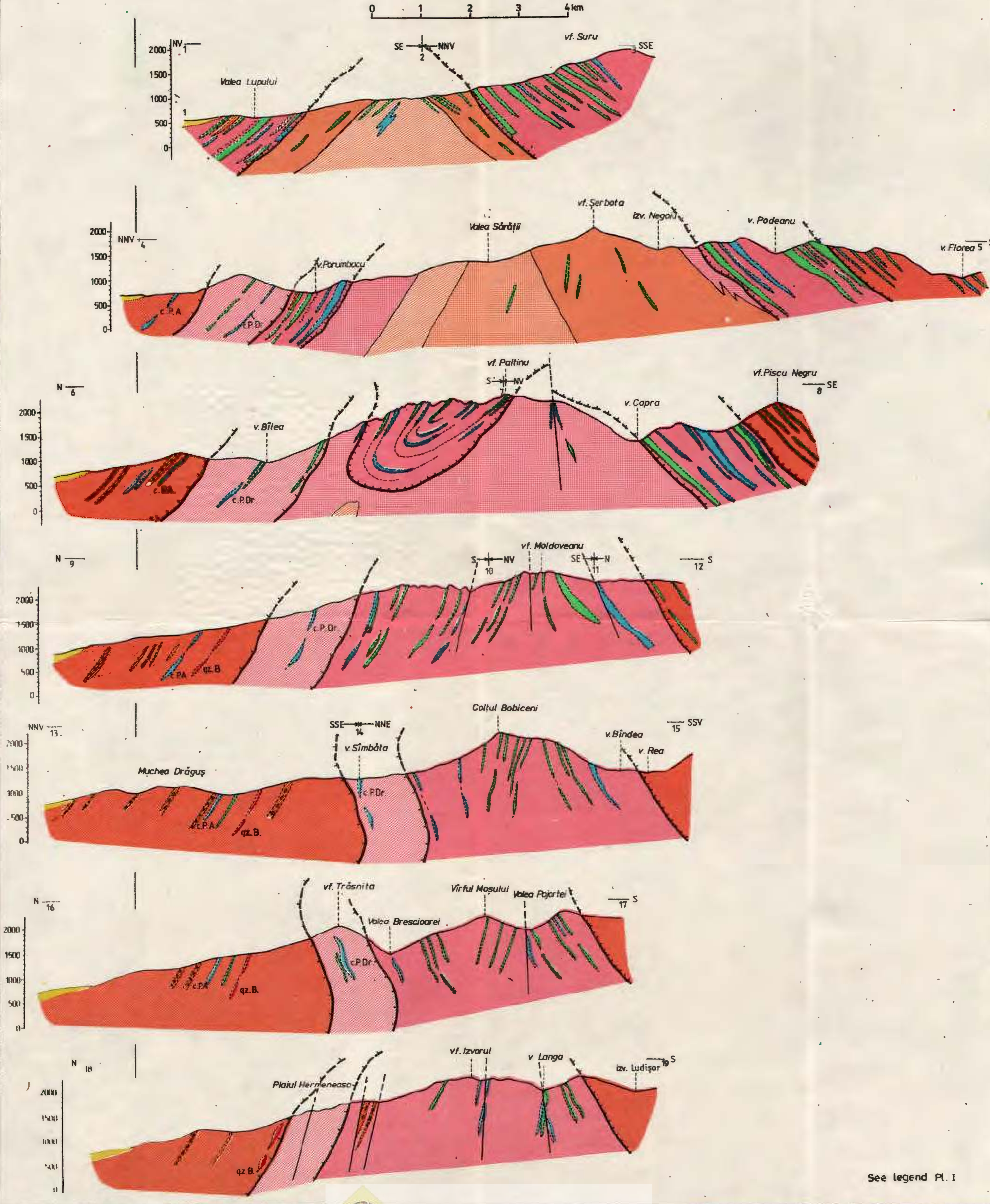
GEOLOGICAL SECTIONS



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5. TECTONICĂ ȘI GEOLOGIE REGIONALĂ

VESTIGES OF A PALEOZOIC OPHIOLITIC SUTURE IN THE SOUTHERN CARPATHIANS AND THE BALKANS[1]

by

H. SAVU [2]

Suture zones. Paleozoic. Ophiolite. Olistoliths. Ultramafics. Oceanic crust. Plate collision. Obduction. Correlation. South Carpathians. Balkans.

Abstract

The vestiges of a Palaeozoic ophiolitic suture in the bend area of the Carpatho-Balkan range are represented by numerous ophiolite slabs and special tectonic relations. The ophiolite slabs occur in the Paring and Țarcu Mountains and in the Iuți-Tisovița, Deli Jovan, Zaglavak and Ciprovci regions. Their geochemical characteristics indicate ultramafic rocks of Alpine type and tholeiitic rocks proceeding from an oceanic crust. The arcuate belt formed of these ophiolite slabs is marked by an ophiolitic suture line situated behind them, which was formed by the collision of the Moesian and Transylvanian convergent plates.

Résumé

Traces d'une suture ophiolitique paléozoïque dans les Carpathes et les Balkans. Les traces d'une suture ophiolitique paléozoïque de l'aire de courbure carpatho-balkanique sont représentées par de nombreux corps de roches ophiolitiques et par des relations tectoniques spéciales. Les corps de roches ophiolitiques se développent dans les monts Paring et Țarcu et dans les régions de Iuți-Tisovița, Deli Jovan, Zaglavak et Ciprovci. Leurs caractéristiques géochimiques indiquent des roches ultramafiques de type alpin et des roches basiques tholéiitiques provenues de la croûte océanique. La ceinture en voûte constituée de ces corps de roches ophiolitiques est marquée par une ligne de suture ophiolitique située derrière celles-ci et s'est formée de la collision des plaques convergeantes moesienne et transylvaine.

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Introduction

In the Carpatho-Balkan area geotectonic and magmatic processes manifested since the Precambrian (Giuşcă et al., 1969). Thus a system of ophiolitic sutures in association with basic and ultramafic rocks occured (Savu, Strusievicz, 1987 — Fig. 1). Such phenomena also took place during the Paleozoic. Numerous vestiges — ophiolite slabs and tectonic relations — which indicate a Paleozoic suture — are to be found in the Southern Carpathians and the Balkans. The present paper deals with the correlation of these data and the mapping out of the line of the old ophiolitic suture.

Ophiolite Slabs

The Palaeozoic bodies occur along the Danubian Autochthon and the Stara Planina Mountains in the Carpatho-Balkan bend zone (Fig. 1). In the Parîng Mountains of the Danubian Autochthon (Southern Carpathians) such rocks were described by Pavelescu and Pavelescu (1965), Schuster (1980), and Berza et al. (1985, unpublished data). These bodies are represented by olistoliths (Savu and Udrescu, 1982; Savu, 1985), of serpentinites and basic rocks assigned to the Latoriţa Formation of Lower Palaeozoic age, metamorphosed in the greenschist facies during the Caledonian or Hercynian metamorphism. This formation consists of metagraywackes, basic metatuffs, quartzites, graphite schists, crystalline limestones, metaquartzkeratophyres and metaradiolarites. The association of these rocks with radiolarites points to a Palaeozoic metamorphosed olistostrome with a volcano-sedimentary matrix.

The ultramafic rocks occur as lenticular bodies maximum 900 m long, in places phacoidized during the folding and metamorphism of the olistostrome. They consist of antigorite-lizarditic serpentinites formed at the expense of the dunites and harzburgites and contain tectonic inclusions of basic rocks, partly rodingitized, an association which indicates olistoliths broken from an ophiolitic mélange (Coleman, 1980). The ultramafic rocks correspond to the Alpine-type peridotites (Fig. 2), a feature which results also from the average contents of MgO, Cr, Ni, and the values of the Cr/Ni and Ni/Co ratios (Savu, Strusievicz, 1987) pointinge to their origin in the peridotitic complex (O_1) of an oceanic crust. The basic rock olistoliths consist of metagabbros and metabasalts which are tholeiitic rocks (Fig. 3). On the diagram in Figure 4 these rocks are situated in the high-Ti ophiolite field, in which the basic rocks from the mid-ocean ridges usually plot (Serri, Saitta, 1980). All these data show that the ultramafic and basic rock bodies from the Palaeozoic crystalline schists in the Parîng Mountains are olistoliths proceeding from a dismembered oceanic crust, obducted and incorporated in a Palaeozoic olistostrome with a volcano-sedimentary matrix, later on metamorphosed.

Such small bodies of basic and ultramafic rocks occur also along the Danubian Autochthon towards the west, some of them even on the margin of the Getic Nappe, as those to the south of Vulcan (Pavelescu, Pavelescu, 1965). In the light of these data, it is not impossible that the ultramafic olistoliths from the Jurassic formations on the Craiu Valley

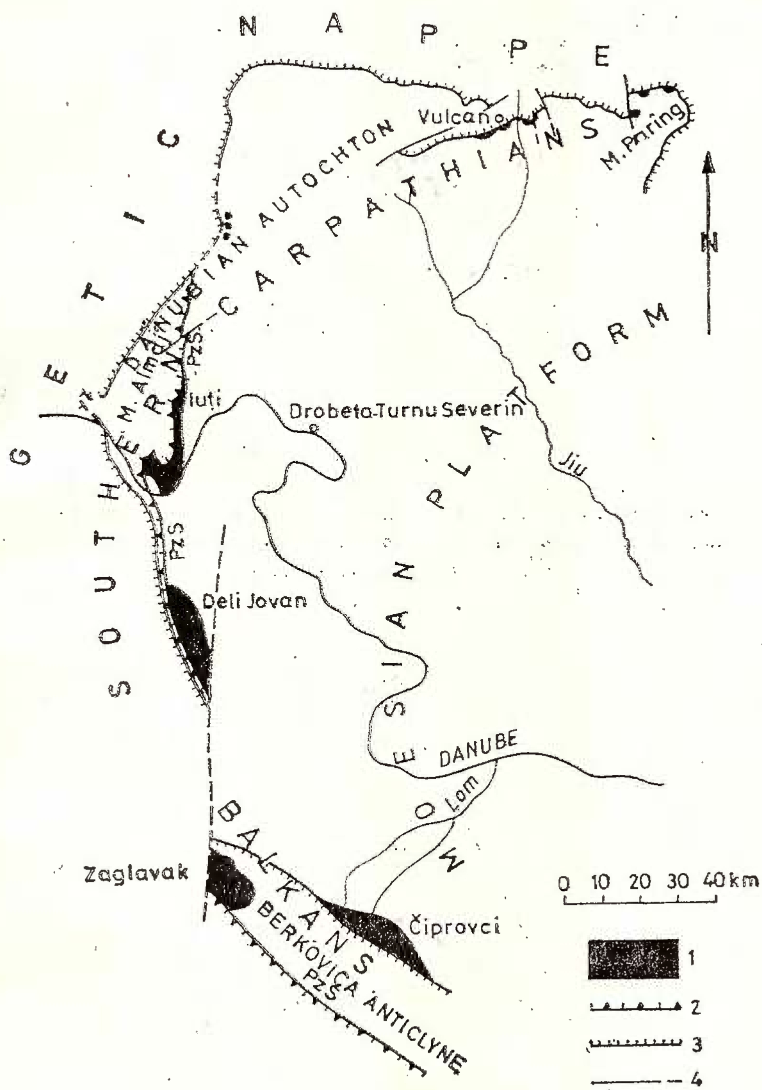

Fig. 1 — Palaeozoic ophiolites and tectonic relations in the Southern Carpathians and the Balkans (according to the Geological map of Romania, Yugoslavia and Bulgaria — completed).
1, ophiolite slabs ; 2, ophiolitic suture line (PzS) ; 3, thrust ; 4, fault.

(Savu et al., 1986) might represent also olistoliths torn from a Palaeozoic ophiolite slab like the Iuţi-Tisoviţa slab (Fig. 1) — situated on the ophiolitic suture — which was removed during the Alpine cycle, broken and completely dismembered.

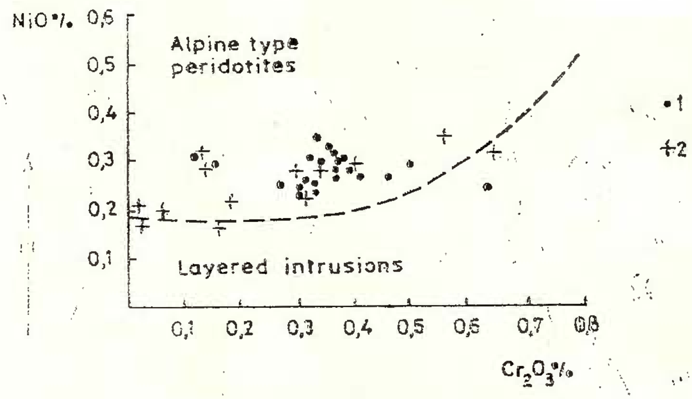

Fig. 2 — NiO—Cr_2O_3 diagram (Malpas, Stevens, 1977).
1, ultramafic rocks from the Paring Mountains (Berza et al., 1985, unpublished data); 2, ultramafic rocks from the Iuţi-Tisoviţa ophiolite slab (Udubaşa et al., 1985, unpublished data; Strusievicz et al., 1986, unpublished data).

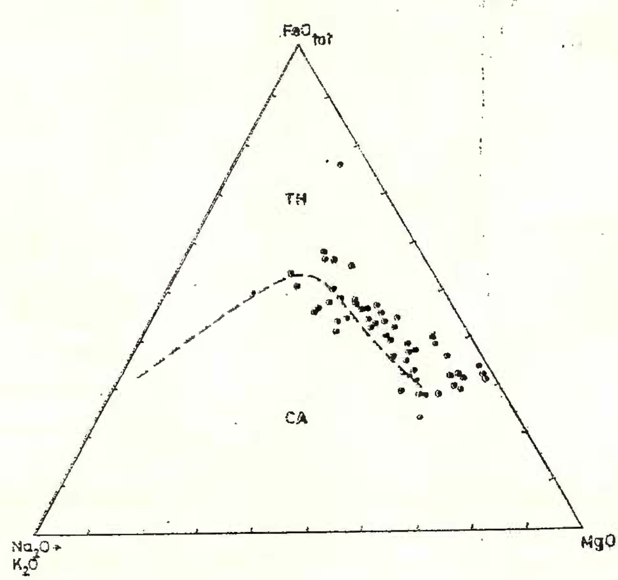

Fig. 3 — FeO tot-MgO—Na_2O + K_2O diagram (Irvine, Baragar, 1971) for the basic rocks of the Paring Mountains (Pavelescu, Pavelescu, 1965), Iuţi gabbros (Bercia, Medeşan, 1976). Deli Jovan gabbros (Terzich-Petkovich, 1960) and the Ciprovci basic rocks (Haidutov et al., 1985).

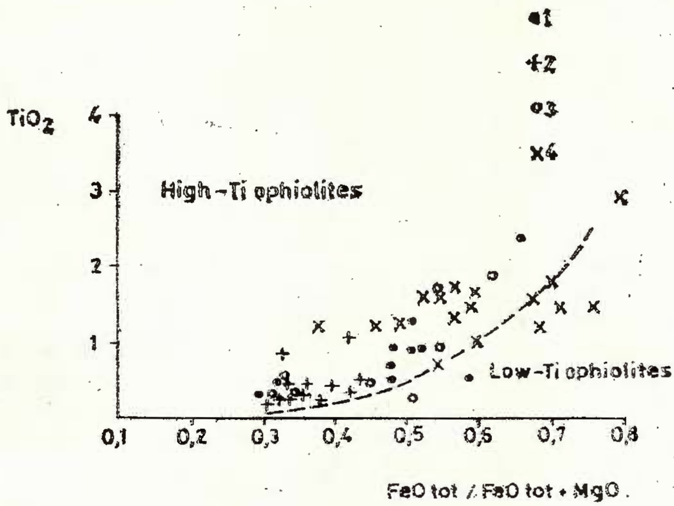

Fig. 4 — TiO_2—FeO tot + MgO diagram (Serri, Saitta, 1980) for the basic rocks from : 1, Paring Mountains ; 2, Iuţi-Tisoviţa ; 3, Deli Jovan ; 4, Ciprovci (the same sources as in Figure 3).

The Iuţi-Tisoviţa ophiolite slab in the South Banat (Fig. 1) constitutes a mass of basic and ultramafic rocks (Iuţi Unit), tectonically situated between the Ogradena Unit towards the east and the Almaj Unit towards the west (S. Năstăseanu and V. Iancu, oral communication). These ophiolites had been considered successively as a Palaeozoic igneous body covered by Permian deposits (Codarcea, 1940 ; Bercia, Bercia, 1962), a mass of ophiolite rocks similar to the ones from the eastern part of Yugoslavia (Karamata et al., 1980), a Caledonian allochthonous ophiolitic complex (Mărunţiu, 1984), a fragment of oceanic crust tectonically emplaced (Udubaşa et al., 1985, unpublished data) or a lens of ophiolites tectonically caught on the thrust plane between the two adjacent tectonic units (Savu, 1985). The ophiolite slab extends on a north-south area of about 30 km. It consists, starting from the east towards the west — that is from the lower part of the slab — of a horizon formed of the Plavişeviţa epigabbro, overlain by a horizon made up of ultramafic rocks showing an obvious magmatic lamination, after which, in the south-west of the slab, the Iuţi gabbros succeed, consisting of diallage gabbro, olivine gabbro, norite gabbro etc.

The Plavişeviţa epigabbro (flaser gabbro) presents a blastomylonitic texture and clinopyroxene uralitization and plagioclase saussuritization phenomena. These features represent the results of the dynamothermal metamorphism in the greenschist facies, which took place during the ophiolite slab obduction from the Palaeozoic oceanic crust over the margin of the Moesian microcontinent to the east, from which the present Ogradena Unit came off.

Different authors (Karamata, 1980; Gealey, 1980; Jamieson, 1980) reported that during the emplacement of the ophiolite slabs or sheets a dynamo-thermal metamorphism of different degrees took place at their bottom. It was brought about by the dynamic effects and by the temperature resulting from the friction heat and the heat content of the obducting ophiolite rocks. The metamorphism in the greenschist facies and the absence of the blueschist facies indicate — according to Ernst (1974) and Karamata (1980) — that the Iuți-Tisovița ophiolite slab could be obducted from a narrow ocean zone characterized by a high heat flow.

In the ultramafic horizon Bercia and Bercia (1962) described dunites and apodunitic serpentinites, apoharzburgitic serpentinites and rarely apowehrlitic serpentinites, rocks which have been transformed in places in talc and listwanites (Udubașa, 1970). The ultramafic rocks present geochemical characteristics resembling the Alpine-type peridotites (Fig. 2). The Iuți gabbros are tholeiitic rocks (Fig. 3) situated on the diagram in Figure 4 in the field of high-Ti ophiolites. All these geochemical features indicate that the ophiolitic rocks of Iuți-Tisovița originate in an oceanic crust. The ophiolite slab structure, in which horizons of layered basic and ultramafic rocks alternate, shows that it includes a part of the transition zone from the peridotitic complex (O_4) up to the gabbroic complex (O_3) of this ocean crust.

South of the Danube, in Yugoslavia, two important bodies of basic and ultramafic rocks occur (Karamata et al., 1980): one of them is situated in the southern extremity of the Carpathians — the Deli Jovan Massif — and the other one in the western extremity of the Balkans, in the Stara Planina Mountains — the Zaglavak Massif.

The Deli Jovan Massif (25 km long and 8 km wide) was studied in petrographic and structural respect by Terzich-Petkovich (1960), who pointed out its tectonic relations with the country rocks and supposed the massif is not in its initial emplacement. Consequently it is an obducted ophiolite slab like the Iuți-Tisovița slab. Like the latter, it presents a layered structure in which layers and laminae of basic and ultramafic rocks alternate, constituting an important petrographic variety.

Terzich-Petkovich (1960) showed that the ultramafic rocks lie on the east and north-east massif border, while the basic ones and especially the diabases are developing in its south-western part. Taking into account these observations and the layered structure of the ophiolite slab, one could admit that it was torn from the gabbroic complex (O_3) of the oceanic crust, because it presents in its base rocks enriched in olivine which make up the transition down towards the peridotitic complex (O_4) and in the upper part gabbros crossed by diabase dykes which announce the beginning of the sheeted dyke complex (O_2) of the oceanic crust. A proof that this ophiolite massif does not represent a layered intrusion is the mode in which its basic rocks plot on the diagrams in Figures 3 and 4, from which it results that they proceed from a tholeiitic magma and have characteristics of high-Ti ophiolites, similar to the basic rocks in the mid-ocean ridges (Serri, Saitta, 1980).

Taking into account the position and structure of the massif, it can be concluded that the ophiolite slab was obducted from south-west towards north-east and it was sliding on the serpentinized ultramafic rocks from the lower horizon. The gabbro layers within the slab were also affected by amphibolization and saussuritization, probably under the influence of the dynamo-thermal metamorphism in the greenschist facies which took place during the ophiolite obduction.

Another Palaeozoic ophiolite slab in the Balkans is represented by the Ciprovci ophiolites (Bulgaria), situated at the springs of the Lom River, in the Stara Planina Mountains (Haidutov et al., 1985). This slab lies at the base of the diabase-phyllitoid complex – probably an equivalent of the Latoriţa Formation – and comprises an almost complete sequence of an ophiolite suite, represented by the ocean floor basalt complex (O_1) in pillow lava facies (Rupska Unit), the sheeted dyke complex (O_2) or the Manastir Unit and a lower complex of layered cumulates (Kopilovici Unit) consisting of layers of gabbros and olivine-clinopyroxene rocks, which might represent the gabbro complex (O_3) of the oceanic crust. The ophiolite rocks have been affected by the ocean floor metamorphism in greenschist facies (Haidutov et al., 1985).

The quoted authors showed that the Ciprovci ophiolites are similar to the ophiolites from the other regions in the world. The basic rocks are tholeiitic rocks (Fig. 3) and show geochemical characteristics very similar to the basic rocks in the present mid-ocean ridges (Fig. 4).

Ophiolitic Suture Line

The Palaeozoic ophiolite slabs are distributed along an arcuate belt which extends from the Southern Carpathians to the Balkans, being conformable to the bend of this range and parallel to the western margin of the Moesian Platform (Fig. 1). Behind these ophiolite slabs, in places in contact with them, there is a system of tectonic lines with thrust characteristics, which might represent the line of the old ophiolitic suture (Fig. 1, PzS). Behind this line the Alpine thrust plane of the Getic Nappe follows from the west (Fig. 1). Măruntiu (1984) reported that in the west of the Southern Carpathians the ophiolitic suture, on which the ophiolite slab of the Iuţi-Tisoviţa lies, has an important northward extension. In our opinion, it followed the arcuate regional structural trend of the Danubian Autochthon. In the north of the Southern Carpathians it probably coincides with the big gravimetric anomaly pointed out by Airinei (1980), which could be determined by an important mass of Palaeozoic ophiolites like the Iuţi-Tisoviţa one, situated under the Getic Nappe.

In the north of the Almaj Mountains the ophiolitic suture is overlain by the Getic Nappe; farther its line – if it does not merge in the thrust plane of this nappe – seems to maintain, however, near its border. Evidences in this respect are the serpentinitic slabs which occur in the Danubian Autochthon, some of them just on the Getic Nappe border and others in the Palaeozoic formations from Ciunget to the big Iuţi-Tisoviţa ophiolite slab on the Danube and farther towards the Deli Jovan slab. In the Danube area the thrust plane of the Getic Nappe moves off the ophiolitic suture line up to west of the Almaj Mountains; in Yugoslavia it again gets near this line (Fig. 1).

South-east of the Deli Jovan ophiolite massif the two tectonic planes seem to be cut by a fault trending north-south and strongly shifted by it (Fig. 1). East of this fault, in the Stara Planina Mountains, the tectonic image changes a little, for it seems that there the ophiolitic suture plane is the one situated south-west of the Berkovica Anticlinorium, on the plane of which the Zaglavak ophiolite massif is situated (Fig. 1), and the tectonic plane in front of this anticlinorium, on which the ophiolites of Ciprovci lie, might be a secondary thrust plane, possibly an Alpine one or an old plane reworked during the Alpine cycle.

Conclusions

The Palaeozoic ophiolite slabs in the Southern Carpathians and the Balkans represent fragments torn from a dismembered oceanic crust of Lower Palaeozoic age.

The cumulate structure with thick monomineral sequences of adcumulus type as well as the dynamo-thermal metamorphism in greenschist facies suggest their origin in a narrow ocean zone of high heat flow.

By the ocean zone closure fragments of unconsumed oceanic crust have been obducted towards the east on the microcontinent, so that due to collision, some ophiolite slabs were caught on the Benioff plane, thus marking the ophiolitic suture plane.

Due to the very complicated geological processes in the bend area of the Carpatho-Balkan range, an imbricate tectonic structure took place. Under these conditions the Palaeozoic ophiolitic suture plane — as well as that of other structures — underwent the following changes: it got a strong dipping, in places it was shifted by different faults and covered by new sedimentary deposits and tectonic units, so that the suture line is very difficult to trace nowadays.

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VESTIGIILE UNEI SUTURI OFIOLITICE PALEOZOICE ÎN CARPAŢII MERIDIONALI ŞI BALCANI

(Rezumat)

Vestigiile unei suturi ofiolitice paleozoice prezente în zona de curbură a catenei carpato-balcanice sînt reprezentate prin numeroase mase (slabs) de ofiolite obduse dintr-o crustă oceanică dezmembrată și prin relații tectonice specifice.

Zona oceanică din care au provenit ofiolitele era situată aproximativ pe amplasamentul actual al catenei carpato-balcanice, la vest de placa moesică. La închiderea acestui ocean fragmente de crustă oceanică neconsumate au fost obduse pe continentul moesic, astfel că datorită coliziunii, unele mase de ofiolite au fost prinse pe planul Beniotf, ele marcînd astfel planul suturii ofiolitice din figura 1.

Masele de ofiolite apar în munții Parîng, Țarcu și în regiunile Iuți-Tisovița, Deli Jovan, Zaglavak și Ciprovci (fig. 1). Aceste ofiolite se prezintă fie ca blocuri exotice (olistolite) înglobate într-o olistostromă metamorfozată (Parîng), fie ca unități tectonice obduse (Iuți-Tisovița, Ciprovci) sau sub formă de corpuri ofiolitice de dimensiuni mai mari sau mai mici. Unele dintre ele (Iuți-Tisovița, Deli Jovan) prezintă în bază fenomene de metamorfism dinamotermic în faciesul șisturilor verzi, care s-a produs în timpul obducției pe continent.

Textura cumulatelor cu secvențe monominerale groase de tip adcumulus, ca și metamorfismul dinamo-termic în faciesul șisturilor verzi arată că zona oceanică în care s-au format ofiolitele a fost îngustă și a avut un flux termic ridicat.

Caracteristicile geochimice ale ofiolitelor bazice și ultrabazice (bazalte, dolerite, gabbrouri, peridotite, dunite, serpentinite etc.) indică roci ultrabazice de tip alpin și roci tholeiitice de fund oceanic (fig. 2, 3, 4), ce provin din crusta oceanului paleozoic menționat.

Centura arcuită schițată de corpurile de ofiolite obduse este marcată și de o linie de sutură ofiolitică situată în spatele lor (fig. 1). Aceasta s-a format datorită convergenței celor două plăci de la est și vest de zona oceanică paleozoică, respectiv plăcile moesică la est și carpatică la vest.

Datorită proceselor geologice foarte complicate care au avut loc în zona de curbură a catenei carpato-balcanice ia naștere o tectonică imbricată, în care condiții planul suturii ofiolitice paleozoice — ca și alte structuri — a suferit următoarele modificări : a căpătat o înclinare mai mare și pe alocuri a fost secționat și decroșat de diferite falii transversale mai noi și acoperit de depozite sedimentare mai recente (fig. 1), astfel că astăzi este foarte dificil să urmărim vechea linie a suturii ofiolitice paleozoice.

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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

STRUCTURE OF DANUBIAN TECTONIC UNITS BETWEEN NOVACI AND THE PARÎNG PEAK WITH SPECIAL REGARD TO THE AMPHIBOLITE DRĂGŞAN SERIES[1]

by

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Danubian Units. Metamorphic rocks. Volcanoclastics. Ultramafics. Amphibolites. Major elements. Trace elements. Retrograde metamorphism. Thrust fault. Tectogenesis. South Carpathians — Crystalline Danubian Realm — Parîng Mountains.

Abstract

There are three major tectonic units in the Novaci-Parîng Peak region from the Danubian Autochthon (South Carpathians) as follows : Crasna, Sadu and Parîng-Vulcan. The latter unit thrusts over the Sadu Unit probably at the end of the Assynthic movements or in a pre-Hercynian movement and both units thrust over the Crasna Unit during the post-Liassic movements. During the Hercynian and post-Liassic movements the infrastructure was retromorphosed and structures (scales) that affected also the pre-Hercynian overthrust plane of the Parîng-Vulcan Unit over the Sadu Unit formed within it. The small ultramafic bodies from the Drăgşan Series show features characteristic of the peridotites from an oceanic crust. The chemical composition of the amphibolites from the Drăgşan Series and their content of trace elements are influenced by several factors, so that the tectonic setting of the pre-metamorphic basic volcanic rocks cannot be established for certain, although these rocks might have represented an oceanic crust.

Résumé

Structure des unités tectoniques danubiennes entre Novaci et le sommet Parîng à considérations sur la série amphibolitique de Drăgşan. Il y a dans la région de Novaci-sommet Parîng de l'autochton danubien (Carpathes Méridionales) trois unités tectoniques majeures : de Crasna,

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de Sadu et de Paring-Vulcan. La dernière unité a charrié sur l'unité de Sadu probablement vers la fin des mouvements assyntiques ou bien pendant un mouvement préhercynien, toutes les deux unités chevauchant l'unité de Crasna durant des mouvements postliasiques. Au cours des mouvements hercyniens et postliasiques l'infrastructure a été rétromorphisée et des structures s'en sont formées ; elles ont affecté également le plan de charriage préhercynien de l'unité de Paring-Vulcan surmontant l'unité de Sadu. Les petits corps d'ultrabasites de la série de Drăgșan présentent des caractéristiques des roches de l'horizon péridotitique (O_4) d'une croûte océanique. La composition chimique des amphibolites de la série de Drăgșan et leur contenu en éléments trace sont influencés par plusieurs facteurs, de sorte qu'on ne peut pas établir avec certitude la situation tectonique des roches volcaniques basiques prémétamorphiques bien qu'elles puissent représenter une croûte océanique.

Introduction

On the occasion of the investigations for the elaboration of the geological map, scale 50,000, sheet Novaci, we revised and partly mapped the region in the Paring Mts. situated between the Gilort Valley and the eastern slope of the Galbenu Valley, from Novaci to the Paring Peak (Savu et al., 1985). The researches in this region represent the eastward continuation of the studies carried out in the Danubian Autochthon by some of the authors of the present paper (Savu, 1972 ; Savu et al., 1973 a ; 1973 b, 1974). We shall further present the new data obtained, especially the tectonic ones, which will provide a complete image of the geological structure of the Danubian Autochthon (Pl. I).

The southern part of the Paring Mts. was previously investigated by Paliuc (1973), Manolescu (1937) and Ghika-Budești (1940). Later on new data were obtained by Bercia and Bercia (1958, 1959), Trifulescu et al. (1964), Dragomir and Arsenescu (1965) and Arsenescu (1966) who carried out prospection works in the region for various mineral resources. Recently Savu (1970, 1972) and Savu et al. (1973 a ; 1973 b ; 1974 ; 1976) remapped the whole region on scale 1 : 10 000 and 1 : 25 000 investigating the rocks from the petrological and geochemical viewpoints. The Mesozoic formations were investigated by Huică and Sîrbu (1964), while the Neogene ones by Marinescu (1965).

Danubian Tectonic Units in the Region

In 1980 Savu and Hann (1982) showed based on the observations on the Muntele Mic area that the imbricated tectonics of the Danubian Autochthon is polyphasic and much more complicated than was previously known, pre-Hercynian, Hercynian and Alpine structures existing in this area. In 1981 Savu (in Savu et al., 1984) establishes the pre-Hercynian overthrust relationships existing between the Drăgșan Series (Cerna-Ciunget Unit) and the Lainici-Păiuș Series (Godeanu-Olteț Unit) ; up to that time all the tectonic accidents known in the region were considered of Alpine age. Later, in the same year, Kräutner et al. ([1]981) publishes a map to the Guide for the Congress of the Carpatho-Balkan Geological Association in which structures varying in age are presented in the Danubian Autochthon, changing the denomination given by Savu et al. (1984) to the two pre-Hercynian mentioned tectonic units.

The structure of the Danubian Autochthon is determined by three tectonic units varying in age, which are facing to SSE, towards the Moesian Platform which is subducted under the Transylvanian Plate (Savu et al., 1972 a ; Savu, 1980), the shearing of which gave rise to the tectonic units from the Carpathians. In the investigated region, the following tectonic units are represented from bottom to top : the Crasna Unit, the Sadu Unit and the Paring-Vulcan Unit (Savu et al., 1985).

Crasna Unit. This is the lowermost unit, which crops out only between Crasna and Drăgoești — west of the region figured on the presented map. It is overthrust from the north by the crystalline schists of the Lainici-Păiuș Series and the tardeorogenic granitoids from the Cărpiniș-Novaci massif (Savu, 1972 a ; Savu et al., 1972 b ; 1973 b) from the Sadu Unit and transgressively overlain from the south by the Neozoic sedimentary deposits of the Pericarpathian Depression. In our opinion, this unit extends eastwards in the Novaci region, but it does not outcrop, being overlain by younger deposits. Also, it extends westwards, under the crystalline schists and the Arșeni, Sadu and Porceni granitoids from the Sadu Unit, up to Schela, farther westwards along the external margin of the Danubian Autochthon, where it does not outcrop from under the Sadu Unit and the more recent sedimentary deposits. As it was found at Drăgoești (Savu et al., 1982 b) the basement of the unit consists of crystalline schists similar to those from the Lainici-Păiuș Series. This basement is overlain by deposits of the Schela Formation bearing anthracite, which are considered to be Liassic in age (Huică, Sîrbu, 1964). The presence of these deposits shows that the overthrust of the Sadu Unit took place after the Liassic.

In its present position, the Crasna Unit appears to be in an autochthonous position. But it is not known, without a checking by drillings, whether the Crasna Unit is the unit of the Transylvanian Plate, which is in collision with the Moesian Plate, in accordance with Savu's model (1985) or it overthrusts another unit tectonically contacting the Moesian Plate as was suggested by Savu et al. (1987) ; at present, these units are overlain by younger sedimentary deposits.

Sadu Unit. This unit, pointed out by Savu et al. (1984) under the name of the Godeanu-Olteț Unit covers most of this region (Pl. I). It comprises also the Schela Nappe outlier, which is thrust over the formation of the same name (Drăghici et al., 1967 ; Savu, 1970). Towards SSE it thrusts over the Crasna Unit and is in its turn overthrust by the Paring-Vulcan pre-Hercynian Unit from NNW. These two tectonic units, which have a common Paleozoic sedimentary cover (Savu et al., 1984) formed together during the Alpine cycle what Berza et al. (1985) named the Lainici Alpine Unit, which in fact overthrusts the Crasna Unit. The southern margin of this overthrust unit can be followed up to Stăncești, but it continues westwards under the younger sedimentary deposits.

This unit consists of the Lainici-Păiuş Series, the granitoids from the Suşiţa synkinematic pluton, those from the Cărpiniş-Novaci tardekinematic pluton, showing a zonal structure as well as the Paleozoic formations from under the scale on the crest west of the Roman Valley (Pl. I). According to the chemical and mineralogical composition, the plutonic rocks represent collision granitoids (Harris et al., 1986) generated by the collision between the Moesian Plate and the Transylvanian Plate (Savu et al., 1972 a). The Lainici-Păiuş Series represents a detrital metamorphic formation of psammitic and pelitic nature. It was folded and metamorphosed during the Assynthic movements, simultaneously with the Drăgşan Series, over which it formed in the Precambrian geosyncline. Although it was affected by retromorphism processes, especially Hercynian (Savu, 1972), from the parageneses indicated by the relict minerals, it was found that it was initially metamorphosed in the conditions of the almandine amphibolite facies to the greenschist facies. This series consists mainly of paragneisses and almandine, biotite and muscovite feldspathic quartzites interbedded with biotite quartzites, graphite schists, paragneisses and micaschists, limestones and crystalline dolomites. Near the granitoid rocks plutons and especially near the Cărpiniş-Novaci one, the metamorphism is more intense, as the PT regional metamorphism conditions combined with those of the synkinematic contact metamorphism resulted in a special, Danubian type metamorphism (Savu, 1970). In the zones affected by this type of metamorphism, especially in the strip from the Arşeni-Olteţ syncline, situated between the two plutons, sillimanite $\pm$ cordierite $\pm$ andalusite gneisses form (Savu, 1972; Savu et al., 1973 a; 1973 b). Silicate limestones (forsterite, salite etc.) also form on the Gilort and Galbenu Valleys under the influence of the Danubian type metamorphism. This series includes very rare amphibolite interbeds such as those on the Teiş Valley and east of the Măgura Mount.

The Lainici-Păiuş Series is intensely migmatized in places. Owing to the arteritic migmatization process brought about by the granitoid solutions, various migmatite types get formed in this series (Savu et al., 1974). On the Aninişu Mare Valley, situated west of our map, the nebulitic migmatites associate with kinzigite rocks enriched in large cordierite crystals, partly pinitized, similar to those described by Härme and Perttunen (1971) in Finland which were investigated by one of the authors of this paper round Helsinki. These rocks are crossed by ptygmatic veins of granitoid rocks (Savu, 1972). They might represent an instance of "basic front" associated with the migmatization process (Savu et al., 1973 b).

The Sadu Unit presents its own internal, pre-Alpine tectonics, formed during the Assynthic folding and metamorphism of the crystalline schists of the Lainici-Păiuş Series and the emplacement of the granitoid plutons, and, later on, during the Hercynian movements that determined the metamorphism of the Paleozoic formations and affected the Assynthic infrastructure. Among the Assynthic structures one can distinguish the Nedeiu-Suşiţa-Tismana anticline, in the axial zone of which, the Suşiţa synorogenic granitoid pluton was emplaced (Savu, 1970; 1978). Farther south the Arşeni-Olteţ syncline consisting of the Lainici-Păiuş Series, intensely migmatized, is formed (Pl. II, Figs. 1 and 2). Farther south, the Olteţ-Novaci-Cărpiniş anticline follows, in the axial zone of which the Cărpiniş-Novaci tardeorogenic granitoid pluton is present (Savu, 1972, 1978; Savu et al., 1973 b). As results from the Novaci map, scale 1 : 50,000, this structure is cut by the overthrust plane of the Sadu Unit over the Crasna Unit.

During the Hercynian movements, in the formed structures, the Assynthic infrastructure, the Lainici-Păiuş Series respectively, was also involved; the crystalline schists of this series were retromorphosed, the granitoid rocks being metamorphosed in the greenschist facies (Savu, 1970). This last process manifested 290 m.y. ago (Soroiu et al., 1970) as shown by the radiometric age of a sericite-chlorite orthoschist formed at the expense of the Suşiţa granite (Savu, 1972). In this way a diagonal scale forms, considered by Berza et al. (1985) as the Vîlcan-Pilugu Unit, which is underlain by rocks belonging to the Schela Formation on the crest between the Romanu and Setea Mare Valleys, a scale which would be linked south-westwards with the structures of the present syncline from Rafailă, in the Jiu Valley. The fact that the Paleozoic formations from this syncline are overlain also by Liassic deposits indicates that the Hercynian structures were partly resumed also during the Alpine movements. In the north-eastern corner of the region (Muşetoaia Peak) the above-mentioned scale affected under a sharp angle also the pre-Hercynian tectonic contact between the Paring-Vulcan and Sadu Units, being underlain by Paleozoic phyllites described by Berza et al. (1985) as belonging to the Schela Formation (Pl. I).

Parîng-Vulcan Unit. This pre-Hercynian Unit pointed out by Savu in 1981 (Savu et al., 1984) in the west of the Parîng Mts. as the Cerna-Ciunget Unit comprises in this region the Drăgşan Series, which is intruded by the granitoids of the Parîng pluton. It overthrusts the Lainici-Păiuş Series from the Sadu Unit.

As results from the Novaci map, scale 1 : 50.000 as well as from the here-presented map, the overthrust nappe prolongs from the Jiu Valley (Savu et al., 1984) south of the Tărtărău Peak, the springs of the Gilort Valley, south of the Păpuşa Peak, the springs of the Galbenu Valley and south of the Muşetoaia Peak, where it is affected by a fault trending NW—SE; farther it follows a NE direction, towards the Latoriţa Valley. A clear argument supporting the tectonic position of the two units is also the fact that the direction of the overthrust line between them forms a sharp angle with the direction of the above-described Assynthic structures from the Sadu Unit. As results from the Schela map, scale 1 : 50,000 (Savu et al., 1984 b) the Paleozoic formations from the Vulcan Mts overlie transgressively the overthrust plane, which indicates the pre-Hercynian age of the overthrust (Savu et al., 1984 a).

Along the overthrust plane, the rocks of both units being in contact, but especially those of the Lainici-Păiuş Series from the Sadu Unit, are intensely mylonitized and laminated. Mylonitization shows various aspects in different rocks. Within the amphibolites from the Drăgşan Series, it gradually advances to their complete lamination, when lenticular structures occur in these rocks. Initially amphibole is chloritized, while plagioclase is sericitized; secondary minerals are oriented in the lamination plane which is parallel to the overthrust plane. This process advances as we get nearer the overthrust plane, giving rise to schistous rocks consisting entirely of chlorite and relict amphibole, impregnated with fine iron oxide grains. Within the laminated amphibolic gneisses the plagioclase and amphibole crystalloblasts are transformed into a strongly foliated groundmass, in which the recrystallized quartz grains are transformed into very elongated lenses; the hard and brittle almandine prophyroblasts pierced by quartz grains remain as such, but are rolled in places and usually crossed by numerous parallel and sometimes broken fissures, leaving behind them a trail of grains varying in size. Within the rocks of the Lainici-Păiuş Series, which do not include garnet porphyroblasts, the rocks are strongly laminated, showing a uniform schistous or lenticular structure. The porphyric granitoids from the apophyses intruding the Drăgşan Series are also affected by lamination, that transforms their groundmass which is granular and consists of quartz, plagioclase, biotite and K-feldspar into a schistous mass formed of fine quartz, sericite and chlorite grains, in which small recrystallized quartz lenses can be noticed. It is worth noting that the K-feldspar megacrysts exhibiting a microcline or microcline-perthite texture are not affected by this lamination process, their deformation being hardly visible, as is shown by their texture as well as by the undisturbed plagioclase and biotite fine crystals.

Later transverse faults affect and shift the overthrust plane of the Parîng-Vulcan pre-Hercynian Nappe.

Structure and Petrography of the Drăgşan Series Amphibolites

The Drăgşan amphibolites Series is the oldest metamorphic series of the Danubian Autochthon geological formation; it was formed in the Upper Precambrian eugeosyncline (Giuşcă et al., 1969) covering approximately the present Carpathian area, before the Lainici-Păiuş Series. It represents a volcano-sedimentary series metamorphosed during the Assynthic movements. This series consists of centimetric to decimetric bands (Pl. III, Figs. 1, 2), formed of orthoamphibolites, amphibolic gneisses, garnet amphibolites and more rarely fine biotite gneisses alternating between them; this leads to a banded structure, very characteristic of this series. In places small serpentinized ultramafic rock bodies are hosted in the Drăgşan Series especially between the Muşetoaia Peak, in the extension of the alignment of serpentinized ultramafic rocks in the western part of the Parîng Mts. (Savu et al., 1984 a). They belong to two petrographic types: harzburgites and lherzolites, the latter type occurring more rarely. The rocks consist of a schistous serpentinite mass, formed of antigorite-lizardite, which is impregnated with fine secondary magnetite grains. Some rocks are marked by a banded structure in which very thin serpentine bands alternate with extremely fine bands bearing frequent magnetite grains. In these rocks the bastite pseudomorphoses are also lenticular and elongated. In the rocks originating from harzburgites bastite pseudomorphoses after an orthopyroxene can be noticed, while in those resulted from lherzolites both bastite pseudomorphoses and deformed clinopyroxene crystals occur. Owing to the granitoid solutions the ultramafic rocks are in places altered in talc.

As the serpentinite bodies are associated with a great mass of orthoamphibolites, they may be considered to represent, as in the case of the ultramafic rocks of the Ielova and Măru Series, small intrusions of ultramafic magma within an oceanic crust, which were later subsequently metamorphosed during the Assynthic movements, together with the Drăgșan Series rocks.

In the upper basin of the Sadu Valley, situated in the north-western part of the region figured on the Novaci map, scale 1 : 50,000, in the Drăgșan banded amphibolites small limestone and dolomite limestones are interbedded (Paliuc, 1937), which represent the same horizon in which the crystalline limestones and the silicate limestones lie, that occur in the Drăgșan Series on the Jiu Valley and were figured on the Schela map, scale 1 : 50,000 (Savu et al., 1984 b).

The Drăgșan Series is crossed by the Parîng granitoid pluton (Savu et al., 1976) which develops in the north (Pavelescu, 1970). Only the southern and south-western apophyses of this pluton reach the investigated region.

The Assynthic metamorphism that affected both the Drăgșan Amphibolite Series and the Lainici-Păiuș detrital Series was achieved approximately at the level of the almandine amphibolite facies. Subsequently both series were affected by retromorphism processes that manifested in three epochs : a) at the end of the Assynthic movements when the thrust of the Parîng-Vulcan Unit over the Sadu Unit took place; b) during the Hercynian movements; c) more rarely during the Alpine movements. The retromorphism processes manifest by the deformation and partial recrystallization of the metamorphic rocks; the primary minerals of the amphibolites such as the green amphibole, plagioclase, quartz, biotite and almandine recrystallize, the melanocrate ones being replaced by chlorite, while the plagioclase feldspar was replaced by albite.

Geochemistry of the Drăgșan Series and Origin of Amphibolites

In order to know the chemical composition of the basic and ultramafic rocks of the Drăgșan Series and the distribution of the trace elements in this metamorphic series, 5 serpentinites and 10 orthoamphibolites (Tab.) were chemically analysed and by emission spectrography.

The MgO contents in serpentinites range between 33.60 and 38.69% which indicates their origin in a primary peridotitic magma (Savu et al., 1986). Due to the transformation in talc of some ultramafic rocks, determined by the synorogenic and tardeorogenic granitoid solutions, modifications in the distribution of some "stable" trace elements occur. Thus Cr, which seems to have been the most affected element, shows a content characteristic of the ultramafic rocks (Turekian, Wedepohl, 1961) in a single sample. It is found in much smaller amounts in the other rocks, being not present in the talc network, the latter secondary mineral being formed at the expense of serpentinite.

The SiO_2 and MgO contents in amphibolites indicate a series of basic rocks which are basaltic in composition. Savu and Udrescu (1982) showed that the Assynthic amphibolites of the Carpathiansas wellas thoseofanother age, exhibit features characteristic of the tholeiitic rocks.

TABLE

Chemical composition of the basic and ultramafic rocks from the Drăgşan amphibolite Series

Sample *	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	37.37	37.79	38.02	38.10	40.95	45.52	45.85	45.89	47.35	47.59	47.76	47.84	48.20	50.07	53.80
TiO_2	—	—	—	—	—	0.16	1.80	0.72	1.89	1.48	0.64	0.16	1.20	1.44	0.44
Al_2O_3	1.85	1.30	2.20	2.05	1.00	17.21	10.50	15.01	13.58	13.16	17.99	15.51	15.56	15.35	15.10
Fe_2O_3	9.75	11.99	9.84	11.50	4.44	2.50	2.31	5.04	4.54	3.09	2.93	0.07	3.86	1.73	2.12
FeO	3.62	4.18	5.46	3.23	1.65	5.87	9.71	7.18	7.61	9.15	7.98	5.36	6.12	8.82	6.60
MnO	0.13	0.14	0.02	0.18	0.08	0.16	0.20	0.20	0.22	0.21	0.24	0.13	0.18	0.22	0.16
MgO	35.84	34.84	33.60	35.04	38.69	9.20	12.82	9.90	8.17	6.46	5.98	11.66	7.94	6.54	6.68
CaO	0.14	0.78	0.54	0.17	0.14	1.56	11.01	10.96	9.46	13.36	9.10	13.76	9.98	10.20	7.86
Na_2O	0.09	0.07	0.07	0.06	0.07	1.59	1.51	1.69	2.69	2.11	2.51	0.82	2.69	2.04	2.74
K_2O	0.02	0.02	0.03	—	0.04	1.18	0.38	0.80	0.91	0.74	0.74	0.90	0.52	0.91	1.41
P_2O_5	0.04	0.02	—	—	0.02	0.04	0.11	0.02	0.10	0.16	0.18	0.05	0.18	0.55	0.11
S	0.48	0.24	0.41	0.28	0.28	0.13	0.20	0.24	0.20	0.26	0.21	0.23	0.17	0.20	0.24
H_2O	9.90	8.74	9.61	9.39	11.78	2.32	3.15	10.65	2.84	1.69	3.26	2.94	2.90	1.97	2.67
Total	99.23	00.11	99.80	00.00	99.14	99.44	99.55	10.30	99.55	99.46	99.51	99.43	99.49	99.63	99.93
(ppm)	120	80	80	270	2800	260	1000	400	400	110	95	620	300	115	135
Ni	710	1100	950	950	1000	60	820	115	240	80	15	75	115	60	46
Co	125	170	140	160	95	44	80	62	70	53	21	40	55	44	26
V	22	16	26	24	23	260	300	370	370	430	330	155	330	420	190
Sc	7.5	6	6.5	9	5.5	40	25	33	30	36	34	29	34	40	28
Cu	337	14.5	95	55	6	10	48	6	180	38	60	29	30	40	60
Pb						2	6	5	6	33	6.5	19	3	20	5
Ga						16.5	16	19	18	19	22	10	23	19	18
Sn						2	2.5	2	3.5	2	2	2	2	2	2
Zr						24	220	32	280	130	60	10	120	95	33
Nb						10	18	10	10	10	10	10	10	10	10
Y						8.5	22	12	32	38	21	5	24	36	17
Yb						0.9	1.5	1.6	2.3	3.6	2.2	0.6	3.1	4.5	2.0
Ba						420	60	90	165	90	200	72	140	160	420
Sr						240	190	380	400	140	320	300	340	300	260

* The analysed rocks represent: serpentinites (1–5): 1, Păpuşa Peak; 2, Setea Mare Valley; 3, and 4, Cioara Peak; 5, Tărtărău Peak; amphibolites (6–15): 6, 9 and 13, Tărtărău Peak; 7, 10, 11, 12 and 15, Păpuşa Peak; 8, Cioara Peak; 14, Galbenu Peak.

The trace elements from the amphibolites are also characteristic of basic rocks. The metamorphic differentiation processes determine higher Cr and sometimes Ni values in the rocks formed prevailingly of green amphibole, concomitantly with the increase of the MgO amount (Tab.). In the case of Zr, the values widely range between 10 and 280 ppm. The high Zr values may be due to the arteritic migmatization of the rocks from the amphibolite series, achieved under the influence of the granitoid solutions. Similar aspects are also noticed in the case of Ba (Tab.).

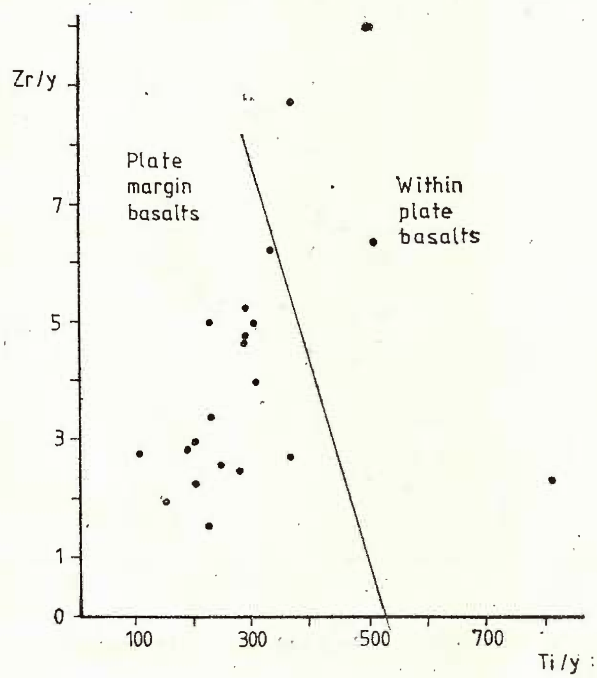

Fig. 1 — Zr/Y—Ti/Y diagram for the amphibolic rocks of the Paring Mts.

As regards the origin of the rocks forming the amphibolites of the Drăgșan Series, as it has already been shown (Savu et al., 1984 a), they might represent ocean floor rocks which underwent several modifications caused by the sedimentary material supply during the deposition of the basic tuffs, the metamorphic differentiation and the arteritic migmatization processes (Savu, Udrescu, 1982). That is why, as shown also by our present investigations, it is very difficult to apply the methods, commonly used for establishing the tectonic setting of the eruptive rocks, unaffected by the above-mentioned phenomena. Nevertheless, on the diagram from Figure 1 (Pearce, Gale, 1977) in which the rocks from Table as well as those previously published for the Parîng Mts. are plotted (Savu et al., 1984 a), most amphibolitic rocks plot in the marginal plate basalt field.

The diagram in Figure 2 (Pearce, Gale, 1977) shows that 50% of these rocks plot in the ocean floor basalt field, which would indicate that the Drăgșan Series amphibolites might have formed in a rift zone, as a result of an intense volcanic activity that manifested simultaneously with intermittent depositions of terrigenous sediments and limestones. The rocks affected more intensely by migmatization, which exhibit high Zr and lower Ti and V contents plot well outside any of the fields on the diagram.

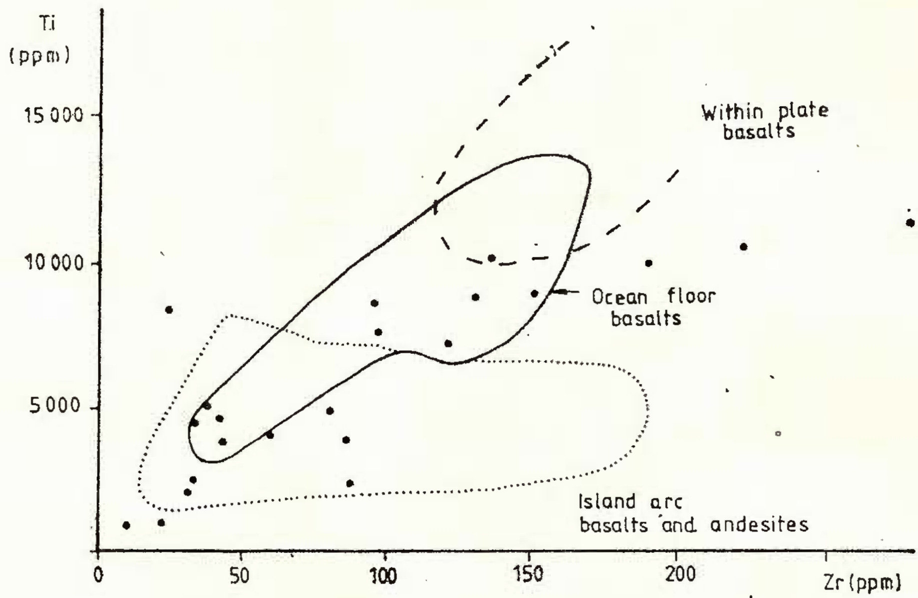

Fig. 2 — Ti—Zr diagram for the amphibolic rocks from the Parîng Mts.

Taking into account the volcano-sedimentary character of the series, we think that the possibility that it represents the product of a volcanism showing a transition tectonic setting of MORB/WPB type (Pearce, 1980) is not excluded.

Conclusions

The conclusions arising from this paper are as follows :

In the Novaci-Parîng Peak region three major tectonic units can be distinguished : Crasna, Sadu and Parîng-Vulcan.

The Parîng-Vulcan Unit thrust over the Sadu Unit probably at the end of the Assynthic movements or in a pre-Hercynian movement and both units thrust over the Crasna Unit during a post-Liassic movement.

During the Hercynian and post-Liassic movements, the Assynthic infrastructure is retromorphosed and forms structures which affect also the pre-Hercynian overthrust plane of the Parîng-Vulcan Unit over the Sadu Unit.

The ultramafic rocks show features characteristic of an oceanic crust.

The chemical composition of the amphibolites from the Drăgşan Series as well as their contents of trace elements are influenced by several factors, so that the tectonic settings of the pre-metamorphic basic volcanic rocks cannot be established for certain, although they might have represented an oceanic crust.

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STRUCTURA UNITĂȚILOR TECTONICE DANUBIENE ÎNTRE NOVACI ȘI VÎRFUL PARÎNG CU PRIVIRE SPECIALĂ ASUPRA SERIEI AMFIBOLITELOR DE DRĂGȘAN

(Rezumat)

În regiunea Novaci-vîrful Parîng din partea de est a Hărții geologice a R.S.R., 1 : 50000, foaia Novaci, se găsesc următoarele unități tectonice, considerate de jos în sus : unitatea de Crasna, unitatea de Sadu și unitatea de Parîng-Vulcan (Savu et al., 1985) (pl. I). Unitatea de Crasna este cea mai sudică din harta Novaci ; ea afloraeză între Crasna și Drăgoești și mai la vest la Schela. În zona Novaci ea este acoperită de depozite sedimentare mai noi. Ea constă într-un soclu cristalin apropiat de seria de Lainici-Păiuș, peste care repauzează discordant formațiuni liasice de tip Schela cu antracit. Unitatea este încălecată de la nord de unitatea de Sadu, pe un plan de șariaj postliasic. Unitatea de Sadu este alcătuită dintr-un soclu assyntic constituit din seria de Lainici-Păiuș de vîrstă precambrian-superioară B, care este migmatizată (pl. II) și străbătută de plutonul granitoid de Sușița — sinorogen, și de plutonul granitoid tardeorogen de Cărpiniș-Novaci cu structură zonară. Acesta din urmă este secționat de planul de șariaj menționat, astfel că din el nu apare la zi decît partea sa nordică. Această unitate a fost încălecată de unitatea de Parîng-Vulcan pe un plan de șariaj assyntic-prehercinic (Savu et al., 1984). Unitatea de Parîng-Vulcan este constituită și ea dintr-un soclu assyntic, format din seria vulcanogen-sedimentară metamorfozată de Drăgșan, care este străbătută de granitul de Parîng. Aceste două unități care au fost sudate în timpul Paleozoicului, au fost acoperite de formațiuni paleozoice și apoi liasice. Acestea fiind în cea mai mare parte erodate, nu se mai păstrează în prezent decît sub doi solzi : unul este situat între valea Romanu și valea Setea Mare și altul în colțul de NE al regiunii, pe vîrful Mușetoaia (pl. I).

În timpul metamorfismului hercinic (290 m.a.) este retromorfozată infrastructura metamorfică assyntică ce formează soclul celor trei unități tectonice descrise, iar granitoidele sînt laminate pînă la ortoșisturi sericito-cloritoase.

Seria de Drăgșan este rubanată (pl. III) și constă în amfibolite, gnaise amfibolice, paragnaise, gnaise cuarțo-feldspatice și calcare cristaline sau cu silicați în care sînt prinse rar mici corpuri lenticulare de serpentinite.

Compoziția chimică a amfibolitelor (tab.) indică roci tholeiitice de margine de placă (fig. 1), dintre care 50% se situează în cîmpul bazaltelor de fund oceanic (fig. 2), indicînd probabil caracterul real al rocilor bazice inițiale ; celelalte roci se disperseză în alte cîmpuri ale diagramei sau în afara lor. Se consideră totuși că, datorită contaminării rocilor sub influența mai multor factori, nu pot fi stabilite cu certitudine condițiile tectonice în care au erupt rocile vulcanice premetamorfice.

EXPLANATION OF PLATES

Plate II

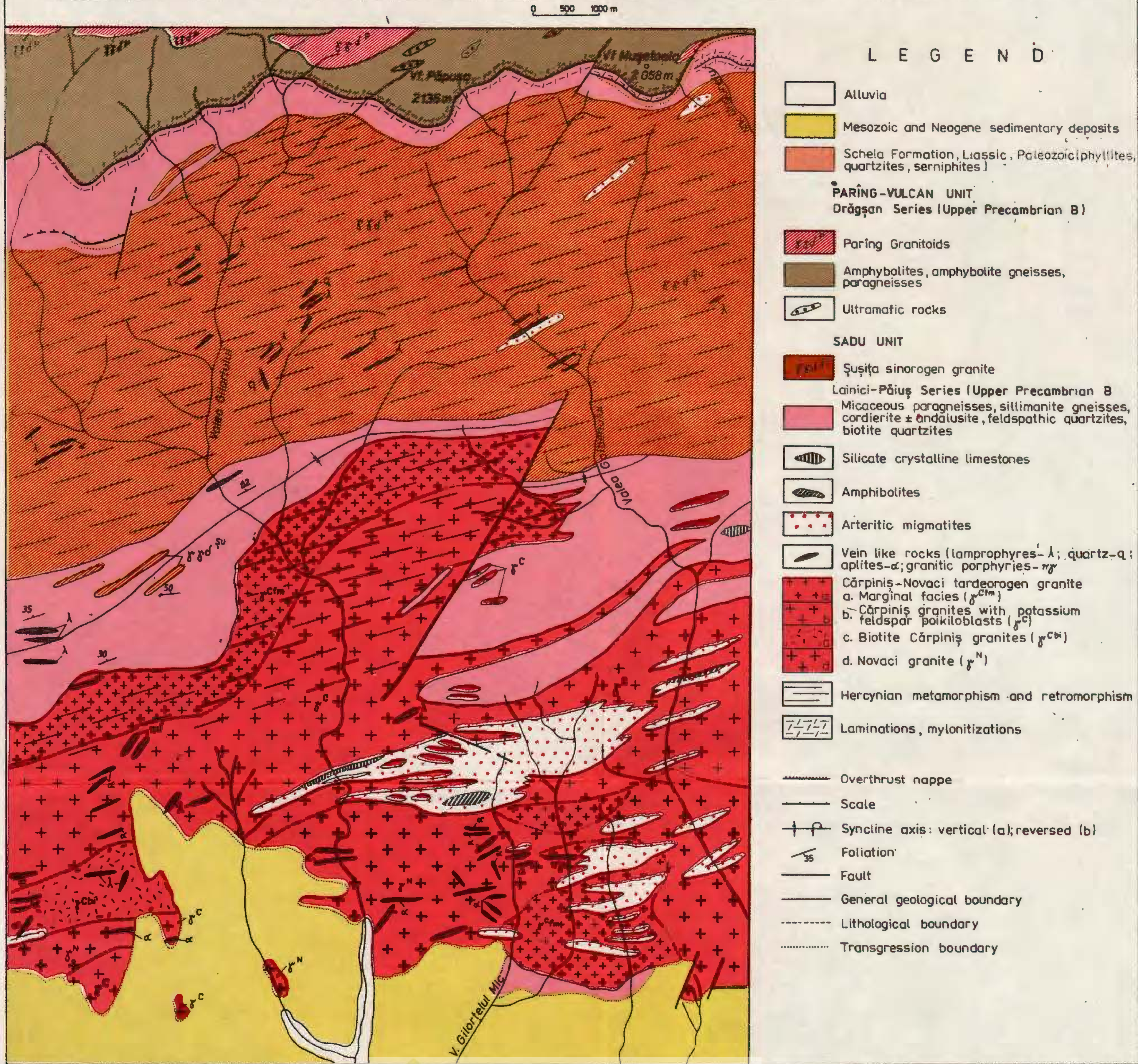
Fig. 1 — Arteritic migmatites lit par lit on the Galbenu Valley.
Fig. 2 — Reticular arteritic migmatites on the Galbenu Valley.

Plate III

Fig. 1 — Banded structure of the Drăgşan Series. Păpuşa Peak.
Fig. 2 — Banded structure of the amphibolites of the Drăgşan Series-Păpuşa Peak.

H. SAVU, H.P. HANN, M. TATU

GEOLOGICAL MAP OF THE REGION BETWEEN THE GILORT VALLEY AND THE GALBEN VALLEY



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5. TECTONICĂ ŞI GEOLOGIE REGIONALĂ

RELATIONSHIPS BETWEEN SEDIMENTARY DEPOSITS AND ERUPTIVE ROCKS IN THE CONSUL UNIT (NORTH DOBROGEA) — IMPLICATIONS ON TECTONIC INTERPRETATIONS [1]

by

IOAN SEGHEDI [2], ALEXANDRU SZAKÁCS [2], ALBERT BALTRES [2]

Limestone. Rhyolites. Basalts. Tuff. Lava flows. Igneous activity. Bimodal magmatism. Structural analysis. Folds. Cleavage. Tectogenesis. Dobrogea — North Dobrogea — Tulcea and Consul-Niculițel zone.

Abstract

In the Consul Unit of North Dobrogea, the Lower Triassic deep water limestones and the eruptive rocks assembled into a single formation (Somova Formation) revealed spatial relations and structural geometry accounting for a two-phase deformational history. The Lower Triassic (Spathian) carbonate rocks are represented by debrites, turbidites and pelagites. The felsic volcanic rocks, except some tuffites and tuffs are effusive in nature (massive, perlitic, fluidal and breccious rhyolites). Their emplacement took place after the deposition of the main mass of calcareous sediments but before their lithification, in subaqueous environment. The mafic eruptive rocks are either interbedded in the sedimentary deposits (as basaltic lava flows, tuffs and tuffites) or as intrusive bodies in the limestones and rhyolites. This magmatism was partly subsequent to the rhyolitic volcanism and they formed together a bimodal system during the Lower Triassic. The sedimentary deposits and the eruptive rocks of the Somova Formation were folded together in two deformational phases. The first event (B_1) (Early Kimmerian) is characterized by a tight folding along the main structural trend of North Dobrogea (NW—SE); the second event (B_2), post-Liassic in age, generated symmetrical E—W trending folds. The combination of these two structural elements resulted in a characteristic structure of dome and basin type.

Résumé

Relations entre les dépots sédimentaires et les roches éruptives de l'unité de Consul (Dobrogea du Nord) — implications sur les interprétations tectoniques. L'ouvrage apporte des précisions portant sur la nature des dépôts sédimentaires et des roches éruptives (Formation de

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Somova), les relations spatiales entre ceux-ci et les déformations tectoniques qui les ont affectées. Les dépôts sédimentaires, principalement carbonatés, d'âge spathien comportent des turbidites, des débrites et des pélagites. Les volcanites acides, excepté certaines intercalations de tuffites de la partie supérieure des calcaires ainsi que des tufs à la limite, sont de nature effusive (rhyolites à textures perlitiques, fluidales et bréchiques en base, rhyolites massives). Leur mise en place s'est produite après le dépôt des sédiments calcaires et avant leur lithification dans des conditions subaquatiques. Les roches éruptives basiques sont soit intercalées dans les dépôts sédimentaires (coulées de laves, tufs), soit traversent les calcaires et les rhyolites (corps éruptifs), en attestant un magmatisme basique qui s'est produit bien avant ou après le volcanisme rhyolitique, dans le cadre d'un système bimodal, pendant le Trias inférieur. Les dépôts sédimentaires et les roches éruptives de la Formation de Somova sont plissés ensemble durant les deux phases déformationnelles. La première (B_I), d'âge kimmérien ancien, se caractérise par un plissement serré (plis déversés et plis-écaille) en concordance à la direction principale des structures nord-dobrogéennes (NV—SE) ; la deuxième (B_2) probablement postliasique, a engendré des plis symétriques orientés EV. De leur jonction a résulté une structure caractéristique de type dômes et bassins.

1. Introduction

The time-space relationships between felsic and mafic igneous rocks and associated sedimentary deposits is one of the key problems for understanding the geological structure of the Consul Unit. This problem was approached by most of the previous researchers of the Consul Unit (Murgoci, 1914 ; Savul, 1935 ; Mirăuţă, Mirăuţă, 1962 ; Mirăuţă, 1966 ; Stiopol et al., 1975 ; Constantinescu et al., 1978, 1981, 1983 ; Vîlceanu et al., 1980 ; Manea et al., 1983 ; Baltres et al., 1984 ; Berbeleac et al., 1985 ; Nedelcu et al., 1986). In spite of the intense investigation and progress in geological knowledge of this area, various problems, mainly connected to the emplacement of rhyolites and basalts, volcanological and genetic aspects and their relations with sedimentary deposits remained partly unsolved. This is fairly due to the complex structure, which is difficult to decipher, as is shown on the previous geological maps.

The relationships between the rhyolites and the Triassic limestones of the Consul Unit are dependent on the interpretation of emplacement mode of rhyolites. Accepting the intrusive nature of rhyolites, they are implicitly subsequent to limestones (Murgoci, 1914 ; Ştefan, in Berbeleac et al., 1985, and in Nedelcu et al., 1986). The hypotheses supporting the effusive (and, possibly, partly intrusive) and/or ignimbritic nature of rhyolites regard them as interlayered in the pile of the Triassic deposits (Savul, 1935 ; Stiopol et al.. 1975 ; Constantinescu et al., 1978, 1981, 1982 ; Caraveţeanu, in Vîlceanu et al., 1980, and in Manea et al., 1983 ; Baltres et al., 1984).

The effusive nature of rhyolites implies their emplacement prior to the tectonic deformations. Their pre- or postdeformational position is not clear in the hypotheses supporting the subvolcanic nature of these rocks, as results from the interpretations of the geological profiles in the Consul Unit belonging to the mentioned authors.

This paper intends to provide additional information concerning these problems, based on the synthesis of all field data and mainly on the detailed geological mapping of the area between Valea Teilor and Iulia, carried out for the elaboration of the geological maps, scale 1 : 50,000, sheets Priopcea and Cataloi (Seghedi et al., 1985, 1986 ; Mirăuță et al., 1986).

2. Geological Setting

The Consul Unit is an Alpine tectonic unit, bordered to the west by the Luncavița-Consul line and to the east by the Valea Teilor-Iulia line ; these lines separate the Consul Unit from the Măcin Unit to the west and from the Niculițel Unit to the east, respectively (Mirăuță, in Patrulius et al., 1973). According to the structural interpretations of North Dobrogea, the Consul and Niculițel units are considered by some authors as digitations of a unique tectonic unit, regarded as a nappe (Patrulius et al., 1973 ; Săndulescu, 1984).

The Consul Unit consists of carbonate rocks (debrites, turbidites, pelagites) and subordinately sandstones, rhyolitic and basaltic volcanics, all constituting the Somova Formation (Baltres, 1982 ; Baltres, in Baltres, Mirăuță, 1987, earlier designed as the Consul Formation by Baltres, in Baltres et al., 1984). The Somova Formation has a large areal development in the Tulcea Unit, where the type locality was established.

The Somova Formation in the Consul Unit commonly overlies low grade metamorphic rocks of the Boclugea Group. At Iulia terrigenous deposits (sandstones and polymictic conglomerates), known from outcrops and subsurface, underlie the Somova Formation. These detrital deposits were assigned to the Bogza Formation (Griesbachian i.e. early Lower Triassic) developed in the Tulcea Unit (Baltres, in Baltres, Mirăuță, 1987).

The Somova Formation was deposited during the Spathian (late Lower Triassic). This age is proved by a foraminifera assemblage containing *Meandrospira iulia* (Premoli Silva), *M. dieneri* Kristan-Tollmann and *Glomospira silensis* Dager (Baltres, in Baltres, Mirăuță, 1987). Conodonts point to the same stratigraphic interval, although several Lower Anisian species have been also found (Mirăuță, in Seghedi et al., 1986).

2.1. *Carbonate Rocks*

The limestones of the Somova Formation are usually banded due to the interlayering of black limestone beds and subordinate black shales. Occurrences of thick argillaceous limestones or marls are uncommon. Typical of these deposits are thick calcirudite beds, more frequent in the upper part of the sedimentary pile.

The thickness of the deposits is difficult to appreciate due to tectonic complications and to deformation of unconsolidated sediments by down-slope sliding. These structures are often responsible for the abrupt changes in dip of the beds in drill cores. In the Consul Hill, the limestone thickness attains 500 m, while in the borehole 5-Iulia the formation exceeds 1,600 metres. The dip of the beds ranges between 20 and 80°.

The limestones of the Somova Formation are deep water debrites, turbidites and pelagites, each of these implying specific transport and sedimentation mechanisms.

The debrites represent coarse grained, resedimented accumulations. They constitute the lower parts of the graded sequences, showing commonly slow or abrupt transition to calcarenites.

The calcarenitic turbidites occur as beds up to ten centimetres thick, consisting of complete or incomplete Bouma sequences. The calcisiltite and calcilutite turbidites are more frequent, consisting of graded and laminated sequences (E_1 and E_2 Piper sequences).

The pelagites are slowly accumulated interturbiditic aphanitic limestones, often intensely reworked by bioturbation. Some pelagites, accumulated under reducing conditions, contain framboidal sedimentary pyrite.

The deep water limestones of the Somova Formation are clinothems, emplaced by sedimentary mechanisms implying essentially gravitational transport pocesses. They built up the continental rise or slope in the Lower Triassic sea.

2.2. *Igneous Rocks*

Rhyolites. The largest and most varied rhyolitic occurrences associated with carbonatic deposits of North Dobrogea are known in the Consul Unit. They occur as tuffites, tuffs and lava flows.

1) The beginning of the Triassic volcanic activity (Consul Hill, Coasta Pășunii) is indicated by the presence of several, centimetric or decimetric, rhyolitic tuffite interbeds at the top of the limestone pile. These consist of a mixture of igneous clasts (weakly devitrified and slightly sericitised volcanic glass, alkali feldspar, plagioclase and quartz crystalloclasts) and carbonatic material (recrystallised lithoclasts). In some coarse tuffites small metamorphic quartz and granite clasts were found. The grain size is variable, from fine to coarse psammite; the amount of volcanic material is variable.

2) In the vicinity of the limestone-rhyolite boundary (Consul Hill, Coasta Pășunii Hill, south of Lozova Hill), the main volcanic body contains metric rhyolitic tuff beds. The tuffs consist of clasts of glass, crystals, and crystalloclasts and pumice within a fine grained, devitrified rhyolitic groundmass. Sometimes they show a slight grading (Consul Hill, Coasta Pășunii Hill). The grain-size range of the tuffs is 2—3 mm. The features described are typical of ash fall tuffs.

The tuffs occurring on the southern slope of the Eschibalîk Hill are apparently massive, showing inhomogeneous groundmass and oriented structure, with features suggesting advanced welding; they can be interpreted as ash flow tuffs. Their genesis may be assimilated with ignimbrite-forming processes, but lack of typical vesiculation and of glass shards suggest a subaqueous depositional environment (Fisher, Schminke, 1984).

3) The main rhyolitic body presents various textural-structural features, including massive, perlitic, brecciated and fluidal aspects. The perlitic, brecciated and fluidal structures are typical only of the basal parts of the rhyolites overlying the limestones, while most of the eruptions show a massive structure. The massive rhyolites show a more pronounced porphyritic texture. The phenocrysts are of alkali feldspar (partly albitized), quartz and plagioclase. The presence of chlorite and opaque minerals with

hexagonal outlines suggests the initial presence of a femic mineral, probably biotite. Accessory minerals are represented by zircon, sphene and apatite. The former glassy or microgranular groundmass is always devitrified. The perlitic and brecciated structural varieties show a strong secondary chloritization. All these features strongly suggest an effusive facies for most of the rhyolites, in agreement with the observations of Savul (1935). Association with tuffs , the large areal extent, the lack of any thermal and/or metasomatic contact phenomena in limestones (perhaps the most intriguing features for the previous researchers) are additional arguments for the effusive origin of most rhyolites of the Consul Unit.

Ignimbrites showing typical features of pyroclastic flow deposits occur south-east of the Consul Hill, in the vicinity of Mihai Bravu and Nicolae Bălcescu. As opposed to the rhyolites in the Lozova and Consul Hills, they are associated with Lower Triassic (Griesbachian) and Permian (?) terrigenous deposits (Mirăuță, in Szász et al., 1981).

Mafic rocks Starting northward from the Malciu Hill, the calcareous deposits of the Consul Unit are associated with basaltic rocks, which occur either as lava flows and tuffs, or as dykes and irregular bodies.

1) Lava flows are interbedded with limestones. The rocks have a greenish colour and show aphanitic, seldom slightly porphyritic texture. The intersertal or subophitic texture are typical. In thin sections these rocks consist of plagioclase, clinopyroxene (altered to calcite, serpentinic minerals and chlorite), opaque minerals and chloritized glass. In the Malciu Hill, the rocks contain calcite or chalcedony filled vacuoles.

The thin beds (centimetric and decimetric) of basic tuffs and tuffites interbedded with limestones consist of albitized plagioclase fragments, pyroxenes (altered to epidote, chlorite, clay minerals) and a variable amount of terrigenous clasts (quartz, quartzites, carbonates).

2) Mafic rocks within dyke (on the left bank of the Pîrlita Creek) generally strike N 35—70° . Within sills (Sultana Creek) and dykes textural variations occur. Usually the rocks have ophitic or intergranular texture and massive structure. Their crystallinity is more pronounced compared to the effusive basalts ; rocks consist mainly of plagioclase, clinopyroxenes and opaque minerals and show various degrees of secondary transformations. Sometimes small amounts of chloritized glass are present.

An intrusive body on the northern slope of the Malciu Hill pierces both limestones and rhyolites and slightly alters them. The rock is holocrystalline, showing intergranular texture and massive structure. It consists of plagioclase feldspars and pyroxenes, transformed into serpentinic minerals and chlorite. The rock may be considered dolerite or microgabbro.

3. Relationships between Eruptive Rocks and Sedimentary Deposits

Excepting tuffites, interbedded at the top of the limestone sequence, most rhyolites overlie the Lower Triassic carbonatic deposits. This relationship is constantly seen along the boundary between limestones and rhyolites in the Lozova, Coasta Pășunii and Consul Hills. In outcrop, the rhyolite/limestone boundary is a sharp, slightly irregular contact surface (Pl. III, Fig. 1). Sometimes it is marked by dissolution voids (Pl. III, Fig. 1).

The aphanitic or porphyritic rhyolites with flow structure (western slope of the Lozova and Consul Hills) show significant aspects regarding their relationships with the limestones. Usually there is a concordance between the flow structure of basal rhyolites and limestone bedding (Pl. II, Figs. 1, 2 ; Pl. III, Fig. 2). The rock structure, its low crystallinity and the presence of perlites (Pl. II, Fig. 3), indicating rapid cooling, suggest subaqueous conditions for the emplacement of the rhyolitic flow. The perlitic parts represent the outer crust of the flow, quenched in contact with water, subsequently disrupted and included in the main flow mass. This explains also the lack of thermal and metasomatic phenomena in limestones next to the rhyolites.

In Coasta Pășunii Hill, detailed observations revealed an inhomogeneous mixture between the rhyolitic and carbonatic material, within a thin band of about 15—20 cm in thickness at the contact. At the boundary with the rhyolitic rock the limestone shows a slight recrystallisation band a few millimeters thick. This peculiar feature indicates that rhyolite emplacement took place soon after the deposition of the carbonatic material and before sediment lithification. The rhyolites next to the limestones may sometimes contain tiny limestone fragments.

Piercing relations of limestones by rhyolites have not been noticed. The limestones have never been found to overlie the main rhyolitic body. Apparent overlying of rhyolites by limestones or repeated interlayering of rhyolites with limestones are exclusively the result of folding.

The above observations suggest that most rhyolites constitute a single important effusive episode, with subaqueous consolidation of the acid volcanic material.

Field relations between limestones and mafic rocks indicate at least two moments for the emplacement of the latter. The basaltic volcanics (lava flows), basic tuffs and tuffites interbedded with limestones have been emplaced prior to or possibly synchronous with the beginning of the acid volcanic activity. The presence of centimetric clasts of basalts, reworked in the limestones from the western slope of the Lozova Creek (in the vicinity of the Malciu Hill, Mirăuță, in Patrulius et al., 1974 ; Mirăuță, in Seghedi et al., 1986) may be correlated with this moment or possibly with a previous effusive moment. The microscopic basalt xenoliths within the rhyolites from the Malciu Hill lead to the same conclusion.

The dolerite dykes and bodies show evidence for emplacement after the deposition of the main limestone sequence and even after the emplacement of the rhyolite flows (Malciu Hill) (Savul, 1935).

The synthesis of the petrographic features of the acid and mafic volcanics and their relationships with the sedimentary deposits suggest several volcanological considerations regarding the eruptive rocks of the Consul Unit.

The beginning of the volcanic activity is related to the deposition of the Griesbachian detrital deposits (at Mihai Bravu, Nicolae Bălcescu). This activity produced typical ignimbrites, attesting a subaerial explosive volcanism.

The presence of the Spathian basaltic lava flows, tuffs and tuffites, as well as of rhyolitic tuffites, interbedded in the deep water calcareous sequence near the limestone-rhyolite boundary, suggests an initially quasi-synchronous manifestation of a mafic, effusive-explosive and of a felsic, explosive volcanism. The local occurrence of air fall or pyroclastic flow tuffs at the same level and the rhyolite-limestone field relations strongly point to the manifestation of an initially explosive, subsequently effusive subaqueous felsic volcanism, simultaneous with and subsequent to the deposition of the last carbonatic sequences of the Somova Formation. The quasi-simultaneity and spatial association of the mafic and felsic eruptive activity suggest a bimodal magmatism in the Consul Unit, previously supposed by Savu (1986).

4. Structural Elements

The deciphering of the geological structure of the Consul Unit is based on observations regarding the nature and cartographic pattern of the rhyolite-limestone boundary, as well as on the observation of the structural elements, especially the axial plane cleavage of the folds, affecting limestones, rhyolites and basalts. Field evidence points to the presence of two generations of structural elements, suggesting two folding events successive in time and different in style, which affected the Somova Formation.

1. The first folding phase (B_1) produced overturned folds and small scale thrusts with NW—SE trends. This folding is accompanied by the development of axial plane cleavages in the hinge zones of the main folds, affecting especially the limestones and, in a lower degree, the rhyolites (Mirăuță, 1960) (Pl. VI, Fig. 2) due to their different response to stress. A more pronounced plasticity and mobility of the limestones favoured their piercing in anticlinal fold cores (see geological sections, Pl. I) while rhyolites tend to occur in syncline cores, due to their lower plasticity and higher stratigraphic level: in a few cases, the rhyolites cover the limestones in the anticline hinges (Consulul Mare Peak, Lozova Hill).

In the hinge zones, the angle between bedding and cleavage (S_0/S_1) ranges between 55—80° (Pl. V, Fig. 1). It is rarely below 30°, due to the folding of the initial bedding of limestones by gravitational folds. The frequency and distribution of the axial plane cleavages in limestones is highly variable, according to the competence of rocks. Thick sequences of calcirudites show penetrative cleavages only in fold axes (Pl. V, Fig. 2). The cleavage planes are curved and sometimes anastomozing, spaced at 5—15 cm (Pl. IV, Fig. 1). Argillaceous limestones show a more pronounced planar, parallel cleavage, seldom anastomozing (Pl. VII, Fig. 1), developed evenly throughout the folds. In this case, the spacing of the cleavage planes ranges between a few millimeters and a few centimeters. The thin calcarenitic beds interbedded in calcirudites or in argillaceous limestones form asymmetrical microfolds, associated with the main folds, with wavelengths ranging between 0,5—3 m (Pl. VII, Figs. 1, 2; Pl. VIII, Fig. 1). Rhyolites rarely show penetrative cleavages, which are usually spaced, rough, accompanied by chlorite and sericite recrystallisation. Cleavage refraction is common at the boundary between tuffites and limestones in the northern and western parts of the Consul Hill.

At least three axes of major folds of this folding system, with a wavelength of about 500 m, were identified. The direction of the axial planes ranges between 145—170°, dipping 75—90° to SW. Reverse dips occur on the eastern slope of the Consul Hill. The main folds are accompanied by overturned folds and minor thrusts, reaching meters or tens of meters in size (Pl. VIII, Fig. 2).

All these features reveal a tight folding of the Somova Formation, as a consequence of strong compressions. The maximum compression zone was attained in the Consul Hill, where the tightest folding occurs. Drilling operations in the iron ore accumulation area at Iulia revealed the presence of rhyolites at depths up to 300 m. In good agreement with field observations, this situation suggests a complex structure of the Consul Unit, including overturned folds and high-angle reverse faults (Băltres, in Baltres et al., 1984).

The main folds (B_1) resulted during the first folding event have the same structural trend (NW—SE) as the entire fold belt of North Dobrogea. The consequence of this folding was the reduction in initial width of the Consul Unit by at least three times. The present maximum width of this area does not exceed 2 km.

2. The existence of a second folding event is suggested by the highly irregular pattern of the limestone-rhyolite boundary, showing pinching and swelling along an E—W trend (Pl. I). Detailed outcrop observations revealed structural elements of a later folding phase, subsequent to that responsible for the development of the NW—SE trending axial plane cleavage. The second folding (B_2) refolds the axial plane cleavages of the main folds (Pl. IX, Figs. 1, 2) ; in some outcrops (Consul Hill, Lozova Hill), a second, crenulation cleavage (S_2) develops, associated to this folding of kink type. The (S_2) cleavages have E—W trends and steep dips to the N or S.

These observations suggest that the second generation folds are open, symmetrical folds, with wavelengths of about 1 km. It is supposed that the intensity of this folding event was lower, having, however, important implications within the Alpine fold belt of North Dobrogea. The angle between the axial planes of the two generations of folds ranges between 45—60° (Pl. IX, Figs. 1, 2, 3). The interference of these two generations of folds resulted in a dome and basin structural pattern of the Triassic deposits in the Consul Unit.

5. Discussion

The presence of two generations of folds affecting the Triassic deposits in the Consul Unit raises the problem of timing of these deformational events.

The first deformation, connected to the development of the main thrusts in North Dobrogea, is considered Early Kimmerian (Mirăuță, 1966 ; Săndulescu, 1984). In the Tulcea Unit, the Mesozoic formations are involved also in two tectogeneses, the second folding being assigned to the Late Kimmerian (intra-Neocomian) or Mesocretaceous events (Săndulescu, 1984). The preservation of the Early Kimmerian structural trends (NW—SE) during the second folding is implicitly accepted. E—W structural trends are known in the Măcin Unit of North Dobrogea, belonging to a main Variscan event (Seghedi, 1985, 1986). E—W trending folds of the Mesozoic deposits occur on some maps of the Tulcea Unit (Mirăuță et al., 1985). The cartographic pattern of the various formations and tectonic units in North Dobrogea suggest that Mesozoic formations of the Tulcea Unit, including Lower-Middle Jurassic deposits, are also involved into this second folding event, subsequent to the Early Kimmerian phase. These considerations indicate a certain post-Triassic, highly probably post-Liassic age of the second folding which affects the Mesozoic rocks of the Consul Unit. Further investigations on the tectonic style of the Mesozoic deposits in different units of North Dobrogea will possibly provide additional information concerning this problem.

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RELAȚIILE DINTRE DEPOZITELE SEDIMENTARE ȘI ROCILE ERUPTIVE DIN UNITATEA DE CONSUL (DOBROGEA DE NORD) — IMPLICAȚII ASUPRA INTERPRETĂRILOR TECTONICE

(Rezumat)

Lucrarea aduce precizări privind natura depozitelor sedimentare și rocilor eruptive din alcătuirea formațiunii de Somova, asupra relațiilor spațiale dintre ele și asupra deformărilor tectonice, care le-au afectat.

Unitatea de Consul este o entitate tectonică alpină, separată la vest de unitatea de Măcin prin linia tectonică Luncavița-Consul, iar la est, de unitatea de Niculițel prin linia Valea Teilor-Iulia. Este alcătuită esențial din depozite carbonatice, roci eruptive acide și bazice, constituind împreună formațiunea de Somova (Baltres, 1982; Baltres et al., 1984; Baltres, Mirăuță, 1987).

Depozitele sedimentare, în esență carbonatice, sînt reprezentate, din punct de vedere sedimentologic, prin turbidite, debrite și pelagite. Ele au construit taluzul de la baza pantei continentale a mării triasice inferioare.

Vulcanitele acide, cu excepția unor intercalații de tufite de la partea superioară a calcarelor și a unor tufuri la limită, sînt de natură efuzivă (riolite cu texturi perlitice, fluidale și brecioase în bază, riolite masive în rest). Riolitele, alcătuind în esență o singură venire, stau în relații de superpoziție față de calcare. Din relația cu depozitele carbonatice reiese punerea în loc a marii mase de riolite după sedimentarea calcarelor și înaintea litificării lor, în condiții subacvatice. Rocile eruptive bazice sînt fie intercalate în depozitele sedimentare (curgeri de lavă, tufuri), fie străbat atît calcarele, cît și riolitele (corpuri, dyke-uri, silluri), atestînd un magmatism bazic desfășurat în parte anterior, în parte ulterior vulcanismului riolitic, în cadrul unui sistem bimodal, în cursul Triasicului inferior.

Depozitele sedimentare și rocile eruptive din formațiunea de Somova sînt cutate împreună în două faze deformaționale. Prima (B_1), de vîrstă chimerică veche, se caracterizează prin generarea de cute deversate și

cute-solzi, cu planele axiale orientate NV—SE. Cutele se evidenţiază prin dezvoltarea accentuată a clivajului plan-axial în zonele de şarnieră ale cutelor majore, afectînd în special calcarele şi mai puţin riolitele, datorită comportării diferite la efortul tectonic. Au fost identificate trei axe de cută majore ale acestui sistem plicativ, cu o lungime de undă de cca 500 m. Ele sînt însoţite pe flancuri de cute deversate şi cute-solzi minore, avînd dimensiuni de ordinul metrilor şi zecilor de metri (pl. VIII, fig. 2).

Cea de a doua fază de cutare (B_2) se manifestă prin cutarea clivajului plan-axial anterior (cute kink) sau prin generarea unui clivaj plan-axial propriu, avînd orientarea aproximativ E—V. Ea a generat cute largi, simetrice, cu lungimea de undă de cca 1 km. Din îmbinarea celor două elemente structurale (B_1 şi B_2) rezultă o structură caracteristică de tip domuri şi bazine. Cea de a doua cutare (B_2) este probabil postliasică; cu toate că a avut o amploare mai redusă, implicaţiile ei se resimt la scara orogenului nord-dobrogean.

EXPLANATION OF PLATES

Plate II

Fig. 1 — Porphyritic rhyolite showing fluidal structure; Lozova Hill.

Fig. 2. — Glassy rhyolite showing fluidal structure and flow microfolds; sample from the Lozova Hill.

Fig. 3 — Perlitic rhyolite; Lozova Hill; N//, 40 ×.

Fig. 4. — Rhyolite showing a breccious texture; Lozova Hill; N//, 40 ×.

Plate III

Fig. 1 — Contact between rhyolites (1) and limestones (2) marked by dissolution voids; Lozova Hill.

Fig. 2 — Contact between rhyolites and limestones; (1) massive rhyolites, (2) fluidal rhyolites, (3) limestones; Lozova Hill.

Plate IV

Fig. 1 — Axial plane cleavage (S_1) in calcirudites; Lozova Hill.

Fig. 2 — Anastomosed cleavage (S_1) in calcirudites; Consul Hill.

Plate V

Fig. 1 — Relationship between bedding (S_0) and axial plane cleavage (S_1) in an anticline hinge of folded limestones; Lozova Hill.

Fig. 2 — Relationship between bedding (S_0) and axial plane cleavage (S_1) in the hinge of a minor overturned anticlinal fold. Note the penetrative cleavage in calcirudites (at the top) only in the fold axis; Coasta Păşunii Hill.

Plate VI

Fig. 1 — Relationship between bedding (S_0) and axial plane cleavage (S_1) in an alternation of calcirudites and limestones. Overturned limb of an anticlinal fold. Note the different character of the cleavage in calcirudites and argillaceous limestones; south-eastern slope of the Consul Hill.

Fig. 2 — Hinge of a major anticline with limestones (1) below, and rhyolites (2) above, in the Consul Peak. Note the relationship between bedding and axial plane cleavage within limestones.

Plate VII

Fig. 1 — General aspect of the B_1 folding in an overturned aticline limb with largely developed axial plane cleavage in argillaceous limestones (1) and asymmetrical overturned microfolds in calcirudites (2); northern slope of the Consul Hill.

Fig. 2. — Minor overturned B_1 folds in the base of the northern slope of the Consul Hill.

Plate VIII

Fig. 1 — B_1 microfolds in calcarenites; northern slope of the Consul Hill.

Fig. 2 — B_1 minor thrust at the foot of the northern slope of the Consul Hill.

Plate IX

Fig. 1 — Folded (B_2) axial plane cleavage (S_1) in argillaceous limestones. The hammer indicates the direction of the S_2 microfold axes; Consul Hill.

Fig. 2 — Kink microfolds (B_2) of the S_1 cleavages in limestones; Lozova Hill.

Fig. 3 — Axial plane cleavage (S_1) in limestones, refolded by B_2 folds; western side of the Lozova Creek, in front of the Malciu Hill.

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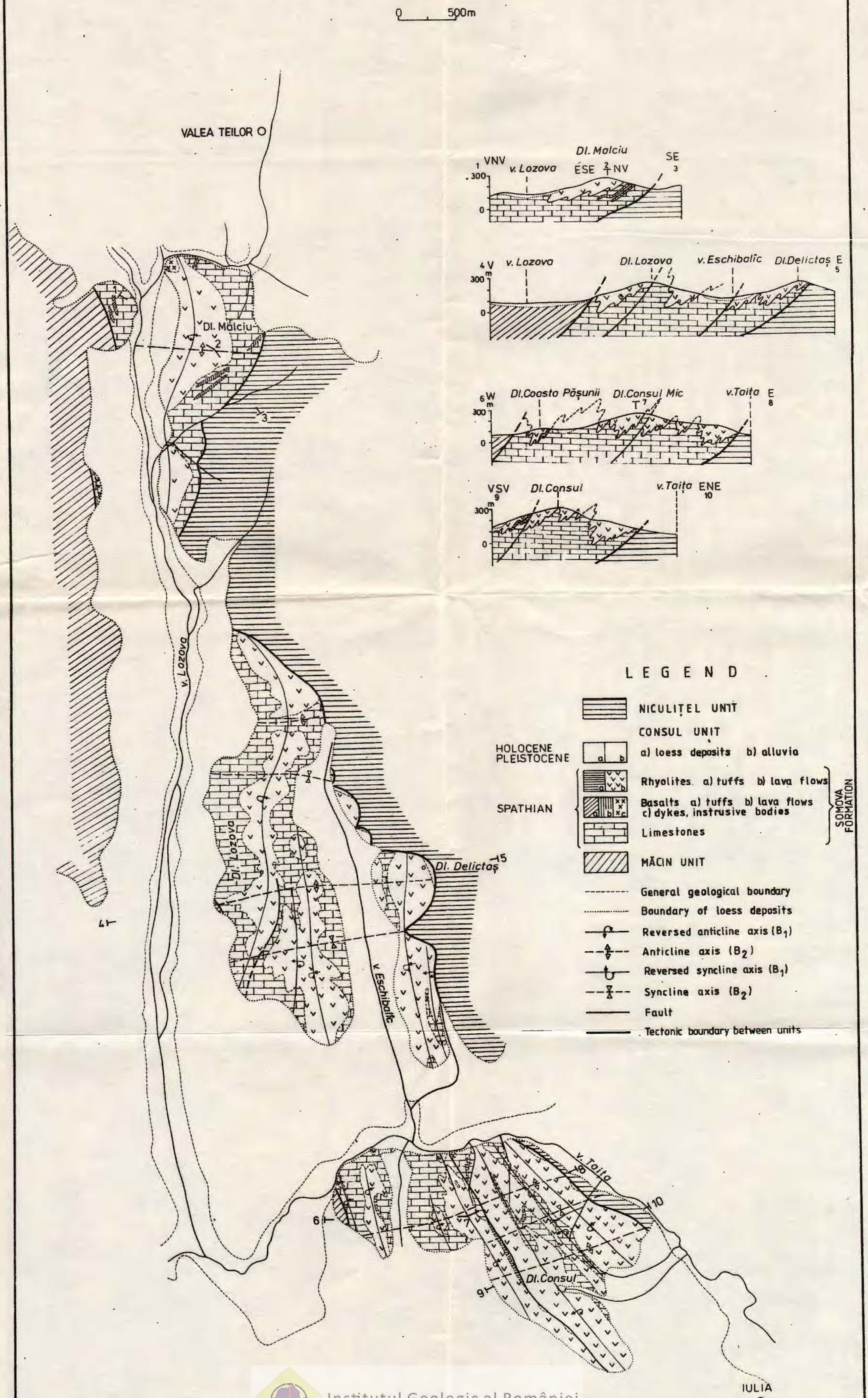
GEOLOGICAL MAP OF THE CONSUL UNIT BETWEEN VALEA TEILOR AND IULIA



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TECTONICĂ ȘI GEOLOGIE REGIONALĂ

STRUCTURE OF THE INNER FLYSCH ZONE IN THE PIETROȘIȚA-RÎUL ALB-BĂRBULEȚU AREA[1]

by

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Flysch. Cretaceous. Nappes. Tectonic wedges. Faults. Right-lateral faults. Thrust fault. Tectogenesis. Correlation. East Carpathians-Internal flysch zone — Comarnic — Pietroșița area.

Abstract

This paper deals with the detailed structure of the internal flysch zone in its westernmost outcropping zone, between Pietroșița and Bărbulețu. Among the structural elements of the study area, the frontal line of the Bratocea digitation and the Rîul Alb fault, with the northern sunk compartment, are worth mentioning.

Résumé

Structure de la zone interne du flysch dans le secteur de Pietroșița—Rîul Alb—Bărbulețu. La note présente la structure de détail de la zone interne du flysch dans son extrémité ouest d'affleurement, entre Pietroșița et Bărbulețu. Parmi les éléments structuraux de ce secteur se remarquent surtout la ligne frontale de la digitation de Bratocea et la faille de Rîul Alb à compartiment nord abaissé.

Introduction and Evolution of Ideas

The well-known structural zonality of the East Carpathians gradually disappears westwards, in their bend zone. The main tectonic lines are difficult to be traced both because of the development of the post-nappe covers and of the secondary faults concealling the major ones. For this reason, several geologists (Mrazec, Popescu-Voitești, 1914; Băncilă, 1958; Dumitrescu et al., 1962; Hristescu et al., 1967; Ștefănescu,

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1969), who investigated the area west of the Prahova Valley, described the major tectonic accidents rather as a necessity of their prolongation from the East Carpathians segment than as a result of the detailed study of the area.

This paper refers only to a small part of this area of the East Carpathians Bend Zone, namely "Pietroşiţa anticline" between Ialomiţa Valley and Bărbuleţu Valley. This structure was formely described by Murgeanu (1930) as the "marginal anticline" of the flysch inner zone. The mentioned author considered that the anticline west of the Ialomiţa Valley is bordered on both flanks by faults. The external one was regarded by some workers as the frontal line of the west-internal unit (= Ceahlău Nappe) (Băncilă, 1958; Hristescu et al., 1967). Ştefănescu (1971) demonstrated that the frontal plane of the Ceahlău Nappe is situated more southwards being concealled and unconformably overlain by Upper Senonian-Lower Miocene posttectonic deposits constituting the Buciumeni syncline.

The available field data made possible several specifications on the structure of the pre-Senonian Cretaceous uplift between Pietroşiţa and Bărbuleţu Valley.

Stratigraphy and Structure

At the Ialomiţa Valley meridian the oldest deposits cropping out are the Sinaia Beds, represented only by their upper level with *Lamellaptychus angulocostatus*. They die out from the Tîţa Valley basin westwards, their exposure place being taken over by the Comarnic Beds (Barremian). The Comarnic Beds are overlain by two different sequences, as follows:

— In the inner part (northwards): a Lower Bedoulian shaly flysch (Vîrfu Rădăcinii Beds), overlain by Aptian deposits developed in a sandy-shaly flysch facies with sandy flysch episodes including Upper Jurassic or Urgonian limestone klippen;

— In the outer part (southwards): the Comarnic Beds continue with a shaly flysch (Podu Vîrtos Beds) covering the whole Aptian.

Both types of Aptian deposits are unconformably overlain by a hemipelagic or pelagic pile of Upper Vraconian-Turonian age (Dumbrăvioara Series) which displays certain particularities in spite of its unitary lithological base. Thus, Dumbrăvioara Series overlying Podu Vîrtos Beds consists of levels of sedimentary breccias as well as of a packet of platy green sandstones at its upper part. These lithological elements do not occur in Dumbrăvioara Series that covers the Aptian sandy flysch of the internal sequences.

Dumbrăvioara Series, or directly the Aptian sandy flysch, is overlain by a common cover that starts with Upper Senonian pelagic deposits (Gura Beliei marls) which pass gradually to a shaly sandy flysch (Şotrile facies), Paleocene-Eocene in age, overlain in its turn by mostly pelitic deposits (Valea Caselor facies), of Oligocene-Lower Miocene age.

Finally, after a discontinuity, Lower-Medium Miocene deposits displaying a typical molasse facies (Doftana molasse) accumulated.

From the Ialomiţa Valley basin eastwards the two sequences differentiated by facies of the Aptian and Upper Vraconian-Turonian deposits are separated by a significant tectonic accident considered (Ştefănescu, 1971) to represent the plane which separates the two digitations of the Ceahlău Nappe, namely Măgura Digitation and Comarnic Digitation. The Măgura Digitation sequence is similar (in places even Albian conglomerates being found) to that of the Bratocea Digitation (= Nappe, Popescu, 1958). Moreover, even the age of the thrusting is the same. For these reasons the first denomination (Măgura) became a synonym (Ştefănescu, 1976).

In the present paper the denomination Bratocea Digitation is used for the internal succession characterized by the more sandy facies of the Aptian. It is of note that north of the Teleajen Valley basin the frontal line of the Bratocea Digitation corresponds to the front of the Ceahlău Nappe.

The thurst plane which separates the two digitations of the Ceahlău Nappe can be traced west of the Ialomița Valley meridian, too. Its course (Pl.) is situated in the very outcropping zone of the Lower Cretaceous deposits, on the segment between Ialomița Valley and Rîul Alb Valley. From the left side of the Rîul Alb Valley to the west the deposits of the Comarnic Digitation die out under the thrust plane, the frontal line of the Bratocea Digitation intersecting the Gura Beliei marls from the internal limb of the Buciumeni syncline, a position maintained up to Bărbulețu Valley basin. In spite of the difficulties of checking lithological packets because of frequent slidings occurring in the area of study one can observe that, approximately west of Rîul Alb, the deposits of the Bratocea Digitation occur as overliers in a higher morphologic position on interfluves. This particular situation is due to the fact that the deposits pertaining to the Bratocea Digitation, those of its cover inclusively, are affected by a vertical fault with its northern block sunk, so that in the higher zones of the relief the frontal part of the digitation is preserved while in valleys the Upper Senonian-Lower Miocene cover from the two digitations comes into direct contact without the interference of Lower Cretaceous, because these deposits have been removed by erosion having an uplifted position. Therefore the main consequence of the fault from the northern flank of the Pietroșita "anticline" is the marked sinking of its northern compartment. However, if the Gura Beliei marls are considered as a guide-mark one can estimate that a shifting along the plane of the fault under discussion took place and consequently the fault is regarded as a right-strike slip fault.

The mentioned vertical fault — called by us Rîul Alb fault — can be traced from the Ialomița Valley up to the summit west of Bărbulețu Valley. The strike of the fault varies from NE to SW in the area between Ialomicioara Valley and the left side of Bărbulețu Valley, and from ENE to WSW in the other zones. From the lower Ialomicioara Valley to the east, after a transversal fault, the movement of the two compartments of the Rîul Alb fault changes its direction; further it can be correlated with the fault traced south of Podu Vîrtos Beds from Ialomița Valley and north of the tectonic window within which the same Podu Vîrtos Beds crop out from Valea fără Nume up to Talea Valley. The simplified course of this fault was figured also on general maps (Săndulescu et al., 1978; Săndulescu, 1976).

In order to achieve a structure similar to the above-mentioned one the following succession of events had to occur :

— the resuming of the thrust of the Bratocea Digitation over the Comarnic Digitation after the accumulation of the whole continuous sequence from Gura Beliei marls to Şotrile-Valea Caselor facies. It enables us to establish the moment when the thrusting took place which, according to the most recent deposits from the above-mentioned sequence, is the Lower Miocene ;

— the genesis of the Rîul Alb fault. Considering the field evidence, it appeared before the Meotian which overlies it unconformably. However, there is no other surface proof for a more precise dating of the "moment" of occurrence of this fault. As shown in Figure 1 it was considered older (before Sarmatian) than the most recent significant movements of the Carpathian Orogen over the Moesian Platform.

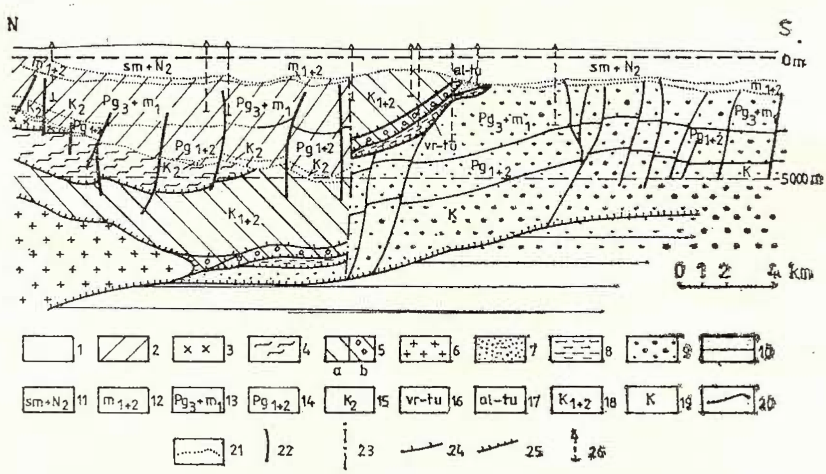

Fig. 1 — Geological section along the Doamna Valley.

1. Sarmatian-Pliocene cover ;2, Upper Cretaceous-Middle Miocene posttectonic cover ; 3, Getic Nappe ; 4, Holbav Unit ; 5, Ceahlău Nappe : a, Bratocea Digitation ; b, Comarnic Digitation ; 6, Danubian Realm ; 7, Teleajen Nappe ; 8, Macla Nappe ; 9, Tarcău Nappe ; 10, Moesian Platform ; 11, Sarmatian-Pliocene ; 12, Lower and Middle Miocene ; 13, Oligocene-Lower Miocene ; 14, Paleocene and Eocene ; 15, Upper Cretaceous ; 16, Vraconian-Turonian ; 17, Albian-Turonian ; 18, Lower and Upper Cretaceous ; 19, Cretaceous ; 20, geological boundary ; 21, unconformity ; 22, dip-slip and upthrow faults ; 23, Rîul Alb fault ; 24, Bratocea Digitation basal thrust plane ; 25, thrust nappe ; 26, drillings carried out whose data are plotted on the section plane.

The fault generated a differentiation of the relief by the uplifting of its southern block. The higher relief from the southern compartment was more intensely eroded, so that the frontal part of the Bratocea Digitation was partially or totally removed. Consequently, the interference of the Rîul Alb fault is one of the main reasons which makes impossible the tracing of the thrustings from the internal flysch zone to west of the Bărbulețu Valley basin.

Extrapolation of Data

The uplift tendency of the southern compartment of the Rîul Alb fault occurs also west of Bărbulețu Valley, so that in the Stîlpeni uplift zone, determined by drillings (Hristescu et al., 1967), the deposits of the Lower Cretaceous flysch occur directly from under the Pliocene cover.

The cores drilled from this uplift point out the existence of deposits which can be referred to several major structural units: Ceahlău Nappe with Bratocea (most of the uplift area) and Comarnic digitations, Teleajen Nappe and Macla Nappe (Fig. 1).

The correlation of the drilling data with the data from the seismic sections effectuated in Doamna Valley by IPGG accounts for the thrusts of the mentioned nappes over the Tarcău Nappe. The thrust plane between them stops suddenly in a vertical fault (Fig. 1); to the north of this fault there are deposits of the Upper Senonian Miocene covers and for this reason the fault is regarded as the western continuation of the Rîul Alb fault. Its main influence, as in the area where it crops out, is that of sinking very much its northern compartment so that the deposits of the Cretaceous flysch nappes have not been intercepted by drillings to the north of this fault and to the west of Stîlpeni rise (Fig. 2).

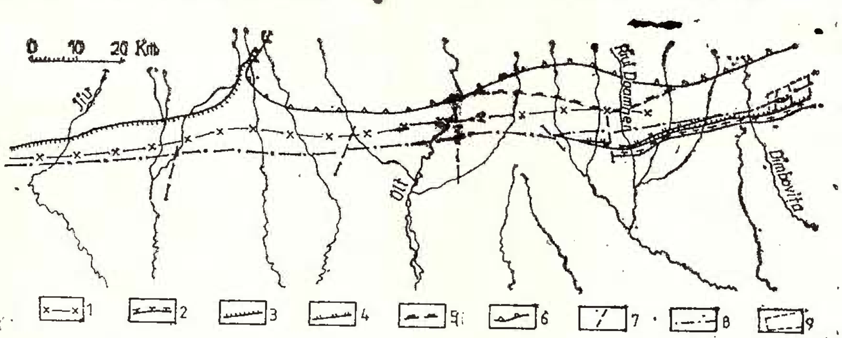

Fig. 2 — Sketch of the Rîul Alb direction.

1, axis of gravimetric minimum; 2, frontal line of the Cretaceous flysch nappe; 3, trace of the Ceahlău Nappe thrust plane; 4, frontal line of the Bratocea Digitation; 5, frontal line of the Holbav Unit; 6, trace of the Getic Nappe thrust plane; 7, traverse fault; 8, Rîul Alb fault; 9, position of the study area.

At the Doamna Valley meridian this tectonic accident occurs on the external flank of the gravimetric minimum which borders the South Carpathians. Considering that its participation in the gravimetric anomaly has the same result west of Doamna Valley too, we extended hypothetically its course in the same position as against the axis of gravimetric minimum up to the Jiu Valley meridian. Although west of the Jiu meridian there is no physical indication in favour of its extension, one cannot but mention that along its strike there are the right-slip faults in the Motru Valley basin.

It is worth mentioning that the structure of the zone between Ialomiţa Valley and Bărbuleţu Valley represents a clear model which has to be taken into account when interpreting the data on the structure of the eastern part of the Getic Depression. We are referring to the western prolongation of the Rîul Alb fault which represents a significant tectonic element which brings into contact the Cretaceous flysch nappes with the western prolongation of the Tarcău Nappe not through a thrust nappe but through a vertical fault which can apparently separate two simple adjoining synclines.

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STRUCTURA ZONEI INTERNE A FLIŞULUI ÎN SECTORUL PIETROŞIŢA-RÎUL ALB-BĂRBULEŢU

(Rezumat)

Nota prezintă structura zonei interne a flişului în extremitatea sa vestică de aflorare, între Pietroşiţa şi Bărbuleţu. În acest sector au fost deosebite două dintre digitaţiile pînzei de Ceahlău, şi anume cea de Bratocea şi cea de Comarnic, separate printr-un plan de încălecare de vîrstă miocen inferioară. Atît depozitele digitaţiei de Bratocea, cît şi cele ale cuverturii post-tectonice a unităţilor cu tectogeneză intrasenoniană sînt afectate de o falie verticală — falia Rîul Alb — cu compartimentul nordic puternic coborît. Modelul recunoscut la zi în sectorul cercetat se continuă spre vest cel puţin pînă la meridianul Văii Doamnei, unde datele de foraj şi seismice indică o structură similară.

QUESTIONS

M. Săndulescu 1. Which are the arguments on the basis of which you figured (on the general section presented) structural elements of the Danubian Realm?

2. On what basis did you considered the Holbav Unit to have such a wide extension?

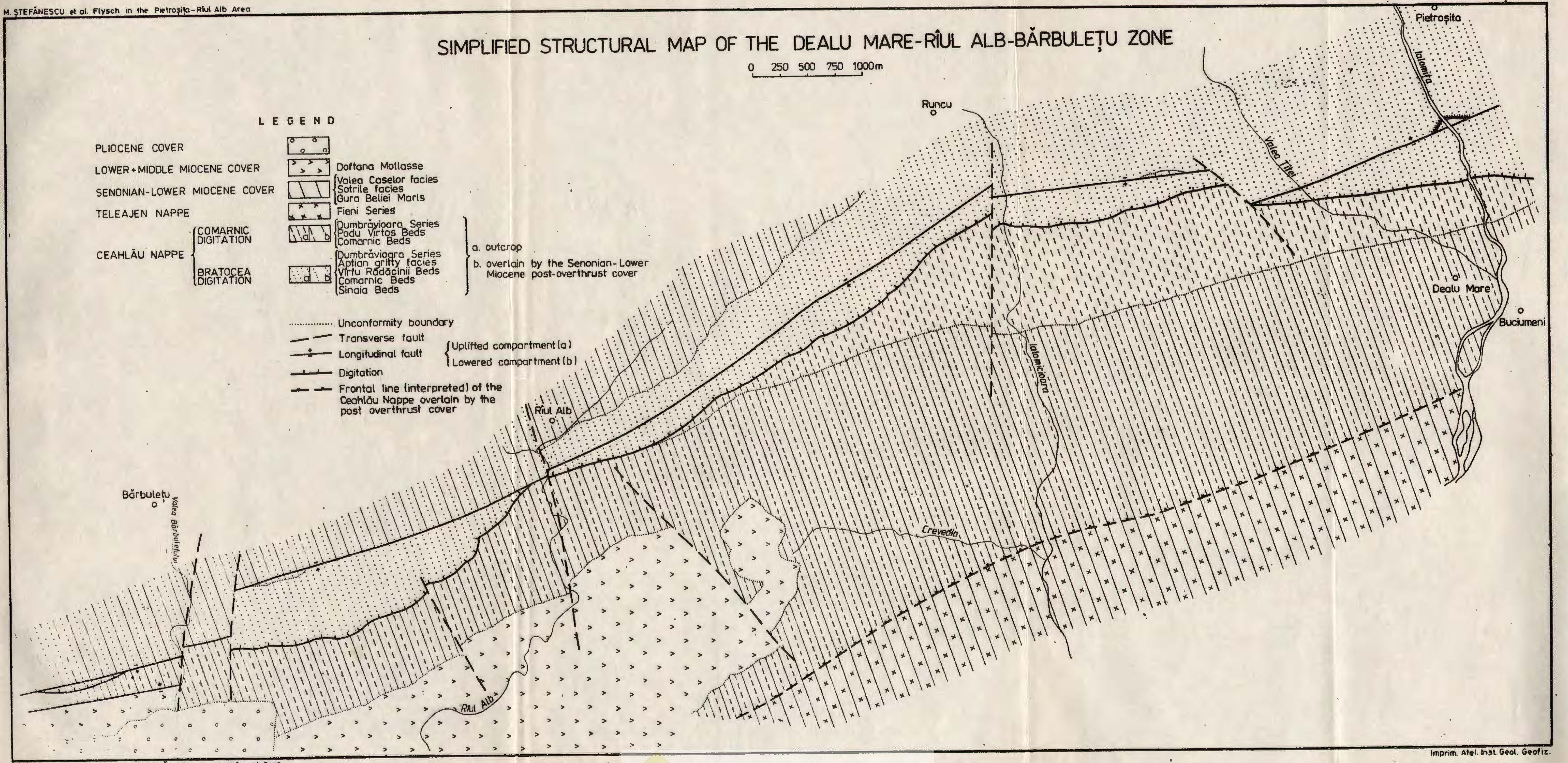
Answers 1. On the section appended to the paper the Getic Realm (a unit corresponding to it) was represented below the internal part of the Ceahlău Nappe and in the interior of the vertical plotting of the frontal line of the Getic Nappe. Its presence at the Doamna Valley meridian was admitted in order to give an interpretation, at least partial, of the increase of the Δg values north of the minimum minimorum zone in the south of the South Carpathians.

2. The presence of the Holbav Nappe was admitted up to the Vîlsan Valley basin area on the basis of data supplied by the drillings in the locality of Vîlsăneşti. This drilling intercepted a sequence of deposits belonging to the posttectonic covers and then it crossed and cut cores from Lower Jurassic (Tithonian) calcareous deposits and Liassic detrital deposits. The latter include also a tuffaceous sandstone, a lithologic element which points to a similarity of the Liassic formation intercepted by drillings with those of the Holbav Nappe.

DISCUSSIONS

M. Săndulescu. This paper deals with problems related to the deep structure of the Carpathian Chain beside new elements based on surface data. In this respect we shall refer first to the extension of the Danubian elements up to the Doamna Valley transversal line and generally to the western part of the Romanian Carpathians Bend Zone. The configuration of the gravimetric anomaly, regional minimum minimorum respectively, cannot be considered an argument in favour of the extension of Danubian elements because it has different causes (bodies with relatively lower densities situated in underthrust platforms). Unlike it, the configuration of the regional magmatic anomaly leads to conclusions contrary to the hypothesis presented (the extension of the Danubian elements east of the intramolasse fault). According to these data (magnetometric and aeromagnetic) in the (underthrust) basement of the flysch nappes, the Ceahlău Nappe inclusively, in the bend zone occur elements similar to those found in Central Dobrogea (Greenschists Zone) and South Dobrogea (Palazu-type formations) well marked in the characters and intensities of the magnetic anomaly. For this reason and taking into account the palinspastic reconstitutions, we consider that there are no arguments for the extension of Danubian-type elements (in spite of specific features of basement, sedimentry formations and tectogenetic evolution of the realm) in the deep structure of the Romanian Carpathians Bend Zone.

Answer : The extension of the Danubian units up to the Doamna Valley meridian was admitted as one of the possible reasons which can explain the increase of the values north of the minimum minimorum zone, not to justify the existence of the gravimetric minimum proper. The interpretations of the data supplied by other geophysical methods are also hypotheses which, for the time being, cannot be verified by drillings at least in the study zone; all this makes us to maintain our hypothesis according to which elements of the Danubian Realm can occur up to the Doamna Valley meridian in an internal position as illustrated on the section appended to the paper.



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D. S. Inst. Geol. Geofiz., vol. 74/5 (1987), 1990, pag. 145—148

TECTONICĂ ȘI GEOLOGIE REGIONALĂ

BOOK REVIEW

N. P. JAMES, P. W. CHOQUETTE (Edit.): *Paleokarst*. Springer-Verlag, New York, Berlin, Heidelberg, London, Paris, Tokyo, 1988, 416 pages, 277 figures.

The volume edited by Noel P. James and Philip W. Choquette has its origins in a symposium entitled "Paleokarst Systems and Unconformities — Characteristics and Significance", organized by the Society of Economic Paleontologists and Mineralogists in Colorado in 1985.

This volume includes mostly reports presented at the Symposium as well as some papers received after the meeting.

In "Introduction" Philip W. Choquette and Noel P. James mention that the book is dedicated to the documentation and interpretation of karst in the geologic record. Several definitions of the paleokarst and the way in which the karst term is used in the book, the conditions of karst formation and intrinsic factors (lithology, formation permeability and fractures) and "extrinsic" factors (climate and vegetation) are also presented.

The book is divided into two sections: Part one consists of 7 papers dealing with aspects of the development, preservation, modification, and recognition of karst terranes. Part two consists of 11 papers documenting examples of paleokarst, Proterozoic to Cretaceous in age, from a wide variety of shelf and platform settings.

Part I — General Karst Features and Processes — deals with several aspects related to paleokarst : subsurface karst, spelothems, cave sediments, karst breccia, surface karst, and "drowned" karst.

The first paper, signed by Derek Ford, "Characteristics of Dissolutional Cave Systems in Carbonate Rocks" is a synthesis of the subsurface dissolutional process. Most of the caves seem to have been formed as a result of the meteoric water circulation. Many caves have multiphase features with sequences of levels pointing to a vadose morphology or a morphology created by shallow or deep phreatic water. Paleokarst is presented not only by karst phenomena generated by dissolution but also by different precipitations which have subsequently filled the voids.

In the paper "Geochemical Patterns of Meteoric Diagenetic Systems and Their Application to Studies of Paleokarst" K. C. Lohmann describes in chemical respect the operation way of the system emphasizing the content of ^{18}O and ^{13}C stable isotopes.

On the basis of geochemical and petrographic analyses, L. A. Gonzales and K. C. Lohmann in their paper "Controls on Mineralogy and Composition of Spelean Carbonates : Carlsbad Caverns, New Mexico" analyse the speleothems from the well-known caverns in the world and made a comparison of the geochemical analyses of the paleokarst speleothems.

The karst breccia can host several types of ore deposits, particularly Mississippi Valley-type deposits. In his paper "Breccia-Hosted Lead-Zinc Deposits in Carbonate Rocks" D. F. Sangster suggests two hypotheses : 1) solution collapse brought about by the incursion of meteoric water into the carbonate strata during the period of emergence ; 2) solution collapse induced by a slightly earlier phase of the ore solutions themselves with no meteoric involvement.

In their paper "Blackened Limestone Pebbles : Fire at Subaerial Unconformities" E. A Shinn and B. H. Linz, on the basis of the studies and tests carried out on Cenozoic and Holocene limestones in the Caribbean, reach the conclusion that the selective blackening of the carbonate pebbles is caused by the "instantaneous" forest fire heating.

The genesis of marine carbonates of drowned karst systems is little known. A step in this respect is presented by P. L. Smart, R. J. Palmer, F. Whitaker and V. P. Wright in their paper "Neptunian Dikes and Fissure Fills : An Overview and Account of Some Modern Examples". They describe the most important factors in the initiation and development of sedimentation in neptunian dikes and in the filling of the fissures. The preliminary results of the sedimentological studies carried out in "blue holes" from the Bahama Banks are given as an example.

From the same area K. A. Rasmussen and C. A. Neumann describe the latest results of a study effectuated in a shallow drowned karst in their paper "Hologene Overprints of Pleistocene Paleokarst : Bight of Abaco, Bahamas". This study demonstrates clearly how a karst surface can be modified immediately after its formation.

Part II — Examples of Paleokarst Terranes — includes papers dealing with different situations of paleokarst terranes. They are presented in chronological order, from the Precambrian till the Cretaceous.

One of the most important questions concerning the paleokarst is how far in the geological eras can the paleokarst terranes be recognized? This question is answered by C. Kerans and J. A. Donaldson in their paper "Proterozoic Paleokarst Profile, Dismal Lakes Group, N.W.T., Canada", which gives a complete spectrum of the dissolution, precipitation and deposition in Precambrian formations.

A. Desroches and N. P. James describe in their paper "Early Paleozoic Surface and Subsurface Paleokarst : Middle Ordovician Carbonates ;Mingan Islands, Québec" Middle Ordovician platform carbonates on the northern margin of the Appalachian orogen. These rocks are characterized by two paleokarst unconformities and numerous paleokarst surfaces.

The paper "Ordovician Knox Unconformity, Appalachians" by W. J. Mussman, I. P. Montanez and J. F. Read describes paleotopographic highs, sinkholes and caves that extend to over 65 m below the unconformity and intrastratal breccias down to 300 m.

C. F. Kahle presents a variety of paleokarst elements in his paper "Surface and Subsurface Paleokarst, Silurian Lockport, and Peebles Dolomites, Western Ohio". They are partially overlain by a paleosol whose composition is similar to the present soils.

Three papers deal with the Mississippian system in North America. In his paper "Madison Limestone (Mississippian) Paleokarst : A Geologic Synthesis" W. J. Sando presents a synthesis of data from several published and unpublished papers and describes the extension of this paleokarst in Wyoming and adjacent states.

In the study "Late Mississippian Paleokarst and Related Mineral Deposits, Lead-ville Formation, Central Colorado" R. H. De Vado describes the extension of the paleokarst presented in the above-mentioned paper southwards, in Colorado. The paleokarst includes numerous dolines, cavities, enlarged joints extending in the Carbonatic deposits overlain by the Devonian and Ordovician Mississippian formation.

W. J. Meyers deals, in his paper "Paleokarstic Features in Mississippian Limestone, New Mexico", with the extension of the paleokarst in Mississippian formations and more southwards in New Mexico.

V. P. Wright in "Paleokarsts and Paleosols as Indicators of Paleoclimate and Porosity Evolution : A Case Study from the Carboniferous of South Wales" uses the paleokarst as an indicator for paleoclimate. Extending the results of this study Wright confirms the observations made by Longman in 1980 according to which most of the oolitic limestones forming the host rock for oil were cemented under conditions of an arid climate; no such situation occurs in a humid climate.

Paleokarst development in a tectonically active region is described by J. A. Vera, M. Garcia-Hernandez, J. M. Molina and P. A. Ruiz Ortiz in the paper "Paleokarst and Related Pelagic Sediments in the Jurassic of the Subbetic Zone, Southern Spain".

Karst phenomena such as caves, speleothemes, breccias are presented in connection with an emergence of the Liassic carbonatic platform which led to the change of the sea level.

The Cretaceous in Mexico is another region where hydrocarbons are hosted in karst reservoirs, e.g. Golden Lane area which yielded up till now 1.6 billion baryls of oil. In his paper "Sedimentation and Diagenesis Along an Island-Sheltered Platform Margin, El Abra Formation, Cretaceous of Mexico" C. J. Minero described the karst phases of an episodically exondated platform, still exposed to exondation.

Gh. Ponta

OGNJEN BONACCI : *Karst Hydrology With Special Reference to the Dinaric Karst.* Springer-Verlag, Berlin, Heidelberg, New York, London, Paris, Tokio, 1987, 184 pages, 119 figures.

Professor Dr. Ognjen Bonacci from the University of Split, Yougoslavia, presents in the volume "Karst Hydrology With Special Reference to the Dinaric Karst" a synthesis on the knowledges on the karst hydrology with numerous practical examples; most of the examples refer to the Dinaric karst areas, but there are numerous examples from other parts of the world also (United Kingdom, France, the Soviet Union, Switzerland, etc.).

This book is primarily hydrological, but it is written so as to build a bridge over the gaps between experts from varoius scientific fields related to karst.

The study of water circulation in karst involves numerous hydrometric measurements and detailed scientific analyses. This book covers 15 years of the author's work.

The volume is divided into 9 chapters.

Chapter 1 — Introduction.

Chapter 2 — Karst Terminology — gives definitions of some terms which show the connection water and carbonatic rocks as well as of terms of karst geomorphology.

Chapter 3 — Principles of Karst Groundwater Circulation — presents several classifications of the way of water circulation in karst, the reference model being that presented by Droguw in 1980, divided into two subchapters: Subcutaneous Zone and Karstification Depth and Karst Capacity for Water Storage.

Chapter 4 — Karst Springs — gives a detailed classification of the karst springs, with examples especially from the Dinaric karst and details on the discharge curves, hydrograph analysis and determination of the catchment area.

Chapter 5 — Swallow Holes (Ponors) — describes several types of ponors, with several examples, and determination of swallow capacity of ponors.

Chapter 6 — Natural Streamflows in Karst — deals particularly with the interaction between groundwater and water in the open streamflows, hydrologic regime of rivers in karst and water losses along the open streamflows in karst.

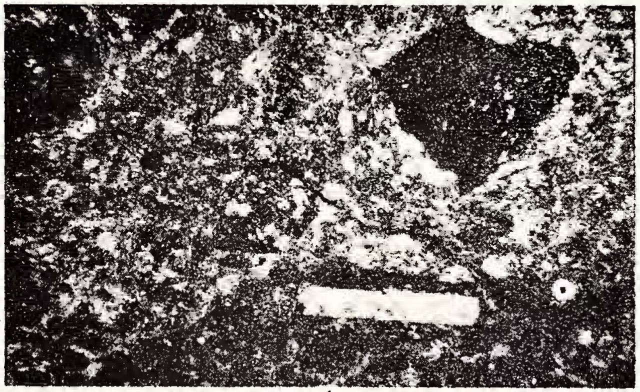
Chapter 7 — Hydrologic Budget for the Poljes in Karst — presents ways of calculation of the hydrologic budget.

Chapter 8 — Water Temperature in Karst — shows that the measurement of the water temperature provides answers to several questions connected with water circulation in karst, its origin, etc.

Chapter 9 — Man's Influence on the Water Regime in the Karst Terrains — lays stress on the catchment and exploitation of the karst waters.

In conclusion, Professor Ognjen Bonacci's book represents a contribution to the knowledge of the karst hydrology, a reference book in this field.

Gh. Ponta



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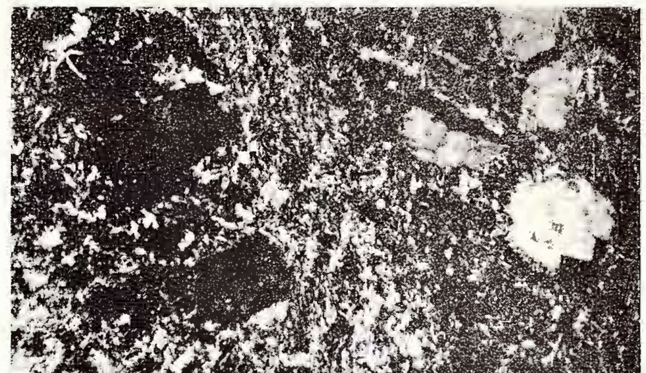

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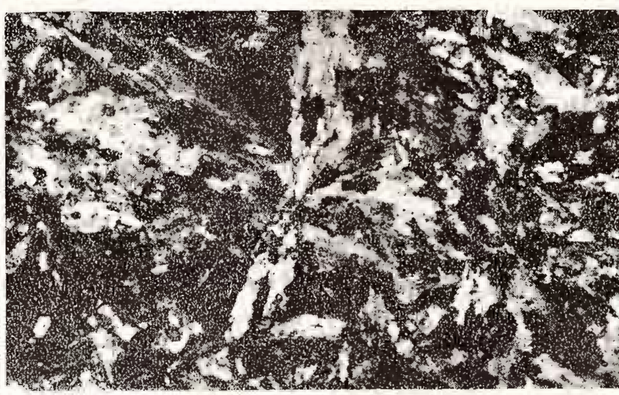

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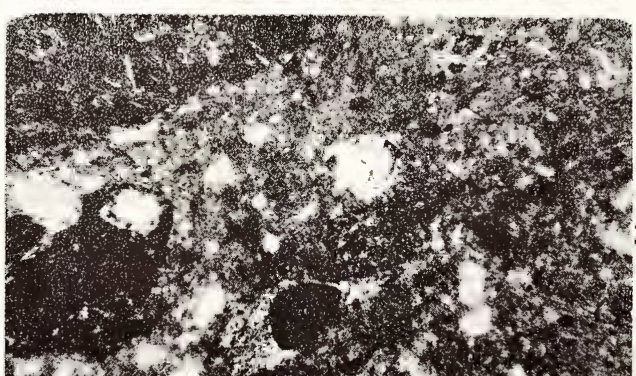

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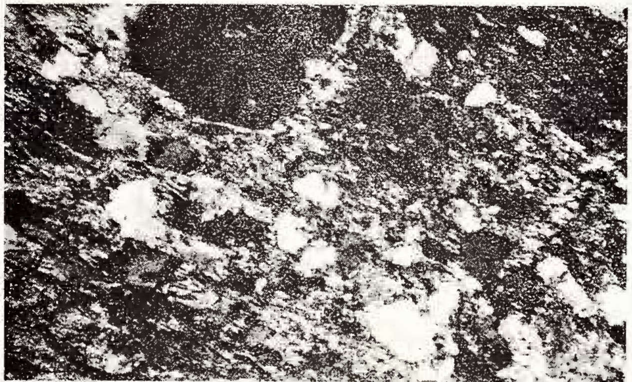

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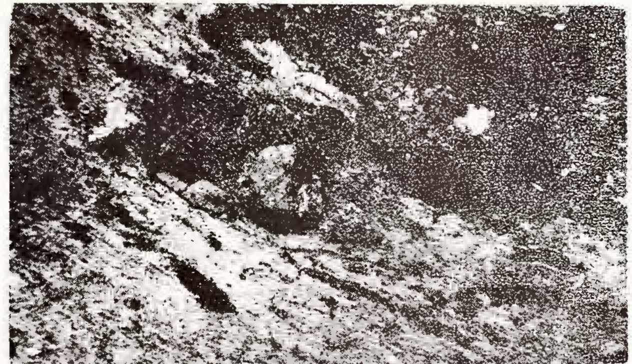

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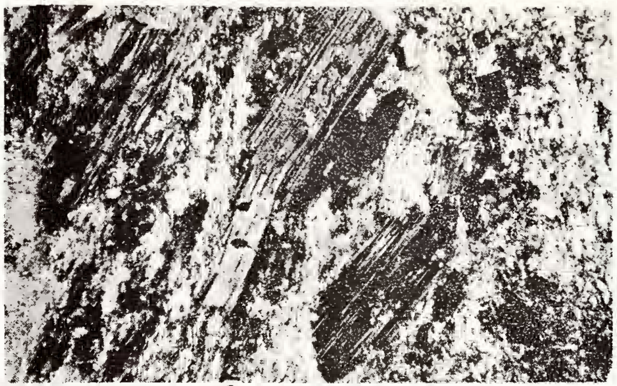

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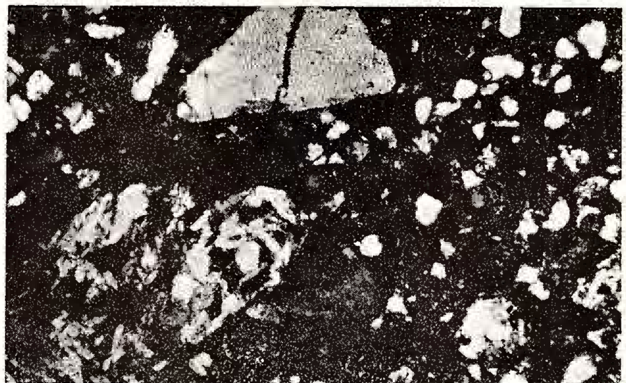
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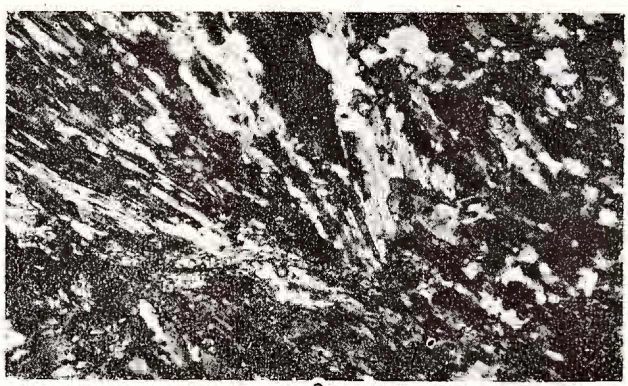
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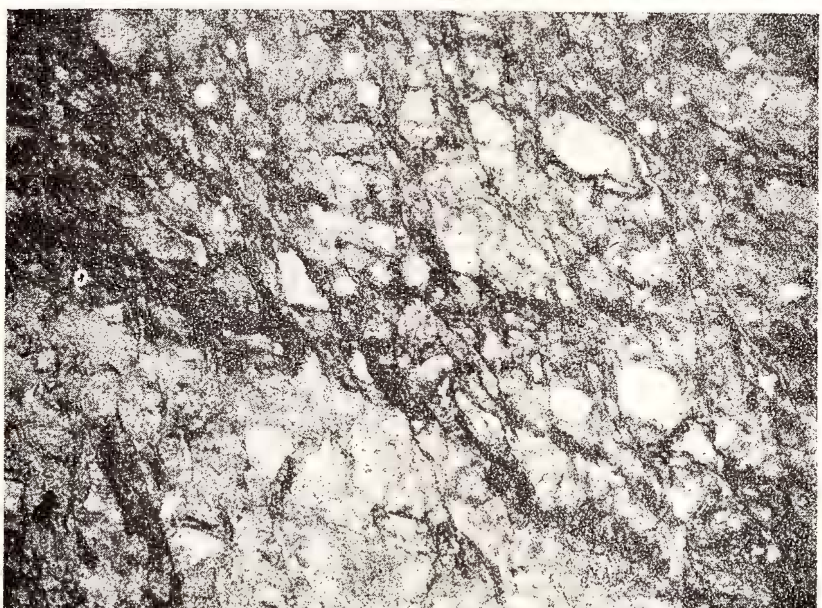

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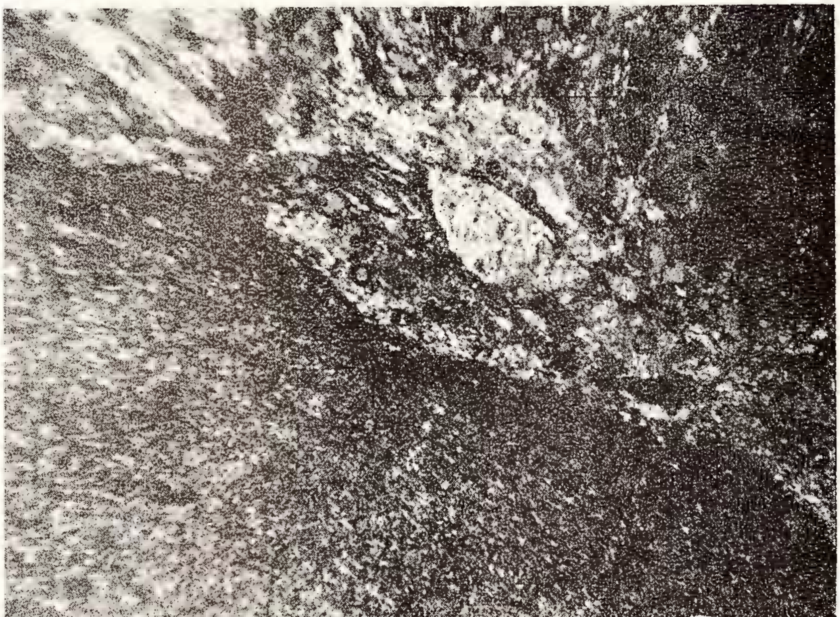

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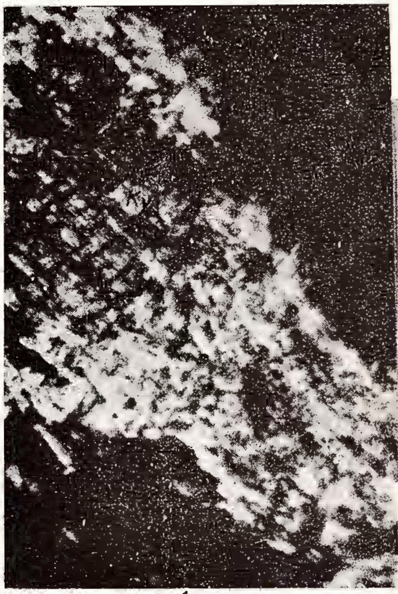

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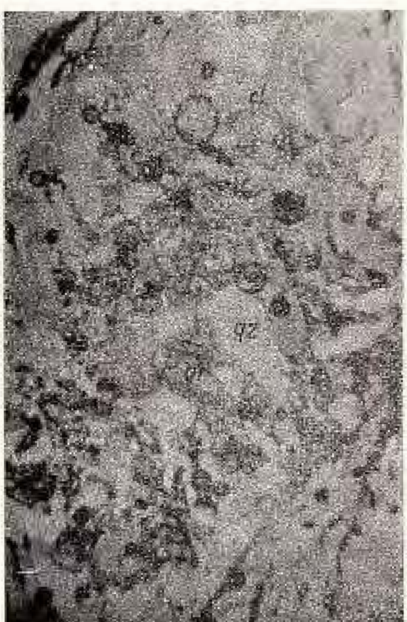

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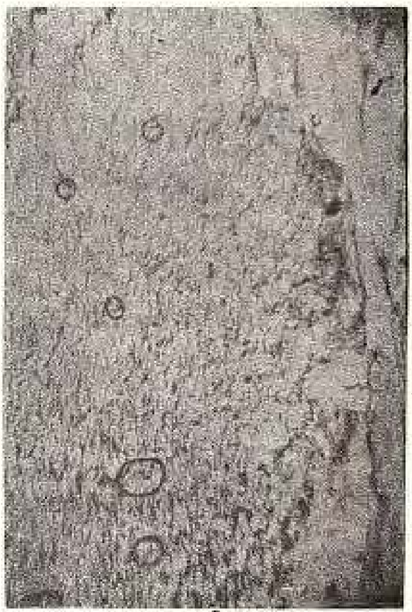

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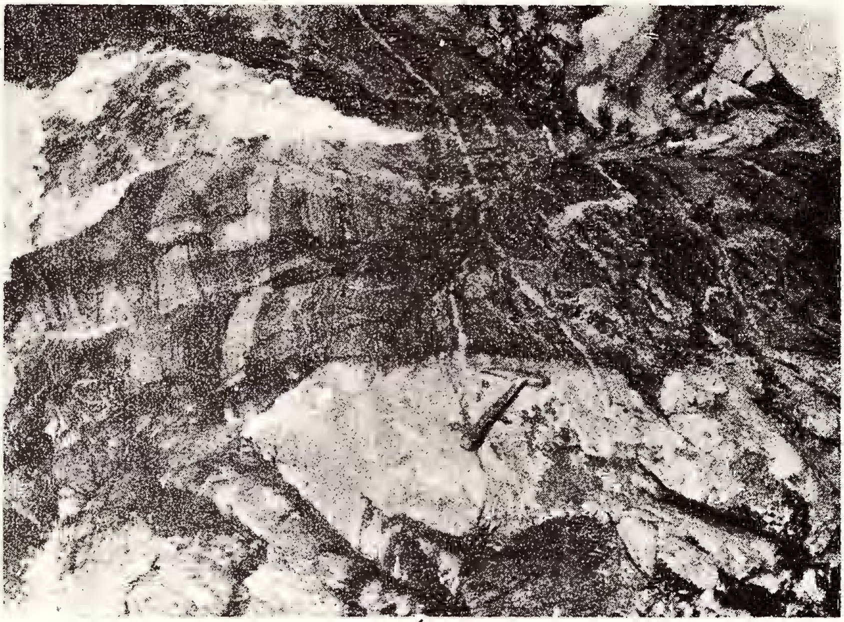

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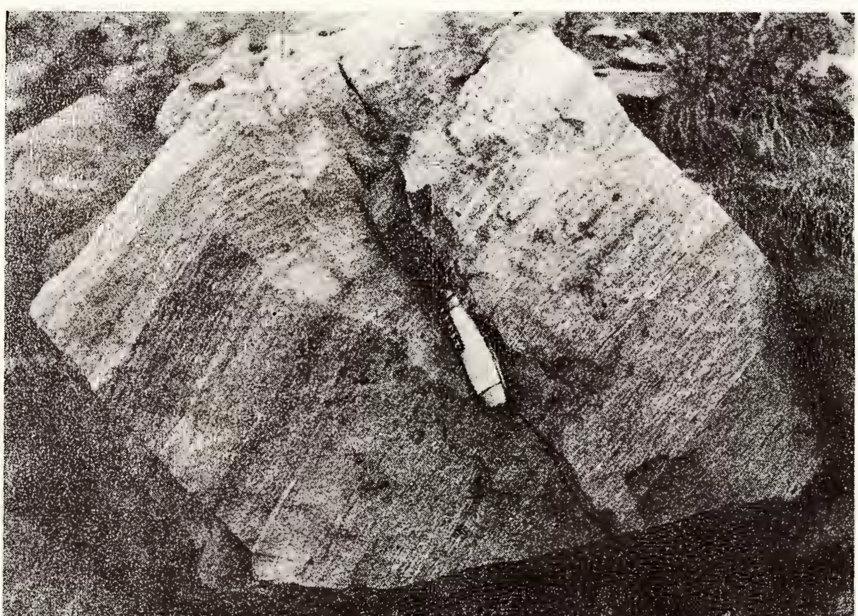

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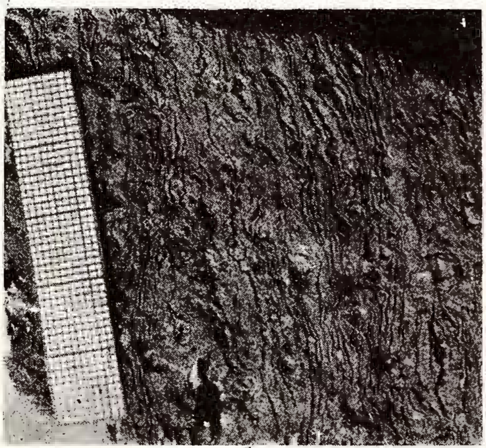
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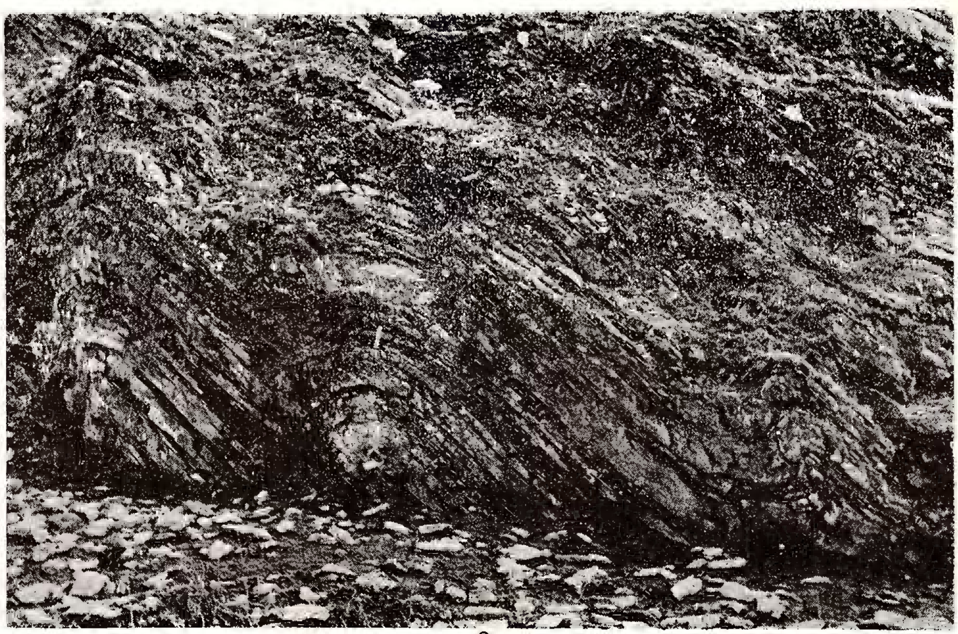

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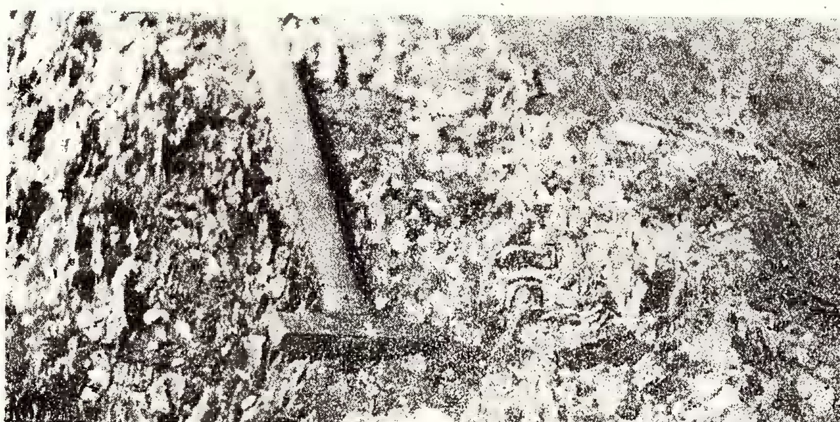

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Coperta : **Cristian Vasile**

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Comitetul Geologic t.XXXVII–LII/1(1953–1966)

Comitetul de Stat al Geologiei t.LII/2–LV/1 (1967–1969)

Institutul Geologic t.LV/2–LX(1970–1974)

Institutul de Geologie și Geofizică – à partir du tome LXI(1975)

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