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CONTENTS

Retezat Granitoid Pluton (South Carpathians) – A Geochemical Approach *Berza T., Macaleţ V., Andăr P., Udrescu C.* 1

Culmea Cernei Granitoids – Structural Setting and Petrochemical Data *Iancu V., Andăr A., Udrescu C.* 19

Characterization of S–I Type Granitoids from the Banat Area, South Carpathians *Stan N., Tiepac I.* 33

New Data Concerning the Origin of the Igneous Leucocratic Rocks in Northern Sebeş Mountains – South Carpathians. Geochemical Evidence *Dobrescu A., Stoian T.* 41

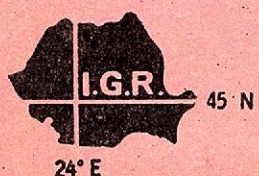
Metasomatitele brecioase asociate zăcămintelor de baritină şi de sulfuri din zona Somova (Dobrogea de nord) *Întorsureanu I.* 51

Metamorphic and Deformation Processes in the Ophiolites of the Severin Nappe and the Associated Stratiform Copper Mineralization (South Carpathians) *Savu H., Vanghelie I., Vâjdea E., Tănăsescu A., Stoian T., Stelea G.* 57

Petrology and Geochemistry of the Effusive Rocks from the Ocean Floor Basalt Complex in the Mureş Zone *Savu H., Udrescu C., Neacşu V., Ichim M.* 67

Petrological Significance of REE, U, Th, K, Rb, Sr, Hf and of the $^{87}Sr/^{86}Sr$ in the Basaltic Complex of the Mureş Zone *Savu H., Stoian M., Tiepac I., Popescu Gh., Grabari G.* 77

Intrusive Breccias Related to Microgranitic Rhyolites in the Dumbrava–Căpuşu Mic Area (NW Transylvania) *Ştefan A., Roşu E., Milu V., Rusu A.* 85



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RETEZAT GRANITOID PLUTON (SOUTH CARPATHIANS) A GEOCHEMICAL APPROACH

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Key words: Granitoids. Major elements. Minor elements. Trend-surface analysis. South Carpathians - Danubian Basement - Retezat Mountains

Résumé: Le massif granitoïde de Retezat, probablement intrus au Precambrien , est constitué d'une masse prédominante de leucogranodiorites et leucotonalites à biotite, muscovite et épidote, avec des bordures dioritiques et des zones centrales tonalitiques ou granitiques. 157 analyses chimiques ont démontré une évolution limitée à caractère calco-alcalin et, par l'analyse de tendance, une distribution plus ou moins centrifuge ou centripète des taux des principaux oxides.

1. Introduction

The highlands of Retezat Mts – one of the most Alpine-like massifs of the South Carpathians – broadly correspond to the outcrop area of a granitoid body, a lithologic feature readily noticeable in the typical granitic geomorphology. With an exposed area around 180 km^2, the Retezat Pluton is one of the important cases of the petrographic province represented by a lot of granitoid intrusions in the crystalline basements of the Danubian Nappes, the lowest structural units of the South Carpathians.

Noticed as a granitic area by early geologists at the beginning of this century, the first petrographic study of Retezat granitoids was published by Gherasi (1937), who mapped and described muscovite-epidote granites in the southernmost part of the body. He was the first to mention epidote as an important phase and to stress the abundance of muscovite.

Pavelescu (1953a, 1953b) mapped the whole outcrop area of the Retezat Pluton and separated "laminated granodiorites" at the borders, "gneissic granodiorites" on the limbs and "massive granodiorites" in the central area. He interpreted this zonal structure as a consequence of the diapiric intrusion of the massif, the external effect being a retrogression of surrounding amphibolites to greenschists.

Lately, Macaleţ (1983, 1991) has studied the northern and central parts of the Retezat Pluton, presenting detailed mineralogic and petrographic observations, and Berza (between 1982–1986) has surveyed the southern area; their samples constitute the material basis for the present work.

2. Geological setting

Following the structural and lithostratigraphic terminology used in recent papers devoted to this area of the South Carpathians (Kräutner et al., in Zoubek et al., 1988; Berza et al., 1988a, 1988b), the Retezat Pluton is emplaced in metamorphic rocks belonging to the Drăgşan Group from the basement of the Lainici Nappe – an important Alpine tectonic unit of the Danubian Nappes, outcropping on large areas in the Retezat, Vâlcan and Parâng Mts.

With a striking dominance of amphibolites, (± garnet), the Drăgşan Group metamorphics are also represented by augen gneisses, biotite gneisses (± garnet), mica gneisses (± kyanite ± sillimanite ± garnet), actinolitic schists and serpentinites. Clearly polymetamorphic, this sequence frequently bears a low-grade look, expressed by abundant chlorite, epidote, fine grained white mica, mostly connected with intensive shearing induced by both pre-Alpine and Alpine tectonics. K/Ar isotopic ages do not exceed 325 ± 9 Ma (Grünenfelder et al., 1983), but regional grounds strongly suggest a late Precambrian age for the regional medium-grade metamorphism of the Drăgşan Group. Due to young, post-nappe, folding in the Retezat Mts, the pre-Alpine basement of the Lainici Nappe outcrops in the core of an E-W oriented antiform, covered southwards and northwards by its

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Alpine sedimentary cover (Lower Jurassic conglomerates and sandstones, Mid-Upper Jurassic - Lower Cretaceous limestones, Eocretaceous flysch) and by upper Danubian Units. Between Râul Bărbat in the East and Râul Mare in the West, the Drăgşan metamorphics are intruded by a NE-SW oriented granitoid body, the Retezat Pluton. With a NW-SE dimension of around 5–7 km, and a NE-SW length of around 30 km (visible between southern and northern covers), the Retezat Pluton covers ca 180 km^2, being one of the main granitoid plutons from the South Carpathians. Considering both the noticeable southwards broadening of the pluton and the petrographic zonality, we assume an at least equal development of the pluton, under the Mesozoic cover and the overlying nappes, in the deep underground of the Godeanu Mountains.

The NE-SW oriented borders of the body are generally sharp, only seldom granitoid apophyses or metamorphic roof pendants being noticed (Nucşoara brook, Mălăieşti brook). Gherasi et al. (1974) considered this contact as tectonic, but Macaleţ (1983, 1991) pointed to the existence of biotite rich schists in the border zone, and interpreted them as "schistose hornfels". On the other hand, Berza (in Berza et al., 1989) has mapped on the eastern border, inside the body, a narrow (meters to tens of meters) strip of biotite rich quartz diorites, much darker than normal biotite-muscovite Retezat granodiorites or tonalites, and, on the western border, a strip (tens of meters to 200m) of hornblende-biotite diorites. This fact enables us to consider these borders as true emplacement limits, all the dynamic effects noticeable here (Pavelescu, 1953a, b) being local effects of pre-Alpine tectonics at the contact between two different rock bodies - the granitoid pluton and enveloping metamorphic rocks. Only on the right slope of the Lăpuşnicul Mare Valley (southern part of the outcrop area), the eastern contact is represented by a young fault which crosses also the Alpine cover.

The diapiric look noticed by Pavelescu (1953a, b) arises, in our opinion, from the effect of a large scale post-nappe folding of the two pre-Alpine units included in the basement of the Lainici Nappe (see Berza et al., 1983), along a NE-SW axis, with the disposition of the Drăgşan Group and Retezat Pluton in a synform. The underlying Lainici-Păiuş Group outcrops to the east and west (here known as "Râuşoru Series") from beneath the Drăgşan belt. A similar situation was described in the Vâlcan Mts by Berza et al. (1988b).

However, the most important effect of pre-Alpine tectonics is the dynamic metamorphism undergone by the granitoids. If in the central area of the body igneous fabric and mineralogy are frequently preserved, most of the Retezat granitoids from the surrounding zone shows a foliation ("gneissic granodiorites" of Pavelescu, 1953a, b) resulted both from preferential reorientation and deformation of igneous minerals (eye-shaped K-feldspar, slipped micas), and from blastesis of oriented sericite, chlorite or quartz layers. The effects of dynamic processes are more evident in the marginal (W and E) parts of the body ("laminated granodiorites" of Pavelescu, 1953 a, b), where sometimes rocks are in fact true low-grade blastomylonites. The dip of the foliation expresses some fluctuations (sometimes even centimetric microfolds), but generally it is SE on the western flank and NW on the eastern one, suggesting the mentioned diapiric look.

In the northern end of the outcrop area, the Retezat granodiorites are overlain by dynamo-metamorphosed Jurassic (according to Berza et al., 1988a) conglomerates, arkoses and limestones. In this area the granodiorites show a well expressed alpine foliation, disposed E-W/N, parallel to the schistosity developed in the cover rocks.

3. Petrography and mineralogy

The Retezat Pluton is striking among the granitoid bodies from the South Carpathians by the large prevalence of biotite-muscovite-epidote granodiorites and tonalites (Pavelescu, 1953 a, b; Macaleţ, 1983, 1991). However, some granitic areas were mapped by the authors of this study in the southern-central part of the exposed pluton, and a narrow (tens of meters) strip of biotite rich quartz diorites was identified by Berza at the eastern border of the massif. The gradual transition, from biotite-muscovite granodiorites, through biotite granodiorites and biotite tonalites, to biotite rich quartz diorites, strongly supports the interpretation of the latter as a more basic, first crystallised, border facies of the pluton (while internal granitic facies represent last consolidated zones). A similar situation was described in the Frumosu granitoid Pluton (also from the basement of the Lainici Danubian Nappe of the South Carpathians), by Berza et al. (1981). This is not the case of a strip of hornblende-biotite diorites, bordering westwards the Retezat Pluton in the right slope of the Râul Mare Valley. From some tens to 200 m thick, this strip, noticed also by Gherasi (1937), has eastwards a neat contact with normal biotite-muscovite granodiorites, representing clearly a distinct (most probably older) intrusion, similar to the small basic precursory intrusions described at the western border of the Vârful Pietrii granitoid Pluton (from the same Danubian province) by Gherasi (1937) and by Andăr (1991).

The modal composition of 157 samples of Retezat granitoids is projected on a standard Streckeisen QAP diagram (Fig. 1A). Most of the samples are projected in the lower part of the granodioritic field, spreading

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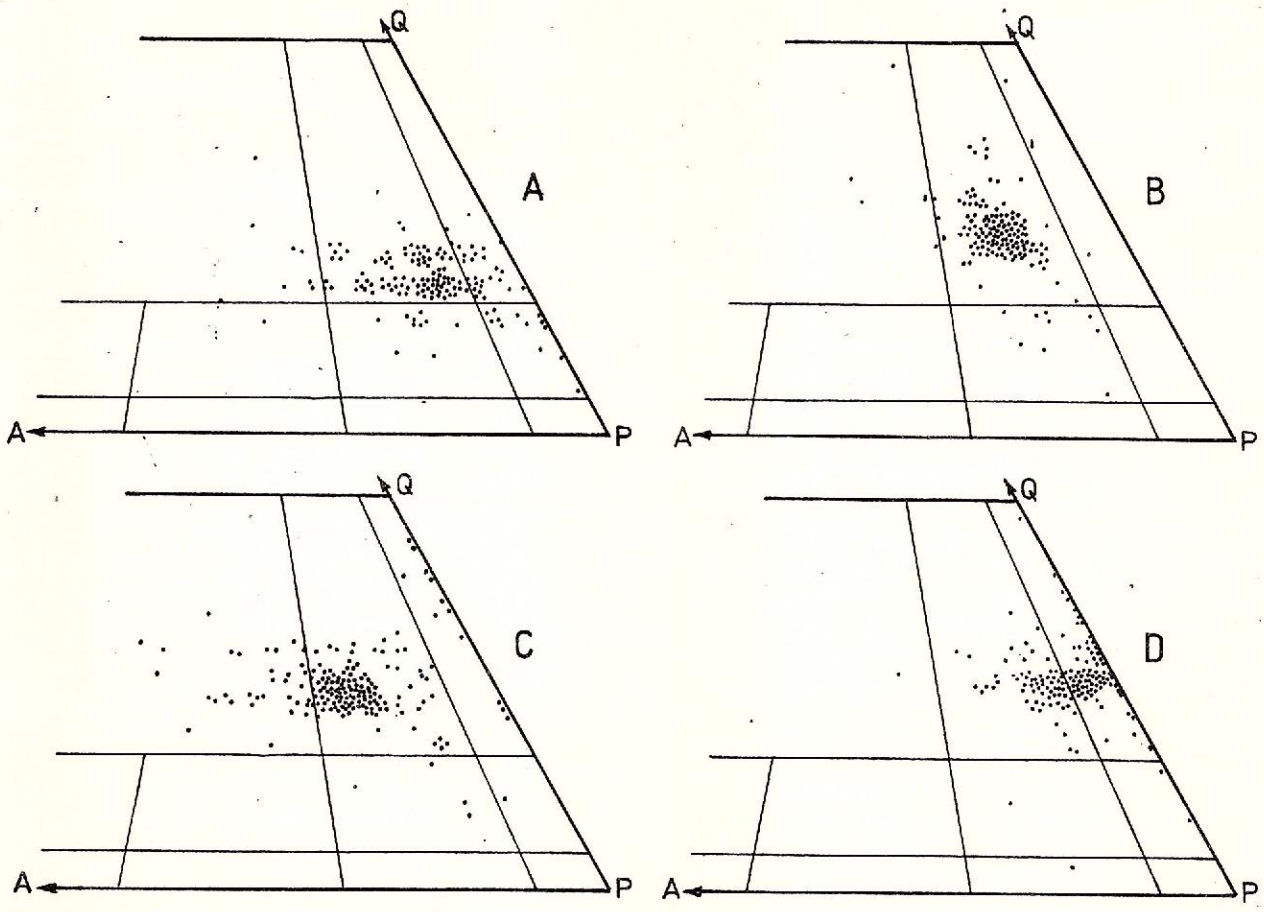

Fig. 1 - Q A P diagrams. A, Modal composition; B, CIPW norme; C, Rittmann norme; D, Mesonorme.

from it both to tonalitic and granitic fields; some samples correspond also to quartzmonzodiorites and quartzdiorites, but the main tendency of differentiation goes almost horizontally, at the Q = 20–30 % level. QAP-diagrams (Fig. 1B, 1C, 1D), resulted from CIPW, Rittmann or mesonorme calculations based on the chemical analyses of these samples, show the same prevalence of the granodioritic composition, but also sensible shifts of the cloud of projection points, each shift being specific for the computation method used.

Hornblende-biotitediorites from the right slope of the Râul Mare Valley are built from the millimetre euhedral plagioclase (too altered to be determined) and hornblende crystals, with some smaller interstitial biotite and quartz; biotite frequently grows on hornblende crystals, sometimes replacing them completely (an effect of the main, granodioritic, intrusion?). A pervasive fracture cleavage gives the rocks a foliation, characteristic of the border zones of the pluton.

Biotite-quartzdiorites bordering eastwards the Retezat Pluton are also foliated, but igneous minerals (euhedral andesine and biotite), 1–5 mm large, can be recognized, even if plagioclase suffers a transition to epidote + albite, or biotite is chloritized. Quartz is here a major component and hornblende was not yet found.

Biotite-muscovite-epidotegranodiorites and tonalites have the same 1–5 mm large euhedral plagioclase (oligoclase) and biotite crystals and intergranular quartz, but K-feldspar also becomes a major component. However, its content is difficult to estimate in hand specimen, only thin section examination revealing the petrographic type. In some areas there is a drop in the quartz content and the rocks correspond to quartzdiorites. The main characteristic of this petrotype (which covers around 80 % of the exposed area of the Retezat Pluton, giving its specificity among the Danubian granitoid plutons) is the constant presence of 1–3 mm large muscovite crystals (3–7 %; sometimes also interlayered with fresh or chloritized biotite) and 0.5–1 mm large euhedral epidote crystals (1–3 %, exceptionally 8 %). These minerals may be easily distinguished from the usual fine-

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grained white mica and epidote developed, especially in sheared rocks, at the expense of plagioclase. Gherasi (1937), Pavelescu (1953 a, b) and Macaleţ (1991) have considered the large crystals of epidote and muscovite as primary minerals, crystallized from the granitoid magma. Epidote represent here the Ca-Fe-Mg mineral phase, taking the place of the usual hornblende of Danubian granodiorites and tonalites; although frequently epidote crystals grow around a small euhedral allanite crystal, no evidence of replacement of hornblende by epidote has been found in the Retezat granodiorites and tonalites. This case supports an older hypothesis of Mrazec and Giuşcă (1934) on the direct crystallization of epidote from acid magma, but to the question of the physical control of this unusual crystallization path, we still have no answer. As regards the color index of these rocks, the sum of epidote, biotite and accessory minerals generally average 3–5 %; this means that most of the Retezat granodiorites and tonalites are, in respect of the Streckeisen classification, leucogranodiorites, respectively leucotonalites.

B i o t i t e - m u s c o v i t e - e p i d o t e g r a n i t e s differ from the rocks described above only by a larger participation of K-feldspar. According to the optical and X-ray study of this mineral done by Macaleţ (1991), it is a perthitic, intermediary, crosstwinned or untwinned, microcline with a D = 80 triclinicity index. Both in granites and granodiorites, microcline appears either intergranular or as up to 2 cm wide ocelli, including more or less corroded former euhedral plagioclase crystals, frequently surrounded by muscovite crystals.

D y k e r o c k s are not characteristic of the Retezat Pluton. Clearly connected to it, as last acid products, aplites and pegmatites are the most frequent. A large (tens of meters) garnet-muscovite aplite outcrops in the Piciorul Slăveiului summit, just over the right angle bent of the Lăpuşnicul Mare Brook. A few porphyric augite bearing microdioritic dykes are by no way proved to be in genetic relation with the host granitoid body.

4. Geochemistry

The true aim of this paper is to present a geochemical survey of the Retezat granitoid Pluton, filling by this approach the last gap in the study of the Danubian granitoid bodies (besides a few old, scattered, analyses, the first 15 chemical analyses of Retezat granitoids have just been published by Macaleţ, 1991; these 15 samples are incorporated in the 157 of the present work). In this way, the work done (mainly in the Geological Institute of Romania) in the last twenty years on the granitoids from the South Carpathians is completed, with a contribution bringing both the highest amount of analytical data and the highest ratio number of analysis on area of exposure. Following similar studies published in many countries and introduced in Romania by Andăr (1991)), we have presented our data also in the form of trend maps, picturing the areal distributions of the main chemical components.

Knowing from previous authors the lithologic monotony of the Retezat granitoid Pluton, we tried to study its geochemistry using a one kilometer spaced sample network, as geometrical as possible in a high zone (mostly over 2000 m) as this, where Quaternary morenae, boulder fields and debris cover important areas. The standard wet chemical analyses of the 157 used samples are listed in table 1 and their trace element contents of 74 samples were determined using emission-spectrometry (table 3).

From the data presented in table 1 and 3, the outsider nature of the western strip of hornblende-biotite diorites is very clear, even if the two analyses show important differences in SiO_2 and MgO content between them. For the main pluton, continuous increase in SiO_2 and K_2O, repectively continuous decrease in CaO and MgO, FeO + Fe_2O_3, TiO_2, P_2O_5, Zr, V, Cr, Ni, Sc and low variation in Al_2O_3 and Na_2O, from quartz diorites to granites, is typical of a moderately differentiated sequence. The mean contents computed for the analysed oxides and trace elements from tables 1 and 3 are presented, for the main petrotypes, in tables 2 and 4. From all these tables it results also a considerable overlap in element contents of the four main petrographic types (quartzdiorites, tonalites, granodiorites and granites), the main element for discrimination being in such cases modal analysis (Fig. 1A), but this one is also subject to the representativity of the thin section in respect of the hand specimen.

These statements are confirmed by variation diagrams. Using the Nockolds-Allen differentiation index (Fig. 2), the limits between the mentioned rock-types fall at 8.5, 10 and 12–13 of (1/3Si + K)–(Ca + Mg), projection clouds and lines illustrating graphically the above described situation. In a ternary SiO_2-CaO+MgO-Al_2O_3+$Fe_2O_{3(t)}$ Korjinski diagram (Fig. 3), projection points show also a clear differentiation trend to higher silica rocks, from quartz diorites to granites. In the FMA diagram (Fig. 4) the concomitant alkali increase and Fe and Mg decrease point to a relatively short differentiation of calc-alkaline trend.

In order to test also the R_1-R_2 Hubert de la Roche diagram, we constructed Figure 5, where most projections lie next to the border dividing the field of leucogranodiorites from that of granites, a few occupying the granodioritic field or contiguous fields. Preferring both local tradition and the international adopted Streckeisen classification, we discard this

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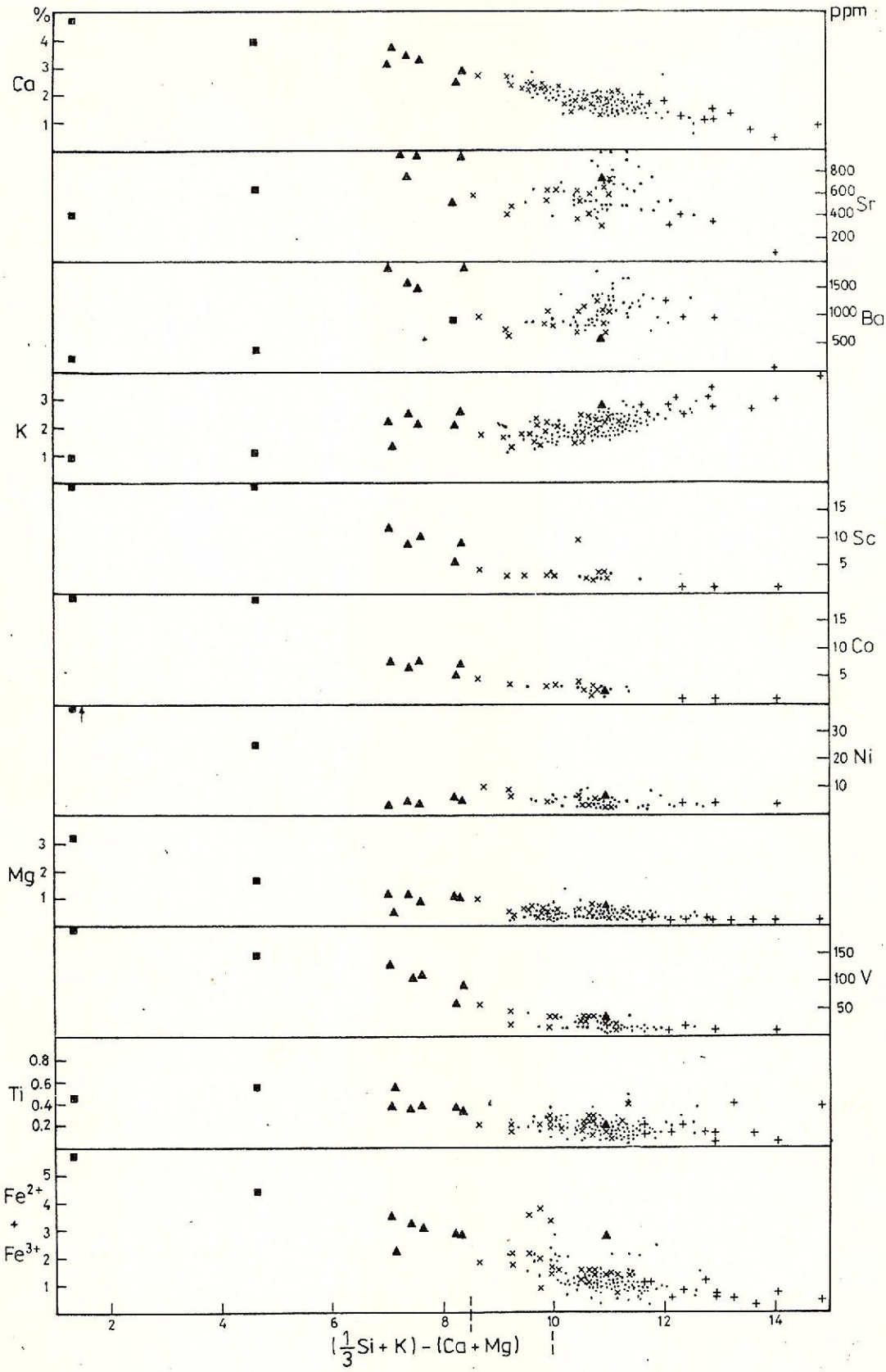

Fig. 2 - Nockolds-Allen variation diagram.

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Table 1 – Chemical analyses (%) of the granitoid

Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO
								DIORI
198	53.50	0.75	17.01	3.22	4.71	0.13	5.34	6.64
185	58.11	0.92	17.31	3.17	3.01	0.12	2.74	5.55
								QUARTZ
376	60.60	0.58	17.06	3.24	1.53	0.11	1.88	4.81
313	60.87	0.59	17.61	3.73	1.31	0.11	1.86	4.90
1848A	61.00	0.43	22.14	1.43	1.57	0.04	0.96	5.16
308	62.49	0.52	17.51	3.69	0.58	0.11	1.76	4.04
309	62.58	0.63	17.51	2.80	1.57	0.10	1.56	4.53
4683	62.65	0.64	18.40	1.91	2.17	0.01	1.81	3.44
1875	63.68	0.90	19.11	2.19	0.92	0.04	0.92	5.08
1892	66.20	0.34	16.79	3.20	0.89	0.04	0.85	2.39
								TONALI
4687	66.06	0.32	17.90	1.17	1.32	0.04	1.43	3.60
160	66.06	0.36	16.71	4.67	0.73	0.06	0.64	3.15
163	66.40	0.38	16.80	4.28	0.84	0.07	0.79	3.12
161	66.70	0.44	17.61	2.30	0.72	0.06	1.06	3.30
160A	66.90	0.38	17.76	2.46	0.58	0.07	0.79	3.12
164	67.10	0.40	16.56	4.18	0.58	0.06	0.80	3.12
4707	68.28	0.34	16.80	1.58	0.55	0.07	0.58	2.98
1895	68.38	0.30	16.94	1.40	0.41	0.02	0.60	3.64
4696	68.41	0.38	16.40	1.80	0.66	0.02	0.79	3.67
1891	68.62	0.34	16.84	1.27	0.76	0.03	0.47	2.88
1837	68.62	0.33	18.51	0.76	0.57	0.04	0.50	3.03
1826	68.85	0.23	15.78	1.78	1.00	0.04	0.87	2.77
1887	68.88	0.24	16.70	2.17	0.86	0.02	0.59	3.33
4708	68.90	0.42	16.40	1.53	0.70	0.06	0.81	3.10
187	68.94	0.24	16.52	0.73	1.01	0.04	0.53	2.73
4690	69.46	0.27	16.40	1.64	0.48	0.02	0.65	2.58
4743	69.52	0.30	16.20	1.08	0.90	0.04	0.87	2.36
1882	69.65	0.24	16.39	1.30	0.62	0.04	0.73	2.56
1852	69.68	0.36	16.39	1.21	0.61	0.04	0.65	2.27
199	69.70	0.23	15.96	1.48	0.49	0.06	0.70	2.31
1823	69.76	0.20	17.00	1.21	0.43	0.06	0.55	2.59
1848	69.84	0.13	18.65	0.04	0.38	0.01	0.35	3.45
4689	69.93	0.39	16.30	1.43	0.66	0.02	0.76	2.35
1824	70.24	0.20	16.68	0.25	0.85		0.44	1.20
1860	70.32	0.64	16.54	1.07	0.70	0.05	0.55	2.34
4616	70.33	0.34	16.80	0.34	0.72	0.02	0.38	2.40
1825	70.69	0.20	17.14	0.41	0.71		0.53	2.58
1822	70.72	0.23	17.22	0.28	0.71	0.06	0.66	1.89
276	70.75	0.19	15.40	1.53	0.32	0.07	0.51	2.39
4735	71.15	0.27	15.60	1.27	0.36	0.05	0.68	2.46
								GRANODIO
162	66.95	0.38	15.96	3.59	0.55	0.07	0.80	2.93
169	66.96	0.28	17.61	1.44	0.35	0.05	0.68	2.19
4692	67.39	0.38	16.30	2.49	0.70	0.04	1.20	2.80
1289	67.44	0.45	18.44	1.93	0.85	0.14	0.40	2.80
4709	67.71	0.28	17.50	2.14		0.05	0.67	3.17
1286	68.20	0.32	18.98	1.33	0.43	0.06	0.10	2.94
1839	68.22	0.36	19.27	0.08	0.71	0.04	0.60	2.73
628	68.35	0.35	14.85	2.28	0.81	0.24	0.60	3.92
1428	68.38	0.40	18.80	2.28	0.71	0.08	0.30	2.80
1288	68.50	0.45	19.10	2.06	0.57	0.08	0.60	2.60
1886	68.60	0.34	16.36	1.39	0.64	0.04	0.79	2.97
1429	68.65	0.42	18.14	2.11	0.85	0.04	0.30	3.36
1897	68.70	0.24	17.49	1.04	0.51	0.04	0.49	2.78
1871	68.78	0.48	16.26	1.26	0.79	0.06	0.73	2.68
167	68.94	0.24	16.86	1.74	0.43	0.07	0.51	2.27
4744	68.99	0.30	15.90	1.10	0.82	0.04	0.76	2.54
1427A	69.00	0.40	17.60	2.41	0.57	0.14	0.60	2.50
30	69.04	0.10	14.96	2.31	0.50	0.06	2.05	1.82
1890	69.09	0.28	15.14	0.90	0.66	0.03	0.70	2.32

Institutul Geologic al României

rocks from the Retezat Pluton

K_2O	Na_2O	P_2O_5	H_2O^+	H_2O^-	CO_2	S	Total
TES							
1.15	3.70	0.21	0.74		2.36	0.16	99.62
1.44	4.31	0.30	0.88		1.56	0.19	99.61
DIORITES							
3.10	4.36	0.31	1.83			0.14	99.55
2.62	3.95	0.32	1.56			0.17	99.60
1.62	4.60	0.19	0.36	0. 15	0.24		99.59
3.14	3.94	0.26	1.24			0.28	99.56
2.48	4.40	0.40	1.13			0.19	99.88
2.47	4.88	0.20	0.90			0.06	99.54
1.69	4.50	0.15	0.47		0.97	0.01	100.63
3.44	4.06	0.11	1.36			0.14	99.81
TES							
2.18	5.10	0.13	0.82			0.06	100.13
2.58	4.08	0.15	1.05		0.20	0.16	100.60
2.40	4.42	0.12	0.99		0.10	0.10	100.81
2.78	4.17	0.13	1.20		0.10	0.20	100.77
2.17	3.98	0.16	1.75			0.13	100.25
2.71	3.90	0.15	0.91			0.17	100.64
2.33	4.95	0.17	0.75			0.08	99.46
1.99	4.71	0.08	1.16			0.18	99.81
2.04	4.63	0.10	0 61			0.07	99.58
1.92	4.95	0.07	1.17			0.13	99.45
1.87	4.66	0.09	0.79	0.12			99.65
2.27	4.20	0.12	1.04	0.30			99.25
1.53	4.40	0.07	0.95			0.16	99.90
2.31	4.30	0.08	0.73			0.06	99.40
2.58	4.37	0.09	0.28		1.35	0.19	99.60
2.30	4.82	0.07	1.06			0.07	99.82
2.60	4.60	0.07	0.88			0.05	99.47
2.31	4.01	0.11	1.58			0.19	99.73
2.28	5.22	0.08	0.39		0.82	0.11	100.11
2.61	4.52	0.10	0.61		0.91	0.20	99.61
2.50	4.20	0.08	1.08	0.18	0.08		99.56
0.93	5.50	0.07	0.30	0.12		0.17	99.70
2.31	4.40	0.07	0.89			0.05	99.56
3.12	5.20	0.07	0.81	0.20	0.44		99.30
2.85	4.52	0.09	0.60		0.36	0.06	100.69
2.53	4.93	0.06	0.61			0.05	99.51
4.80	1.75	0.08	0.78	0.14			99.53
2.87	4.05	0.07	0.95	0.11			99.60
2.40	4.57	0.07	1.34			0.17	99.71
2.60	4.63	0.06	0.46			0.08	99.67
RITES							
2.59	4.28	0.12	0.97		0.10	0.16	99.45
3.58	4.65	0.06	1.35			0.25	99.45
2.70	4.10	0.16	1.42			0.07	99.75
1.95	3.37	0.09	0.42		0.13	0.10	98.51
2.02	5.00	0.09	0.62			0.06	99.49
1.92	3.34	0.08	0.60		0.83	0.08	99.21
2.37	4.36	0.09	0.92	0.11			99.64
2.62	3.26	0.13	1.39		0.63	0.09	99.52
1.87	3.37	0.07	0.35		0.44	0.05	99.90
1.50	3.20	0.08	0.36		0.48	0.04	99.62
1.99	4.79	0.09	1.28			0.18	99.46
1.37	3.77	0.07	0.43		0.44	0.05	100.00
2.70	4.22	0.07	1.04			0.20	99.88
2.31	4.89	0.08	0.54		0.41	0.42	99.69
2.81	4.63	0.09	1.31			0.08	99.98
2.83	4.26	0.06	2.12			0.07	99.79
1.85	3.17	0.08	0.80		0.35	0.04	100.51
2.45	4.60	0.04	1.34		0.40		99.67
2.13	4.57	0.08	3.79			0.17	99.86

Institutul Geologic al României

Table 1

Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO
							GRANODIO	
219	69.32	0.20	12.70	4.40	0.79	0.05	1.16	2.80
1858	69.41	0.14	16.45	1.42	0.36	0.02	0.65	3.00
1427B	69.50	0.42	18.00	1.74	0.85	0.10	0.50	2.70
4554	69.57	0.42	16.50	0.21	1.28	0.03	0.40	2.99
1851	69.61	0.13	16.75	1.55	0.57	0.07	0.53	2.77
4614	69.65	0.36	17.80	0.77	0.75	0.04	0.60	2.20
4613	69.67	0.35	17.60	0.57	0.71	0.03	0.46	2.45
1893	69.67	0.24	15.84	0.37	1.37	0.04	0.51	2.85
1864	69.69	0.72	13.78	2.85	0.27	0.05	0.98	2.38
1829	69.72	0.20	17.00	0.26	0.57		0.49	2.30
4705	69.76	0.36	16.74	1.07	0.64	0.04	0.66	2.50
1894	69.80	0.26	15.74	0.95	0.66	0.04	0.50	2.96
1817	69.85	0.23	17.11	0.39	0.57	0.08	0.59	2.04
1868	69.85	0.60	16.39	1.24	0.92	0.04	0.61	1.24
1883	69.88	0.24	15.94	1.04	0.56	0.06	0.56	2.94
1821	69.90	0.22	17.30	0.42	0.85	0.07	0.71	2.26
1896	69.90	0.20	15.94	1.09	0.41	0.04	0.52	2.85
1841	69.95	0.28	18.29	0.16	0.57		0.49	2.49
165	69.96	0.18	15.96	1.27	0.41	0.04	0.44	2.50
1888	69.96	0.20	15.80	3.10	0.47	0.03	0.43	1.51
1849	70.01	0.33	18.57		0.40	0.06	0.48	2.46
1828	70.02	0.20	17.49	0.28	0.85		0.47	2.56
1859	70.04	0.10	15.84	0.98	0.79	0.05	0.56	2.79
4552	70.09	0.61	16.80	0.54	0.60	0.07	0.60	2.28
4728	70.10	0.33	17.00	1.38	0.30	0.06	0.68	2.51
117	70.16	0.10	15.20	1.29	0.64	0.05	1.45	2.38
1867	70.16	0.24	15.59	1.16	0.55	0.05	0.59	2.56
1877	70.18	0.20	16.24	0.91	0.75	0.03	0.67	3.05
654	70.19	0.60	14.28	2.95	0.40	0.06	1.10	2.66
18	70.24	0.16	16.21	1.24	0.40	0.04	0.49	2.41
1830	70.25	0.29	17.78		0.62	0.06	0.51	2.39
159	70.27	0.24	16.24	0.94	0.70	0.06	0.49	2.41
1818	70.30	0.23	16.00	0.27	1.14	0.07	0.53	2.71
4553	70.32	0.36	16.50	0.33	0.95	0.04	0.50	2.34
1873	70.33	0.24	16.09	1.26	0.49	0.06	0.59	2.20
19	70.36	0.10	16.31	1.04	0.59	0.04	0.56	2.38
4759	70.37	0.35	15.90	0.81	0.32	0.03	0.69	0.78
1820	70.42	0.23	16.68	0.21	0.85	0.07	0.49	2.68
1827	70.55	0.20	16.74	0.62	0.57		0.60	1.97
1889	70.60	0.18	17.69	1.10	0.41	0.02	0.64	2.28
166	70.60	0.24	15.86	1.80	0.27	0.05	0.60	2.17
4727	70.60	0.32	15.90	1.06	0.32	0.04	0.50	2.35
1842	70.69	0.18	16.89	0.87	0.64		0.34	2.55
1840	70.69	0.14	16.52	1.43	0.85	0.05	0.75	1.41
4697	70.70	0.28	15.80	0.94	0.37		0.48	2.32
4699	70.73	0.22	15.60	0.89	0.62	0.03	0.63	2.40
5	70.76	0.18	15.70	1.42	0.20	0.05	0.54	2.41
1879	70.86	0.26	16.31	1.36	0.38	0.05	0.49	3.05
3	70.88	0.18	15.64	1.06	0.36	0.03	0.48	2.39
1874	70.89	0.46	13.64	2.09	0.50	0.05	0.56	2.80
1866	70.90	0.12	15.44	1.90	0.23	0.04	0.59	2.31
558	70.93	0.20	14.22	1.14	0.81	0.08	0.90	2.66
2	70.94	0.19	15.64	1.42	0.20	0.06	0.63	2.43
1869	70.94	0.28	13.89	2.42	0.54	0.05	0.57	2.78
4747	70.95	0.30	15.80	1.40	0.40	0.04	0.70	2.10
4612	70.96	0.39	15.80	0.18	0.85	0.03	0.31	2.00
97	71.01	0.04	16.24	0.89	0.33	0.03	0.46	2.23
4742	71.03	0.21	16.00	1.01	0.44	0.03	0.57	2.42
1836	71.07	0.29	17.15	0.40	0.71	0.03	0.63	1.75
96	71.09	0.12	15.79	1.02	0.38	0.03	0.53	2.12
1865	71.12	0.34	15.94	1.72	0.44	0.05	0.47	2.30
1876	71.12	0.08	15.64	0.92	0.47	0.04	0.49	2.70

Institutul Geologic al României

continued

K_2O	Na_2O	P_2O_5	H_2O^+	H_2O^-	CO_2	S	Total
RITES							
2.04	4.41	0.04	0.84		0.20	0.12	99.07
2.06	4.86	0.09	0.58		0.51	0.04	99.59
1.75	3.26	0.08	0.24		0.22	0.04	99.40
2.09	4.89	0.09	0.91			0.05	99.38
2.12	4.74	0.09	0.87	0.12			99.68
2.63	4.10	0.05	0.81			0.05	99.81
2.49	4.51	0.04	0.92			0.05	99.85
2.43	4.51	0.06	1.39			0.19	99.47
3.38	3.78	0.11	0.64		0.76	0.05	99.44
2.87	4.65	0.80	0.69	0.27	0.08		99.36
2.17	4.93	0.08	0.66			0.07	99.68
2.17	4.58	0.07	1.63			0.12	99.48
3.00	4.65	0.07	0.86	0.22	0.04		99.98
3.60	5.08	0.07	0.43		0.46	0.13	100.66
3.13	3.85	0.10	1.09			0.17	99.56
2.62	4.25	0.04	0.81	0.29	0.04		99.20
2.26	4.90	0.07	1.15			0.11	99.44
2.25	4.52	0.09	0.72	0.09			99.72
2.62	4.91	0.06	1.01			0.15	99.51
2.76	4.19	0.04	0.86			0.12	99.47
2.31	4.52	0.09	0.62	0.10			99.75
2.50	4.52	0.08	0.46	0.29	0.04		99.18
2.42	4.82	0.08	0.62		0.55	0.05	99.69
2.22	4.70	0.07	0.64			0.05	99.27
2.52	4.65	0.06	0.52			0.09	100.20
2.65	4.25	0.04	0.72		0.10	0.10	99.13
2.05	5.01	0.08	0.68		0.77		99.49
2.14	4.91	0.08	0.57		0.36	0.16	100.25
1.87	4.58	0.09	0.38		0.51		99.67
2.32	4.66	0.07	0.22		0.92	0.17	99.55
2.70	4.44	0.05	0.57	0.10	0.09		99.65
2.58	4.42	0.06	0.77		0.20	0.09	99.47
2.75	4.82	0.09	0.64	0.14	0.04		99.45
2.63	4.50	0.03	0.65			0.05	99.20
3.24	4.50	0.08	0.50		0.70	0.02	100.30
2.45	4.37	0.10	0.25		0.80	0.19	99.54
3.10	5.08	0.03	2.29			0.07	99.82
2.75	4.25	0.08	0.88	0.16			99.43
2.87	4.57	0.07	0.69	0.27	0.13		99.31
2.01	4.25	0.06	1.07			0.17	100.48
2.754	4.59	0.04	1.31			0.22	100.50
2.54	4.84	0.66	1.02			0.06	100.21
2.12	4.74	0.07	0.56	0.15	0.02		99.52
2.37	4.60	0.09	0.85	0.09	0.08		99.74
2.38	4.89	0.07	1.33			0.06	99.62
2.27	4.58	0.06	1.11			0.09	99.23
2.60	4.66	0.10	0.94			0.14	99.70
2.19	4.67	0.08	0.35		0.53	0.14	100.72
2.52	4.69	0.07	1.09			0.14	99.53
2.58	4.57	0.09	0.37		0.57	0.30	99.47
2.84	4.59	0.05	0.60		0.36	0.09	100.06
2.60	3.77	0.08	0.61		0.59	0.24	98.83
2.32	4.63	0.09	1.00			0.13	99.68
3.36	4.48	0.09	0.55		0.82	0.06	100.83
2.69	4.42	0.03	0.64			0.08	99.55
2.69	5.04	0.04	0.94			0.05	99.45
2.63	4.46	0.08	0.45		0.70	0.20	99.75
2.60	4.92	0.05	0.74			0.07	100.09
2.37	4.46	0.07	0.85	0.06			99.72
2.26	4.62	0.05	0.15		1.14	0.21	99.51
2.68	4.39	0.08	0.36		0.74	0.15	100.78
2.44	4.74	0.06	0.40		0.70	0.04	99.84

Institutul Geologic al României

Table 1

Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO
								GRANODIO
1898	71.14	0.22	15.39	1.06	0.48	0.03	0.48	1.83
1834	71.16	0.20	16.75	0.06	0.38		0.42	2.12
1844	71.18	0.29	16.51	0.20	0.71		0.65	2.37
4	71.21	0.18	15.24	1.52	0.09	0.03	0.60	2.23
275	71.26	0.19	15.29	1.52	0.15	0.04	0.55	2.42
1833	71.33	0.23	17.54	0.16	0.57	0.04	0.37	2.03
1863	71.36	0.36	16.29	0.98	0.37	0.05	0.43	1.63
1858	71.38	0.82	13.97	0.98	0.70	0.05	0.57	2.58
4494	71.39	0.32	15.60	1.06		0.02	0.50	1.82
4740	71.40	0.25	15.70	0.87	0.35	0.03	0.50	2.15
272	71.40	0.22	15.24	1.44	0.15	0.05	0.71	2.07
4782	71.52	0.22	15.80	1.03	0.30	0.03	0.49	1.95
273	71.54	0.16	15.30	1.04	0.60	0.05	0.47	2.39
1846	71.56	0.29	17.52		0.35	0.03	0.37	2.02
1832	71.60	0.23	17.54		0.54	0.04	0.43	2.22
4739	71.63	0.31	15.20	0.73	0.70	0.03	0.49	2.18
1880	71.68	0.48	14.69	1.15	0.36	0.03	0.63	2.31
6	71.69	0.16	15.54	1.35	0.04	0.03	0.42	2.23
1845	71.72	0.23	17.23	0.12	0.42		0.51	1.82
24	71.74	0.14	15.72	0.90	0.44	0.03	0.49	1.77
1	71.89	0.16	15.34	1.03	0.30	0.04	0.46	2.33
274	71.93	0.19	14.84	1.28	0.27	0.03	0.54	2.20
1881	71.98	0.46	15.69	1.05	0.34	0.04	0.54	1.58
1884	72.94	0.18	14.89	0.53	0.44	0.05	0.49	1.58
								GRANI
1854	70.20	0.32	15.57	1.02	0.56	0.04	0.44	2.63
168	71.20	0.22	15.06	1.36	0.30	0.05	0.46	1.63
1835	71.84	0.11	16.75	0.24	0.50	0.04	0.38	1.72
1878	71.86	0.26	15.59	1.01	0.59	0.04	0.53	2.44
4555	72.05	0.30	17.70	0.32	0.50	0.04	0.22	2.56
1843	72.21	0.13	15.83		0.71	0.03	0.30	2.13
4678	73.12	0.28	15.44	0.73	0.41		0.38	1.80
4686	73.35	0.01	15.40	0.62	0.25	0.02	0.33	1.52
1861	73.82	0.60	15.46	0.42	0.33	0.05	0.16	1.25
4738	74.60		14.50	0.88	0.20	0.16	0.33	0.71
1857	74.95	0.61	12.04	0.70	0.18	0.03	0.23	1.86
1872	76.70	0.18	12.79	0.32	0.16	0.07	0.09	1.02

Analysts: A. Medeşan, V. Neacşu, E. Călinescu, D. Necşulescu

Table 2 – Range of variations and average values of the oxides in the

	Diorites n=2			Quartzdiorites n=8			Tonalites n=30		
	X-min	X-max	X	X-min	X-max	X	X-min	X-max	X
SiO_2	53.50	58.11	55.80	60.60	66.20	62.50	66.05	71.15	68.96
TiO_2	0.75	0.92	0.84	0.13	0.90	0.58	0.13	0.64	0.31
Al_2O_3	17.01	17.31	17.11	16.79	22.14	18.27	15.40	18.65	16.76
Fe_2O_3	3.17	3.22	3.19	0.04	3.73	2.77	0.04	4.67	1.55
FeO	3.01	4.71	3.86	0.38	2.17	1.32	0.32	1.32	0.67
MnO	0.12	0.13	0.12	0.01	0.11	0.07	0.01	0.07	0.04
MgO	2.74	5.34	4.04	0.85	1.88	1.45	0.35	1.43	0.67
CaO	5.55	6.64	6.09	2.39	5.16	4.29	1.20	3.67	2.77
K_2O	1.15	1.44	1.29	1.62	3.44	2.57	0.93	4.80	2.44
Na_2O	3.70	4.31	4.00	3.94	4.88	4.34	1.75	5.50	4.45
P_2O_5	0.21	0.30	0.25	0.11	0.40	0.24	0.06	0.17	0.09
H_2O	0.74	0.88	0.81	0.36	1.83	1.11	0.28	1.75	0.88

Institutul Geologic al României

continued

K_2O	Na_2O	P_2O_5	H_2O^+	H_2O^-	CO_2	S	Total
RITES							
2.98	4.55	0.06	1.08			0.15	99.45
2.91	5.00	0.06	0.65	0.15			99.56
2.25	4.66	0.09	0.76	0.13			99.54
2.64	4.62	0.10	0.84			0.29	99.59
2.31	4.72	0.07	1.15			0.21	99.92
2.50	4.52	0.06	0.40	0.11	0.02		99.66
2.83	4.98	0.05	0.55		0.42	0.06	100.36
2.88	4.58	0.09	0.68		0.18	0.06	99.52
3.20	4.52	0.08	0.66			0.05	99.22
2.85	4.93	0.03	0.51			0.08	99.65
2.23	4.70	0.07	0.96			0.28	99.52
2.71	4.88	0.04	0.68			0.07	99.72
2.75	4.46	0.09	0.56			0.21	99.62
2.70	4.44	0.06	0.42	0.07			99.69
2.50	4.16	0.07	0.37	0.16	0.11		99.65
2.67	4.90		0.82			0.08	99.74
2.45	4.51	0.07	0.74		0.34	0.09	99.53
2.51	4.75	0.07	0.79			0.16	99.74
2.37	4.74	0.05	0.51	0.07	0.09		99.74
2.69	4.63	0.06	0.23		0.68	0.19	99.71
2.80	4.50	0.07	0.71			0.19	99.82
2.53	4.62	0.07	0.82			0.24	99.56
3.12	4.64	0.05	0.57			0.04	100.10
3.12	3.99	0.01	1.10				99.32
TES							
3.43	4.34	0.07	0.48		0.88	0.07	100.05
3.78	4.21	0.06	1.35			0.09	99.77
3.12	4.44	0.06	0.55	0.11	0.11		99.75
3.10	4.23	0.05	0.62		0.24	0.04	100.60
3.47	4.40	0.05	0.32			0.05	101.98
4.10	3.92	0.05	0.36	0.11	0.02		99.68
3.02	4.38	0.04	0.21			0.10	99.91
3.35	4.60		0.46			0.07	99.98
5.22	3.12	0.03	0.13		0.09	0.11	100.79
3.81	4.40		0.21			0.06	99.86
3.68	4.06	0.03	0.82		0.15	0.11	99.45
2.98	4.84	0.03	0.17		0.11	0.06	99.52

granitoid rocks from the Retezat Pluton

Granodiorites n=105			Granites n=12		
X-min	X-max	X	X-min	X-max	X
66.95	72.92	70.23	70.20	76.70	72.99
0.04	0.90	0.29	0.01	0.61	0.27
12.70	19.27	16.25	12.04	17.70	15.18
0.06	4.40	1.02	0.24	1.36	0.69
0.04	1.37	0.55	0.16	0.71	0.39
0.02	0.24	0.05	0.02	0.16	0.05
0.10	2.05	0.59	0.09	0.53	0.32
0.78	3.92	2.39	0.71	2.63	1.77
1.37	3.60	2.52	2.98	5.22	3.59
3.17	5.08	4.49	3.12	4.84	4.24
0.01	0.80	0.08	0.03	0.09	0.05
0.15	3.79	0.81	0.13	1.35	0.47

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Table 3 – Trace elements (ppm) in the granitoid rocks

Sample	Pb	Cu	Ga	Sn	Ni	Co	Cr	V	Sc
									DIORI
198	4.5	70.0	13.0	< 2	80.0	29.0	100.0	220	21.0
185	2.0	52.0	17.0	< 2	26.0	19.0	38.0	145	22.0
									QUARTZDIO
376	7.0	< 3.0	22.0	< 2	4.5	7.0	3.5	100	9.0
313	9.0	< 3.0	28.0	< 2	4.0	8.0	5.0	130	12.0
308	8.0	< 3.0	20.0	< 2	4.5	7.5	4.0	90	9.0
309	9.0	< 3.0	28.0	< 2	4.0	8.0	5.0	110	10.5
4683	15.0	4.0	15.5	< 2	7.0	5.5	5.5	60	5.5
1892	17.0	< 3.0	23.0	< 2	7.5	2.5	<10.0	32	2.0
									TONALI
4687	20.0	< 3.0	15.0	< 2	10.0	4.5	8.0	55	4.0
4707	19.0	< 3.0	13.0	< 2	6.0	3.5	5.0	37	3.0
1895	11.0	< 3.0	18.0	< 2	3.5	< 2.0	< 10.0	20	< 2.0
4696	17.0	3.0	13.0	< 2	8.5	3.5	5.5	46	3.0
1891	16.0	< 3.0	18.0	< 2	5.5	< 2.0	< 10.0	17	2.0
1887	17.0	4.0	16.0	< 2	7.0	< 2.0	< 10.0	20	< 2.0
4708	18.0	< 3.0	13.0	< 2	5.0	3.0	4.0	37	3.0
187	9.0	< 3.0	15.0	< 2	< 2.0	2.5	2.0	18	2.5
4690	16.0	3.0	17.0	< 2	6.0	4.0	3.5	33	2.0
4743	18.0	< 3.0	16.0	< 2	5.5	3.0	6.5	36	3.5
1882	15.0	< 3.0	18.0	< 2	7.5	< 2.0	< 10.0	24	9.5
199	9.0	< 3.0	19.0	< 2	3.5	2.5	3.0	20	3.5
4689	17.0	3.0	17.0	< 2	5.0	2.5	5.0	36	2.5
4616	13.0	4.5	16.0	< 2	3.0	< 2.0	3.0	6	< 2.0
276	9.0	5.5	25.0	< 2	< 2.0	< 2.0	2.0	17	3.0
4735	15.0	< 3.0	16.0	< 2	4.0	2.0	3.5	18	2.0
									GRANODIO
4692	19.0	< 3.0	13.0	< 2	6.5	3.5	5.5	37	3.0
4709	19.0	< 3.0	15.0	< 2	6.0	3.0	5.0	40	3.0
1886	19.0	< 3.0	21.0	< 2	5.5	< 2.0	< 10.0	17	< 2.0
1897	16.0	< 3.0	18.0	< 2	6.0	< 2.0	< 10.0	18	2.0
4744	19.0	< 3.0	14.0	< 2	5.0	3.5	7.0	38	3.0
1890	13.0	< 3.0	17.0	< 2	3.0	< 2.0	< 10.0	17	< 2.0
4554	13.0	4.0	15.0	< 2	4.0	< 2.0	2.0	12	< 2.0
4614	13.0	4.0	15.0	< 2	3.0	< 2.0	2.5	13	< 2.0
4613	10.0	3.0	16.0	< 2	< 2.0	< 2.0	2.0	10	< 2.0
1893	18.0	< 3.0	18.0	< 2	8.5	< 2.0	< 10.0	17	2.0
4705	15.0	4.0	11.0	< 2	6.0	3.0	5.0	36	3.0
1894	20.0	< 3.0	23.0	< 2	7.0	< 2.0	< 10.0	15	< 2.0
1883	16.0	< 3.0	16.0	< 2	6.0	< 2.0	< 10.0	23	3.5
1896	14.0	< 3.0	17.0	< 2	7.0	2.0	< 10.0	28	2.0
1888	17.0	< 3.0	18.0	< 2	9.0	< 2.0	< 10.0	15	< 2.0
4552	17.0	18.0	14.0	< 2	4.0	< 2.0	2.5	12	< 2.0
4728	20.0	4.0	16.0	< 2	4.0	2.0	3.5	25	2.0
18	11.0	< 3.0	19.0	< 2	5.0	< 2.0	3.0	10	< 2.0
4553	15.0	5.0	13.0	< 2	4.0	< 2.0	3.0	14	< 2.0
19	7.5	< 3.0	16.0	< 2	< 2.0	< 2.0	< 1.0	8	< 2.0
4759	8.5	3.0	16.0	< 2	4.0	2.0	4.0	18	< 2.0
1889	20.0	3.5	23.0	< 2	10.0	< 2.0	< 10.0	12	< 2.0
4727	15.0	< 3.0	14.0	< 2	5.0	2.0	3.5	20	2.0
4697	18.0	< 3.0	12.0	< 2	5.0	2.5	4.5	18	< 2.0
4699	16.0	< 3.0	13.0	< 2	4.5	2.5	4.0	22	2.0

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from the Retezat Pluton

Y	Yb	La	Zr	Nb	Ba	Sr	Be	Li
TES								
15	1.3	< 30	90	< 10	250	420	< 1	–
30	2.4	32	130	< 10	420	650	1.2	–
RITES								
24	1.9	55	210	< 10	1600	750	2.0	–
29	2.8	65	210	< 10	2300	> 1000	1.8	–
24	2.4	62	205	< 10	1900	1000	1.8	–
24	2.0	120	360	< 10	1500	> 1000	1.8	–
11	< 1.0	48	190	10	950	530	2.1.	72
< 10	< 1.0	< 30	110	13	650	750	3.2	–
TES								
< 10	< 1.0	< 37	180	< 10	1000	600	1.8	38
< 10	< 1.0	< 30	150	< 10	850	650	1.7	45
< 10	< 1.0	< 30	125	13	1000	750	4.0	–
< 10	< 1.0	30	160	< 10	650	420	2.1	60
< 10	< 1.0	< 30	125	12	900	650	3.4	–
< 10	< 1.0	< 30	145	16	750	500	2.2	–
< 10	< 1.0	< 30	145	< 10	1100	550	2.5	50
< 10	< 1.0	< 30	130	< 10	1200	570	2.2	–
< 10	< 1.0	< 32	135	< 10	1100	600	1.8	34
< 10	< 1.0	< 30	115	< 10	1300	620	1.4	34
< 10	< 1.0	< 30	125	< 10	750	400	1.6	–
10	< 1.0	30	190	< 10	900	340	2.0	–
< 10	< 1.0	< 30	150	< 10	900	440	2.0	37
< 10	< 1.0	30	70	< 10	1100	750	1.6	20
19	1.6	< 30	150	< 10	750	750	1.8	–
< 10	< 1.0	< 30	135	< 10	1150	620	1.8	19
RITES								
< 10	< 1.0	32	115	12	920	400	2.2	32
< 10	< 1.0	< 30	140	< 10	900	530	1.8	45
< 10	< 1.0	< 30	110	16	900	650	4.4	–
< 10	< 1.0	< 30	100	12	850	540	4.8	–
< 10	< 1.0	< 30	125	< 10	1800	550	1.2	28
< 10	< 1.0	< 30	120	10	1000	600	3.2	–
< 10	< 1.0	< 30	62	< 10	1400	700	1.8	25
< 10	< 1.0	< 30	50	< 10	1200	500	1.7	34
< 10	< 1.0	< 30	55	< 10	1100	440	1.5	28
< 10	< 1.0	< 30	100	12	800	600	3.8	–
< 10	< 1.0	< 30	145	10	900	520	1.8	29
< 10	< 1.0	< 30	100	13	900	640	3.8	–
< 10	< 1.0	< 30	90	16	1300	500	2.1	–
< 10	< 1.0	< 30	150	12	850	500	3.5	–
< 10	< 1.0	< 30	100	13	750	450	3.0	–
< 10	< 1.0	< 30	70	< 10	1400	900	1.5	25
< 10	< 1.0	< 30	115	< 10	850	460	1.5	19
< 10	< 1.0	< 30	150	< 10	1050	650	1.4	–
< 10	< 1.0	< 30	48	< 10	1500	750	1.8	25
< 10	< 1.0	< 30	95	< 10	1050	480	1.6	–
< 10	< 1.0	< 30	120	< 10	1350	400	1.4	20
< 10	< 1.0	< 30	100	14	750	550	3.4	–
< 10	< 1.0	< 30	125	< 10	1100	500	1.5	35
< 10	< 1.0	< 30	125	< 10	900	500	1.9	40
< 10	< 1.0	< 30	110	< 10	950	380	1.7	46

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Table 3 – continued

Sample	Pb	Cu	Ga	Sn	Ni	Co	Cr	V	Sc
5	11.0	5.0	14.0	< 2	< 2.0	2.0	2.5	14	2.0
3	14.0	6.5	18.0	< 2	< 2.0	2.0	2.5	15	2.0
2	15.0	24.0	14.0	< 2	2.0	2.5	3.5	22	2.5
4747	9.5	< 3.0	13.0	< 2	5.5	2.5	4.0	33	3.0
4612	13.0	6.0	19.0	< 2	2.5	< 2.0	2.5	8	< 2.0
97	8.5	< 3.0	16.0	< 2	< 2.0	< 2.0	2.5	12	< 2
4742	15.0	< 3.0	17.0	< 2	4.0	2.0	2.5	18	2.0
96	6.5	< 3.0	17.0	< 2	< 2.0	2.0	2.0	10	2.0
1898	12.0	< 3.0	19.0	< 2	7.5	< 2.0	< 10.0	14	< 2.0
4	8.5	< 3.0	15.0	< 2	< 2.0	3.0	2.0	12	< 2.0
275	9.0	< 3.0	19.0	< 2	< 2.0	< 2.0	2.0	12	< 2.0
4494	16.0	3.0	17.0	< 2	3.5	< 2.0	3.0	7.5	< 2.0
4740	17.0	< 3.0	12.0	< 2	3.5	2.0	3.0	18	2.0
272	5.5	< 3.0	18.0	< 2	< 2.0	< 2.0	1.5	20	< 2.0
4782	15.0	< 3.0	16.0	< 2	3.5	< 2.0	2.5	16	2.0
273	11.0	< 3.0	15.0	< 2	< 2.0	2.0	2.0	14	2.0
4739	13.0	< 3.0	16.0	< 2	3.0	2.0	2.5	15	< 2.0
6	10.0	< 3.0	11.0	< 2	< 2.0	< 2.0	2.0	12	< 2.0
24	10.0	< 3.0	14.0	< 2	< 2.0	2.0	1.5	12	< 2.0
1	24.0	65.0	15.0	< 2	3.0	< 2.0	2.0	11	2.5
274	7.0	< 3.0	15.0	< 2	< 2.0	< 2.0	1.5	9	< 2.0
									GRANI
4555	15	4.0	14.0	< 2	4.0	< 2.0	3.0	5.5	< 2.0
4678	19	< 3.0	16.0	< 2	4.5	2.0	3,5	15	2.0
4686	20	< 3.0	13.0	< 2	4.0	< 2.0	3.0	9.5	2.0
4738	29	< 3.0	26.0	< 2	4.0	< 2.0	2.5	5	< 2.0

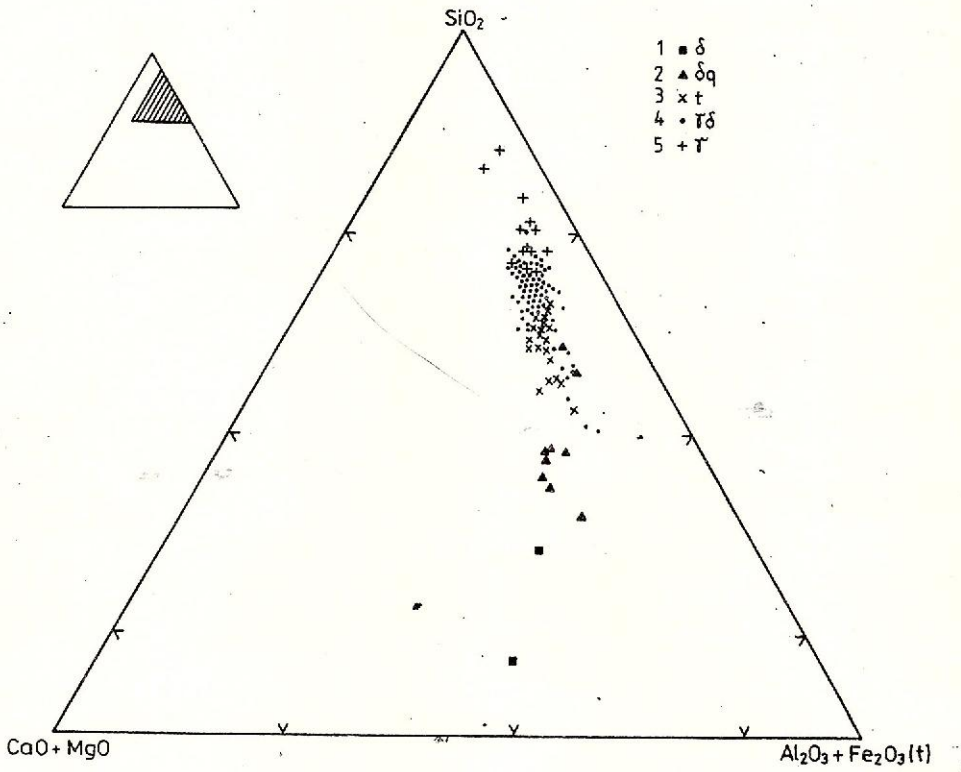

Fig. 3 - Korjinski diagram. 1, diorites; 2, quartzdiorites; 3, tonalites; 4, granodiorites; 5, granites.

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Y	Yb	La	Zr	Nb	Ba	Sr	Be	Li
< 10	< 1.0	< 30	150	< 10	1600	> 1000	1.2	–
< 10	< 1.0	< 30	120	< 10	1350	700	1.4	–
< 10	< 1.0	< 30	130	< 10	1400	850	1.8	–
< 10	< 1.0	< 30	150	< 10	1000	500	1.6	19
< 10	< 1.0	< 30	62	< 10	1400	700	1.7	25
< 10	< 1.0	< 30	135	< 10	1050	650	1.2	–
< 10	< 1.0	< 30	135	< 10	1300	800	1.3	28
< 10	< 1.0	< 30	160	< 10	1300	750	1.6	–
< 10	< 1.0	< 30	100	14	1000	530	4.4	–
< 10	< 1.0	< 30	125	< 10	1250	670	1.4	–
< 10	< 1.0	< 30	185	< 10	1400	1000	1.2	–
< 10	< 1.0	< 30	60	< 10	900	550	2.0	20
< 10	< 1.0	< 30	120	< 10	1200	480	1.1	30
< 10	< 1.0	30	210	< 10	1200	750	1.4	–
< 10	< 1.0	30	140	< 10	1300	700	1.8	26
< 10	< 1.0	< 30	150	< 10	1150	500	1.2	–
< 10	< 1.0	< 30	125	< 10	1250	620	1.2	38
< 10	< 1.0	< 30	170	< 10	1700	1000	1.2	–
< 10	< 1.0	< 30	95	< 10	1350	750	1.6	–
< 10	< 1.0	< 30	92	< 10	1300	850	1.8	–
< 10	< 1.0	< 30	145	< 10	1700	900	1.4	–
TES								
< 10	< 1.0	< 30	40	< 10	1300	330	2.2	16
< 10	< 1.0	< 30	90	< 10	1000	440	2.0	45
< 10	< 1.0	< 30	65	< 10	1000	360	1.6	20
36	3.2	< 30	32	19	38	56	3.5	24

Table 4 – Average values of the trace elements in the granitoid rocks from the Retezat Pluton

	Diorites n=2	Quartz diorites n=6	Tonalites n=16	Granodiorites n=46	Granites n=4
Pb	3.2	10.8	15.3	13.9	20.7
Ga	15.0	22.7	16.7	15.9	17.2
Ni	2.0	5.3	5.6	4.2	4.1
Co	24.0	6.4	2.7	2.2	2.0
Cr	69.0	5.5	5.9	4.6	3.0
Sc	21.0	8.0	3.1	2.2	2.0
V	98.0	80.3	28.0	17.2	8.7
Zr	110.0	214.0	140.0	116.0	56.7
Be	1.1	2.1	2.1	2.0	2.3
Li	–	–	37.0	29.4	26.2
Ba	335.0	1483.0	946.0	1152.0	834.0
Sr	535.0	230.0	576.0	623.0	296.0

approach, but not without recognizing its merit in stressing the leucocratic nature of the Retezat granodiorites and tonalites.

In order to check the areal variability of major elements, we made an analysis of trend surfaces. The trend maps represented in Figure 6A, 6B, 6C and 6D picture some characteristics of the areal distribution of oxides. SiO_2 and Al_2O_3 show a main tendency to increase from borders to the central part of the exposed area, but TiO_2, Fe_2O_3, CaO and MgO show a reverse trend. The variation of alkalies is less evident, but the content isolines show also areal inhomogeneity. In general, these trend maps of major elements are in acceptable agreement with the petrographic map (based on modal analysis) of the Retezat sheet (Berza et al, 1989).

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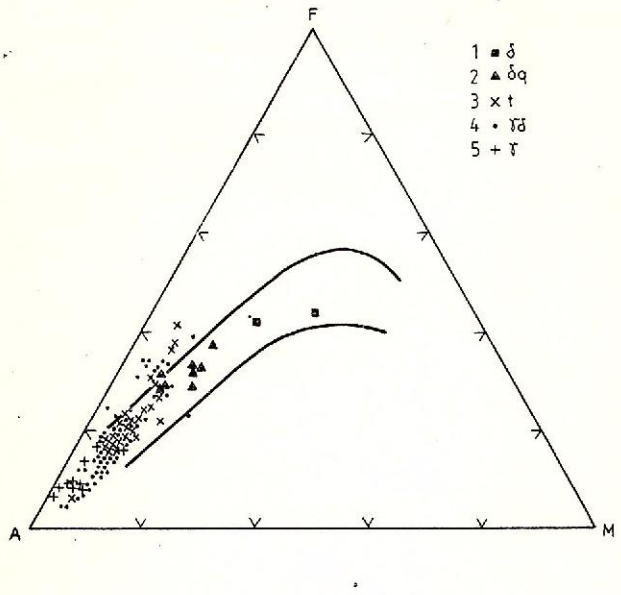

Fig. 4 - FMA diagram. See legend Fig. 3.

Trend surface analysis confirms the emplacement nature of the eastern and western borders; at the southern and northern ends, content isolines are latitudinally crossed by the projection of the overlying Alpine cover.

5. Conclusions

Our study of the Retezat granitoid Pluton has shown, in a body known for its petrographic homogeneity, several orders of variability.

Easy to notice, but exposed on the very steep right slope of the Râul Mare Valley an elongated, (2 km/200 m) precursory (?) intrusion of biotite-hornblende diorites borders westwards the granodioritic body. Its genetic relation with the main pluton is not clear, but similar situations have already been reported for Danubian granitoids.

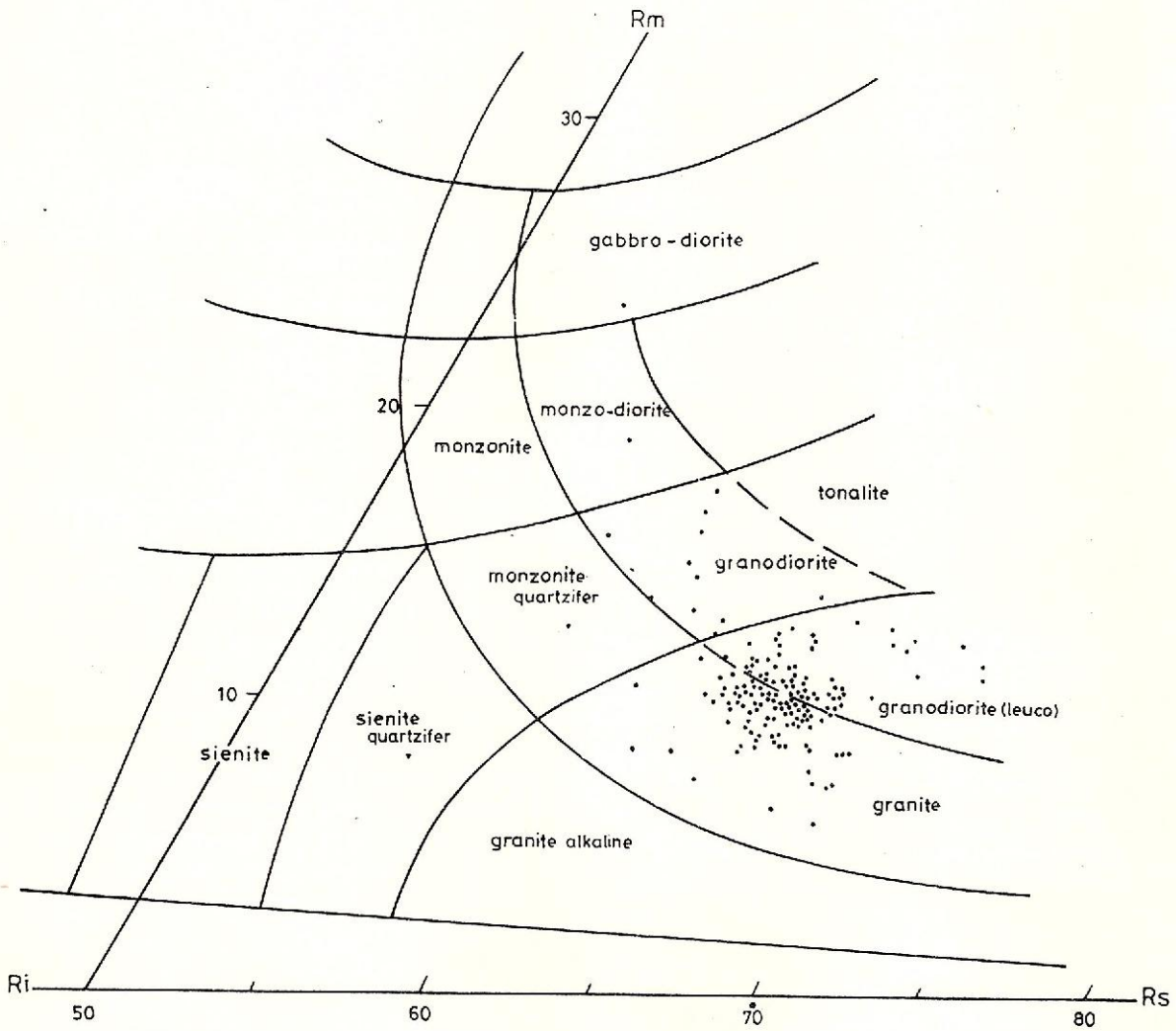

Fig. 5 - R1-R2 diagram.

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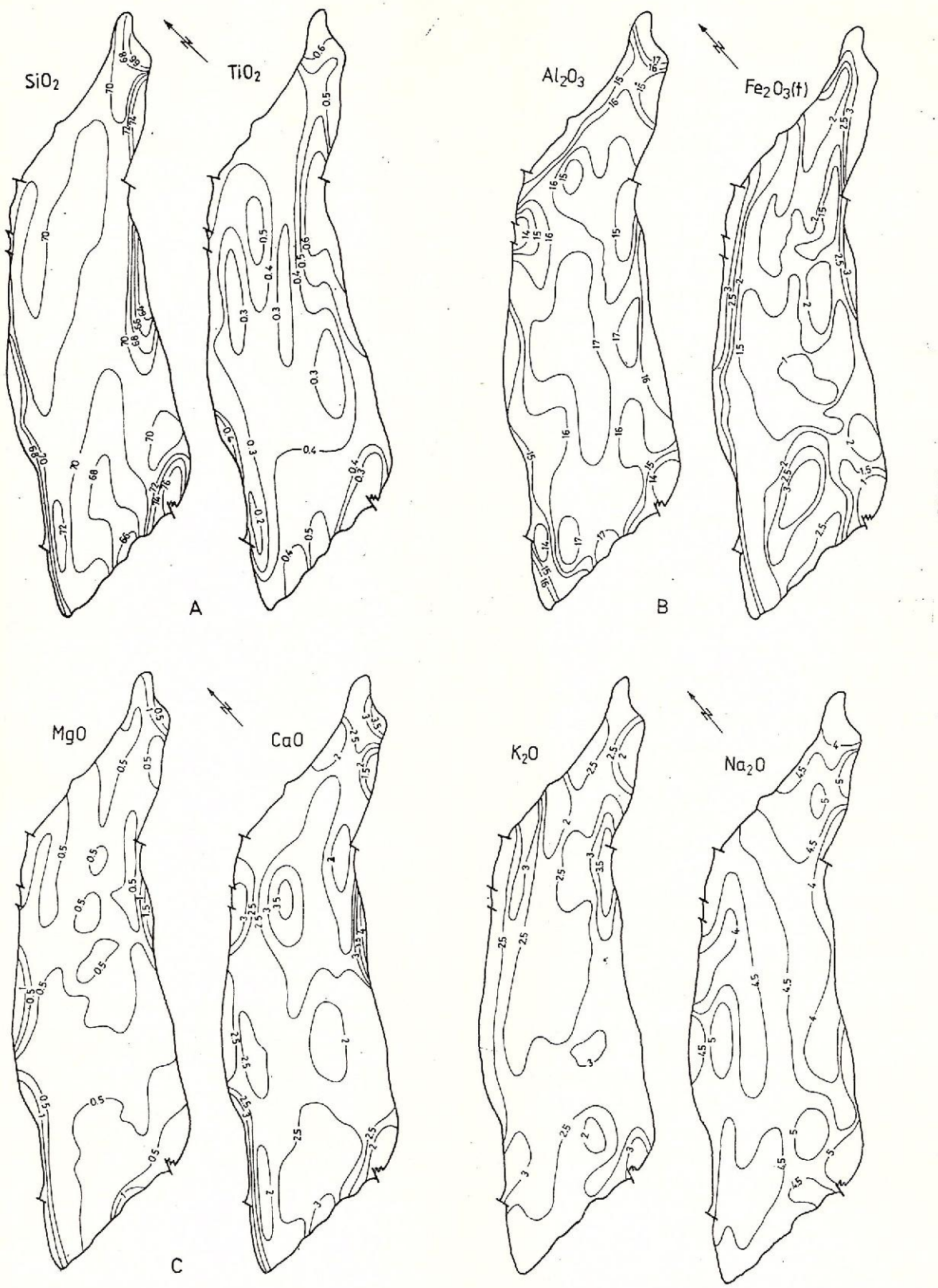

Fig. 6 - a, Areal distribution of the SiO_2 and TiO_2 contents. b, Areal distribution of the Al_2O_3 and $Fe_2O_{3(t)}$. c, Areal distribution of the MgO and CaO. d, Areal distribution of the K_2O and Na_2O contents.

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Not so striking, but still noticeable in the field, the eastern border shows a more or less continuously thin (tens of meters) strip of darker granitoids, this aspect resulting from a higher content in biotite (at the expense of muscovite), a lower content in quartz and the absence of K feldspar. This quartzdiorite border may be interpreted as resulting from a cooling effect, similar to situations previously described in the Danubian.

However, the main discovery resulting from our systematic sampling is the presence, in the granodioritic body, of kilometric areas in which the deficit or the excess of K-feldspar shifts the rocks to tonalitic, granitic respectively, compositions. The tendency of granitic rocks to develop in the southern-central part of the exposed area, along with the noticeable broadening southwards of the NE-SW elongated pluton, enable us to consider that at present only the northern part of the intrusion is exposed, the southern one running under the Mesozoic cover and the overlying nappes, in the deep underground of the Godeanu Mts.

The main characteristic of the muscovite-biotite granites, granodiorites and tonalites of the Retezat Pluton is the constant presence of 1–3 % euhedral millimetric epidote, representing indeed a specific mineral of rocks. We agree with previous researchers to consider it as a magmatic mineral phase.

Our geochemical study revealed the major and trace element limits of variability of the Retezat granitoids, confirming both the extraneous nature of the western diorites and the very gradual transitions from quartz diorites, through tonalites and granodiorites, to granites. The trend maps built for the main components in the exposed area stress the field and microscopical observations, featuring more basic borders and a more acid central zone. The main geochemical characteristic of the Retezat granitoids is a short range calc-alkali differentiation. As isotopic data are lacking, we believe that a tonalitic-granodioritic magma of unknown source (deep crust?), emplaced in special (H_2O?) pressure conditions, allowing the presence of epidote (instead of hornblende), as a primary Ca-Fe-Mg mineral, underwent a thermally controlled centripet consolidation, leading to an early consolidated basic border (quartzdiorites), a main granodioritic volume and a more acid central zone (granites), representing more acid differentiated melts.

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Institutul Geologic al României

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CULMEA CERNEI GRANITOIDS – STRUCTURAL SETTING AND PETROCHEMICAL DATA

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Institutul Geologic al României.Str. Caransebeş nr. 1, 78344 Bucureşti 32



Key words: Granitoids. Granodiorites. Monzogranites. Magmatic differentiation. Calc-alkalic composition. Majore elements. Minore elements. South Carpathians.

Abstract: The Culmea Cernei allochthonous granitoids (Danubian Realm) are spatially associated with the polymetamorphic formations of the Drăgşan Group. Temporally, they can be related to the "Assyntic" (Upper Proterozoic) or "Caledonian" (Lower Paleozoic) orogenic events. These granitoids form a composite massif constituted of granodiorites and monzogranites and, subordinately, tonalites and diorites. The geochemical data revealed a medium-potassic calc-alkaline series with a differentiation trend specific to the suites generated in a compression geotectonic medium. The Nb contents (less than 40–50 ppm) are characteristic of the island arc magmas. On the basis of the Na_2O/K_2O ratio, these rocks are assigned to I type granites. They can be regarded as M type granitoids, that is originating in the mantle, or with a mixt genesis - mantle + crust.

1. Introduction

The Culmea Cernei granitoids have been relatively less studied up till now (Iancu, 1974,1977) and no age dating has been made on them.

Due to their geological relationships with the Precambrian formations, on the one hand, and with the Paleo-Mesozoic cover formations, on the other hand, their age ranges between the Upper Precambrian ("Assyntic" granitoids) and Lower Paleozoic (granitoids associated with "Caledonian" events).

The Culmea Cernei granitoids are cropping out in the left slope of the Cerna Valley, from the confluence of the Cerna Valley with the Cărbunelui Valley to Vârful lui Stan. Southwards, they occur discontinuously from under the Mesozoic cover formations up to north of the town of Herculane.

This paper deals with the geological-structural framework of the Culmea Cernei granitoids, internal structures and relationships with the surrounding rocks. The petrochemical characterization is based on forty chemical analyses and twenty five spectral analyses (Tabs. 1 and 2). Mineralogical and petrographical studies carried out by us, partly published (Iancu,1974) have been also taken into account.

The petrochemical characterization has been made on the basis of the analytical data processing and the computation of the CIPW, Rittmann (1973) and la Roche et al.(1980) norms, in view of their plotting on diagrams useful for the study of the granitoid rocks.

The petrogenetical diagrams used by us have been elaborated by different authors on the basis of numerous statistical studies and analyses carried out on REE and isotopes. We have attempted to infer the geodynamic framework within which the study granitoids have been generated, taking into account the fields in which the analytical data available are plotting.

2. Geological-structural setting

The features of the Culmea Cernei granitoids and their geological setting are rendered on Plate.

2.1. The granitoid body displays a strongly elongated shape (ca 20 km), its outcropping area being of maximum 3–4 km wide. Eastwards it is delimited by the Precambrian metamorphics of the Drăgşan Group, which at the time of intrusion represented the host rocks, and westwards by Mesozoic formations (by means of Meso-Cainozoic tectonic contacts).

Institutul Geologic al României

The primary structures, well represented in the eastern and northern part, points to an oval-shaped batholithic body, initially enrooted, with a marked regional structural control imposed by his emplacement in a stress field induced by orogenic compression.

2.2. The regional-structural setting is based both on geological mappings and on regional data. The Culmea Cerneı granitoids enter into the constitution of the basement of the Alpine Lainici Nappe (Berza et al., 1988), belonging to the Lower (or External) Danubian Units. The tectonic division of the Danubian Realm, Laramian according to the mentioned authors, was considered polyphase by Iancu et al.(1990), as a result of repeated shearings that occurred during the Austrian and Laramian (low-angle shearings) and intra-Miocene (generally high-angle shearings) tectogeneses, the latter especially in the outermost part of the Danubian Realm.

The nappes (Supragetic, Getic, Infragetic and Danubian) generated during the Cretaceous tectogeneses are tectonically superposed and folded. During the Miocene they were thrust together over the Moesian Platform (Iancu et al., unpublished Bala and Bălvăneşti sheets, sc.1:50.000 and Iancu,Mărunţiu,1994 in this vol.). Thus, the Danubian basement, including also the Culmea Cernei granitoids, underwent repeated shearings along different discontinuity planes. The postnappe faultings of the Cerna-Jiu system fragmented the Culmea Cernei granitoids. Along the Cerna Valley they were intensely cataclased and shifted both vertically (Iancu, 1977) and horizontally (Bercia, 1975; Berza,Drăgănescu,1988).

2.3. The geological setting and the granitoids relationships with the host rocks are rendered on Plate. In direct relationships with the granitoids are formations of the Drăgşan Group, represented in the area of study by a mostly amphibolitic sequence (amphibolites and amphibolic gneisses), including a terrigene (plagiogneisses, micaschists) sequence (subformation), geometrically situated at the upper part, in a suite of narrow and elongated synclinal structures; the whole sequence belongs to a pre-Mesozoic anticlinal regional structure, with the Culmea Cernei granitoid body in the hinge (Iancu,1977).

A pre-Alpine tectonic contact, marked by an epimetamorphic blastomylonitic zone, cropping out on a width of about 1–1.5 km, makes the transition to formations belonging to the Lainici-Păiuş Group, represented by quartzous metapelitopsammitic and carbonatic rocks, with intercalations of amphibolic or graphitous rocks. Two subformations are outlined, the carbonatic and amphibolitic rocks cropping out in synclines; they are situated in an upper lithostratigraphic position as against the quartz-feldspathic and micaceous rocks. On a regional scale, the Drăgşan and Lainici-Păiuş "Series" have been divided by Berza et al.(1988) into several subdivisions regarded as formations. According to the mentioned authors the whole sequence belonging to the Lainici-Păiuş "Series" in the study area could pertain to a lower, "carbonate-graphitous" formation. The terrigene megasequence of the Lainici-Păiuş Group is spatially associated with: synkinematic and latekinematic Precambrian granitoids (Savu et al., 1978; Berza,1978), which do not crop out in the area presented on Plate; Arsasca diorites (blastomylonitic diorites cf. Iancu, 1977); numerous pre-Silurian veins and dykes of microdiorites and microgranodiorites (± porphyric) (Berza, Seghedi, 1975 a).

According to Iancu et al.(1990), Iancu,Mărunţiu (1990) the two "groups" – Drăgşan and Lainici-Păiuş – represent contrasting lithofacial associations coming from areas with a different geotectonic position in the Middle-Upper Proterozoic. The Drăgşan Group (like the Ielova and Măru crystalline massifs) contains mostly amphibolitic sequences and ocean crust-type rocks, whereas the Lainici- Păiuş Group (like the Neamţu Series) represents a continental crust-type association.

The primary metamorphism of the two units took place under different baric conditions: medium pressure - Drăgşan Group (Iancu, 1974; Berza, Seghedi, 1975 b); low pressure – Lainici-Păiuş Group (Savu, 1970).

In our opinion both megasequences display structural and mineralogical proofs in favour of a polydeformational evolution under thermal conditions specific to the amphibolitic facies, followed by dynamothermal and dynamic reactivations at the greenschist facies level.

The common evolution of the two "groups" of formations after the emplacement of the Culmea Cernei granitoids and of the intermediary rock veins, is proved by: the "transition" of the two groups, marked by chlorite, epidote-albite, stilpnomelane bearing blastomylonites (Iancu, 1977), which in fact points to a pre-Alpine tectonic contact; the retrograde dynamic adaptation to the greenschist facies level both of the polymetamorphic rocks of the Drăgşan and Lainici-Păiuş groups and of the associated granitoids; the metamorphism of the more recent vein rocks which also took place at the greenschists facies level.

2.4. Relationships of the granitoids with the host rocks prove their allochthonous features, on the one hand, and the existence of a regional structural control imposed by a subsequent deformational event of the host rocks, already metamorphosed at the level of the amphibolitic facies, on the other hand.

We shall present briefly field evidence which is in favour of the allochthonous character, post-

Institutul Geologic al României

156529

metamorphic (post M_1, M_2) but synchronous with a more recent "orogenic" (D_3) event, of the Culmea Cernei granitoids:

– existence of the sharp contacts of the granites with the rocks of the Drăgşan Group and the secant character of the aplitic and pegmatitic veins in comparison with the structures of the surrounding metamorphites;

– high frequency of the xenoliths and septa of the metamorphic rocks as well as of the agmatitic migmatites with angular fragments of mesometamorphic rocks, and scarcity of xenoliths of basic magmatites and mafic schlierens;

– existence of a spectacular migmatic aureole, on an outcropping area wide 2-3 km, including mixed rocks (granitic, leucogranitic leucosome and paleosome) with structures either pseudoconformable with the metamorphic foliations ("lit-par-lit"), or unconformable (secant) with them;

– presence of the effects of a thermal event (M_3) superposed on the M_1,M_2 structures and parageneses of the Drăgşan Group, an event which can hardly be detectable due to the isofacial character of the thermal contact parageneses which contain: sillimanite, muscovite, garnet, chloritoid (in micaschists and paragneisses);

– regional structural control illustrated by:intrusion of the granitic body on a crustal discontinuity enscribed in a megaanticlinorium structure; pseudoconformity of migmatic neosome in respect with the S_2 regional foliation; occurrence of some granitic apophyses in the axial zones of B_3 mesoscopic folds.

2.5. Structure of the Culmea Cernei granitoid body refers both to the geometry of the whole body, represented on the map (Plate), and to the internal structure. The latter, although complex, can be hardly emphasized due to the low degree of outcropping and the reduced degree of accessibility of the zone. At the massif scale, one can observe an asymmetrical structure with the eastern border well-preserved (the western part is not observable, due to tectonic accidents) and a terminal northern zone, narrower, with numerous apophyses, arteritic and agmatitic migmatites. The metamorphic septa and enclaves are frequently found in the eastern and northern part; they are both banded amphibolites and slightly hornfelsed plagiogneisses.

The massif is mostly characterized by massive structures, granodiorites clearly prevailing. Towards the marginal zones it becomes tonalitic, with mafic schlierens and xenoliths of consanguineous magmatites, usually more basic (amphibolic diorites, rarely with pyroxenes). Also in the marginal zones (eastern and northern), microgranular up to aplitic facies occur, as a result of the magma cooling processes.

In the "central" zone (left side of the Cerna Valley), ovoidal "nuclei" are to be found, evident in the field by granitic rocks (monzogranites, acc. to Rittmann norm) bearing biotite and poikilitic (pink) potash feldspar (up to 2–3 cm). Monzogranites, although do not present visible contacts with the granodioritic rocks, seem to represent subsequent intrusions within the granodioritic calc-alkaline series (Fig. 10). They have a correspondent also in some veins or apophyses crossing the granodiorites (which locally display magmatic flow texture).

The structural aspects and relationships between different petrographic types point out the imperceptible passing from tonalites to granodiorites and the clear contact (discontinuity) between more basic granodiorites and xenoliths, on the one hand, and granodiorites and monzogranites, on the other hand. All this suggests the existence of a composite body within which the outermost petrographic terms can also represent different stages or moments of intrusion during a long evolutive process, when a magmatic differentiation was possible by fractional crystallization, induced by the existence of thermal gradients (gradual transition from granodiorites to tonalites).

Culmea Cernei granitoids include, both inside and in the marginal zones, genetically associated veins of aplites, microgranites and microgranodiorites (± porphyric), as well as pegmatites.

Primary structures, correlatable with the emplacement of the magmas intrusion, can be rarely observed, due to the outcropping conditions and the effects of the superposed, locally repeated, deformational processes.

The massive, isotrope structures clearly predominate. In some cases planary orientations of the prismatic minerals (feldspars, amphiboles) or of the mafic minerals schlieren (biotite, amphiboles) have been noticed. These planary dispositions (flow textures) are quasivertical in the vicinity of the eastern contact, locally forming an angle with the contact plane. In the central zone, the massive, equigranular (in granodiorites) or inequigranular (in monzogranites) structures are prevailing; locally joints which are likely to be primary (of cooling), with a subhorizontal disposition, are also observed.

Primary structural aspects of interest are preserved in the northern "apex" or terminal zone where the amphibolite angular fragments (host rocks) have been turned round and shifted upwards.

The primary structures from the marginal zones, like the effects of the "structural control" of the apophyses in the migmatic aureole and the elongation (flattening) of the septa parallel with the contact as well as the planary structures of the host rocks, point out the granitoids intrusion under stress.

Secondary structures, subsequent to the magma cooling, give a "gneissic" or "blastomylonitic" aspect to the granitoid rocks, especially on the eastern margin

Institutul Geologic al României

Table
Chemical composition of the Culmea

No.	Sample	Petrographic Type	Location	SiO_2	Al_2O_3
1	94	Granite(monzo)	Cerna Valley	76.03	13.85
2	2002	Granite (monzo)	Capra Valley	74.32	14.83
3	2059	Granodiorite	Cerna Valley	73.26	15.12
4	170	Granodiorite	Cerna Valley	72.30	15.70
5	92	Granite (monzo)	Cerna Valley	71.97	18.13
6	18	Granite (monzo)	Cerna Valley	71.81	14.89
7	247	Granodiorite	Arsasca Valley	71.69	16.14
8	2391	Granodiorite	Dobrota Valley	71.68	16.19
9	82	Granite (monzo)	Seaca Valley	71.54	15.73
10	2010	Granodiorite	Paltinei Brook	71.51	16.06
11	2416	Granite (monzo)	Dobrota Valley	71.46	15.43
12	MI-1	Granodiorite	Cioaca Frasinului	71.20	17.38
13	21	Granodiorite	Cerna Valley	71.15	15.08
14	84	Granite (monzo)	Seaca Valley	70.50	15.81
15	75	Granodiorite	Cerna Valley	70.31	16.65
16	2124	Granodiorite	Arsasca Valley	70.24	16.74
17	76B	Granodiorite	Cerna Valley	69.87	16.49
18	26	Granite (monzo)	Cerna Valley	69.46	16.91
19	172	Granodiorite	Arsasca Valley	69.40	15.40
20	171	Granite (monzo)	Seaca Valley	69.39	15.90
21	2067	Granodiorite	Cerna Valley	69.39	17.23
22	2522	Granodiorite	Capra Valley	69.13	15.57
23	2467	Granodiorite	Dobrota Valley	69.07	14.91
24	2051	Granite (monzo)	Paltinei Valley	68.92	16.12
25	100	Granodiorite	Cerna Valley	68.75	16.68
26	199	Tonalite	Arsasca Valley	68.20	15.83
27	149	Granodiorite	Cerna Valley	68.17	16.50
28	2080	Granite (monzo)	Cerna Valley	67.67	16.58
29	2139	Granodiorite	Capra Valley	67.43	16.63
30	103	Granodiorite	Rîmnuţa Valley	67.10	15.00
31	147	Quartz monzodiorite	Milcanul Valley	66.51	17.20
32	2126	Granodiorite	Cerna Valley	66.00	18.24
33	3	Granodiorite	Cerna Valley	65.31	17.76
34	2022	Tonalite	Paltinei Valley	65.23	13.75
35	119	Granodiorite	Cerna Valley	64.40	17.40
36	2058	Granodiorite	Cerna Valley	64.02	19.38
37	87	Granodiorite	Seaca Valley	62.85	18.13
38	8	Quartz monzodiorite	Cerna Valley	58.70	19.83
39	2127	Porphyric granodiorite	Arsasca Valley	69.50	17.20
40	198	Porphyric granodiorite	Arsasca Valley	68.80	17.08

where the dynamic metamorphism which marks the pre-Alpine contact with the rocks of the Lainici-Păiuş Group affected the granitoids, too. These superimposed structures proved to be generated by: subgranulation, ductile dynamic deformation and epimetamorphic blastesis in a shear zone. The late dynamic deformation and the retrograde mineralogical adaptation were common to the Culmea Cernei granitoids, the host rocks of the Drăgşan Group, the Arsasca diorites and the rocks of the Lainici-Păiuş Group. They also affected the pre-Silurian dyke rocks, without always distroying their primary structures. In the maximal deformed zones nearby the tectonic plane, the dynamic ultrametamorphism determined a strong convergence of characteristics of the blastomylonites formed on rocks with similar petrographic composition.

The brittle deformations (cataclases, etc.) in the western part of the massif have been obviously generated by Alpine movements.

Institutul Geologic al României

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Cernei granitoid rocks

Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	H_2O^+	H_2O^-	Total
0.97	0.30		0.43	1.27	3.41	3.15	0.10	0.03	1.00	0.24	100.63
0.27	1.05	0.02	0.20	1.53	3.41	4.03	0.12	0.12	0.47	0.20	100.37
0.17	1.42	0.03	0.41	2.36	3.91	2.20	0.28	0.04	0.50	0.17	99.70
0.80	0.90	0.03	0.64	2.62	4.30	2.10	0.39	0.05	0.55		100.38
1.05	0.94	0.02	0.43	1.78	3.07	3.90	0.20	0.08	0.97	0.11	102.54
0.97	1.01	0.03	0.41	2.04	3.59	3.15	1.16	0.12	0.87	0.16	100.05
0.79	1.13	0.02	0.65	2.29	3.23	2.45	0.21	0.14	1.01	0.31	99.75
0.60	1.15	0.01	0.30	2.25	4.22	1.53	0.02	0.09	1.14		99.65
1.03	1.05	0.02		2.55	3.61	3.53	0.22	0.16	0.56	0.17	100.00
0.65	1.26	0.15	1.05	1.80	3.88	1.98	0.25	0.08	0.83		99.50
0.70	1.12	0.01	0.66	2.46	4.19	3.21	0.27	0.08	0.62	0.09	100.21
0.90	0.72		0.35	3.46	3.28	1.47	0.17	0.07	0.45	0.15	99.45
2.64	0.80	0.05	0.63	3.01	4.57	1.52	0.25	0.07	0.80	0.20	100.57
1.35	1.12	0.02	0.19	2.03	3.38	3.34	0.24	0.25	1.23	0.21	99.46
1.28	1.23	0.03	0.63	2.29	3.64	2.55	0.25	0.14	0.75	0.19	99.75
0.59	1.61	0.01	0.93	2.75	3.74	2.52	0.28	0.11	0.87		99.59
0.90	1.62	0.05		3.05	3.91	2.05	0.26	0.22	0.73	0.19	99.15
1.78	0.87	0.03	0.42	3.06	3.60	2.83	0.24	0.18	0.96	0.15	100.34
1.13	1.33	0.05	0.91	2.50	4.37	2.56	0.59	0.08	1.19		99.51
0.57	1.47	0.05	0.67	2.23	3.77	3.77	0.50	0.08	0.73		99.13
1.11	1.36		0.83	3.90	3.42	1.30	0.35	0.09	1.09		100.07
0.62	2.08	0.04	1.06	2.62	4.58	1.89	0.40	0.14	1.26	0.13	99.39
1.26	2.38	0.01	1.32	2.88	3.37	2.33	0.59	0.24	1.15	0.06	99.51
0.64	1.70	0.02	1.31	2.62	4.06	3.34	0.35	0.05	0.55	0.11	99.68
1.40	1.56	0.04	1.31	2.80	3.64	2.36	0.27	0.22	1.06	0.23	100.09
1.84	1.55	0.05	2.42	2.52	3.85	1.45	0.48	0.07	0.85	0.35	99.11
1.57	1.87	0.02	1.32	2.81	3.52	2.74	0.40	0.24	1.69	0.24	100.85
1.11	2.34	0.02	1.08	3.41	3.75	3.83	0.46	0.20	0.47	0.23	100.92
0.91	2.34	0.04	1.52	3.17	3.75	2.71	0.51	0.12	1.27	0.24	100.40
2.42	1.55	0.05	1.12	4.40	2.55	2.47	0.54	0.08	2.80	0.20	100.08
1.97	0.41	0.02	0.85	2.10	5.25	3.23	0.64	0.19	1.13		99.50
1.41	1.83	0.01	1.41	3.30	3.54	1.57	0.52	0.16	1.09		99.08
1.96	2.28	0.04	0.62	4.59	3.64	2.27	0.44	0.27	0.10	0.13	99.28
1.44	3.18	0.06	4.61	5.51	3.44	1.89	0.33	0.02	1.16	0.33	100.62
1.61	2.36	0.05	1.97	4.05	3.51	2.55	0.45	0.31	1.15	0.27	99.81
0.93	1.86	0.02	1.16	4.20	3.65	2.64	0.47	0.13	1.46		99.92
1.80	2.56	0.02	2.23	3.57	3.91	2.17	0.57	0.31	1.36	0.28	99.48
2.09	2.84	0.06	2.64	4.50	4.25	2.83	0.62	0.30	0.39	0.20	99.05
0.83	1.61	0.01	1.04		3.44	1.20	0.33	0.12	1.46		96.74
1.10	1.70	0.05	1.37	2.45	3.52	1.80	0.42	0.08	1.15		99.52

3. Petrochemical data and genetic implications

Several diagrams have been obtained on the basis of the mentioned petrographic data and the available analytical data (Tabs. 1 and 2), which make possible significant characterizations. Thus on the variation diagram of alcalinity after Wright (1969) the main petrographic types constituting the Culmea Cernei massif fall in the field of the calc-alkaline rocks, except for two samples which plot in the field of the alkaline rocks (Fig. 1).

The QAP diagram, after Streckeisen (1967) (Fig.2) shows the plotting of most of the samples in the field of granodiorites and monzogranites, and subordinately in the field of tonalites and monzodiorites.

On the diagram AbAnOr, used by Kleeman (1965; after Husch and Moreau,1982), the samples analysed by us (Fig. 3) are obviously grouped between the curves of 700 and 750°C (under a pressure of 5 kb). The samples are clearly individualized according to the fields of acid (A) and basic (B) rocks, being focused in a unitary field of intermediary rocks which belong to a calc-alkaline series (see Fig. 1).

Institutul Geologic al României

Table 2
Trace elements (ppm) in the Culmea Cernei granitoids rocks

No.	Pb	Cu	Ga	Sn	Ni	Co	Cr	V	Sc	Y	Yb	La	Be	Zr	Nb	Ba	Sr
1	21.0	6.0	14	2.0	4.0		1.5	3.5	2.0	10	1.2	65	1.2	70	10	800	160
2	17.0	8.0	19	2.5	4.0	4.0	4.5	14.0	4.0	22	3.2	30	4.0	100	20	450	160
3	13.0	16.0	18	2.0	15.0	6.0	5.0	22.0	4.5	12	1.3	30	1.3	240	14	700	300
4	6.0	19. 0	11	2.0	3.5	2.0	2.5	17.0	2.0	10	1.0	30	1.3	48	10	900	600
5	21.0	6.0	16	2.5	3.5	5.0	6.0	21.0	4.0	16	2.2	48	1.9	150	11	700	150
6	25.0	6.0	15	2.0	4.0	5.0	6.0	23.0	4.0	15	2.1	46	2.0	130	15	600	150
7	11.0	9.0	22	2.5	6.0	5.0	9.0	30.0	5.5	13	1.6	37	2.3	180	10	780	330
8	18.0	5.5	19	2.5	5.5	7.0	9.0	38.0	6.0	14	1.2	30	2.5	150	10	520	460
9	12.5	4.5	17	2.0	5.5	7.0	9.5	42.0	5.0	11	1.5	46	2.0	145	10	470	220
10	38.0	5.0	21	3.0	4.5	6.5	8.0	28.0	6.0	14	1.8	40	2.2	265	12	600	380
11	7.0	200	14	3.0	13.0	3.0	8.0	30.0	4.0	11	1.0	38	1.8	75	10	560	260
12	13.0	5.0	15	2.0	5.0	2.5	5.0	16.0	2.5	16	1.0	40	2.4	85	10	630	260
13	14.0	11.0	23	3.5	9.0	8.5	17.0	38.0	6.0	13	1.2	46	2.0	230	10	290	420
14	12.0	14.0	22	2.5	10.0	11.0	17.0	60.0	5.0	13	1.2	36	1.6	330	10	500	500
15	16.0	25.0	21	3.0	15.0	10.0	24.0	44.0	7.0	16	1.8	45	2.0	210	15	570	300
16	13.0	17.0	21	3.0	11.0	7.0	10.0	32.0	5.0	14	1.4	38	1.9	170	10	570	310
17	15.0	40.0	20	2.5	36.0	9.0	15.0	45.0	6.5	20	1.8	40	2.3	230	10	650	230
18	19.0	26.0	19	3.0	11.5	10.5	13.0	42.0	6.0	14	1.3	50	2.1	245	10	600	450
19	17.0	13.0	19	3.0	11.0	12.0	15.0	60.0	12.0	20	2.2	30	2.2	225	10	550	350
20	7.0	6.0	20	2.0	5.0	2.0	4.0	18.0	2.0	10	1.0	30	1.7	95	10	1000	1000
21	33.0	5.5	17	2.5	11.0	10.0	20.0	42.0	5.0	13	1.2	30	1.8	85	10	550	250
22	5.5	190.0	18	3.5	280.0	37.0	550.0	92.0	22.0	20	1.7	30	1.9	155	15	440	180
23	18.0	16.0	22	3.0	10.0	11.0	23.0	55.0	11.0	26	2.7	30	2.2	250	10	650	440
24	13.0	11.0	20	2.5	5.5	11.0	9.0	55.0	10.0	19	1.7	30	1.9	210	10	600	500
25	13.0	30.0	18	3.0	10.0	14.0	23.0	76.0	17.0	36	2.9	30	2.8	315	13	700	560

for number of sample see Table 1.

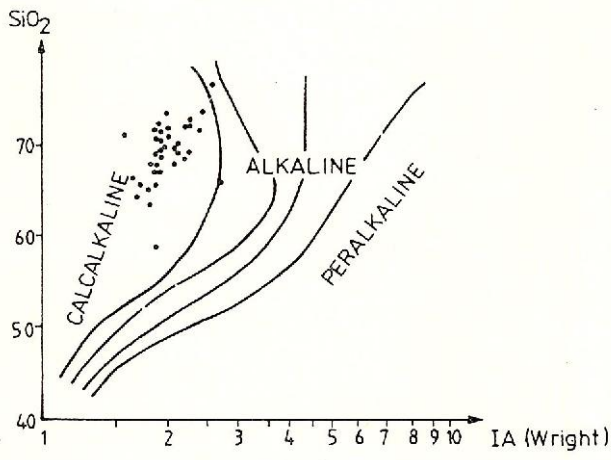

Fig. 1 – Diagram of the alkalinity variation after Wright (1969).

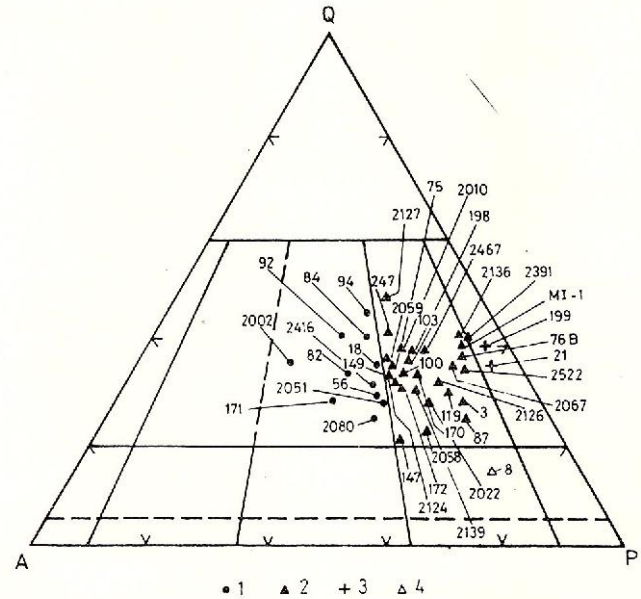

Fig. 2 – QAP diagram (Streckeisen, 1967), Rittmann norms. 1, granites; 2, granodiorites; 3, tonalites; 4, quartz monzodiorites.

The used diagrams helped us to estimate the geodynamic background of the Culmea Cernei granitoids that can be inferred from the differentiation trend of the samples plotting on a AFM diagram elaborated by Irvine and Baragar (1971). According to Petro et al. (1979), the calc-alkaline trend, disposed about perpendicular to the FM side, points to a generation medium specific to a compression geotectonic context (Fig. 4). In this context the calc-alkaline suites show different trends in relation to the alkaline ones and the oceanic tholeiites.

CNK diagram (Fig. 5), after Lameyre and Bowden (1982), displays the samples falling below 50 % K, which makes possible the characterization of the Cul-

Institutul Geologic al României

mea Cernei granitoids as medium-K granitoids within the calc-alkaline series, within which the trondhjemitic rocks are "poor-K rocks" and the monzonitic rocks are "high-K" rocks, as rendered in Figure 10. Such delimitations are clearly influenced by the position of the granitoid generating zones in relation to the Benioff plane during the subduction process. A comparison between our data and those used by Lameyre, Autran (1980) for the Variscan granitoids in France points out the plotting of the " medium-K" granitoids behind the tonalitic line (low K), therefore towards the continental area.

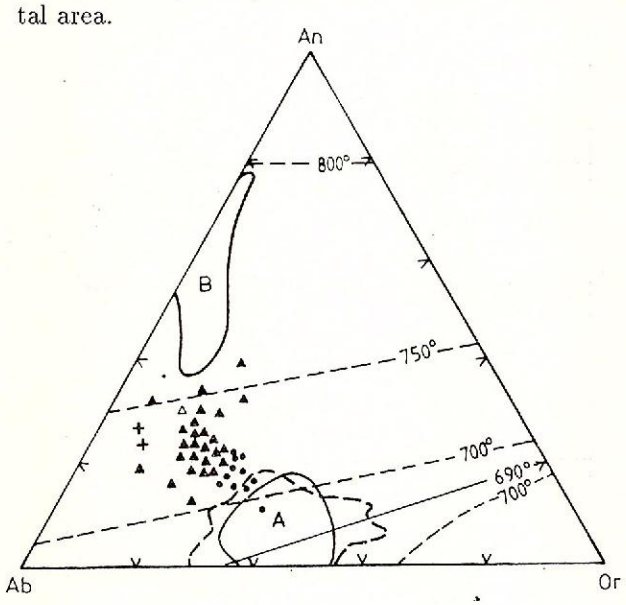

Fig. 3 – AbAnOr diagram (Kleeman, 1965) for T, P intervals: P = 5 kb; granitoids+rhyolites field (dotted line). Superposed, acid (A) and basic (B) rocks fields (after Husch, Moreau, 1982). Legend of symbols as in Fig. 2.

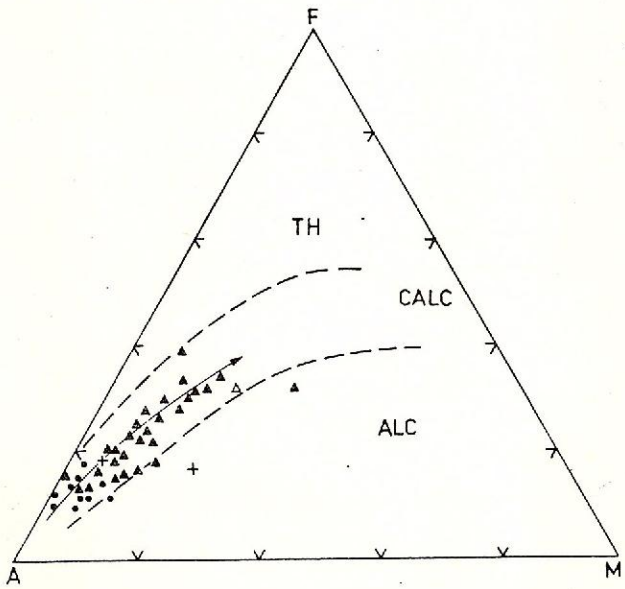

Fig. 4 – AFM diagram. Legend as in Fig. 2.

On the basis of the analyses for minor elements (Tab. 2) represented on the diagram in Figure 6, the Culmea Cernei granitoids could be characterized according to the average contents for Sr, Ba, considering the average values of the standard granitoids (Kraft, Schindler, 1961).

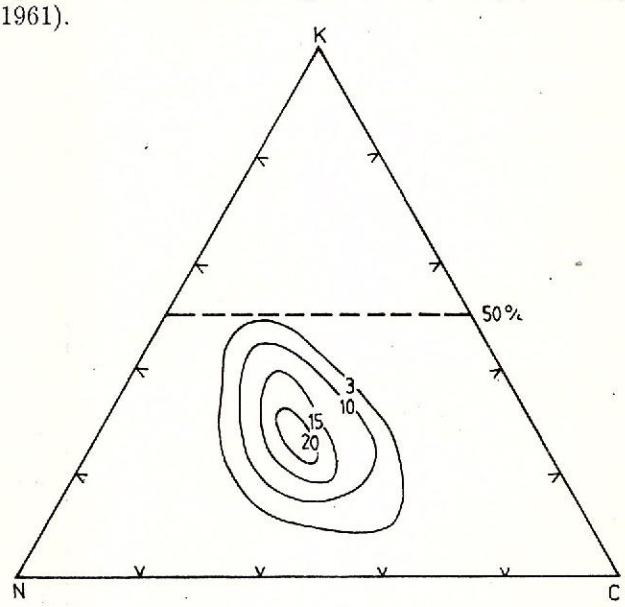

Fig. 5 – CNK diagram (frequency isolines: 3, 10, 15 and 20 %).

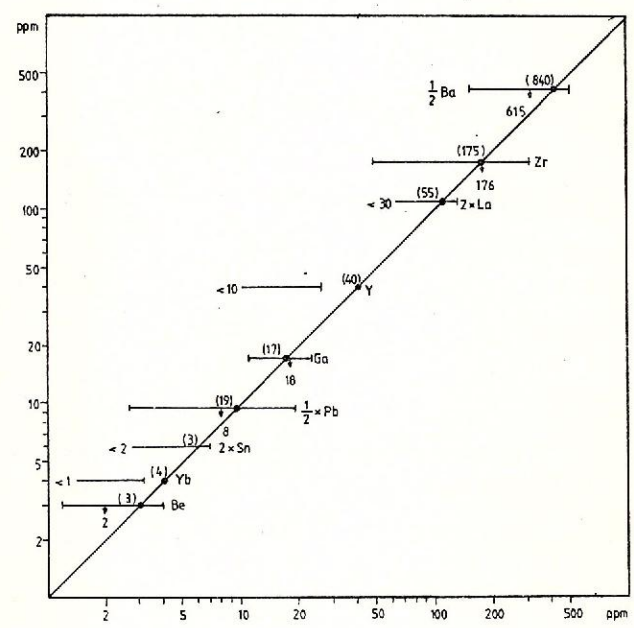

Fig. 6 – Diagram of comparison of the trace elements (ppm) contents in the granitoid rocks of the Culmea Cernei massif with the average values of the trace elements of the standard granitoid rocks (after Kraft and Schindler, 1961).

Sr and Ba variations (Fig. 7), in comparison with Ca, Na and K point to a positive Sr correlation with Ca,

Institutul Geologic al României

and of Ba with Na and K, although the variation trends are relatively less obvious.

Sc, Co, Cr, Ni and V variation diagrams are presented in Figure 8 in correlation with Mg and Fe (total); they point out an obvious positive correlation of Cr and V variation simultaneously with the variation of the mentioned major elements.

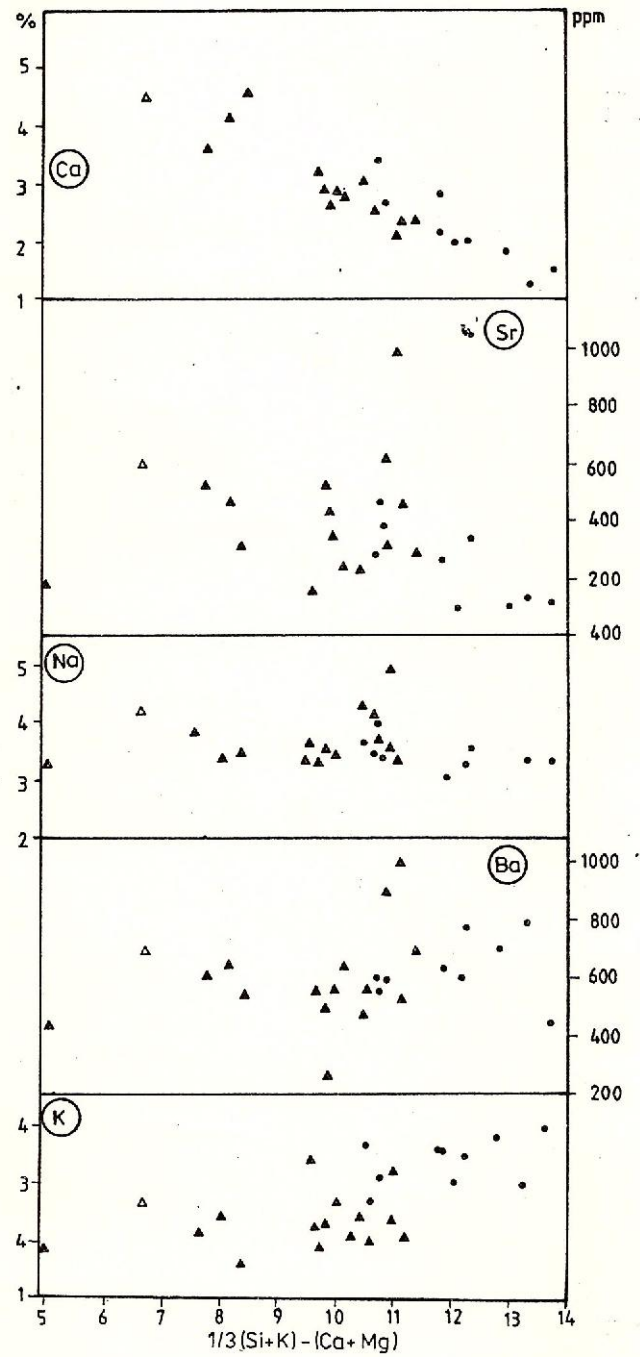

Fig. 7 Sr, Ba variation versus Nockolds Allen index.

On the Nb-SiO_2 diagram in Figure 9 (acc. to Pearce and Gale,1977) the contents of Nb (less than 40–50 ppm) of the Culmea Cernei granitoids fall in the field of volcanic arc magmas, which proves to correspond with the main petrochemical and structural data.

In our attempt to specify the genesis of the magmas which yielded the Culmea Cernei granitoids we used the Na_2O/K_2O diagram, after Chappell and White (1974), according to which one can differentiate I type magmas generated by the anatexis of the magmatic rocks, and S type magmas generated by the anatexis of the sedimentary rocks.

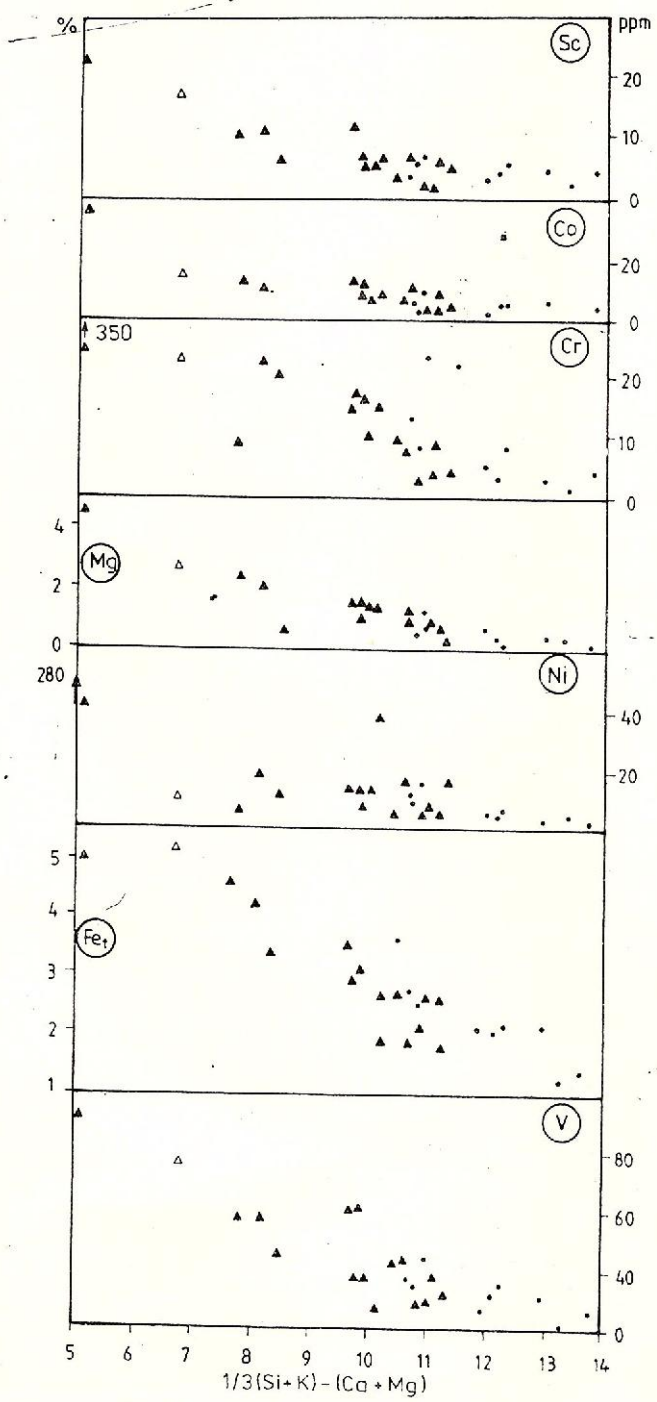

Fig. 8 Sc, Co, Cr, Ni, V variation versus Nockolds Allen index.

Institutul Geologic al României

It is of note an obvious concentration of the study samples in field "I" (Fig. 10); they represent "igneous" granitoids, called by Ishihara (1977) "magnetite series", based on mineralogical criteria.

Further arguments concerning the genesis of the primary magmas which yielded the Culmea Cernei granitoids were provided by the elaboration of Table 3 which presents for comparison granitoids of type I and S belonging to magmatic series described by Didier et al (1982) and Hussein et al (1982). A complex argumentation makes possible a more veridical genetical characterization that can replace the lack of REE and isotopes analyses.

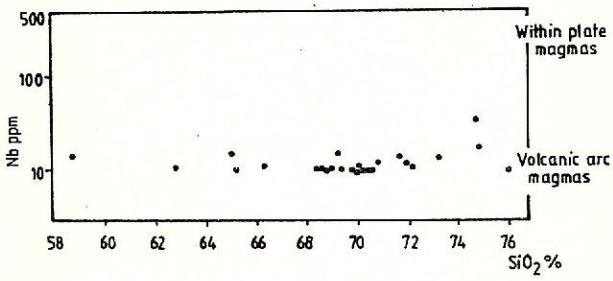

Fig. 9 – Nb/SiO_2 diagram (after Pearce, Gale, 1977).

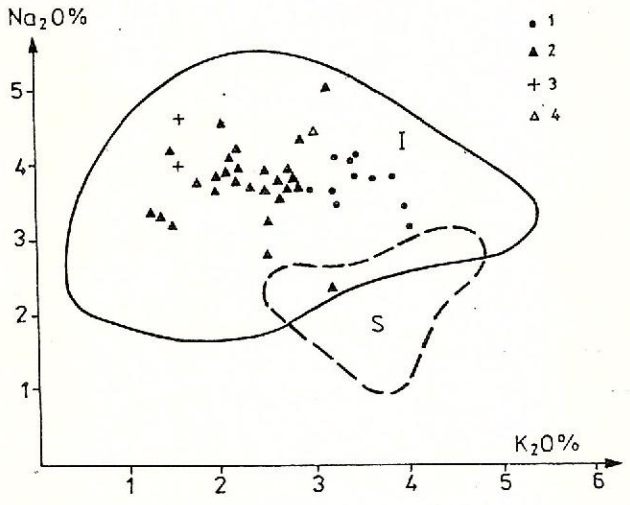

Fig. 10 – Na_2O/K_2O diagram (after Chappell, White, 1974).

According to Table 3, Culmea Cernei granitoids, considering the field relationships and petrographic prevalence (Fig. 2), are quite close to I type granites. The mineralogical criteria point out a prevalence of the magnetite normative value in the hornblende granitoids and subordinately in the biotite granitoids, as well as a relatively high colour index: 11.6. All the data are in correlation with the low content of accessory minerals and the normal values in minor elements.

In chemical respect, the Culmea Cernei granitoids also present similarities with the I type granites. As shown in Table 3, SiO_2=67.55 per cent, Na_2O=3.76 per cent, CaO=2.90 per cent (close to the value 2.97 per cent given by Chappel and White), K_2O=2.56 per cent. Furthermore, there is the uniform variation of the minor elements in relation to the differentiation index.

In more recent papers, e.g. Didier et al. (1982), a genetic classification of the granitoids is proposed into two main types: M- coming from the mantle or mixed material and C - or crustal which, in their turn, can be of two types – I and S – according to the origin of the primary material which yielded them.

Taking into account the existence of the basic magmatic enclaves, as well as the more obvious basic character (tonalites) and the prominence of the microgranular structures in the marginal zones of the body and in some xenoliths, one can ascertain that the Culmea Cernei granitoids are of type M (acc. to Didier et al., 1982), that is they originate from the mantle or crust + mantle.

Further arguments in this respect are: common presence of hornblende in granodiorites and trondhjemite-tonalites, the content of biotite less than 10 per cent, elements specific to the granodiorites from magmas originating from mantle material.

QAP diagram (Fig.11) elaborated by Lameyre and Bowden (1982) was used in view of the differentiation of the "granitoid series" (generated either by differentiation from mantle magmas or by crustal fusion). Fields 2, 3, 4 (Fig. 11) illustrate the variation trends of the calc-alkaline plutonic series, all the samples studied showing a variation trend given by "medium-K" granodiorites, within which the highest acid terms of differentiation are represented by the monzogranites from the central "nuclei".

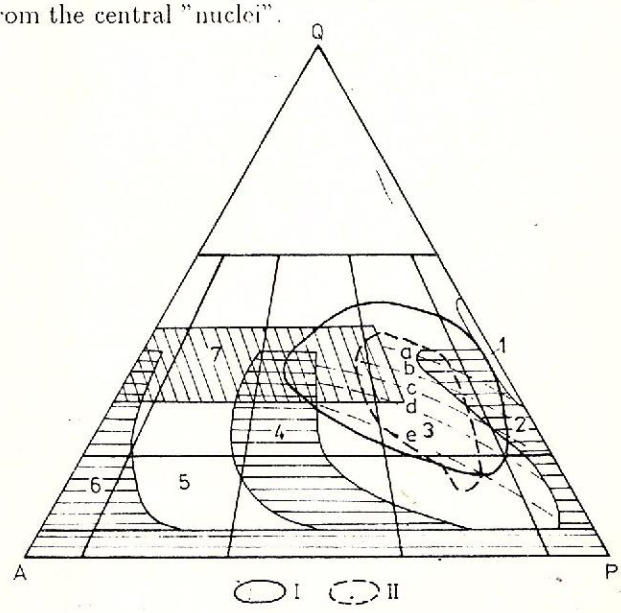

Fig. 11 – Plotting of the Culmea Cernei granitoids on the QAP diagram (after Lameyre, Bowden, 1982).

Institutul Geologic al României

Table 3
Normative classification of the Culmea Cernei granitoids

No.	Sample no.	Classification		
		QAP(CIPW)	H. de la Roche (1984)	Rittmann (1973)
1	94	granodiorite	granite	monzogranite
2	2002	monzogranite	granite	monzogranite
3	2o59	granodiorite	granodiorite	granodiorite
4	170	"	"	"
5	92	monzogranite	"	monzogranite
6	18	granodiorite	"	"
7	247	"	"	granodiorite
8	2391	"	"	"
9	82	"	"	monzogranite
10	2010	"	"	granodiorite
11	2416	"	"	monzogranite
12	MI-1	"	"	trondjhemite
13	21	"	"	"
14	84	"	granite	monzogranite
15	75	"	granodiorite	granodiorite
16	2124	"	"	"
17	76B	"	"	"
18	26	"	"	monzogranite
19	172	"	"	granodiorite
20	171	"	granite	monzogranite
21	2067	"	granodiorite	granodiorite
22	2522	"	"	"
23	2467	"	"	"
24	2051	"	"	monzogranite
25	100	"	"	granodiorite
26	199	"	"	tonalite
27	149	"	"	granodiorite
28	2080	"	"	monzogranite
29	2139	"	"	granodiorite
30	103	"	"	granogabbro
31	147	"	quartz monzodiorite	quartz granodiorite with bi, hb
32	2126	"	granodiorite	granodiorite
33	3	"	"	"
34	2022	"	tonalite	granodiorite with hb, bi
35	119	"	granodiorite	granodiorite
36	2058	"	"	"
37	87	"	"	"
38	8	quartz monzodiorite	quartz monzodiorite	quartz monzodiorite
39	2127	granodiorite	granite	incorrect analysis
40	198	"	granodiorite	granodiorite

As shown on the diagrams presented in this paper, the Culmea Cernei granitoids offer geochemical and petrochemical arguments in favour of their generation during subduction (group I, Andine type, of Hussein et al., 1982). Geochemically, these granitoids differ from the granites of group II, in relation to sutures (formed in a thicker crust in collision zones) and from the granites of group III (anorogenic or within plate type, generated in the mantle in relation to riftogenesis process).

This genesis is in accordance with both the petrochemical data which make possible the characterization of the Culmea Cernei granodioritic calc-alkaline series as a "medium-K" series and their allochthonous, synorogenic character.

The absence of radiogene datings makes difficult the estimation of the tectogenetic events with which these granitoids can be correlated in space and time.

Institutul Geologic al României

Table 4 – Characteristic features of the Culmea Cernei granitoids – comparable criteriawith I and S types

Criteria	I type granitoids	S type granitoids	Culmea Cernei granitoids
Field relationships	– Big complex intrusions – Wide composition limits: gabbro-diorites – 15 % granodiorites – 50 % granites – 35 % – Genetically and spatially related to volcanics	– Small-sized intrusions – Small composition limits: gabbro-diorites – 2 % granodiorites – 18 % granites – 80 % No contemporaneous volcanic rocks	– Big complex intrusions – Compositions: – QAP diagram – Fig. 2 – Table 3 monzogranites – 26 % granodiorites – 66 % monzodiorites – 3 % tonalites – 5 % – No volcanics associated
Mineralogical	– magnetite (common oxide) – hb prevalence versus bi – few muscovite (felsic varieties) – poor in accessory minerals – colour index = 12.3 (5–27) (for granodiorites and monzogranites cf. Didier et al., 1982)	– ilmenite (common oxide) – bi, hb prevalence – in mafic varieties – occurrence of accessory minerals – colour index = 4(0–17) (for leucogranites, cf. Didier et al., 1982)	– average value magnetite = 1.7 CIPW norm – average value ilmenite = 0.7 CIPW norm – magnetite > ilmenite – granitoids with hb + bi or bi – poor in accessory minerals (Fig. 8) – colour index = 11.6 (3–25)
Chemical	– SiO_2 = 67.4 (54–73 %) – $Al_2O_3/(Na_2O+K_2O+CaO)(mole) \leq 1.1$ – Na_2O = 3.31 %; Na_2O 3.2 % – for felsic rocks – CaO = 1.86 % (2.977, cf. Chappel) – K_2O = 4.13 % (= 2.13 % acc. to Chappel and White) – $\Sigma(Fe_2O_3+FeO+MgO+MnO+TiO_2)$ = 5.66 (2–11) – $Fe_3^{3+}/(Fe^{3+}+Fe^{2+})>0.2$ – uniform variations on the minor elements diagrams/differentiation index	– SiO_2 = 72.7 (66–79 %) – $Al_2O_3/(Na_2O+K_2O+CaO)>1.1$ – Na_2O = 3.35 % – CaO = 0.91 % – K_2O = 4.42 % – $\Sigma(Fe_2O_3+FeO+MgO+MnO+TiO_2)$ = 2.61 (0–6) – $Fe^{3+}/(Fe^{3+}+Fe^{2+})<0.2$ – irregular variations on the minor elements diagram/differentiation index	–SiO_2 = 67.55 (58.7 – 76.03 %) – generally ≤ 1.1 samples: 2416, 21, 2080, 103, 2022 have values >1.1 – Na_2O = 3.76 % (>3.2 %) – CaO = 2.90 % – K_2O = 2.56 % – Nb < 10 ppm – $\Sigma(Fe_2O_3+FeO+MgO+MnO+TiO_2)$ = 4.06 – $Fe^{3+}/(Fe^{3+}+Fe^{2+})$ = 0.42 (>0.2) – uniform variations on the minor elements/differentiation index
Isotopic	$Sr^{87}/Sr^{86}<0.706$	$Sr^{87}/Sr^{86}<0.706$	not given

Institutul Geologic al României

In this respect, they can be referred to the Late Proterozoic ("Assyntian")and synchronous with the granitoids associated with the Lainici-Păiuş Group (Frumosu, Tismana, etc) or to the pre-Ordovician Lower Paleozoic ("Caledonian").

In the former case, a polyphase metamorphic evolution of the Drăgşan and Lainici-Păiuş megasequences took place with a change of the physical conditions (baric and thermal) during the second event, the intrusion of the Culmea Cernei granitoids being late-orogenic, post-metamorphic.

In the latter case, the evolution of the host metamorphites could be considered as a polycycle one, that is a reactivation of a Proterozoic basement in the framework of "island arc-type"; in this case the allochthonous intrusion of the granitoids from the Culmea Cernei calc-alkaline suite and the magma generation by subduction could be assigned to Paleozoic events.

Conclusions

Taking into account the field observations and the petrological and geochemical data (Tab. 4) the Culmea Cernei granitoids can be characterized as follows:

- medium-K, calc-alkaline granitoids (below the curve 50 % K on the CNK diagram);
- granodiorites with monzogranite "nuclei" predominate, tonalites and diorites being subordinated quantitatively;
- the whole series plots in the field of intermediary rocks (AbAnOr diagram), which at pressures of 5 kb would have required temperatures of magmas formation varying between 700 and 750 ^{0}C;
- the differentiation tendency of the AFM diagram, perpendicular on the FM side, and the contents of Nb less than 40-50 ppm, illustrate the existence of a compressional geotectonic medium, the rocks falling in the field of volcanic arc magmas (Nb-SiO_2 diagram);
- the contents of minor elements close to those of the standard granitoids point to positive correlations of Sr with Ca, Ba with Na and K and of Sc, Co, Cr, Ni, V with Mg and Fe (total).

The field relationships with the host rocks - polymetamorphic Proterozoic formations of the Drăgşan Group - indicate the allochthonous character of the granitoids, post-metamorphic in relation to the metamorphic events (M_1, M_2) but synorogenic with a regional deformational event (D_3).

Genetically, the Culmea Cernei granitoids belong to type "I" (Na_2O/K_2O diagram), therefore "igneous" I type granitoids cf. Chappel and White (1974). Considering the field evidence, the mineralogical and geochemical data, as well as the presence of the consanguineous magmatite enclaves, the Culmea Cernei granitoids can be regarded as M type granitoids (cf. Didier et al., 1982), that is originating from the mantle or with a mixed genesis - mantle + crust.

The tectonic medium in which the Culmea Cernei granitoids of the medium-K calc-alkaline series were generated seems to have been controlled by subduction processes, considering their appartenance to the infrastructure of an island arc type complex (group I, cf. Hussein et al., 1982).

Temporally, the Culmea Cernei granitoids can be associated with "Assyntic" (Upper Proterozoic) or "Caledonian" (Lower Paleozoic) orogenic events.

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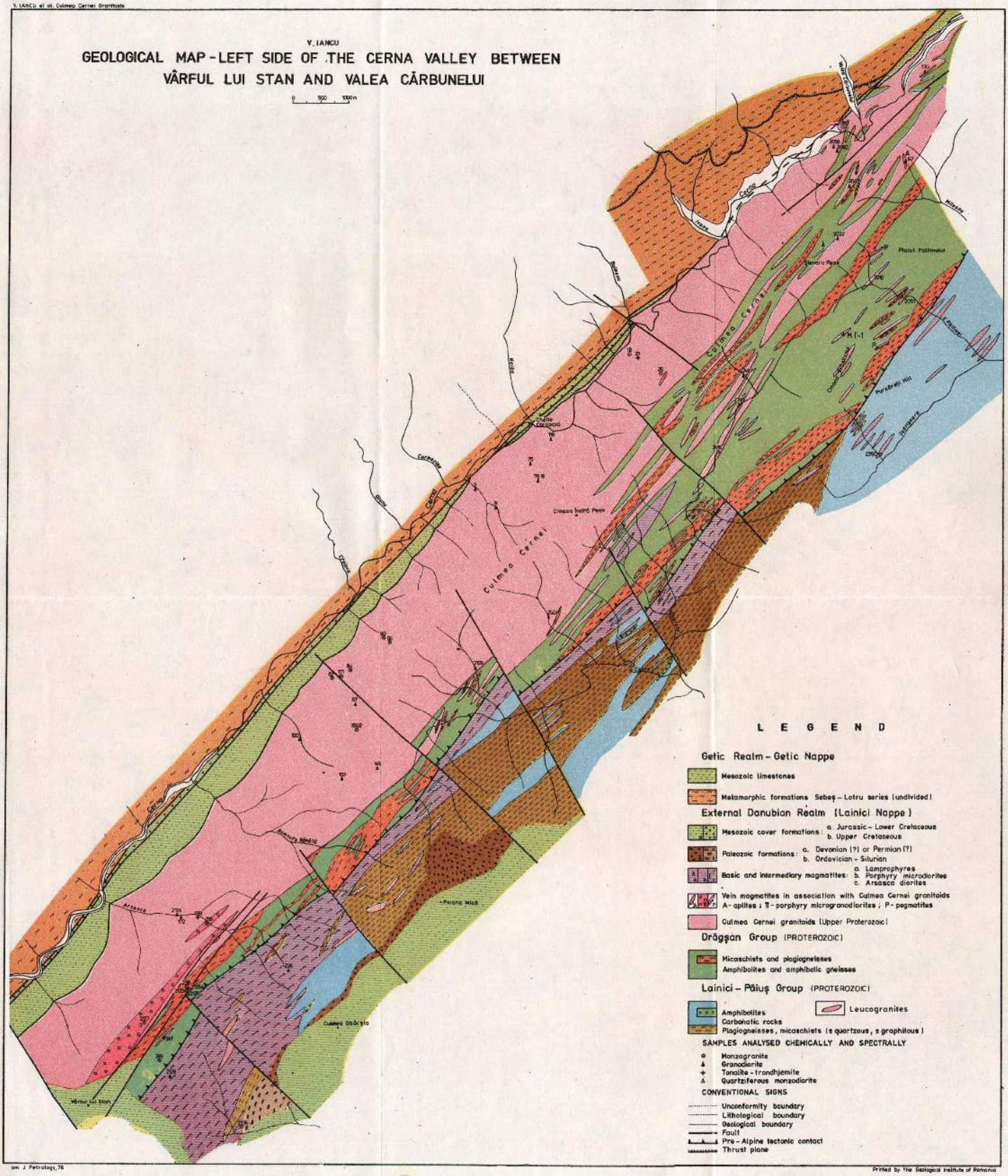
V. IANCU

GEOLOGICAL MAP - LEFT SIDE OF THE CERNA VALLEY BETWEEN VÂRFUL LUI STAN AND VALEA CĂRBUNELUI

0 500 1000m


Culmea Cernei
Cleopa Înaltă Peak
Stevaru Peak
Piatra Paltinului
Porcarii Hill
Culmea Dârâia
Vârful lui Stan
Cerna
Cleopa
Stevaru
Paltinu
Mizgan
Porcarii
Dârâia
Carpen
Stan
Cârbunelui
Valea Cârbunelui
Valea Mizgan
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Valea Cernei
Valea Paltinului
Valea Carpenului
Valea Stevarului
Valea Cleopa Mică
Valea Cleopa Înaltă
Valea Cleopa Joasă
Valea Cleopa Mijlocie
Valea Cleopa Superioară
Valea Cleopa Inferioară

Rom. J. Petrology, 1994, 76, p. 33–39

CHARACTERIZATION OF S–I TYPE GRANITOIDS FROM THE BANAT AREA, SOUTH CARPATHIANS

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Institutul Geologic al României. Str. Caransebeş nr.1, R 78344 Bucureşti 32.



Key words: Granites. Granodiorites. Tonalite. Anatexis. Subduction. South Carpathians.

Abstract: The S–I discriminating parameters (mineralogical, petrochemical, field relationships) show that the analysed granitoids are mainly of crustal-anatectic origin. However, the granitic magmas underwent also the influence of the melts coming from the oceanic crust and/or the upper mantle. The tectonogenetic evolution of granites started with the I-Andine type ocean/continent subduction, continued through the S-Hercynian type (main phase) continent/continent collision and ended with the post-closing intrusion of I-Caledonian type.

1. Introduction

The acid intrusions of Banat are represented by the Sicheviţa-Poniasca (Getic Domains), Cherbelezu-Sfârdin and Ogradena-Sesemin (Autochthonous Domain) granitoids (Fig. 1).

The primary data for the characterization of the granite types have been taken over from the papers of Savu and Vasiliu (1969), Bârlea (unpublished data), Stan et al.(1985), Stan et al. (1992), Stan (in press).

The specific parameters for distinguishing the S I granite types refer to the field relationships and the tectonic setting, petrographic nature, mineralogy and chemism of granitoids.

2. Field relationships; tectonic setting

The granitoid rocks occur as intrusions in the crystalline schists within the pre-alpine orogenesis of the South Carpathians.

Magmatites are represented by plutons (15–40 km long, 5–15 km wide), often associated with small-sized bodies.

The prograde regional metamorphism of the crystalline rocks of Barrovian type took place in the almandine amphibolite facies.

The crystalline schists were retromorphosed in many cases on alignments marked by reverse faults at the level of the greenschist facies, prior to the granitoid intrusion.

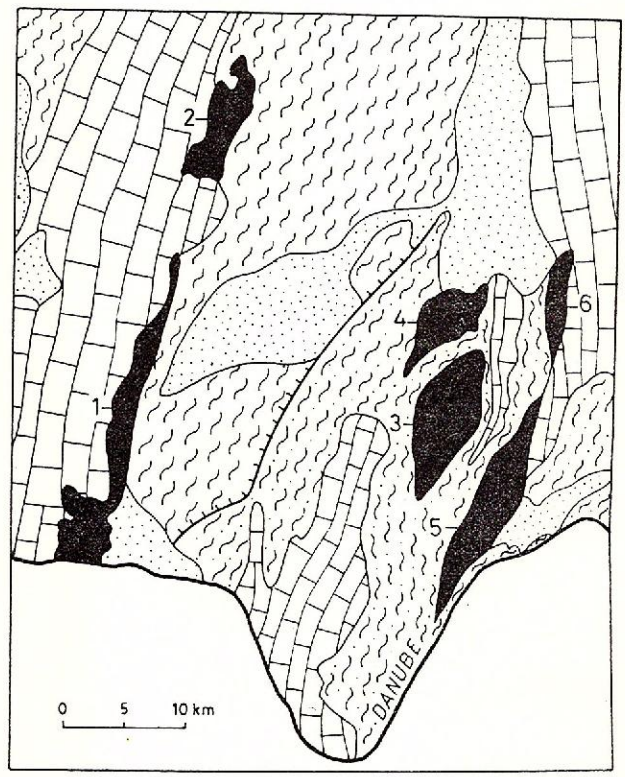

Fig. 1 – Location of granitoids: 1, Sicheviţa; 2, Poniasca; 3, Cherbelezu; 4, Sfârdin; 5, Ogradena; 6, Sesemin.

Crystalline schist enclaves very often appear in the Ogradena granitoid; they occur more rarely in the Sicheviţa-Poniasca and very rarely in the Sesemin intrusions. The Cherbelezu-Sfardin granitoids are de-

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Table 1
Chemical Parameters and Colour Index (CI)

Chemical elements and ratios	Name of the granitoid rocks					
	Sichevița n = 27	Poniasca n = 19	Cherbelezu n = 21	Sfârdin n = 30	Ogradena n = 15	Sesemin n = 6
SiO_2 %	67.75 / 63.52–75.67	68.50 / 59.36–75.96	71.52 / 69.81–73.89	67.67 / 63.32–71.62	74.51 / 73.62–75.68	74.53 / 72.50–75.50
Na_2O %	3.84 / 3.19–4.41	4.01 / 3.36–4.68	4.52 / 3.97–4.92	3.72 / 3.67–4.29	4.17 / 3.03–4.95	4.08 / 3.27–4.76
K_2O %	2.85 / 1.69–4.60	3.55 / 2.30–4.82	3.00 / 2.02–4.30	3.91 / 3.11–4.68	3.29 / 2.08–4.85	4.06 / 3.24–5.80
Mol.Al_2O_3/Na_2O+K_2O+CaO	1.23 / 1.16–1.41	1.16 / 1.02–1.25	1.16 / 1.10–1.29	1.15 / 1.00–1.38	1.07 / 1.00–1.19	1.08 / 1.02–1.20
$Fe^{3+}/Fe^{3+}Fe^{2+}$	0.58 / 0.23–1.00	0.31 / 0.17–0.51	0.60 / 0.43–0.71	0.36 / 0.25–0.54	0.52 / 0.0–0.80	0.63 / 0.72–0.86
Y (ppm)	10-21	10-23	6.5-13	7-15.5	10-26	10-26
Nb (ppm)	10-16	10-20	10-28	10-19	10-14	10-22
CI	9.80 / 4.5–12.98	10.15 / 3.20–14.25	7.54 / 2.31–13.68	13.7 / 7.18–20.5	4.85 / 1.70–9.78	3.00 / 1.16–5.61

n = sample number $\frac{67.75}{63.52-75.67} = \frac{meanvalue}{minimum-maximum}$

Table 2
I-S Discrimi

Characteristics of the granitoid types according to various authors*		
Criterion	I-type (Caledonian) magnetite series	S-type (Hercynian) ilmenite series
Field relationships	Generally large-sized intrusive complexes (batholites): granites 35%, granodiorites 50%, gabbro-diorites 15% Genetically and spatially associated with contemporaneous volcanics Mixed xenoliths Post-kinematic intrusions	Generally small-sized intrusions (plutons) and medium-sized bands: granites 80%, granodiorites 18%, gabbro-diorites 2% Devoid of contemporaneous volcanics Dominant metasediment xenoliths syn- and post-kinematic intrusions
Mineralogical	Hornblende>biotite; quantitatively insignificant muscovite K-feldspar occurs interstitially or invasively Poor in accessory minerals: with allanite and sphene but without cordierite, garnet, andalusite and sillimanite With magnetite	Biotite>hornblende muscovite and biotite common minerals K-feldspar often occurs as megacrystals, locally autometasomatic Frequent accesory minerals : monazite, apatite, zircon and garnet with ilmenite
Chemical	SiO_2 < 65% Na_2O > 3.2% Mol.Al_2O_3/(Na_2O+K_2O+CaO) < 1.1 $Fe^{3+}/(Fe^3+Fe^2)$ > 0.2	SiO_2 > 65% Na_2O < 3.2% Mol.Al_2O_3/(Na_2O+K_2O+CaO) > 1.1 $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ < 0.2
Other parameters	Normative diopside CI > 13	Normative corundum CI < 13

* References: Chappell, White (1974), Didier et al. (1982), Hine et al. (1978), Hussein, Pitcher (1982, 1987), White, Chappell (1977, 1983).

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void of enclaves. The Poniasca granitoid contains melanocrate enclaves.

The granitoid/crystalline schist contacts are always sharp. The contact metamorphism is marked by specific minerals such as andalusite, biotite, garnet, sillimanite and staurolite.

3. Petrographic and mineralogical parameters

The intrusive magmatites of Banat are mostly represented by granites (monzogranites>syenogranites) and to a less extent granodiorites. But the Sichevița-Poniasca granitoids (Getic Domain) consist moreover of trondhjemites and tonalites (Fig. 2).

The potassium feldspar phenocrysts (microcline, microcline-perthite) occur on large areas in certain zones in all the rock types; they occur more rarely in the Ogradena granitoids and very rarely in the Sesemin granitoids. The interstitial potassium feldspar is omnipresent.

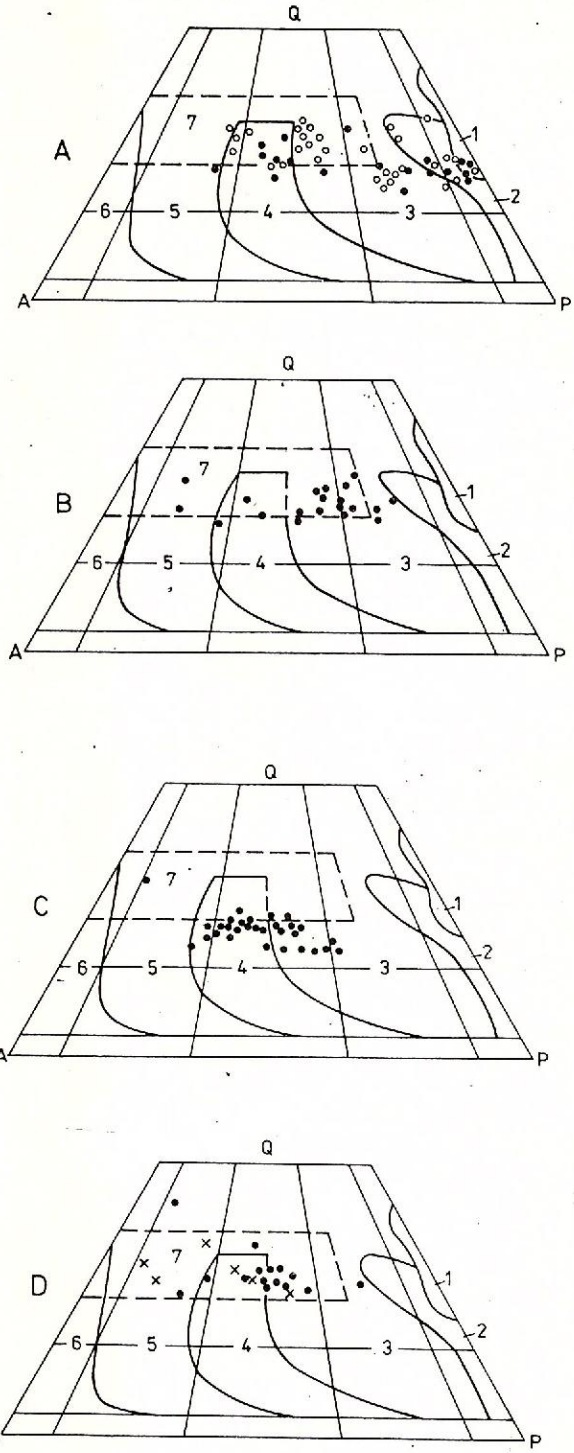

Fig. 2 – Q-A-P Streckeisen diagram showing the variation fields of the granitoid series after Lameyre and Bowden (1982). 1, tholeiitic series; 2, calc-alkaline trondhjemitic series (low K); 3, calc-alkaline granodioritic series (medium K); 4, monzonitic calc-alkaline series (high K); 5, aluminous granodiorites from alkaline provinces; 6, alkaline and peralkaline granites; 7, granitoids formed by crustal fusion. A, Sichevița (circles)-Poniasca (dots); B, Cherbelezu; C, Sfârdin; D, Ogradena (dots)-Sesemin(crosses).

nating Parameters

Characterization of the Banat granitoids	
Comparative data	I-S
Frequent plutons, often associated with medium-sized bands: granites >80%, granodiorites < 10%, tonalites < 2%	S
Devoid of contemporaneous volcanics	S
Frequent metasediment xenoliths; for the Poniasca granitoid also melanocrate separations	S-I
syn- and post-kinematic intrusions	S-I
Muscovite and biotite common minerals; hornblende occurs in small amounts only in the Sichevița and Poniasca granitoids	S
K-feldspar often occurs as megacrystals but also invasively	S-I
Accessory minerals: monazite, apatite, zircon and garnet	S
with dominant ilmenite	S
SiO_2 > 67%(67.67–74.52%)	S
Na_2O > 3.2% (3.72–4.52%)	I
Mol. Al_2O_3/Na_2O+K_2O+CaO >< 1.1 (1.07–1.23)	I-S
$Fe^{3+}/Fe^{3+} + Fe^{2+}$ > 0.2 (0.31–0.63)	I
Normative corundum	S
CI < 13 (3.00–13.7)	S

Ramly (1982) Ishihara (1977), Maniar, Piccoli (1989),

Institutul Geologic al României

Muscovite and biotite are characteristic components of granitoids. Hornblende has been only identified in the tonalitic varieties of the Sicheviţa-Poniasca granitoid.

The accessory minerals such as: monazite, apatite, zircon, tourmaline are more abundant in the Sicheviţa-Poniasca and Cherbelezu-Sfârdin granitoids; they occur sporadically in the Sesemin granitoids. Garnet appears more or less accidentally everywhere.

Ilmenite quantitatively prevails over magnetite in the Sicheviţa-Poniasca and Cherbelezu-Sfârdin granitoids. Magnetite and ilmenite have not been identified in the Ogradena-Sesemin granitoids.

4. Chemical parameters

The chemical features of granitoids may be synthetized as follows (Table 1, average values):

High SiO_2 values with a restricted variation: 67.67–74.53 %;

High Na_2O values: 3.72–4.52 %;

Normal to medium K_2O values: 2.85–4.06 %;

$Na_2O > K_2O$ (Fig. 3);

Mol $Al_2O_3/Na_2O + K_2O + CaO$: 1.07–1.23;

$Fe^{3+}/(Fe^{3+} + Fe^{2+}) = 0.31–0.63$;

Y : 6.5–26 ppm;

Nb : 10–28 ppm;

CI : 10–13.7.

Most of these values are characteristic of the S-type granites; some parameters, however, are specific for the I-type granites (Tab. 2).

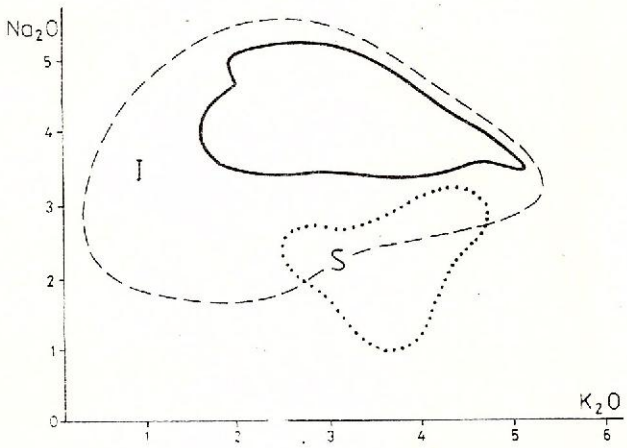

Fig. 3 — Na_2O/K_2O diagram (Chappell, White, 1974) for discriminating the I-S types. Solid line: the Banat granitoid field.

5. Compensational-extensional parameters

Granitoids formed under compressional conditions because:

In the FAM diagram (Petro et al., 1979) which illustrates also the calc-alkaline characters of the rocks, the plotted points follow an approximately perpendicular direction to the FM side (Fig. 4).

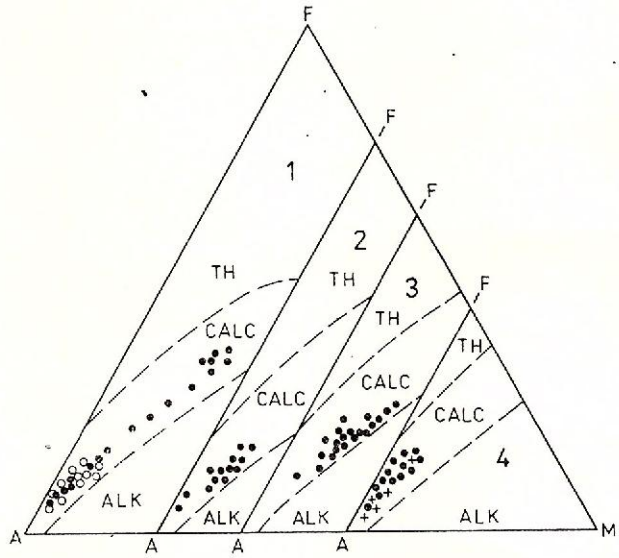

Fig. 4 — AFM diagram (Pedro et al., 1979) for discriminating the extensional/compressional tectonic conditions. 1, Sicheviţa (circles)-Poniasca (dots); 2, Cherbelezu; 3, Sfârdin; 4, Ogradena (dots)-Sesemin (crosses).

On the R_1-R_2 diagram (Batchelor & Bowden, 1985) the values of the parameters plot in the fields 2, 3, 4 (Fig. 5).

Nb and Y values (Pearce et al., 1984) are below 40–50 ppm (Tab. 1).

Mol $Al_2O_3/Na_2O + K_2O + CaO$ ratio (Maniar, Piccoli, 1989) is higher than 1.15 for most of the Banat granitoids (Tab. 1).

6. Data interpretation

According to the classification proposed by Chappell and White (1974), most of the parameters show that the Banat granitoids belong to the S-type (Tab. 2), except for the Na_2O and $Fe^{3+}/Fe^{3+} + Fe^{2+}$ values (Fig. 3, Table 2 respectively) which distribute the granitoid rocks to the I-type. These facts could be interpreted perhaps as an influence of the magmatic products coming from the ocean crust and/or the upper mantle over the crustal anatectic magma. The influence of the basic magma is also suggested by tonalitic and trondhjemitic rocks from the Sicheviţa and Poniasca granitoids; moreover, the melanocrate separations from the Poniasca granitoids (Savu, Vasiliu, 1969) also support

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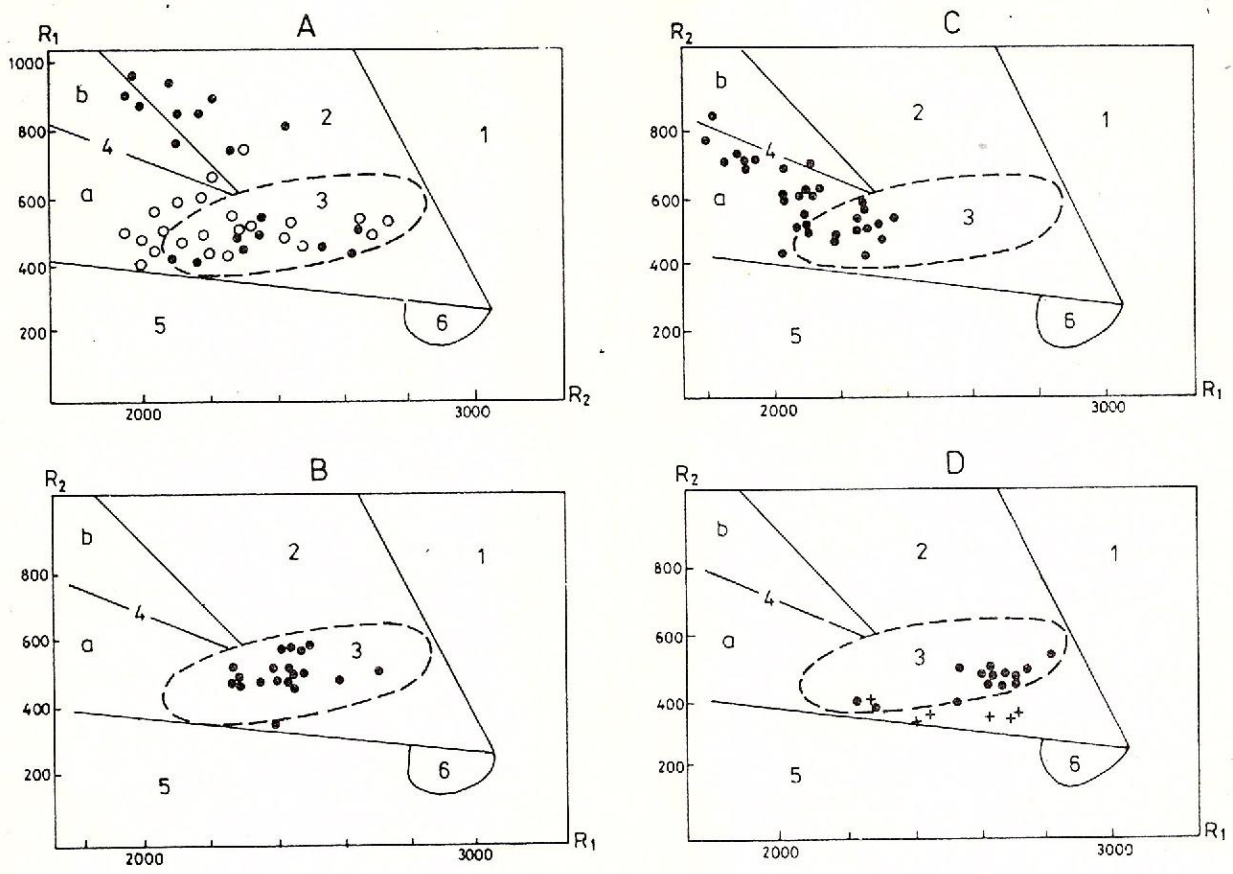

Fig. 5 – R_1–R_2 Diagram (Batchelor, Bowden, 1985) for tectonically discriminating the granitoids: 1, ocean ridge granites; 2, Andine type granites: ocean/continent subduction; 3, Hercynian type granites: continental syncollision; 4, Caledonian type granites: late orogenic (a) and post-collision (b); 5, anorogenic granites; 6, post-orogenic granites. A, Sicheviţa (circles)-Poniasca (dots); B, Cherbelezu; C, Sfârdin; D, Ogradena (dots)-Sesemin (crosses).

this view. The melanocrate enclaves, like those from the granitoid rocks of the Central French Massif, are interpreted by Didier et al. (1982) as an effect of the contamination of the crustal-granitic magmas with melts coming from the mantle. The influence of the basic magmas on the genesis of the Cherbelezu-Sfârdin and Ogradena granitoids (Autochthonous Domain), Poniasca granitoid (Getic Domain) is supported by Robu, Robu (1992), by the morphologic study of the zircon crystals.

Most of the magmas from which the granitoid rocks of Banat consolidated, formed, however, by anatexis of the continental crust as suggests also the Q-A-P diagram (Lameyre, Bowden, 1982; Fig. 2).

From the tectonic point of view there are also elements showing that the granitoids belong to both the S and I-types. Most of the rocks on the R_1-R_2 diagram (Fig. 5) plot in the field 3 (continent/continent collision of the S-Hercynian type); some Sfârdin, Sicheviţa-Poniasca granitoids plot in the fields 4a and 4b (late orogenic, post-collision respectively, both belonging to the Caledonian I-type). Moreover, the tonalitic and trondhjemitic rocks of the Poniasca granitoid plotted in the field 2 suggest their genesis in an ocean-continent subduction process (I-Andine type).

Field data show that the Banat granitoids were intruded on extended regional faults, and in many cases these granitoid rocks intruded the retromorphosed crystalline schists; therefore they are late tectonic or even post-tectonic granitoids (Stan et al., 1985, 1992). The post-closing uplift on the tensional faults of some discordant plutons is characteristic of the I-Caledonian type, while the oblique continental collision with the formation of anatectic melts, often autometasomatically regenerated, is specific for the S-Hercynian type (Pitcher, 1982, 1987).

7. Conclusions

The Banat granitoids formed in several stages during a long period of time. The first stage is marked by the formation of I-Andine type (Fig. 6 A) by the ocean/continent subduction. The continent/continent collision took place through the advancement of the

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subduction process. Thus anatectic magmas, mostly of crustal origin, formed in an S-Hercynian type tectonic setting (Fig. 6 B). The previous Andine-type granitoid rocks were recycled. Besides, the basic melts coming from the lower lithosphere and/or from the mantle, modified the initial composition of the crustal magmas. Finally, the consolidation of the granitoid magmas took place mostly under a post-closing tectonic regime of I-Caledonian type (Fig. 6 C).

The co-existence of the S-I granite types is decisively determined by the thickness of the crust that gets fused. This influences the pressure conditions, the magmatic differentiation processes, the circulation of the autometamorphic solutions and the tectonic regime (Pitcher, 1987).

The Sicheviţa-Poniasca granitoids from the Getic Domain are petrogenetically more differentiated (granites-granodiorites-tonalites ± trondhjemites) than the Cherbelezu-Sfârdin and Ogradena-Sesemin granitoids belonging to the Autochthonous Domain (granites-granodiorites).

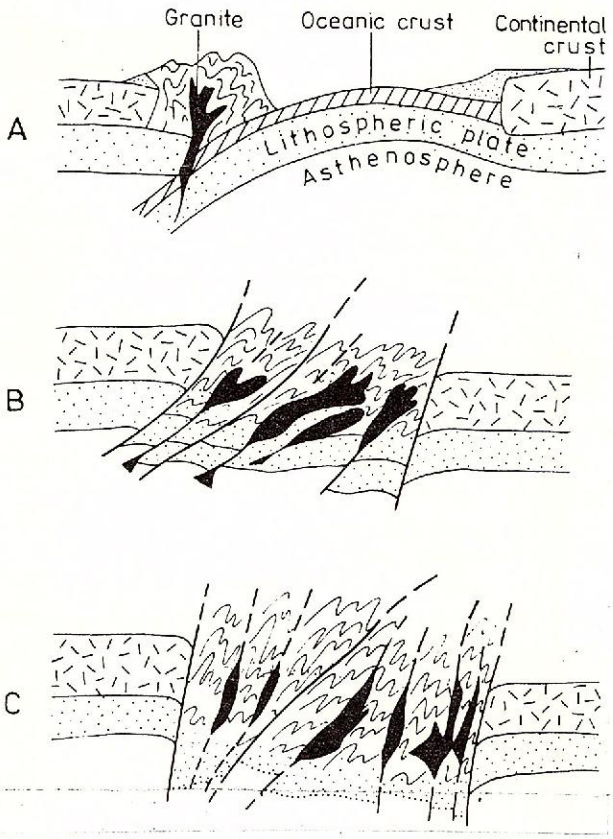

Fig. 6 – Tectonogenetic evolution of granitoids: A, ocean/continent subduction of I-Andine type; B, continent/continent collision of S-Hercynian type; C, continental collision with the uplift of the post closing granitoids of I-Caledonian type.

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NEW DATA CONCERNING THE ORIGIN OF THE IGNEOUS LEUCOCRATIC ROCKS IN NORTHERN SEBEŞ MOUNTAINS – SOUTH CARPATHIANS. GEOCHEMICAL EVIDENCE

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Key words: Igneous rocks. Trondhjemite. Major elements. Minor elements. Rare earths. Partial melting. Sebeş Mountains. South Carpathians.

Abstract: The igneous leucocratic rocks of the northern area of the Sebeş Mountains occur along a major fault that affects the metamorphics of the Sebeş-Lotru Series. The geochemical and mineralogical investigations characterize them as trondhjemitic rocks. The REE distribution patterns, as well as the trace elements behaviour, suggest a partial melting process involving a basic source material, such as amphibolites or Q-eclogites. The lower crust partial melting process led to a trondhjemitic-tonalitic-granodioritic melt which ascended rapidly and consolidated under hypabyssic conditions, without any significant crustal contamination.

1. Introduction

Igneous leucocratic rocks occurring as dykes and bodies in the northern part of the Sebeş Mountains are widespread along a major NE–SW fault that affects the crystalline terrain. We deal here with the leucocraric magmatites which outcrop between Sibişel Valley and Şugag village(Fig. 1). These igneous rocks are called Q-porphyres or porphyric granodiorites and have been studied mainly petrographically in the context of the metamorphic terrains. Chivu (1970) characterized them by Niggli's parameters as differentiation products of some plagioclase bearing granites.

The petrographic composition was considered granodioritic, varying from granites to tonalites, dacites and syenites (Cioflică et al., unpublished data). They were supposed to be consolidated in a hypabissic regime, being the result of a partial melting process that occurred under crustal conditions upon a granitoidic body, without any differentiation or assimilation effects.

2. Geotectonic setting

The leucocratic igneous bodies occur mainly in the Sibişel Series, composed of retrograde mesometamorphites and in the Sebeş-Lotru Series. The lithostratigraphy of this latter series comprises a lower level of augen gneisses, the two micas gneisses level and the micaschists level. The upper part of the middle level consists of augen gneisses, amphibolites, eclogites, metaperidotites and pegmatites. The basic-ultrabasic sequences are considered to represent metamorphosed olistoliths of an ophiolitic melange (Savu, Strusievicz, 1987).

The Sibişel Series which is the main host area for the acid igneous dykes and bodies is supposed to represent in fact a tectonic facies of the Sebeş-Lotru Series, a crustal shearing zone that lies on almost 90 km/8 km area between Subcetate and Răşinari (Stelea et al., 1992, unpublished data). The igneous bodies occur along this shearing zone, offering the access ways for magma.

3. Petrographic features

A detailed petrographic description was made in a previous study (Lupulescu et al., in press), the rocks we refer to were called porphyric microgranodiorites, with some medium-grained exceptions. The texture is porphyric, the mesostasis crystals size ranging from 0.01 to 1 mm and even (very rare) to a cryptocrystalline texture; a modal mineralogical composition of these rocks cannot be estimated for that matter. There are also some mesogranular rocks. The common feature of these rocks consists of a remarkable petrographical homogeneity. Their mineralogical composition is simple and generally constant.

The plagioclase (albite-oligoclase-andesine) is the main mineral and appears as phenocrysts (58–66 %), together with quartz (25–35 %), biotite (5–8 %) and

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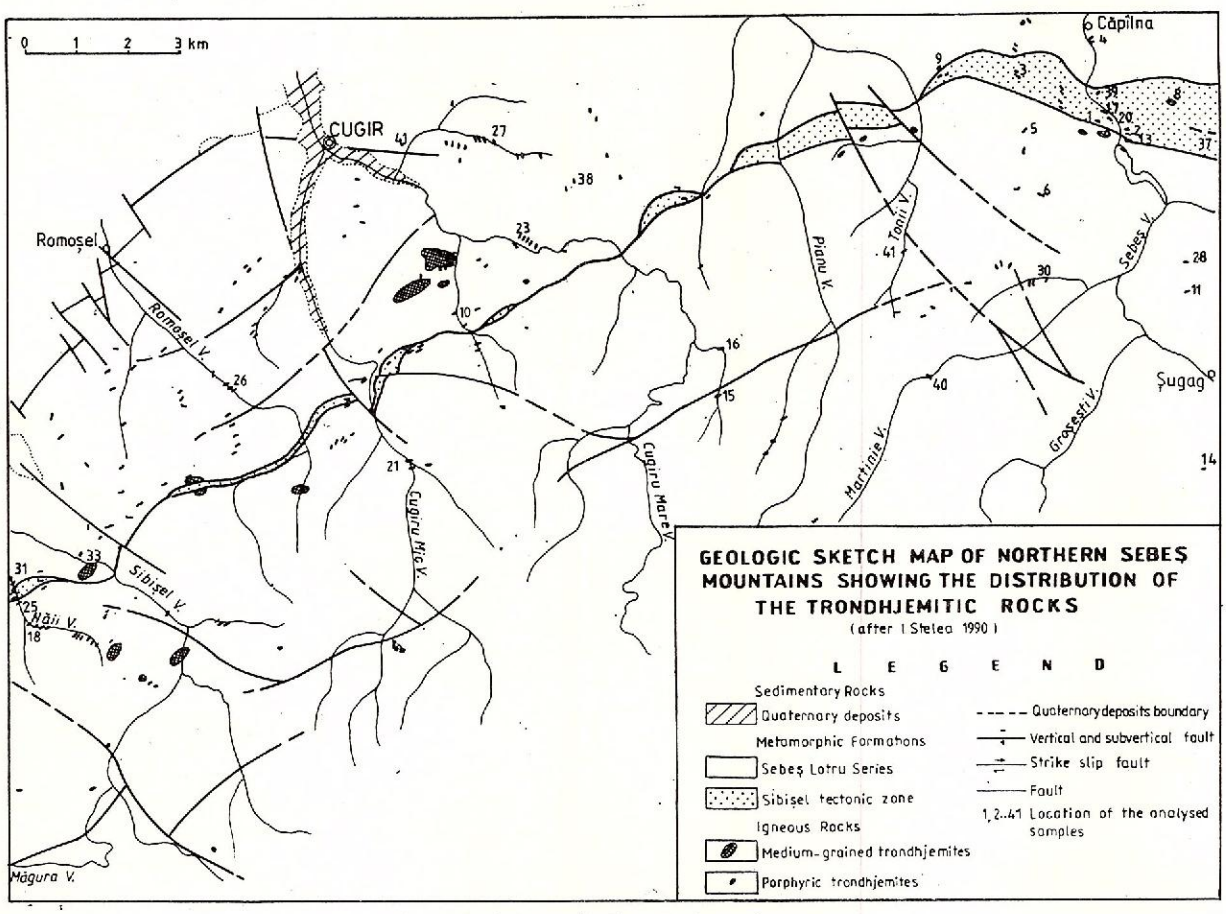

Fig. 1

hornblende (0–1 %). K-feldspar crystallizes only in the groundmass. Accessory minerals are apatite, rare sphene and zircon. Secondary phenomena have yielded chlorite, calcite, sericite, zoisite and epidote.

Recently (Lupulescu et al., in press), some physiographic aspects were described, as suggesting effects of a partial melting process. We agree especially with the large euhedral zoned plagioclase crystals, developed on anhedral supports, sometimes with undulatory extinction. However, most of other aspects previously described could even be interpreted as the effect of a normal crystallization of a melt.

This mineralogical composition, in which plagioclase plus quartz amounts to almost 90 % of the rock, and biotite (sometimes hornblende) represents the single mafic minerals (its weight is less than 10 %), characterizes the trondhjemites. Within the IUGS classification, trondhjemites are defined as leucotonalites with oligoclase-andesine plagioclase, quartz more than 20 % of the leucocratic minerals, K-feldspar up to 10 % of the total feldspars and the colour index less than 10 %.

4. Geochemistry

4.1. Major elements

Major elements composition of 41 rock samples are reported in Table 1.

Their main features are:

- SiO_2 > 68 %, up to 75 %
- Al_2O_3 exceeds 15 % in samples with less than 70 % SiO_2 (on an average), and is less than 14 % in those with more SiO_2.
- $(FeO^*+MgO)<3.4$ % and $FeO^*/MgO \geq$ 2-3, ($FeO^*=FeO+ 0.9\ Fe_2O_3$)
- CaO varies between 0.5 % and 3 %.
- Na_2O varies between 4.5 % and 7 %.
- K_2O is less than 2.5 %.

From this point of view a chemical homogeneity is obvious; the major elements values vary within the trondhjemitic limits of variation (according to Barker, 1979).

Mineralogically, trondhjemites have the same composition as tonalites and granodiorites. From the chemical point of view, FeO^*, MgO and K_2O contents

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differentiate them as follows: trondhjemites have lower (FeO*+MgO) contents than tonalites (less than 3.4 %) and lower K_2O content than granodiorites (less than 2-2.5 %).

For a more accurate definition we used O'Connor's (1965) classification based on normative proportions Ab-Or-An (Fig. 2). Our samples plot strictly in the trondhjemitic field (although this classification does not deal with the mafic components or colour index).

In QAP classification (Fig. 3) the majority of samples plot in granodiorites and tonalites field. This confirms the previous diagnosis, as in this diagram the tonalitic and even the granodioritic fields are quite analogous to that of the trondhjemites (we used normative values, in the QAP-diagram as the modal values are not available because of the microcrystalline structure of the mesostasis).

Harker variation diagrams (Fig. 4) show the following trends:

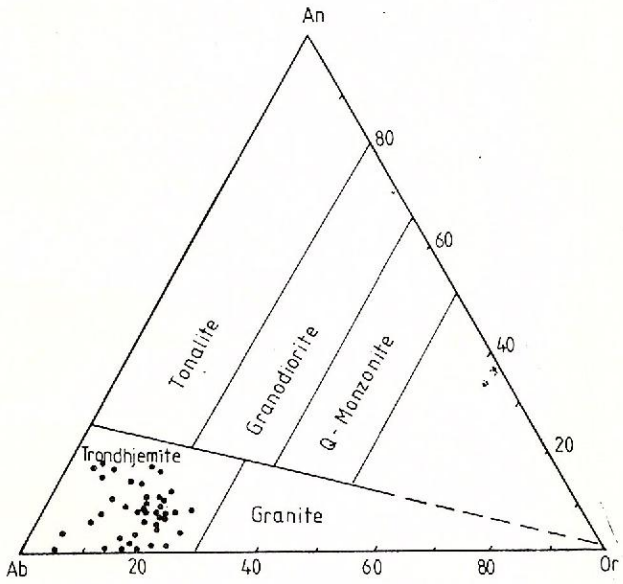

Fig. 2 – Normative plots of Ab-Or-An showing fields of some siliceous plutonic rocks (after O'Connor, 1965).

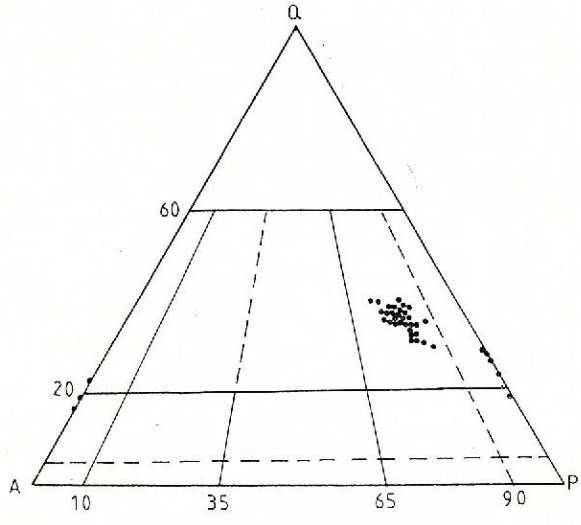

Fig. 3 – QAP-normative diagram.

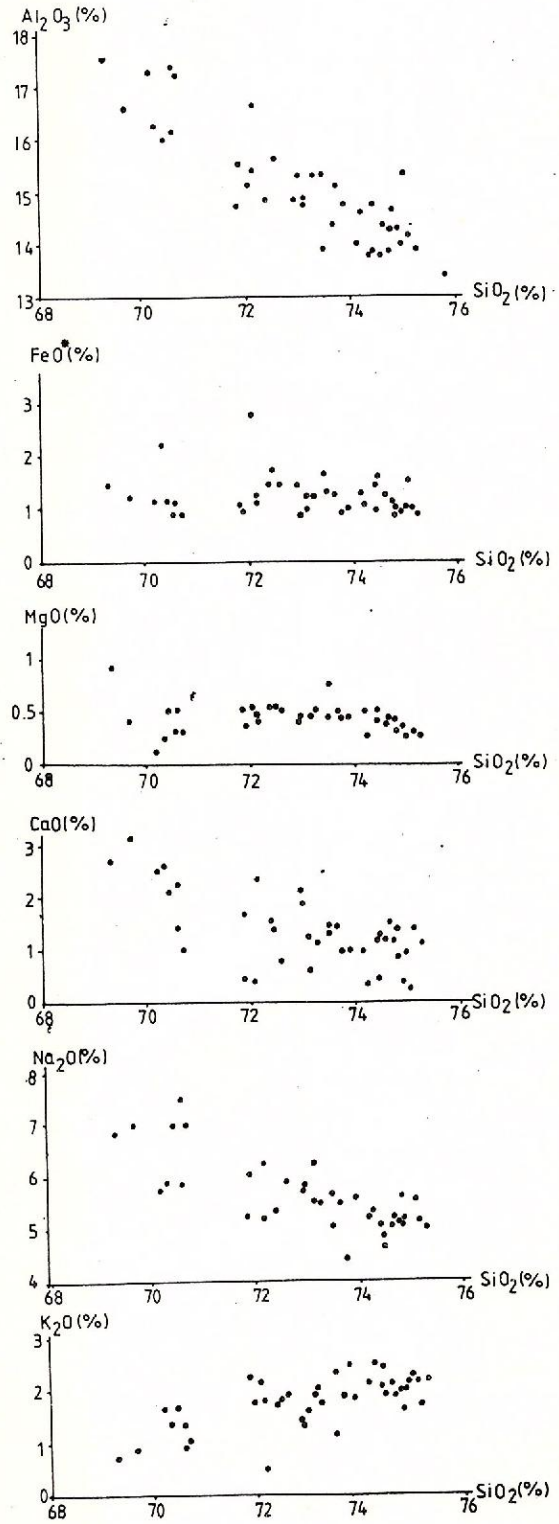

Fig. 4 – Harker diagrams.

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Table
Chemical composition

No.*	Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO
1	5207C	69.28	0.16	17.55	0.47	1.02	0.03	0.90
2	5206D	69.72	0.16	16.59	0.55	0.73	0.03	0.40
3	363	70.20	0.16	17.29	0.26	0.95	0.03	0.10
4	227	70.33	0.12	16.26	0.89	1.41	0.01	0.23
5	364	70.47	0.09	16.00	0.78	0.44	0.03	0.50
6	369	70.61	0.16	16.15	0.65	0.29	0.02	0.30
7	5207A	70.62	0.03	17.35	0.42	0.73	0.02	0.50
8	354	70.70	0.16	17.23	0.54	0.44	0.03	0.30
9	372	71.85	0.16	14.72	0.56	0.58	0.03	0.50
10	529	71.91	0.16	15.54	0.58	0.44	0.02	0.36
11	7098	72.04	0.16	15.09	2.54	0.52	0.11	0.53
12	7018	72.17	0.14	15.39	0.73	0.60	0.02	0.44
13	5205A	72.17	0.03	16.66	0.46	0.73	0.02	0.40
14	7092	72.40	0.26	14.84	0.82	0.74	0.01	0.53
15	525	72.47	0.16	15.53	0.56	1.27	0.03	0.52
16	524	72.59	0.16	15.66	0.53	1.02	0.03	0.50
17	5209A	72.96	0.19	14.86	0.16	1.32	0.03	0.40
18	608	72.98	0.13	15.27	0.44	0.51	0.03	0.44
19	7095	73.12	0.12	14.79	0.79	0.55	0	0.45
20	5208	73.12	0.03	14.86	0.62	0.44	0.02	0
21	536	73.27	0.12	15.33	0.43	0.97	0.03	0.49
22	1733B	73.48	0.15	13.89	0.66	1.08	0.02	0.74
23	507	73.49	0.16	15.31	0.69	0.74	0.03	0.43
24	7096	73.63	0.10	14.34	0.95	0.35	0.01	0.48
25	607	73.77	0.08	15.12	0.52	0.49	0.02	0.43
26	531B	73.88	0.16	14.74	0.44	0.62	0.03	0.43
27	6748	74.15	0.08	13.99	0.87	0.53	0.02	0.48
28	7099	74.22	0	14.59	0.95	0.22	0.03	0.25
29	1733C	74.42	0.10	13.79	0.96	0.59	0.02	0.50
30	6787	74.45	0.02	13.84	1.64	0.13	0.09	0.40
31	601	74.45	0.12	14.73	0.69	0.37	0.02	0.39
32	1786	74.60	0.07	13.79	0.83	0.52	0.02	0.37
33	515	74.64	0.12	14.36	0.40	0.74	0.03	0.39
34	1733B	74.73	0.09	13.84	0.80	0.45	0.01	0.40
35	607B	74.79	0.12	14.27	0.59	0.50	0.02	0.30
36	607A	74.81	0.04	14.64	0.51	0.42	0.02	0.29
37	678	74.89	0.12	14.27	0.83	0.19	0.01	0.34
38	6757	74.97	0	13.99	0.94	0.18	0.05	0.25
39	262	75.06	0.12	15.36	0.74	0.87	0	0.05
40	6970	75.12	0.14	14.14	0.82	0.29	0.01	0.29
41	6721	75.27	0.03	13.89	0.79	0.20	0.02	0.24

* The location of the analysed samples is expressed by order numbers
Mountains (Fig. 1). (Analyst:

– Al_2O_3 decreases with the increasing SiO_2 (showing a stressed slope versus SiO_2);

– FeO* and MgO decrease with increasing SiO_2, but their decreasing slope is much attenuated;

– CaO decreases with the increasing SiO_2, but its dispersion is greater, with an inflexion point at SiO_2 = 73 %;

– the alkalies present opposite trends: while K_2O increases with the increasing SiO_2, Na_2O presents reverse trend.

Institutul Geologic al României

1
of the rocks

CaO	Na_2O	K_2O	P_2O_5	CO_2	H_2O^+	H_2O^-	S	Total
2.66	6.80	0.75	0.10	0	0.26	0.08	0.02	100.08
3.08	7.00	0.90	0.05	0	0.27	0.10	0.02	99.60
2.52	5.75	1.65	0.10	0	0.98	0.14	0.02	100.15
2.61	5.90	1.43	0.09	0	0.28	n.d.	0.25	99.81
2.10	7.00	1.65	0.06	0	0.91	0.23	0.03	100.29
1.40	7.50	1.35	0.10	0	0.75	0.20	0.02	99.50
2.24	5.85	0.90	0.06	0	0.95	0.01	0.02	99.70
0.98	7.00	1.05	0.05	0	1.13	0.15	0.02	99.78
1.68	5.25	2.25	0.08	0	1.79	0.42	0.02	99.89
0.44	6.06	1.77	0.08	0.44	0.36	n.d.	0.22	99.72
0.39	5.08	2.14	0.07	0.34	0.87	n.d.	0.26	100.14
2.35	5.18	1.80	0	0.16	0.26	n.d.	0.30	99.54
2.38	6.25	0.50	0.06	0	0.40	0.26	0.03	100.35
1.57	5.34	1.72	0.07	0.33	0.78	n.d.	0.25	99.66
1.41	5.42	1.81	0.10	0.40	0.08	n.d.	0.27	100.03
0.81	5.91	1.89	0.06	0.11	0.46	n.d.	0.25	99.98
2.10	5.75	1.35	0.04	0	0.23	0.12	0.02	99.53
1.85	5.82	1.62	0.02	0	0.43	n.d.	0.24	99.78
0.60	5.52	1.96	0.04	0.39	0.89	n.d.	0.30	99.52
1.26	6.25	2.00	0.04	0	1.26	0.06	0.03	99.93
1.12	5.49	1.76	0.04	0.17	0.73	n.d.	0.24	100.19
1.29	5.04	2.33	0	0.25	0.32	n.d.	0.33	99.58
1.47	5.64	1.16	0.03	0.13	0.26	n.d.	0.25	99.79
1.42	5.52	1.84	0	0.16	0.53	n.d.	0.27	99.60
0.96	4.44	2.47	0.04	0.03	0.93	n.d.	0.25	99.55
0.98	5.62	1.81	0.08	0.50	0.30	n.d.	0.26	99.85
0.98	5.19	2.13	0.02	0.33	0.49	n.d.	0.29	99.55
0.32	5.36	2.50	0.01	0.09	0.77	n.d.	0.27	99.58
1.20	5.06	2.04	0.04	0.38	0.34	n.d.	0.25	99.69
1.23	4.85	1.90	0.01	0.29	0.71	n.d.	0.29	99.85
0.42	4.67	2.42	0.03	0	1.42	n.d.	0.23	99.96
1.19	5.03	2.10	0.04	0.31	0.38	n.d.	0.24	99.58
1.50	5.23	1.87	0.08	0	0.43	n.d.	0.28	100.07
1.15	5.12	1.98	0	0.10	0.56	n.d.	0.31	99.54
1.41	5.10	2.00	0.06	0.04	0.49	n.d.	0.23	99.92
0.84	5.62	1.61	0.04	0.40	0.29	n.d.	0.24	99.69
0.37	5.21	2.16	0	0.24	0.69	n.d.	0.24	99.56
0.92	5.13	2.22	0	0.07	0.56	n.d.	0.28	99.56
0.24	5.55	2.18	0.30	0	0.30	n.d.	0.28	100.78
1.41	5.15	1.71	0.04	0.10	0.65	n.d.	0.27	100.14
1.10	4.99	2.20	0.04	0.21	0.44	n.d.	0.27	99.69

and is figured on the Geologic Sketch Map of Northern Sebeş
Iosipenco N., "Prospecţiuni" S.A.)

The resembling behaviour of CaO and Al_2O_3 is explained by their presence in plagioclase.

The low and almost constant contents of FeO* and MgO versus SiO_2 trend (the decreasing is very slight) point to a trondhjemitic rock feature and might constitute an argument against a significant differentiation process by fractional crystallization.

Institutul Geologic al României

Another important feature for trondhjemitic rocks is that they are corundum-normative, feature that is common to almost all acid magmatites from northern Sebeş Mountains. In semilog diagram of SiO_2 versus K_2O our rocks plot in the continental trondhjemitic area (Fig. 5).

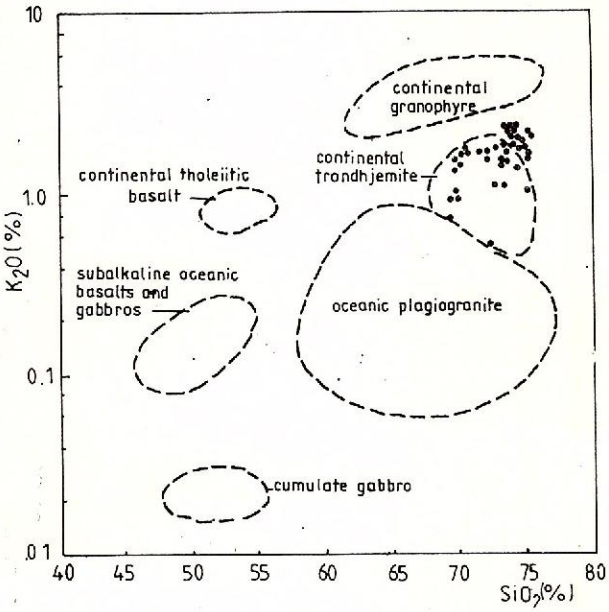

Fig. 5 – Semilog plot of SiO_2 vs K_2O comparing oceanic plagiogranites with continental granophyres, trondhjemites, tholeiites and subalkaline oceanic basalts and gabbros.

4.2. Trace elements

The 41 samples were analysed for some minor elements shown in Table 2. The rocks are very poor in minor elements, their contents ranging between normal values for acid rocks.

Ba and Sr show small contents (between 100-500 ppm), according to the small K and Ca content.

Zr varies around 60 ppm, small contents that could suggest a lack or a very little crustal contamination.

Small contents of Ti and Y indicate the possibility of being captured by hornblende, biotite, sphene and ilmenite.

Ni, Cr, Co and V have very small contents that could show either the remaining of pyroxenes in the residue, or a depleted source-material.

4.3. REE

REE content of the 8 analysed samples are shown in Table 2 and their pattern in Figure 6. These show low abundance for Σ REE, varying between 8 and 60 ppm. LREE/HREE expressed by $(Ce/Yb)_{cn} \approx 1.5$–37.2 is moderate to high.

Two different patterns can be distinguished: five samples show a strong fractionated pattern (mainly of the HREE) and no Eu anomaly, and are defined as granodiorites in QAP classification.

The other three display a slighter fractionation of REE and positive Eu anomalies, and plot in the tonalite field in QAP diagram. The last three sample patterns seem to be characteristic just for some of these rocks, as they belong to a rather limited zone in the eastern part of the region.

The "granodioritic" group has a lower LREE (less than 40) content and MREE much lower (these elements being retained generally by clinopyroxenes, apatite and hornblende). Ce/Yb $\geq$ 11 and the lack of Eu anomaly suggests the absence of plagioclase in residue.

Table

Trace elements contents

No.	Sample	Pb	Cu	Ga	Cr	V	Zr	Ba	Sr
1	5207C	20.0	5.0	25.0	14.0	25.0	130	180	123
2	5206D	10.0	5.0	24.0	0	15.0	110	265	122
3	363	11.0	22.0	18.0	35.0	14.0	60	200	290
4	227	9.0	18.0	17.0	16.0	14.0	62	235	670
5	364	9.0	7.0	15.0	4.5	11.5	47	230	280
6	369	36.0	4.5	12.0	8.5	12.5	65	170	400
7	5207A	15.0	3.0	25.0	0	21.0	125	275	124
8	354	2.0	45.0	14.0	100	12.0	57	280	310
9	372	7.5	7.0	21.0	5.5	11.0	55	240	100
10	529	8.0	3.0	11.0	4.5	13.0	70	340	480

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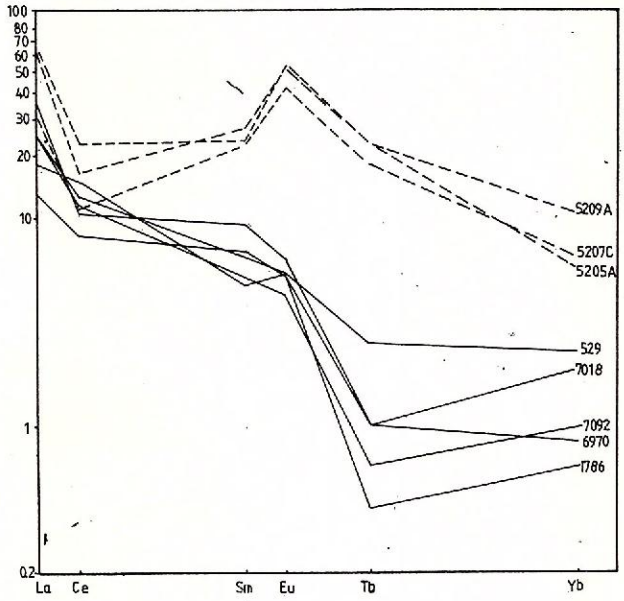


Fig. 6 – REE patterns of the analysed leucocratic rocks from the Sebeş Mountains.

In brief, the HREE depletion, the Yb content eight times lower than the chondritic values and no Eu anomalies or little (+) Eu anomalies define high Al_2O_3 trondhjemitic rocks, or generally continental trondhjemites.

Using the δ Eu=Eu/Eu^x as a criterion for the classification of the granitoids and the relationship between the source and their genesis (Zhonggang, 1982), these rocks belong to the δ Eu >0.7 group, that includes tonalites, granodiorites and biotite-bearing granites (generally called "lower crust group"), presumably being formed by partial melting process of the lower crust.

2
of the rocks

Rb	La	Ce	Sm	Eu	Tb	Yb
–	10.63	9.48	4.58	3.37	0.85	1.50
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
38	6.00	12.90	0.97	0.42	0.12	0.50

The low $^{87}Sr/^{86}Sr$ ratio of 0.702 to 0.704 (samples 515 and 5209 A)[1] could be taken as an evidence for magmas derivation from a Rb-depleted source in the lower crust, or even in the upper mantle, only if the Rb contents (between 25 and 45) were considered low enough.

5. Petrogenetic discussions

The geochemical features previously described, clearly define the igneous leucocratic rocks from the northern Sebeş Mountains as trondhjemites (as they were reviewed by Barker, 1979). Lately, these rocks were considered to be the result of a partial melting process of a granitoidic rock, at crustal levels (Lupulescu et al., in press).

Based on the features described above, we try to suggest a different genetic model, in accordance with the modern ideas explaining the origin of trondhjemites and their related tonalite-granodiorite rocks.

Mineralogically and geochemically homogeneous, the investigated rocks seem to be continental trondhjemites, according to SiO_2 vs K_2O diagram and to the high Al_2O_3 content.

The REE patterns display a remarkable resemblance to those of the Saganaga Tonalite (Northern Minnesota) (Fig. 7), indicating a rather similar model for the genetic process that produced the rocks from northern Sebeş Mountains. Hanson and Goldich (1972) suggested that the Saganaga Tonalite was

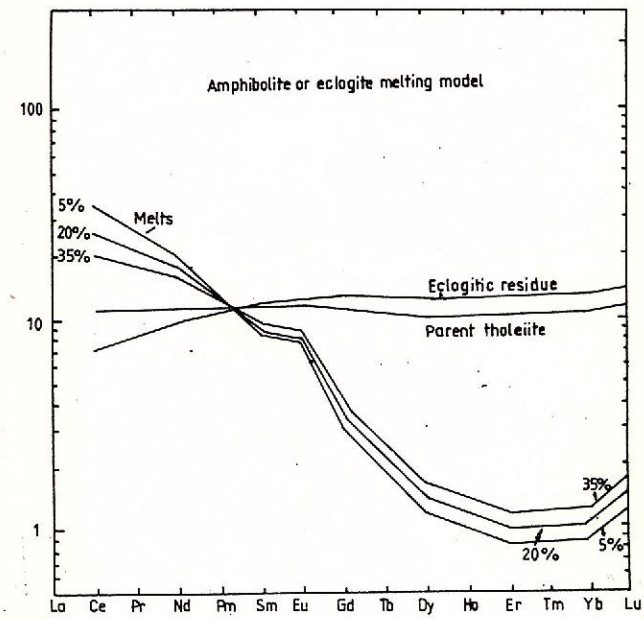


Fig. 7 – Chondrite normalized rare-earth plot showing the patterns predicted in the melting model proposed by Hanson and Goldich (1972) and Arth and Hanson (1972, 1975) to explain the origin of trondhjemites, tonalites and dacites from northeastern Minnesota.

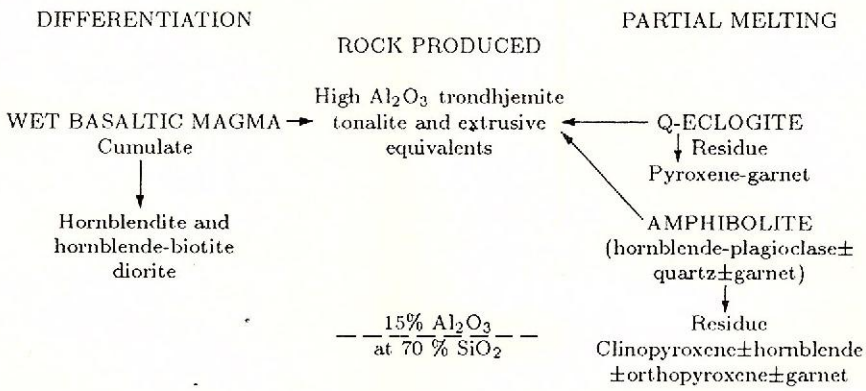
[1]Analyst: Gh. Popescu – I.F.I.N. Măgurele

Institutul Geologic al României

Table 2

No.	Sample	Pb	Cu	Ga	Cr	V	Zr	Ba	Sr
11	7098	3.5	2.0	8.5	3.5	7.0	26	280	340
12	7018	4.5	2.0	8.0	3.0	7.5	38	280	550
13	5205A	40.0	50.0	23.0	0	31.0	120	315	126
14	7092	4.0	3.0	12.0	7.0	11.5	33	310	340
15	525	32.0	14.0	12.0	5.0	9.5	36	260	550
16	524	12.0	5.5	14.0	5.5	11.0	46	240	620
17	5209A	18.0	7.0	25.0	13.0	28.0	180	455	116
18	608	16.0	2.0	8.0	3.5	8.0	36	320	500
19	7095	4.0	2.0	10.0	2.5	6.0	24	250	340
20	5208	5.0	3.0	22.0	0	31.0	95	390	113
21	536	43.0	20.0	10.5	8.5	9.0	66	270	750
22	1733A	5.0	3.0	12.0	7.0	22.0	48	410	200
23	507	22.0	4.0	10.0	5.0	11.0	60	230	630
24	7096	2.0	2.0	8.0	3.0	10.0	38	270	270
25	607	7.0	3.0	15.0	2.0	13.0	52	830	440
26	531B	22.0	2.5	10.5	3.5	9.5	46	250	500
27	6748	2.0	2.0	6.0	2.5	5.0	30	300	270
28	7099	4.5	2.0	10.0	0	2.0	14	340	150
29	1733C	6.0	2.0	13.0	3.5	8.5	38	430	320
30	6787	3.5	2.0	8.5	2.5	7.0	30	380	350
31	601	15.0	2.5	10.0	6.0	8.5	48	390	450
32	1786	6.5	3.5	13.0	3.0	7.5	38	370	250
33	515	27.0	14.0	15.0	4.0	11.0	115	320	360
34	1733B	5.5	2.0	13.0	2.5	9.0	35	380	260
35	607B	29.0	9.0	9.5	2.5	6.0	42	350	400
36	607A	15.0	2.0	9.0	2.0	3.5	34	350	700
37	678	5.0	2.5	12.0	3.0	8.0	30	370	340
38	6757	4.0	3.0	9.0	5.0	2.0	19	400	240
39	262	5.0	6.0	16.0	4.0	9.0	55	250	315
40	6970	2.0	2.0	5.5	1.5	3.0	30	380	350
41	6721	4.5	2.0	6.5	3.0	2.0	28	400	250

Amalysts:



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continued

Rb	La	Ce	Sm	Eu	Tb	Yb
–	–	–	–	–	–	–
41	4.40	7.1(	1.35	0.41	0.05	0.19
–	20.66	14.21	5.51	4.02	1.06	2.30
25	8.30	10.00	1.04	0.33	0.03	0.22
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	22.49	19.30	4.82	4.15	1.06	2.30
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
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–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
45	8.30	11.00	1.33	0.42	0.02	0.14
34	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
–	–	–	–	–	–	–
45	11.90	9.10	1.88	0.49	0.05	0.19
–	–	–	–	–	–	–

Constanţa Udrescu, Maria Stoian, I.G.R.

derived by partial melting of Q-eclogite or garnet-bearing amphibolite, at depths greater than 30 km.

Generally it is thought that such rocks could form by 5 to 35 per cent partial melting of basalts or garnet-bearing amphibolites, or mafic granulites, or Q-bearing eclogites (leaving a mafic residue), or even of some old sedimentary rocks approximating the lower crust composition.

The generation of trondhjemitic-tonalitic liquids (with high Al_2O_3 content) has been briefly summarized by Barker and Arth (1976) in a schematic diagram.

The investigated rocks show a medium-high fractionation of HREE that suggests the presence of garnet and/or hornblende as residual phases in the source. MREE depleted contents suggest their retaining by clinopyroxene, apatite and hornblende. The lack of negative Eu anomaly shows that plagioclase was absent in residue. However, the hornblende presence could repeal the plagioclase presence effects as a residue, so its presence depends on the source material nature and on the partial melting degree. These high Al_2O_3 trondhjemites have positive fractionated REE patterns that show (+) or no Eu anomalies and low HREE contents.

The low K_2O content precludes graywacke as source-material.

δ Eu>0.7 value confirms the assumption of a partial melting model implying a lower crust material, formed by amphibolites, eclogites or basalts.

The high-silicic nature of the rocks eliminates the alternative of an ultramafic or SiO_2-undersaturated parent rock. The most likely source material seems to be common amphibolites.

Hence, the trondhjemitic-tonalitic liquids could have been derived by 15-30 per cent partial melting of amphibolites at the base of a thick pile, along steep geotherms arriving at about 8 Kb (consistent with low TiO_2 content – according to Rapp et al., 1991), at a range of 850 ^{0}C–1000° C (Halloway & Burnham, 1972; Heltz,1973, 1976 in Barker, 1979), at almost 50–60 km depth (Wyllie, 1971; Lambert & Wyllie, 1972, in Barker, 1979).

Since the ascent of the melt seems to have occurred rapidly, there was not enough time for a differentiation (by fractional crystallization) in situ, or proper conditions for contamination.

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METASOMATITELE BRECIOASE ASOCIATE ZĂCĂMINTELOR DE BARITINĂ ŞI DE SULFURI DIN ZONA SOMOVA (DOBROGEA DE NORD)

Ion ÎNTORSUREANU

Institutul Geologic al României. Str. Caransebeş nr. 1, 78344 Bucureşti 32



Key words: Metasomatic rocks. Mineralization. Barite deposits. Sulfides. Metasomatism. North Dobrogea.

Abstract: *Breccia Metasomatites Associated with Baryte and Sulphide Deposits in the Somova Zone (North Dobrogea).* The hydrometasomatic processes of barytization and carbonation associated with the Jurassic (?) calc-alkaline (acid) magmatism have a wide spreading in the ore deposits in the Somova Zone. The altered rocks show frequently breccia characteristics, more or less obvious, according to the intensity and uniformity of the metasomatic substitutes. The breccia metasomatites have formed by the substitution of the sedimentary rocks mainly belonging to the Somova Formation, of Lower Triassic age, and of the rhyolites; they occur as nests, lenses, veins or columns. According to the mineralogical and petrographical composition, three main types can be distinguished: 1) baryte-bearing breccia metasomatites; 2) carbonate ± sulphide breccia metasomatites; 3) breccia metasomatites with carbonates, rhyolite elements ± sulphides. The metalliferous mineralizations are represented especially by sphalerite, galena, chalcopyrite and pyrite. The presence of the breccia metasomatites and altered rocks (carbonated, barytized, silicified, adularized etc.) are in favour of prospecting and exploration works.

Introducere

Procesele hidrometasomatice de baritizare şi de carbonatare au avut o largă dezvoltare în cadrul zăcămintelor de baritină şi de sulfuri din zona Somova, iar rocile transformate se caracterizează printr-o mare variaţie de aspecte structural-texturale, mineralogice şi petrografice.Studiul rocilor afectate sau formate prin aceste procese hidrometasomatice prezintă o importanţă esenţială pentru cunoaşterea genezei zăcămintelor respective, în prezent controversată.

Rocile brecioase, pe care noi le denumim metasomatite brecioase, au fost semnalate şi de alţi autori, dar le-au atribuit origini diferite. Astfel, Gurău şi Gridan (1975) sunt de părere că breciile din cariera Bechir (Mineri) s-au format prin tectonizarea baritinei metasomatice, luînd aspectul unor "diapire". Aceleaşi roci sunt interpretate de Baltreş (1982, raport) Baltreş şi Mirăuţă (1987, raport),ca "baritină sedimentară", formată prin "alunecările submarine" ale unor ipotetice strate ce ar fi conţinut acumulări primare de baritină sedimentară. De asemenea, Berbeleac (în Berbeleac et al., 1985, raport) şi Berbeleac (1988) arată că "breciile carbonatice cu sulfuri ± baritină" din unele zăcăminte de la Somova (Cortelu, Gura Zmeului, Movila Săpată)" reprezintă concentraţii de origine vulcanogen-sedimentară.

Cercetările noastre (Întorsureanu, în Nedelcu et al., 1989, raport), efectuate pentru elaborarea Hărţii metalogenetice a României scara 1 : 50.000, foaia Somova-Cataloi, ne-au permis să demonstrăm natura hidrometasomatică a acestor roci brecioase. În nota de faţă prezentăm datele care relevă originea metasomatică a acestor produse, fapt ce argumentează geneza hidrotermal-metasomatică a mineralizaţiilor de baritină şi de sulfuri din zona Somova. Această concluzie concordă, de altfel, cu majoritatea cercetărilor anterioare (Ianovici et al., 1957; Giuşcă et al., 1967; Stiopol et al., 1976 etc.).

Date geologice

La alcătuirea geologică a zonei Somova participă formaţiuni sedimentare şi roci magmatice.

Formaţiunile sedimentare sunt preponderent de vîrstă triasică şi au fost investigate recent cu foraje structurale. Studiul sedimentologic şi litologic al acestor depozite (Baltreş, 1982, raport; Baltreş şi Mirăuţă,

Institutul Geologic al României

1987, raport) a condus la separarea unor formaţiuni distincte, a căror vîrstă este cuprinsă între Spathian şi Carnian. Aceste formaţiuni alcătuiesc o structură anticlinală majoră (orientată VNV-ESE) denumită "anticlinalul Somova". In axul anticlinalului aflorează formaţiunea de Somova, de vîrstă spathiană, caracterizată printr-un facies turbiditic, fiind interpretată ca un depozit de mare adîncime. Litologic este constituită din alternanţe de calcare, calcare argiloase sau argilite, toate de culoare cenuşie-negricioasă şi cu texturi rubanate foarte pronunţate.

In formaţiunea de Somova apar uneori (cariera Bechir, Gura Zmeului etc.) concreţiuni carbonatice cu forme discoidale, avînd diametrul pînă la 10-12 cm. Baltreş (1982, raport), Baltreş şi Mirăuţă (1987, raport) au emis ipoteza că aceste concreţiuni au fost alcătuite iniţial din baritină sedimentară, care ulterior ar fi fost substituită, în mare parte, prin calcit. Observaţiile microscopice şi analizele chimice efectuate (Intorsureanu, în Nedelcu et al., 1989, raport) relevă compoziţia exclusiv carbonatică a acestor concreţiuni şi absenţa totală atât a baritinei şi a oxidului de bariu, cît şi a proceselor de substituţie. In consecinţă, considerăm că aceste concreţiuni nu constituie o dovadă pentru ipoteza prezenţei baritinei sedimentare în formaţiunea de Somova.

Rocile magmatice sunt reprezentate prin vulcanite bazice sau acide şi au fost cercetate de Stiopol et al. (1974, 1975, rapoarte), Savu et al. (1985), Ştefan şi Roşu (în Berbeleac et al., 1985, raport; Nedelcu et al., 1989, raport).

Vulcanitele bazice, de vârstă triasică, au chimism calco-alcalin şi formează curgeri de bazalte în facies de pillow-lava, iar subordonat apar anamesite şi dolerite.

Vulcanitele acide, mai larg dezvoltate, aparţin aliniamentului Isaccea-Somova-Mineri şi sunt localizate, de obicei, în axul anticlinalului Somova sau pe flancul de NE al acestuia. Ele sunt reprezentate prin riolite şi riodacite calco-alcaline. Aceste roci par a avea, cel mai frecvent, forme de corpuri neregulate, filoane, apofize sau silluri ce străpung formaţiunea de Somova şi vulcanitele bazice. Berbeleac (1988, p.129) admite prezenţa unor domuri de riolite, însă din interpretarea datelor, obţinute prin foraje şi lucrări miniere subterane, nu rezultă asemenea structuri (Braşov, 1988; Braşov et al., 1989, proiecte nepublicate; Nedelcu et al., 1989, raport).

Este foarte probabil ca magmatismul acid să fie generat prin subducţia marginii plăcii Eurasiatice sub microplaca Moesică în intervalul Carbonifer inferior-Triasic superior (?) când s-au format şi granitele alcaline şi calco-alcaline din unitatea de Măcin (Intorsureanu, 1987; Intorsureanu et al., 1989), iar magmatismul bazic să fie legat de un bazin marginal.

Mineralizaţii

Mineralizaţiile de baritină şi de sulfuri (blendă+galenă+ calcopirită+pirită) aparţin cîmpului metalogenetic Somova-Mineri (Cîşla), orientat VNV-ESE, cu lungimea de aproximativ 8 km şi lăţimea pînă la 1,5 km. Acesta include mai multe zăcăminte, dintre care unele au fost exploatate (Dealul Carierii, Bechir, Trifan, Dobrişan), iar altele sunt în fază de exploatare (Cortelu) sau de explorare (Movila Săpată, Gura Zmeului).

Majoritatea cercetătorilor (Ianovici et al., 1957; Giuşcă et al., 1967; Gurău şi Gridan, 1975; Stiopol et al., 1976 ş.a.) au considerat că aceste mineralizaţii sunt de origine hidrotermal- metasomatică, fiind asociate intruziunilor de riolite. Ulterior s-au formulat alte două ipoteze genetice : 1) sedimentară, mineralizaţiile considerându-se de tipul Mississippi Valley-Alpin (Vlad, 1978); 2) vulcanogen-sedimentară, produsele fiind încadrate într-un model genetic asemănător tipului Kuroko (Berbeleac, în Berbeleac et al., 1985, raport; Berbeleac, 1988). O variantă a ipotezei vulcanogen-sedimentare, dar în condiţii de mare adâncime,au propus Baltreş şi Mirăuţă (1987, raport) pentru a explica presupusele "acumulări stratiforme de baritină".

Cercetările recente (Nedelcu et al., 1989),coroborate cu rezultatele obţinute prin lucrările de explorare (Braşov et al., 1988; 1989, proiecte nepublicate),susţin însă ipoteza genezei hidrotermal- metasomatice pentru mineralizaţiile din zona Somova, în sprijinul căreia se aduc noi argumente.

Transformări hidrometasomatice

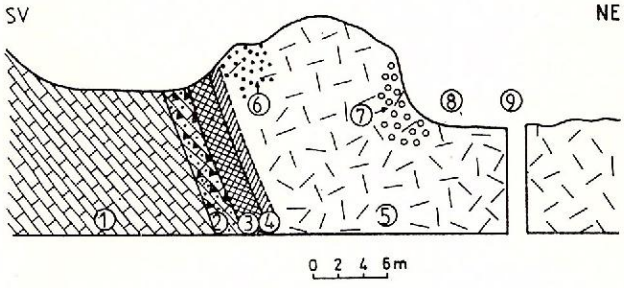
In stadiul postmagmatic s-au desfăşurat intense procese hidrometasomatice,care au afectat aproape în totalitate corpurile riolitice şi în mare parte rocile înconjurătoare situate în imediata vecinătate a corpurilor de minereu. Deşi prezenţa transformărilor hidrometasomatice a fost remarcată de majoritatea cercetătorilor anteriori, la data actuală nu dispunem de un studiu detaliat al rocilor alterate, care ar fi deosebit de util pentru dirijarea lucrărilor de prospecţiuni şi explorări.Intrucât studiul acestor roci transformate depăşeşte scopul lucrării de faţă, vom face doar câteva observaţii asupra lor.

Rocile înconjurătoare, aparţinând formaţiunii de Somova, au fost intens afectate, aşa cum rezultă şi din unele studii anterioare (Ianovici et al., 1957; Giuşcă et al., 1967; Stiopol et al., 1975, raport). Aspectul cel mai frapant, ce se remarcă în cariere,este dat de culoarea cenuşie-deschisă sau galbenă-cărămizie şi textura masivă ori brecioasă a rocilor transformate, în contrast cu calcarele formaţiunii de Somova (paleosom), de culoare cenuşie-negricioasă şi textură rubanată foarte

Institutul Geologic al României

evidentă. De asemenea a fost afectată, pe scară mai redusă (în versantul nordic al Dealului Carierii) şi formaţiunea calcarelor nodulare, de vârstă spathiană-anisian-inferioară.

Procesele metasomatice au determinat, în rocile înconjurătoare, apariţia unor aureole de transformare, cu forme aproximativ ovoidale şi cu dimensiuni ce ajung la ordinul zecilor sau al sutelor de metri. Aceste aureole sunt alcătuite, îndeosebi, din roci carbonatate, baritizate, silicifiate, adularizate, metasomatite brecioase ± mineralizaţii de baritină ± sulfuri. Datele de observaţie arată că procesele metasomatice au fost controlate, în general, de aceleaşi sisteme de fracturi care au favorizat procesele metalogenetice. Astfel, controlul structural este evident în zona fracturii dintre Dealul Carierii şi Bechir cu poziţia N 40^0V/70^0NE şi lungimea de aproximativ 650 m, care a fost urmărită cu lucrări miniere subterane (fig.). Fractura a fost mineralizată cu sulfuri de Pb, Zn ± Cu, iar spre suprafaţă prezintă o zonă de oxidare care a fost cercetată de Giuşcă et al., (1967). Roca din culcuşul fracturii este brecifiată şi slab carbonatată, în timp ce roca situată în acoperiş este puternic carbonatată, baritizată ± adularizată şi conţine, pe alocuri, relicte din paleosom.



Secţiune geologică în Dealul Carierii (Mineri). 1, calcare rubanate, cenuşii-negricioase, netransformate; 2, calcare rubanate, brecifiate şi slab carbonatate; 3, fractură mineralizată cu sulfuri; 4, zonă limonitizată ± malachit; 5, roci carbonatice masive albicioase ± baritină ± sulfuri; 6, adularizări; 7, metasomatite brecioase cu baritină; 8, baritină metasomatică exploatată în carieră; 9, puţ de explorare.

Prezenţa rocilor carbonatate şi baritizate pe scară largă în jurul corpurilor de minereu este, după Kurec (1954), o trăsătură importantă a zăcămintelor de Pb şi Zn, de temperaturi medii sau scăzute, afiliate proceselor postmagmatice. În consecinţă considerăm că încadrarea mineralizaţiilor din zona Somova la tipul Mississippi Valley-Alpin (Vlad,1978) este foarte discutabilă.

Metasomatite brecioase

Rocile care rezultă în urma substituţiil[illegible] metasomatice se caracterizează frecvent, după [illegible] un mai menţionat, prin prezenţa unor variate aspecte brecioase. Principalele particularităţi ale metasomatitelor brecioase din zona cercetată sunt: 1) s-au format prin substituţia rocilor înconjurătoare ori a riolitelor; 2) alcătuiesc corpuri cu aspect de cuib, lentile, filoane sau coloane cu dimensiuni variate; 3) sunt localizate în apropierea zonelor mineralizate sau constituie ele însele corpuri de minereu; 4) caracterul brecios poate fi mai mult sau mai puţin evident; 5) aspectul eterogen al rocii metasomatice, lipsită practic de fenomene cataclastice.

După compoziţia mineralogică şi petrografică se deosebesc trei tipuri principale de metasomatite brecioase: 1) cu baritină; 2) cu carbonaţi ± sulfuri; 3) cu carbonaţi, elemente de riolite ± sulfuri.

Metasomatitele brecioase cu baritină apar în peretele de NE al carierei Bechir, sub formă de cuiburi sau lentile cu dimensiuni metrice, precum şi în Dealul Carierii (fig.), unde formează o zonă neregulată. Corpurile de metasomatite sunt localizate în formaţiunea de Somova, puternic carbonatată ± baritizată. Contactele dintre metasomatitele brecioase şi roca gazdă sunt de obicei sinuoase şi gradate, dar uneori se observă şi contacte nete, marcate de vinişoare metasomatice, de compoziţie carbonatică, identică cu cea a matricei metasomatitului brecios. Macroscopic, se caracterizează prin textura brecioasă foarte evidentă, îndeosebi pe suprafeţele de alteraţie supergenă, datorită levigării parţiale a matricei carbonatice, în timp ce "elementele" cu baritină rămîn în relief (pl. I, fig. 1). "Elementele" de baritină au culoarea cărămiziegălbuie, neuniformă, dimensiuni foarte variate, care ajung până la 15–20 cm, cu contururi angulare şi sunt prinse într-o matrice fină de culoare cenuşie deschisă. În cariera Bechir, un cuib de metasomatit brecios cu baritină conţine o reţea de filonaşe de baritină albă, depusă într-un stadiu ulterior, după formarea metasomatitului brecios.

În secţiuni subţiri se constată că "elementele" metasomatitului brecios sunt de fapt cuiburi metasomatice, constituite preponderent din baritină, la care se pot adăuga, în cantităţi variate, calcit ± dolomit ± cuarţ ± pirită. Prin oxidare, pirita se transformă în hidroxizi de fier care imprimă rocii o culoare cărămizie pătată. Baritina formează cristale prismatice alungite, cu dimensiuni reduse, alcătuind adesea agregate cu aspect penat sau radiar. Calcitul apare sub formă de cristale xenomorfe, bine dezvoltate şi este substituit de baritină. Dolomitul este subordonat, apărînd cel mai frecvent în cristale mici, rombice. Matricea, cu structură micro-sau criptocristalină este alcătuită preponderent din carbonaţi la care se adaugă, subordonat, baritina sub formă de mici cristale izolate sau agregate radiare. Contactul dintre "elementele" sau cuiburile de baritină şi matrice este foarte sinuos (pl. I, fig. 2), fiind

Institutul Geologic al României

dat de agregatele de baritină care substituie matricea (pl. II, fig. 1). Se observă de asemenea vinişoare de baritină care traversează matricea (pl. II, fig. 2).

Datele de mai sus dovedesc în mod clar că aceste roci sunt de origine metasomatică. Această concluzie diferă de unele interpretări anterioare, după care rocile brecioase cu baritină s-au format fie prin mişcări tectonice postmineralizare, care au dus la deformarea baritinei metasomatice (Gurău şi Gridan, 1975), fie prin fragmentarea, în condiţii marine, a unor ipotetice strate de baritină sedimentară şi apoi resedimentarea blocurilor de minereu (Baltreş, 1982, raport; Baltreş şi Mirăuţă, 1987, raport).

Metasomatitele brecioase cu carbonaţi ± sulfuri au o răspândire mai largă fiind deschise în cariere (Cortelu, Dealul Carierii) şi prin lucrări miniere subterane (Gura Zmeului, Movila Săpată). Ele s-au format de asemenea prin substituţia rocilor din formaţiunea de Somova. Aspectele texturale sunt variate în funcţie de natura paleosomului (carbonatic sau argilitic) şi de intensitatea substituţiilor. Într-un prim stadiu, paleosomul este străbătut de o reţea de vinişoare anastomozate de calcit alb, iar elementele relicte, angulare şi omogene, par a proveni dintr-un calcar argilos sau din argilite, slab afectate (dolomitizate ?). În stadiul mai avansat, datorită substituţiilor neuniforme,apar zone neregulate, cu dimensiuni decimetrice, formate din roci carbonatice cu textură masivă, de culoare cenuşie deschisă,dar se mai păstrează şi zone cu filonaşe de calcit din primul stadiu (pl. II, fig. 3). În ultimul stadiu de substituţie roca prezintă o textură masivă, omogenă, în care aspectele brecioase sunt estompate în totalitate.

Aceste metasomatite brecioase sunt frecvent mineralizate cu sulfuri (Pb,Zn,Cu), aşa cum se remarcă în cariera Cortelu şi în unele lucrări miniere subterane (Gura Zmeului, Movila Săpată). La Cortelu breciile mineralizate cu sulfuri sunt localizate în acoperişul corpului riolitic intens hidrotermalizat şi alcătuiesc zone discontinue sub formă de cuiburi sau cu contururi neregulate şi cu dimensiuni care ajung la ordinul zecilor de metri. Aspectul brecios prezintă variaţii datorită impregnaţiilor şi cuiburilor de sulfuri, neuniform răspândite în masa rocii. Cuiburile de sulfuri au contururi sinuoase şi sunt alcătuite din blendă, galenă, pirită, calcopirită. Partea sterilă este formată preponderent din carbonaţi de neoformaţie, dar uneori se mai păstrează şi relicte din paleosom.Într-o etapă finală breciile mineralizate sunt uneori traversate de filonaşe de calcit alb, cu grosimi centimetrice.

La microscop aceste metasomatite brecioase apar constituite din calcit şi dolomit, ca minerale dominante de neoformaţie,la care se adaugă, în proporţii variate, baritină, cuarţ şi sulfuri. În paragenezâ cu sulfurile apare şi adularul (Cortelu).

Analiza chimică parţială a unei probe de metasomatit brecios cu carbonaţi, provenită din Dealul Carierii, arată un conţinut de 29,72 % CaO şi 20,05 % MgO, în timp ce paleosomul este caracterizat printr-un conţinut de 47,70 % CaO şi 0,30 % MgO (Intorsureanu, în Nedelcu et al., 1989, raport).

Observaţiile şi datele prezentate mai sus arată că metasomatitele brecioase cu carbonaţi ± sulfuri au fost generate prin procese de substituţie. Amintim că Berbeleac (1988) atribuie acestor roci o geneză vulcanogen-sedimentară şi consideră că s-au format în cupola unui ipotetic dom riolitic, asemănător tipului Kuroko.

Metasomatitele brecioase cu carbonaţi, elemente riolitice ± sulfuri au fost întâlnite la Cortelu, Gura Zmeului şi Mineri. Prezenţa constantă a elementelor relicte de compoziţie riolitică sugerează că metasomatitul s-a format prin substituţia vulcanitelor acide. Aceste metasomatite se prezintă sub formă de coloane sau zone filoniene, cu grosimi de ordinul zecilor de metri, care pe înclinare au fost explorate până la adâncimea de 200–250 m.

Formarea acestor roci, denumite în practica de şantier "brecii riolitice" (Braşov et al., 1989), a fost controlată de fracturile direcţionale cu poziţia VNV-ESE/70^0-75^{0}SV, care au afectat corpurile riolitice şi care au constituit căi de acces pentru fluidele postmagmatice. Frecvent sunt mineralizate cu sulfuri ce apar ca impregnaţii distribuite neuniform în rocă. La Cortelu aceste metasomatite brecioase trec spre forme de stockwork, având în ansamblu aspectul unei coloane mineralizate considerată de Stiopol et al. (1976) ca o structură tipică ce "breccia pipe".

Macroscopic sunt caracterizate prin relicte riolitice intens alterate, de culoare verzuie închisă şi dimensiuni până la 1–2 cm, incluse într-o masă de carbonaţi de culoare albă. La microscop roca este alcătuită predominant din calcit ± dolomit+ magnezit, cu structuri heterogranoblastice, la care se adaugă, subordonat, sulfuri, baritină, cuarţ. Relictele sunt constituite din agregate criptocristaline, carç par a proveni dintr-o rocă riolitică.

Concluzii

Din datele prezentate rezultă că metasomatitele brecioase sunt localizate în aureolele rocilor de transformare hidrotermală şi s-au format prin substituţia rocilor înconjurătoare sau a riolitelor. Aspectul brecios este uneori foarte evident, iar alteori mai estompat datorită substituţiilor neuniforme. Aceste roci se prezintă sub formă de cuiburi, lentile, filoane sau coloane, localizate în zone de fracturi. După compoziţia mineralogică şi petrografică se disting trei tipuri principale: metasomatite brecioase cu baritină,

Institutul Geologic al României

metasomatite brecioase cu carbonaţi ± sulfuri şi metasomatite brecioase cu carbonaţi, elemente riolitice ± sulfuri. Metasomatitele brecioase, ca şi mineralizaţiile de baritină şi de sulfuri din zona Somova, sunt asociate genetic magmatismului acid calco-alcalin, care a fost generat, foarte probabil, printr-un proces de subducţie.

Sub aspect practic este de reţinut că prezenţa metasomatitelor brecioase şi a rocilor transformate constituie un indice favorabil pentru lucrările de prospecţiuni şi de explorări. Perspective economice prezintă mai ales metasomatitele brecioase cu carbonaţi, elemente riolitice ± sulfuri, care se dezvoltă adesea sub formă de filoane sau coloane (stockwork-uri).

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Fig. 1 – Metasomatite brecioase cu baritină în formă de cuib (5x6 m). Cariera Bechir.

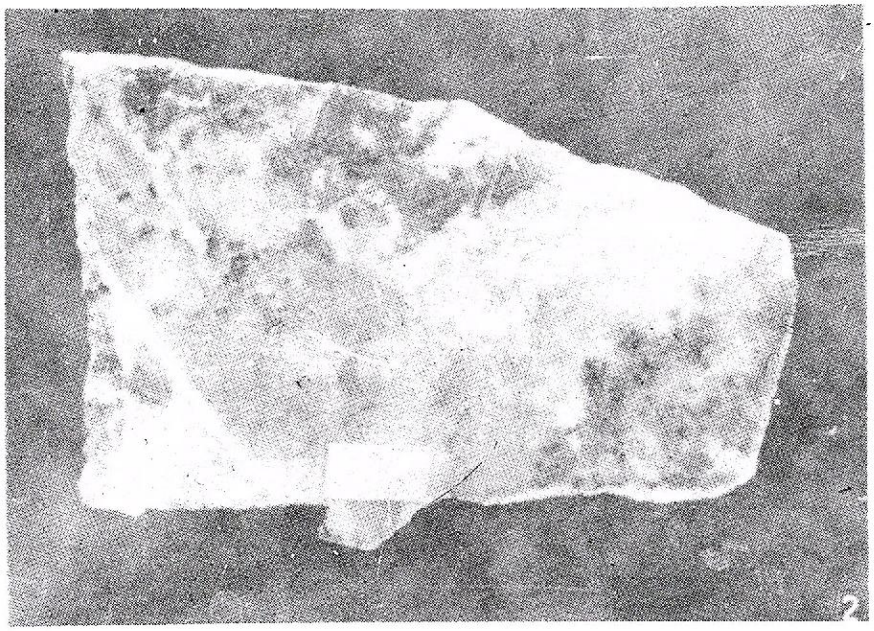

Fig. 2 – Contacte metasomatice între cuiburi de baritină (cenuşiu închis) şi matrice (cenuşiu deschis). Suprafaţă lustruită. Dealul Carierii.

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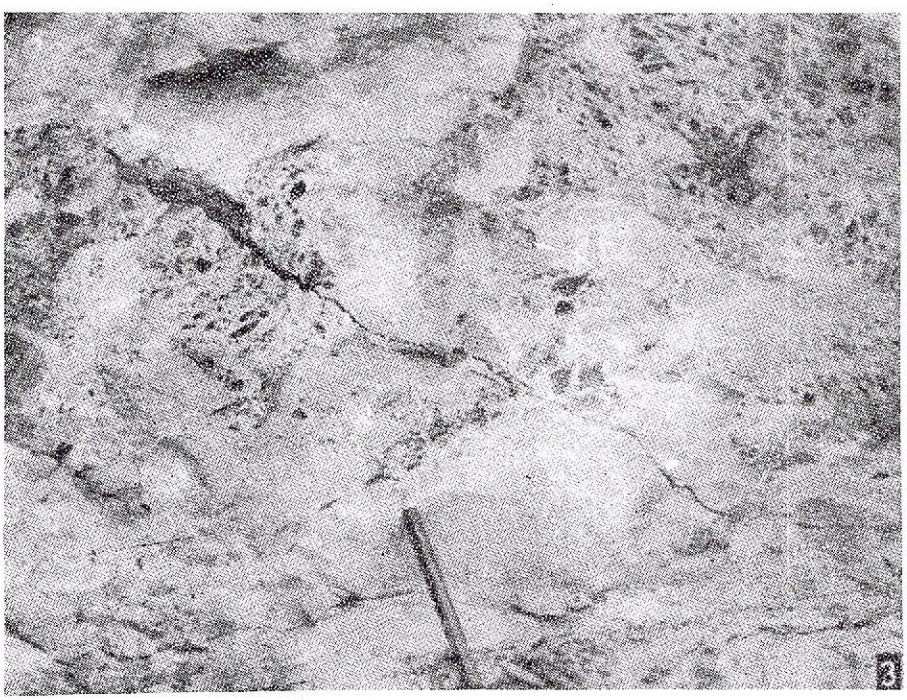

Fig. 1 – Limită de substituţie a matricei de către cuiburi de baritină. N+; x 60, Dealul Carierii.

Fig. 2 – Filonaş de baritină care străbate matricea metasomatitului brecios cu baritină. N+; x 60, Cariera Bechir.

Fig. 3 – Metasomatit brecios cu carbonaţi; se observă substituţia neuniformă şi relicte din paleosom (cenuşiu închis). Cariera Cortelu.

Institutul Geologic al României

Rom. J. Petrology, 1994, 76, p. 57–65

METAMORPHIC AND DEFORMATION PROCESSES IN THE OPHIOLITES OF THE SEVERIN NAPPE AND THE ASSOCIATED STRATIFORM COPPER MINERALIZATION (SOUTH CARPATHIANS)

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Key words: Ophiolite. Deformation. Anchimetamorphism. Mineralization. Stratiform deposits. Absolute age. K/Ar. Mehedinţi Plateau. South Carpathians.

Abstract: The Severin Nappe from the Mehedinţi Plateau was affected by several deformation and very low-grade metamorphism processes. The Late Kimmerian anchimetamorphism – the major one – affected mostly the lower part of the nappe, especially the Azuga Beds and little of the base of the Sinaia Beds. The age of the anchimetamorphism (K/Ar) is estimated to ca 130 Ma. The anchimetamorphism is the result of a very low-grade metamorphism which manifested at low pressure and temperature ranging between 260^0 and 345^0 C. These values indicate the stability limits of the prehnite bearing parageneses, prehnite being the marker which occurs also in the mineral assemblages from the anchimetamorphic rocks of the Severin Nappe. The anchimetamorphism influence on the Cyprus-type copper stratiform mineralization, located in the basalts megaslab manifested in two ways: (a) recrystallization and concentration of the copper mineralization in more or less compact ore bands, which alternate with schistous rocks bands and (b) partial disperssion of mineralization on fissures from the surrounding basic rocks.

Introduction

The sedimentary deposits and the ophiolitic rocks included into these deposits, the Azuga Beds and partially the Sinaia Beds respectively, which constitute the Jurassic-Neocomian olistostrome of the Severin Nappe in the Mehedinţi Plateau, were influenced by several processes of deformation and very low-grade metamorphism. Among them the Late Kimmerian anchimetamorphism and the associated deformation phenomena were the most significant processes. The effect of the metamorphism were pointed out by Codarcea (1940), who mentioned that the Severin Nappe consists of phyllites intruded by ultramafic rocks. Recently, these phenomena have been studied by Savu (1985), Savu et al. (1985 a, 1985 b: 1986) and Popescu et al. (1985), who have emphasized different aspects.

The present paper deals with the effects of these processes on ophiolites and on the Baia de Aramă copper stratiform mineralization associated with the basic rocks (see Savu et al., 1989)[1].

[1]The conclusions are based on those presented by the authors in other papers mentioned here as well as in the report from March 1989 (Savu et al., 1989).

Geological Evolution of the Area and Succession of the Metamorphic and Deformation Processes

The Severin Nappe has a paraautochthon position. It lies between the Danubian Autochthon and the Bahna Outlier, a vestige of the Getic Nappe (Codarcea, 1940)[2], and extends from the Motru Sec Valley up to Brza Polanka (Yugoslavia). The nappe mostly consists of a Jurassic-Neocomian olistostrome (Savu, 1985), represented by the Azuga Beds ($J_2 - J_3$) and partially by the Sinaia Beds (Cr_1) with olistoliths of basic and ultramafic ophiolites, crystalline schists and granites. The olistostrome originates in the Severin trough.

The geological evolution of the Mehedinţi Plateau area developed during two main stages: (1) Jurassic-Lower Cretaceous stage and (2) Upper Cretaceous stage (Savu, 1985; Savu et al., 1985 a). At the beginning of the first stage, probably in the Liassic, the Carpathian Ocean occurred by spreading processes (Savu, 1985), whose crust was influenced by an ocean-

[2]For cartographic details see also the maps of the papers elaborated by Savu (1985) and Savu et al. (1985 a,b; 1986).

Institutul Geologic al României

floor metamorphism. During the Middle and Upper Jurassic the oceanic zone began to close, due to the subduction phenomena. The unconsumed oceanic crust was broken into numerous slabs of basic and ultramafic rocks, obducted in the Severin trough and included into the mentioned Jurassic-Neocomian olistostrome. The olistostrome represents a subduction complex, from which the Severin Nappe was torn off at the end of the Upper Cretaceous.

In the basaltic rocks megaslab lying between Obîrşia Cloşani and Ponoare a Cyprus-type stratiform mineralization, with several levels of pyrite and chalcopyrite, is intercalated (Savu et al., 1985 b). The mineralization bands alternate with schistous rocks bands, consisting of quartz, calcite and sericite. The mineralization probably represents the distal (marginal) facies of the primary ore deposit formed on the ocean floor, most of it being consumed by subduction concomitantly with the oceanic crust, or eroded after the emplacement of the Severin Nappe.

Regional metamorphic processes occurred in the Severin Nappe, which influenced both the pelitic matrix of the olistostrome and the olistoliths of basic and ultramafic rocks, the associated copper mineralization, inclusive. The metamorphism corresponds partly to the Jurassic metamorphism in the Western Alps (Frey et al., 1974) and partly to the Lower Cretaceous one in the other Alps, as well as to the Alpine metamorphism in the NE Hungary (Arkai, 1983).

During the second evolution stage of the Mehedinţi Plateau area, the subduction process was resumed in a new cycle (Savu et al., 1985 a); in the beginning it was accompanied by weak tectonic movements which finally culminated with the Laramian orogenesis. Under conditions of the weak tectonic movements, the Senonian flysch formed which contains olistoliths from the older formations of the Danubian Autochthon. A calc-alkaline arc volcanism, with basalts and basaltic andesites, developed concomitantly (Savu et al., 1987). The emplacement of the Cerna and Severin nappes took place, the first one as an olistonappe, followed by the eastwards thrust of the Getic Nappe, which overlies all the previous structures. There followed the refolding of the whole tectonic edifice, which gave rise to synforms and antiforms (Savu et al., 1985 a). The Bahna Outlier and the underlain Severin Nappe preserved their present-day form being situated into the Bahna synform, whose structure was established on the basis of drilling data.

Age of the Main Anchimetamorphic and Deformation Process

Radiometric (K/Ar) datings of the anchimetamorphism were presented by Lemne et al. in 1983, who indicated values of 143.7 – 127.9 ± 5Ma, indicating

Table

Primary K/Ar datings obtained on phyllitic rocks from the Severin Nappe

No.	Sample	Rock denomination and location	K %	$^{40}Ar_{radx}$ 10^{-10} mol/g	^{40}Ar rad %	Apparent age Ma
1	27	Phyllite Malareca Valley	2.11	3.5361	62.94	94.1±4.0
2	34	Phyllite (olistolith), Vârtopu Brook	3.03	7.3851	73.37	13.3±5.4
3	29	Phyllite, Coşuştea Valley	2.94	5.5787	72.16	106.2±4.2
4	26	Phyllite, Malareca Valley	2.68	6.4844	20.79	134.4±15.1
5	36	Phyllite in basalts, Brebina Valley	1.69	1.9523	55.14	65.4±3.3
6	167*	Phyllite, Valea Verde	2.67	6.1368	86.78	127.9±4.9
7	145	Phyllite in basalts. Dragu Brook	0.93	2.4110	82.14	143.7±5.5

*Samples no. 167 and 145 have been taken over from the report elaborated by Lemne et al. (1983).

Institutul Geologic al României

the Late Kimmerian movements. Recent researches effectuated by us brought further evidence in this respect.

Radiometric (K/Ar) datings were made on the whole rock by means of the isotopic dilution, using ^{38}Ar as a tracer. Radiogenic argon was measured whith the mass spectrometer AEI – MS – 20 in a static regime and the potassium content was determined by the flamephotometric method. The desintegration constants recommended by Steiger and Jäger (1977) were used. The K/Ar apparent ages calculated according to the closed system hypothesis are rendered in Table, with a complex analytical error, determined at the thrust level of 68 per cent, after Cox and Dalrymple's formula (1967). According to the closed system hypothesis, for the individual age datings no transformation occurred since the beginning of the event which is to be dated: consequently, radiogenic argon represents the total argon corrected for the content of contaminating argon with an isotopic composition equal to that of the atmospheric argon. It is a rigid system, but in reality partial losses of radiogenic argon could occur, as shown by the decreasing of the apparent age values. However, a violation of the closed system hypothesis is allowed in the following two cases:

determined by its intersection with the ordinate represents the initial argon ratio (Fig. 1 a).

The existence of an argon inherited in a constant amount which is emphasized representing the samples, in a correlation diagram $^{40}K - {}^{40}Ar_{rad}$ (Fig. 1b), by the isochrone shifting with a distance on the y axis equal to the external radiogene argon. Shafiqullah and Damon (1974), discussing the interpretations that can result from different physical models, considered the concordant radiometric ages obtained from the two systems of representation as a further index of safety. Taking into account all this, samples 34(2), 26(4), 167(6), 145(7) are situated on an isochrone, whose gradient calculated after York's (1969) formula indicated, in the first case, a radiometric age of 129.7 ± 6 Ma with a ratio of the initial argon ($^{40}Ar/^{36}Ar$=333) close to the ratio of the atmospheric argon ($^{40}Ar/^{36}Ar$=295.5) and, in the second case, an isochrone of 128.9 ± 5.1 Ma, with an unsignificant radiogenic argon excess of 0.197 x 10^{-10}mol/g. This age is close to that of the Atlantic opening (135–130 Ma).

Within the limits of the determination errors, the age values obtained by means of two systems of representation are practically identical, and they can be regarded as significant in geological respect. Samples

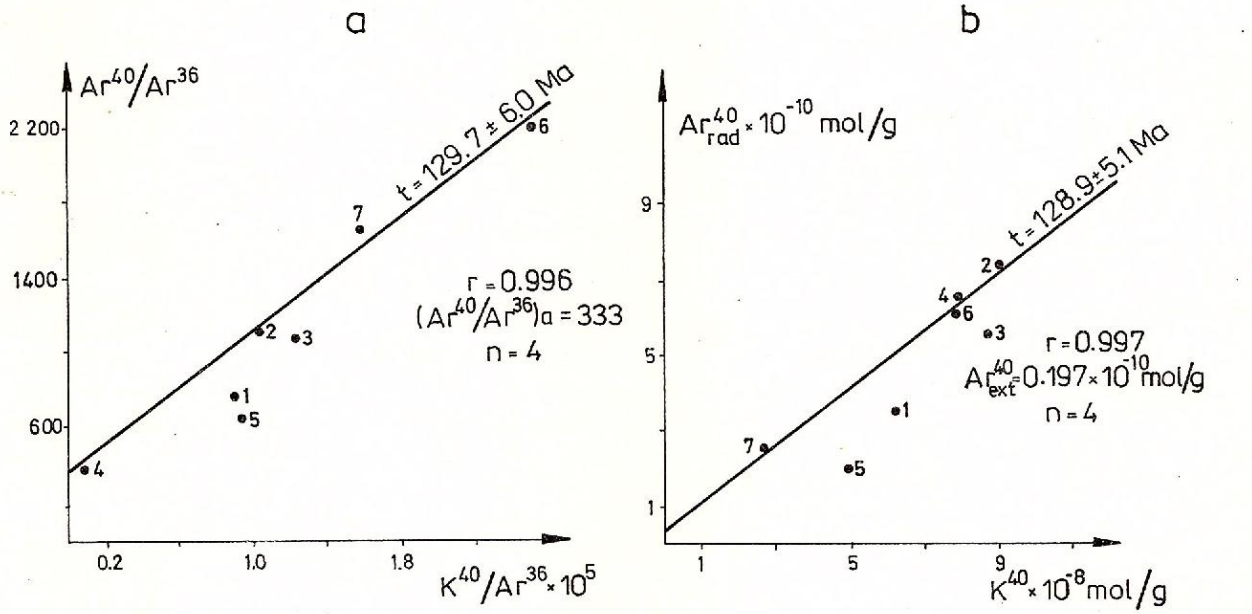

Fig. 1 – Isochrones for establishing the age of the anchimetamorphism.

The existence of an initial argon with a constant $^{40}Ar/^{36}Ar$ ratio for a suite of cogenetic samples which, represented in a system of $^{40}K/^{36}Ar - (^{40}Ar/^{36}Ar)_{tot}$ coordinate, forms an isochrone whose slope is proportional to the age of the sample suite and the segment

27(1), 29(3) and 36(5) (Table) show partial losses of radiometric argon, due to the rocks refolding during the Laramian movements. They are situated below the isochrones (Fig. 1).

Institutul Geologic al României

Structural Aspects of the Anchimetamorphic Rocks

The processes which generated deformations and low-grade metamorphism in the geological formations in the lower part of the Severin Nappe are, as follows: ocean-floor metamorphism, friction of the convergent plates during subduction, dynamo-thermal metamorphism in the base of the obducted ophiolite slab, Late Kimmerian anchimetamorphism, Severin Nappe thrust, Getic Nappe thrust and Late Laramian final refolding. Some of the deformational elements were obliterated by the subsequent ones and others were only partially imprinted in the rocks. These aspects will be presented in the succession of the geological formations.

The pelitic matrix of the Jurassic-Neocomian olistostrome included a primary foliation S_1, approximately corresponding to the bedding S_0, in which the lineation L_1 was imprinted. A lineation L_2 and a cleavage plane S_2 are perpendicular to L_1 (Fig. 2). The first two structural elements were formed during the Late Kimmerian anchimetamorphism and the other probably occurred during more recent movements.

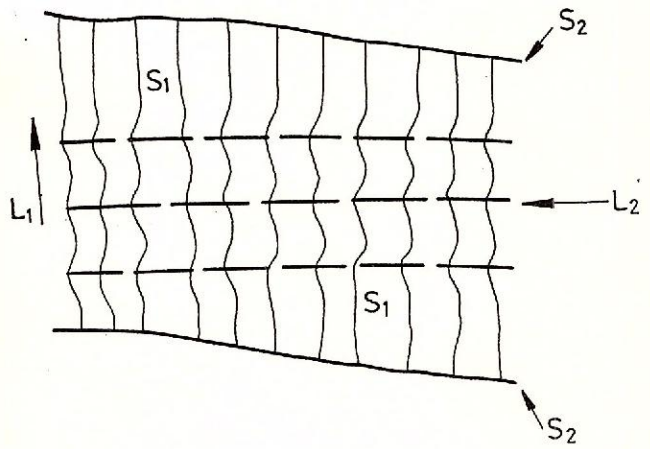

Fig. 2 Structural elements in the matrix of the Jurassic-Neocomian olistostrome (Azuga Beds) in the spring area of the Malareca Valley, under the Ocnei Hill, in the southeastern flank of the Bahna synform.

The ultramafic rock olistoliths were almost entirely serpentinized, the water necessary for the alteration of the primary minerals coming from the pelitic matrix of the olistostrome. Olistoliths are flattened and elongated like any pebble (Cloos, 1946), phenomena which became more obvious with the progress of the serpentinization process, the rocks being ever more plastic, thus increasing their volume. Some of the ultramafic olistoliths were phacoidized, others elongated, stretched and divided into smaller lenses, e.g. those in the spring area of the Coşuştea Valley. The flattening of the olistoliths is concordant with the foliation in the olistostrome matrix. On the margins, the lenses are intensely laminated (Pl. I, Fig. 2), a process determined by mechanical reaction between the two rock types with a different competence to deformation: sedimentary rocks and serpentinites. These are the major structural elements. They formed during the Late Kimmerian anchimetamorphism and obliterated almost entirely the previous structural elements.

Vestiges of the formation stage of the primary ultramafic rocks at the base of the oceanic crust are only localy preserved and they are of two kinds: (1) magmatic and (2) determined by the ocean floor metamorphism. The former ones refer to the layered structure of the rocks, which occurred during the crystallization of the ultramafic rocks by processes of differentiation and crystallization (Wager, Brown, 1968). Such structures can be observed on the crest west of the Turcu Valley, where wehrlite levels with pyroxenite bands occur in serpentinites (Pl. I, Fig. 1). Serpentinites locally display the primary bedded cryptostructure, visible on the surface of the blocks exposed to weathering. Millimetric or centimetric small bands, richer in serpentine and olivine remnants, alternate with thin bands richer in pyroxene, which are more resistant to erosion. Pyroxene form locally isolated schlieren.

The effects of the ocean floor, metamorphism are hardly noticed, the process being rather a submarine autometamorphism devoid of deformational elements (Savu, 1967; Coleman, 1977). It is certain that the serpentinization process began on the ocean floor, as one can notice at the layered ultramafic bodies intruded into the unmetamorphosed basaltic complex of the Mureş Zone (Savu, 1962).

The major moment of the ophiolite deformation, and particularly of the ultramafic rocks, as well as their advanced serpentinization, took place during the subduction, when the ophiolitic mélange formed on the Benioff plane in the presence of the marine water and under conditions of the extraordinary friction between the Moesian oceanic plate and the Transylvanian continental plate (Savu, 1985). The levels of wehrlites and pyroxenites were deformed concomitantly with the peridotitic rocks – which were serpentinized – as shown on Plate I, Figure 3 (Savu et al., 1985). At that time were included in the serpentinites the exotic blocks of igneous basic rocks from the Moesian Plate (basalts, gabbros) and of crystalline rocks from the Transylvanian Plate (crystalline schists, amphibolites, granites, etc.), thus resulting the ophiolitic (serpentinitic) mélange.

The definitivation of the rock serpentinization was, however, due to the Late Kimmerian metamorphism, which affected the olistoliths in their secondary place

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in the Severin trough. It determined both the serpentinization and the mentioned structural aspects superposed over the primary ones.

The tectonic movements which led to the emplacement of the Getic Nappe, and probably of the Severin Nappe, too, determined the formation of the cleavage planes S_2(Pl. I, Fig. 4), as well mylonitization, brecciation and faulting phenomena in the ultramafic olistoliths, as in the matrix of the olistostrome. The Late Laramian refolding probably contributed to these deformations as well.

The basic rocks olistoliths are constituted of basalt flows, often in pillow lava facies (Pl. I, Fig. 5), rarely intercalated with basic tuffs. The remnants of the ocean floor metamorphism are represented by epidote- and calcite-bearing quartz veinlets, often found in the basalt megaslab situated between Obârşia Cloşani and Ponoare and in the olistoliths south of the Cerna Vârf locality, as well as by the chloritization of the initial glassy matrix between the pillow lava separations. Due to the Late Kimmerian anchimetamorphism, which was accompanied by stress, the pillow lava separations were flattened (Pl. I, Fig. 6). It led to their movement one against the other, so that the chloritized matrix got a schistous aspect.

The anchimetamorphic effects over the basic tuffs intercalations are quite obvious. These rocks display a clear foliation (Pl. II, Fig. 1) and they recrystallized into chlorite and albite. Quartz veinlets penetrate these rocks.

A general result of the Late Kimmerian anchimetamorphism was the more advanced chloritization of all the basic rocks of the Severin Nappe, as compared with the basic ophiolites in the Mureş Zone for example, which had not been influenced by such a process. These differentiations can be observed under the microscope, as well as with the naked eye, looking at the two rocks on the polishing plane or their thin sections: the Severin Nappe basic ophiolites are green.

Due to the other tectonic events, the basic rocks were faulted, brecciated and laminated, aspects that can be observed along the dip faults at Brebina and Izverna. In the galleries, in the Cauna mining area, striking fractures can be noticed, which determined the sinking of the basaltic blocks towards the Bahna synform axis; the rocks are also mylonitized, brecciated and, locally, laminated on the fault plane.

The siliceous rocks (anchimetamorphosed jaspers) on the Brebina Valley were clearly affected by the Late Kimmerian anchimetamorphism. The anchimetamorphism determined the recrystallization of the siliceous mass with the concentrical sliding of different rock parts during the folding, so that the quartz veins which unconformably penetrate the rock got the aspect of ptygmatic veinlets (Pl. II, Fig. 2). Likewise, the flattening of their radiolaria took place (Pl. II, Fig. 3). They occurred as flattened micropebbles elongated perpendicular to the stress direction.

The schistous (pelitic) rocks, of a white-greenish or reddish colour, which accompany the radiolarite level on the Brebina Valley, and the black and chocolate-like shales from the marker level, lying under the stratiform mineralization, display an obvious schistosity, the rocks showing a slate character.

The Late Kimmerian anchimetamorphism influenced strongly the stratiform mineralization from the basaltic rocks. According to Macdonald (1982), such a mineralization was deposited on the ocean floor round the chimneys or hydrotherms as sulphuric muds and siliceous or siliceous-argillitic-carbonatic gels. Due to the anchimetamorphism, the mineralization recrystallized, getting a banded structure (Pl. II, Fig. 4), determined by the mineral recrystallization and the metamorphic differentiation. A clear evidence in this respect are the parallel pyrite crystals rows (Pl. II, Fig. 5), frequently showing pressure shadows (Pabst, 1931) formed round them. These phenomena were controled by an obvious stress, as the feather quartz from the pressure shadow developed on preasureless faces and it was elongated in the schistous rocks foliation plane, according to the lineation L_1 (Pl. II, Fig. 6). During the anchimetamorphism some pyrite crystals were fissured, the chalcopyrite and sphalerite grains deformed, and the copper mineralization levels slightly folded. The anchimetamorphism also led to the remobilization of the mineralization on fissures (Savu et al., 1985 b). Therefore the opposite phenomena manifested. The recrystallization and concentration of the substances from the primary mud as bands richer in pyrite and chalcopyrite, which alternate with schistous bands took place during the anchimetamorphism. Later on, when fissures occurred in the anchimetamorphosed rocks, a partial disperssion of the mineralization occurred along them, yielding veinlets which penetrate the rocks round the bands of layered mineralization.

During the thrust of the Severin Nappe over the sedimentary deposits of the wildflysch from the Danubian Autochthon, phenomena of mylonitization, breaking and lamination of the geological formations from both tectonic units into contact took place on the sliding plane, as shown by the ages presented in Table (94-95 Ma). Thus, the tectonic mélange occurred on the north-western (Dragu and Băroaia brooks) and south-eastern (under Ocnelor Hill-Cracul Gorunului, Cracul Joiţei) margins of the Severin Nappe. The mélange included all the rocks which came into contact during the thrust as blocks, often tectonically rolled. These blocks are represented by Azuga Beds, Senonian wildflysch, serpentinites, basalts, stratiform mineralization, in-

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cluding massive pyrite from the Cyprus-type ore deposit, as well as crystalline schists and pre-Alpine granites, which nowadays do not crop out in the study area (Savu et al., 1985 b). In the two mentioned areas the mélange blocks are included into a pelitic-granular matrix of a grey colour, determined by the friction of the blocks on the thrust planes.

It is of note that the Precambrian crystalline schists in the base of the Bahna Outlier were strongly sheared on the thrust plane, yielding sericite-chlorite schists and phyllonites with lenses of relict quartz from the earlier metamorphic rocks (Nadanova, Canicea).

Mineral Parageneses and Metamorphic Conditions

Although the anchimetamorphism was a very low-grade metamorphism concomitantly with the deformation process the composition of the rocks was altered in mineralogical respect, resulting parageneses typical of each type of geological formation. As the minerals were quite often very fine, the parageneses were determined by microscopic, X-ray, DTA and infrared analyses (Figs. 3-7).

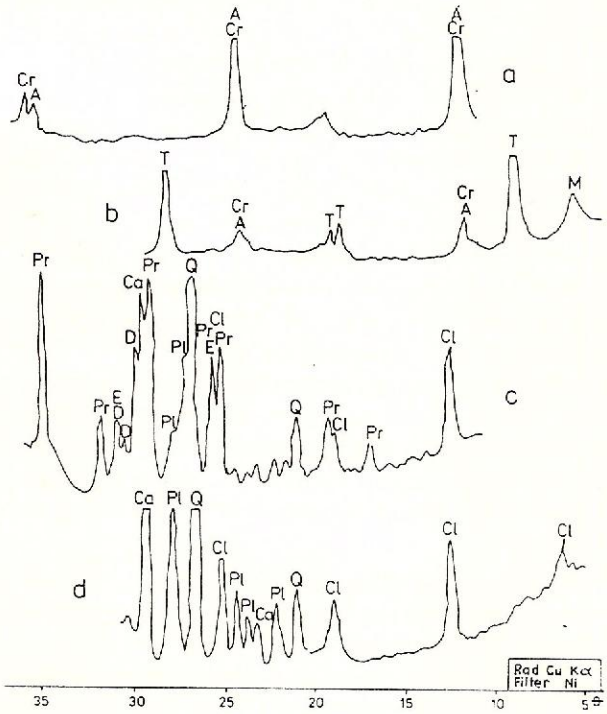

Fig. 3 Diagrams of the X-ray analysis. a, serpentinite from the Bradu Brook (Obârşia Cloşani): A, antigorite; Cr, chrysotile; b, hydrothermalized serpentinite from the Bradu Brook (Obârşia Cloşani): A, antigorite; Cr, chrysotile; T, talc; M, montmorillonite; c, basalt in the spring area of the Malareca Valley; Pr, prehnite; D, diopside (clinopyroxene); E, epidote; Ca, calcite; Q, quartz; Pl, albite; d, greenschist associated with the radiolarite level in the Brebina Valley: Ca, calcite; Pl, albite; Q, quartz; Cl, chlorite.

The paragenesis: quartz-sericite-chlorite-albite formed in the pelitic rocks in which the matrix of the Jurassic-Neocomian olistostrome is prevailing.

The olistoliths of serpentinized ultramafic rocks contain three characteristic parageneses: antigorite-magnetite; chrysotile-magnetite; antigorite-chrysotile-magnetite (Figs. 3a, 5, 6, 7a,b).

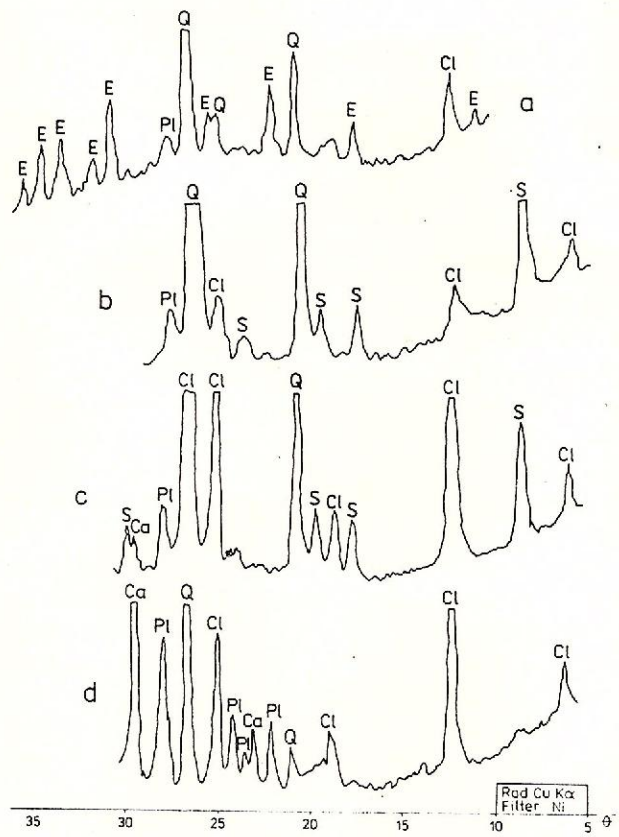

Fig. 4 Diagram of X-ray analysis. a, basic tuff from Ocnelor Hill: E, epidote; Pl, albite; Q, quartz; Cl, chlorite; b, red schist associated with the radiolarite level from Brebina Valley: Q, quartz; Pl, albite; Cl, chlorite; S, sericite; c, black schist from the level overlain by the Cauna stratiform copper mineralization: Ca, calcite; S, sericite; Pl, albite; Cl, chlorite; d, schist associated with the copper stratiform mineralization from the gallery in the spring area of the Malareca Valley: Ca, calcite; Pl, albite, Q, quartz, Cl, chlorite.

Magnetite from these parageneses is secondary, resulting from the iron released from the destroyed network of olivine and pyroxene. The ultramafic rocks influenced by synmetamorphic solutions present the paragenesis: talc-antigorite-montmorillonite (Fig. 3b). The presence of these minerals, especially of montmorillonite, in serpentinites points to a supply of SiO_2 and Al_2O_3 introduced simultaneously with the water used for the substitution of minerals from the primary peridotite. Hovorka et al. (1985) determined in the West Carpathians serpentinized rocks, in

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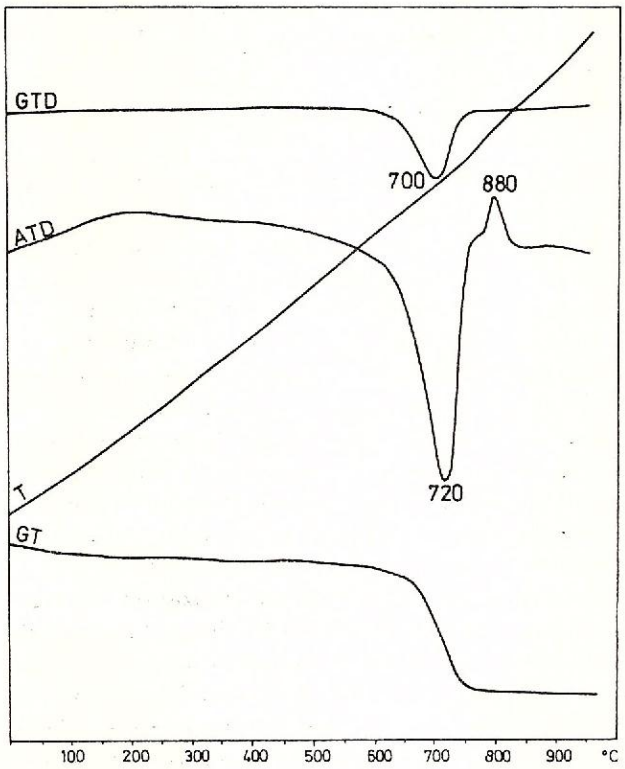

Fig. 5 — DTA diagram of a serpentinite (Antigorite+chrysotile) from the Bradu Brook (Obârşia Cloşani).

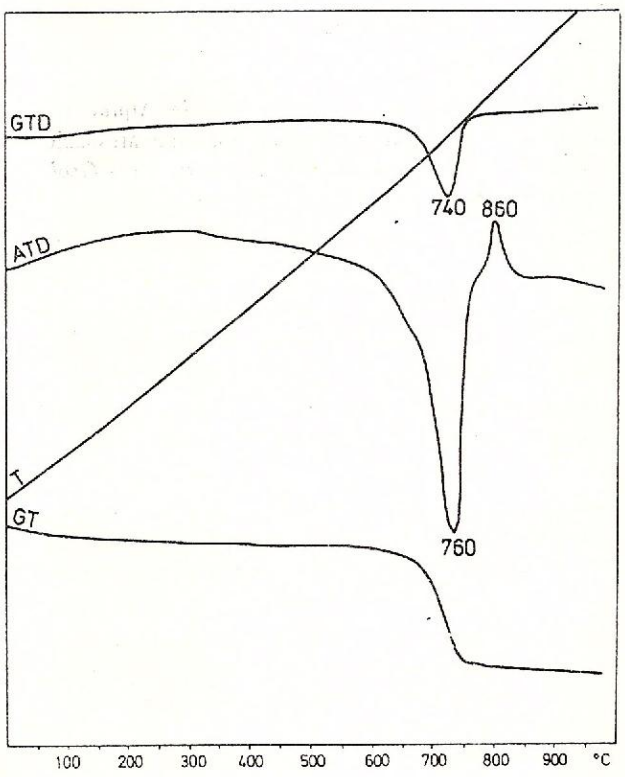

Fig. 6 — DTA diagram of a serpentinite (antigorite+chrysotile) from the crest between the Borloveanu Brook and the Turcu Valley.

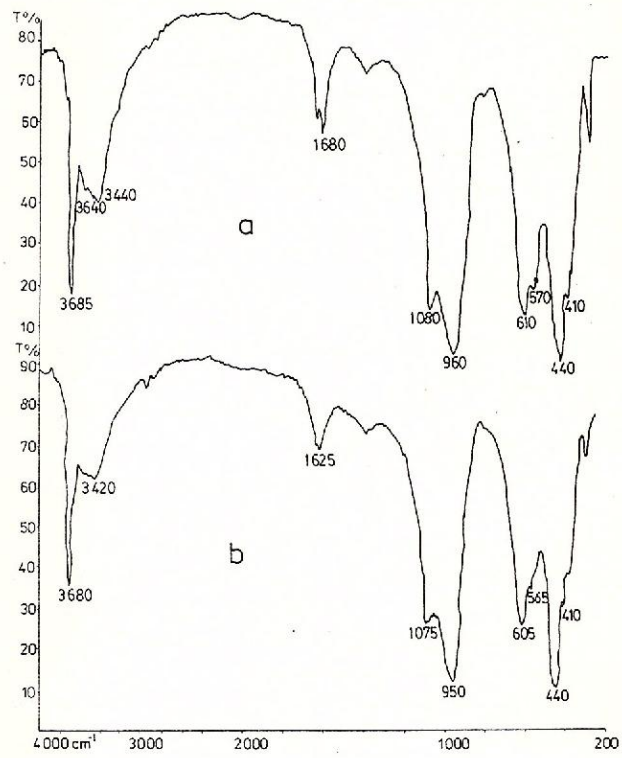

Fig. 7 — Diagram of the infrared analysis. a, serpentinite (antigorite+ chrysotile) from the Bradu Brook (Obârşia Cloşani); b, serpentinite (antigorite+chrysotile) from the crest between the Borloveanu Brook and Turcu Valley.

association with more intensely metamorphosed crystalline schists, which contain parageneses formed of the above-mentioned minerals, beside lizardite, a mineral which points to a higher degree of metamorphism.

The olistoliths of basic rocks present the following association: quartz-chlorite-prehnite-epidote-clinopyroxene-albite (Fig. 3 c). It is a composite association, being constituted of relict minerals from the magmatic primary association (pyroxene), vestiges of the ocean-floor metamorphism in albite-epidote-amphibolite facies and minerals formed under the influence of the Late Kimmerian anchimetamorphism, among which chlorite and prehnite are index minerals. The schistous basaltic tuff from the Măgura Brook include the paragenesis: albite-chlorite (Fig. 4 a). The low-grade metamorphosed reddish schists, coming from the red abyssal argillites, which occur beside the radiolarite intercalated in basalts on the Brebina Valley, present the following paragenesis: quartz-sericite-chlorite-albite-hematite (Fig. 4b). The paragenesis: quartz-chlorite-albite-calcite is formed in the greenschists occurring in association with the same jasper horizon on the Brebina Valley (Fig. 3d).

The black shales from the marker level overlain by the copper stratiform mineralization present

Institutul Geologic al României

the paragenesis: quartz-chlorite-sericite-albite-calcite-carbonaceous material (Fig. 4c). The researches carried out by Frey et al. (1974) on the carbonaceous material from the very low-grade metamorphic rocks from the Central Alps pointed out that the carbonaceous material is represented by anthracite and metaanthracite. Landis (1971) emphasized that under conditions of low-grade metamorphism the coaly material shows a higher degree of crystallization than in the diagenesis process, reaching the graphite d_2-d_1. The crystallization degree seems to be also specific to the carbonaceous material from the Baia de Aramă black shales, because a crystallized graphite is found in the rocks of the albite-epidote-amphibolite facies of the regional metamorphism.

As regards the conditions of anchimetamorphism, its type and the subfacies in which it developed, several factors are to be considered. First, the lacking of zeolites from the above-mentioned parageneses excludes the possibility that the rocks alterations should have been determined by a burial or ocean floor metamorphism. Such a metamorphism developed in the Alps, where parageneses with laumontite (Kubler et al., 1974) and with pumpellyite (Coombs et al., 1976) were found. Secondly, the very low-grade metamorphism, which influenced the geological formations in the lower part of the Severin Nappe, was accompanied by a synkinematic deformation pointing to a very low-grade regional anchimetamorphism (Dunoyer de Segonzac et al., 1968).

Taking into account the parageneses from the silico-aluminous rocks, the anchimetamorphism of the geological formations in the Severin Nappe is located in the very low-grade metamorphic zone of Barrovian type, quartz-albite-muscovite (sericite)-chlorite subfacies (Winkler, 1967). The pressure and temperature conditions of this type of metamorphism were experimentally determined by Nitsch (1971), who pointed out that the paragenesis prehnite-(pumpellyite)-chlorite-quartz was established at a pressure of 2 kb and a temperature of $345^0 \pm 20^0$ C, and 7 kb and $260^0 \pm 20^0$ C, respectively. These conditions of metamorphism explain the mineralogical modifications in the rocks belonging to the Jurassic-Neocomian olistostrome of the Severin Nappe, as well as the quartz mobilization in lenses and veinlets occurring on fissures during the final stage of metamorphism.

Considering the temperature conditions under which the hydrothermal metal sulphide deposits are emplaced (Schneiderhöhn, 1955), the anchimetamorphism temperatures ranging between 260^0 and 345^0 correspond to the mesothermal stage (200^0-350^0 C). These thermal conditions explain quite well the recrystallization of the stratiform sulphidic ore which, under these stress control, was arranged as parallel bands in alternation with schistous rocks containing the paragenesis quartz-sericite-albite-calcite (Fig. 4 d).

Conclusions

The Late Kimmerian anchimetamorphism and the metamorphic differentiation processes that accompanied it influenced the formations of the lower part of the Severin Nappe, particularly the Azuga Beds and, to a less extent, the base of the Sinaia Beds. The radiometric (K/Ar) dating points to about 130 Ma.

The anchimetamorphic processes were determined by a very low-grade metamorphism achieved at a low pressure and temperatures ranging between 260^0 and 345^0 C, which represent the stability limits of the prehnite-bearing parageneses, a marker mineral found also in the mineral associations from the anchimetamorphosed rocks of the Severin Nappe.

The Cyprus-type stratiform mineralization from the basalt megaslab was influenced by the anchimetamorphism as regards (a) the recrystallization and concentration of the mineralization as ore bands, more or less compact, alternating with schistous rock bands and (b) the partial disperssion of the mineralization on the fissures from the surrounding basic rocks.

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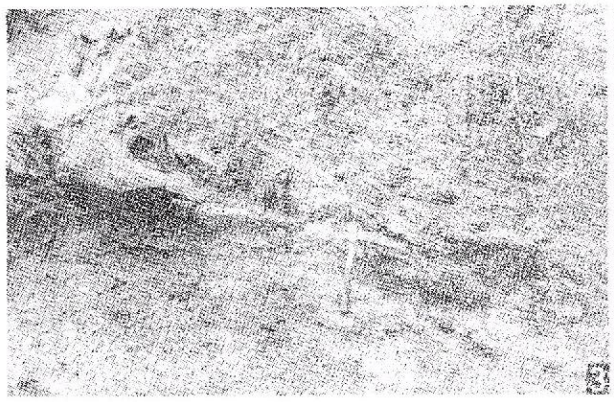

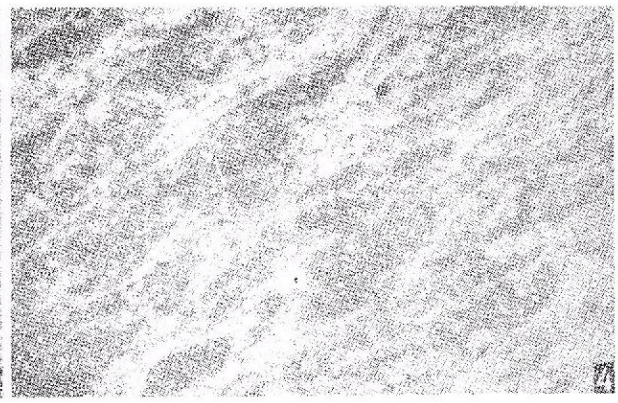

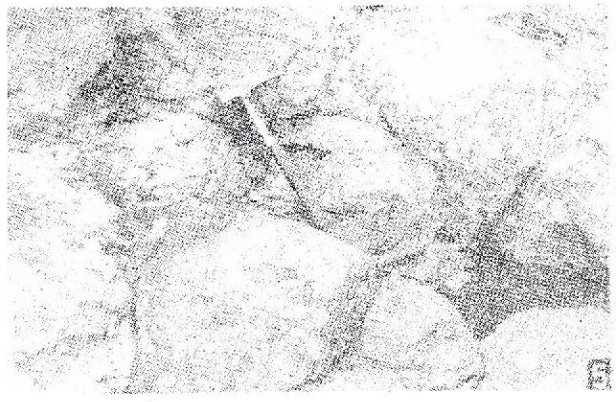



Fig. 1 – Wehrlite intercalation in serpentinite, with a thin band of pyroxenite in the middle. Crest between the Borloveanu Brook and Turcu Valley.

Fig. 2 – Laminated serpentinite. Borloveanu Brook. x 30, crossed polarizers.

Fig. 3 – Wehrlite intercalation broken during the anchimetamorphism. Crest between the Borloveanu Brook and Turcu Valley.

Fig. 4 – Serpentinite with two visible foliations (S_1 and S_2). Hoanca Valley. x 30, crossed polarizers.

Fig. 5 – Pillow lava separations. Turcu Valley.

Fig. 6 – Flattened pillow lava separations. Turcu Valley.

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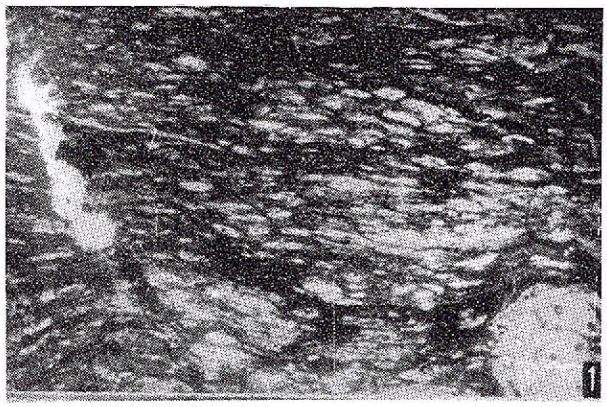

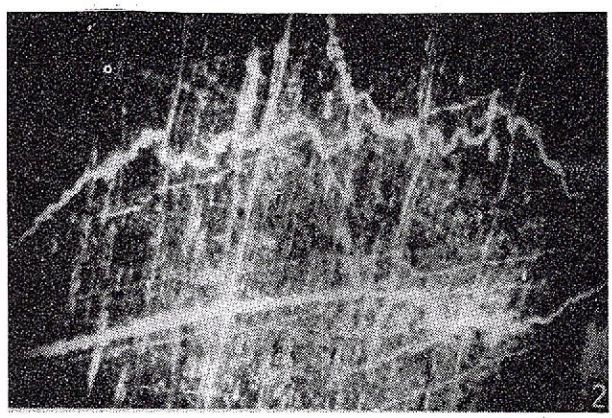

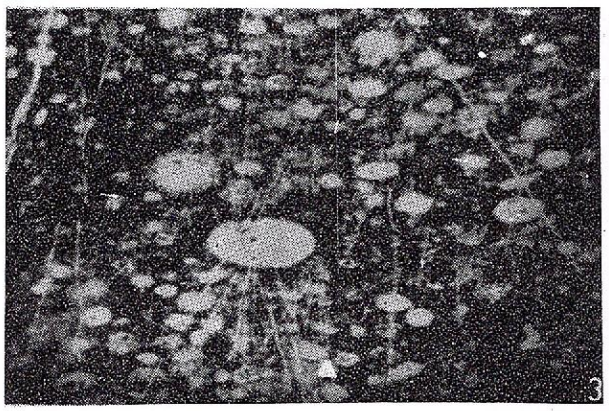

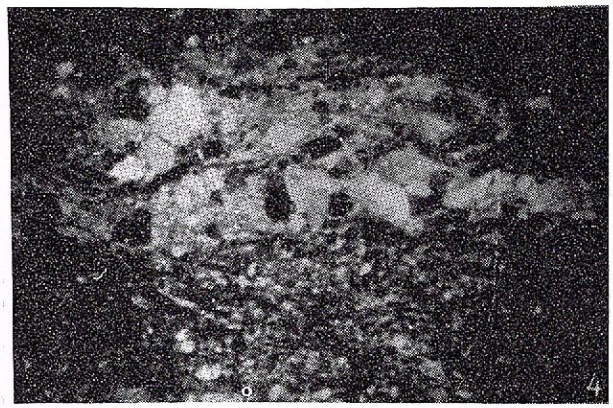

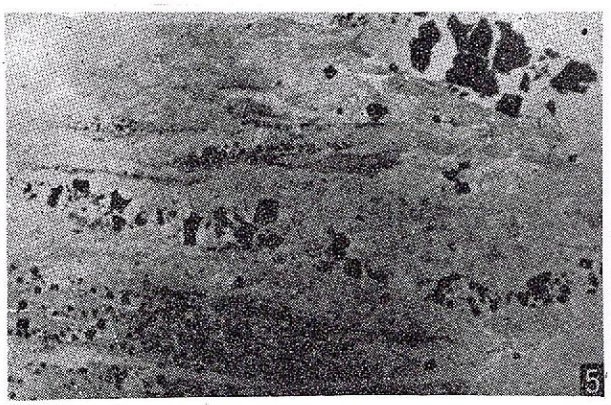

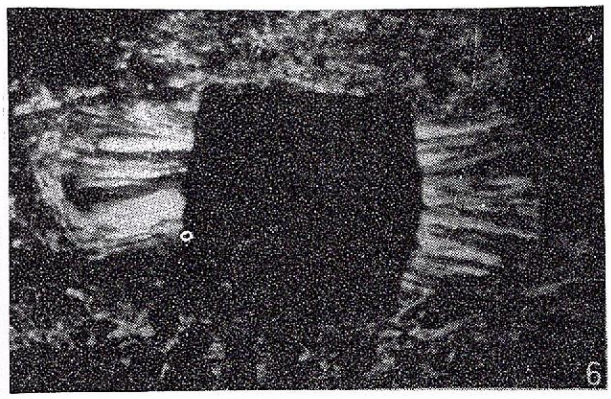

Fig. 1 – Basic tuff in the Brebina Valley. x 30, crossed polarizers.

Fig. 2 – Ptygmatic vein of exuded quartz in the siliceous rocks from the Dragu Valley. x 30, crossed polarizers.

Fig. 3 – Deformed radiolaria in Brebina Valley jaspers. x 30, crossed polarizers.

Fig. 4 – Schistous rock covering the copper stratiform mineralization, with a band of quartz grains with sulphides (black crystals). Cauna Mine. x 30, crossed polarizers.

Fig. 5 – Parallel rows of pyrite crystals with pressure shadow in the schistous rock which accompanies the stratiform mineralization. Cauna Mine. x 30, crossed polarizers.

Fig. 6 – Elongation of the feather quartz from the pressure shadow parallel with the schistous rock foliation. Cauna Mine. x 120, crossed polarizers.

Institutul Geologic al României

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PETROLOGY AND GEOCHEMISTRY OF THE EFFUSIVE ROCKS FROM THE OCEAN FLOOR BASALT COMPLEX IN THE MUREŞ ZONE

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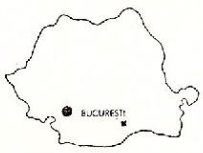

Key words: Basalts, Spilite. Magmas. Tholeiitic composition. Magmatic differentiation. Major elements. Minor elements. Mureş Zone. Apuseni Mountains.

Abstract: The ocean floor basaltic complex (O_1) from the megaslab present in the Mureş ophiolitic suture (Mureş Zone) formed two wide areas of ocean floor basalt flows, situated on either side of the ocean ridge, represented now by the sheeted dyke complex (O_2). The magma from which these complexes resulted was a tholeiitic one, formed in the mantle at the depth of about 150 km. It differentiated in a magmatic chamber at depth. The lavas from which the rocks of this magmatic complex resulted flowed on the Mureş Ocean floor. They derived from a normal, tholeiitic magma, from which the common basalt and dolerite sheeted dykes resulted in the first stage of the initial tholeiitic magma differentiation. Therefore the basaltic rocks show features characteristic of normal (common) tholeiitic basalts.Except for some spilites richer in Na_2O, the basaltic complex is devoid of the differentiated rocks (mafic and acid) of the sheeted dyke complex, which form in the second stage of the evolution of the magmatic chamber.

1. Introduction

Basaltic complexes (O_1) constitute a very important member of the ocean crust. Their presence was established both in the ophiolite masses situated on the oceanic plates, as is the case in the Pacific Ocean (Griffin, Varne, 1980) and in those obducted on continents (Coleman, 1977). Parts of such a basaltic complex are found also in fragments of ophiolitic rocks broken from the initial megaslab from the ophiolitic suture of the Mureş Zone (Savu, 1983). In the Căpălnaş Unit the basaltic complex is associated with the sheeted dyke complex (O_2), as shown in Figure 1. This ophiolitic megaslab was scraped from the upper part of the Mureş Ocean crust, so that it seems to include only these two ophiolite complexes that can be observed. The age of the basaltic complex as well as of the other alpine ophiolites from the Mureş Zone is Liassic (180 Ma – Hertz et al., 1974).

The petrology and geochemistry of the sheeted dyke complex was presented in several papers (Savu, 1984 b; 1985; Savu et al., 1985 a; Savu, Stoian, 1988). The present paper comprises a similar study on the ocean floor basaltic complex from the Mureş Zone. Several data regarding these aspects of the basaltic rocks were previously presented (Savu et al., 1970, 1985; 1988 a; 1989) but a synthetic study has not yet been achieved, this paper attempting to meet this necessity.

Structure of the Basalt Complex and Its Relationships with Other Geological Formations

The ocean floor basalt complex (O_1) is present in all the tectonic units from the Mureş Zone, where the allochthonous Alpine ophiolites occur. Its rocks are part of either the ophiolite masses situated in the basement of these units, as in the Drocea Mts, or appear as olistoliths in mélanges with Tithonian–Neocomian volcanic matrix and in the flysch deposits from the Metaliferi and Trascău Mountains (Savu, 1984 a; Savu et al., 1990 a). This complex shows a complete and very characteristic development in the oceanic crust fragment from the base of the Căpălnaş–Techereu Nappe in the Drocea Mts and the western part of the Meta-

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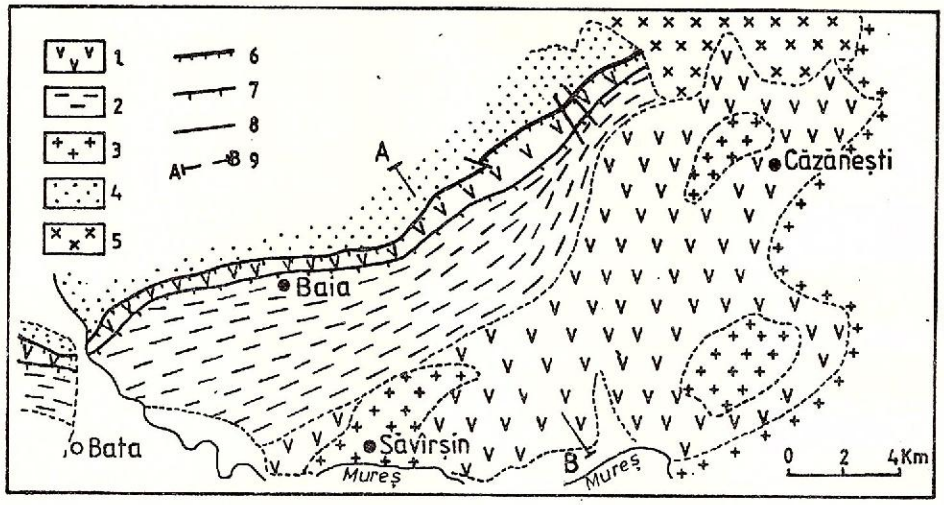

Fig. 1 — Sketch map of the distribution of the Alpine ophiolites from the Drocea Mts (Mureş Zone). 1, Ocean floor basalt complex (O_1); 2, sheeted dyke complex (O_2); 3, Late Kimmerian volcanics and intrusions ($J_3 - Cr_1$); 4, Tithonian–Neocomian flysch; 5, Neogene volcanics; 6, overthrust; 7, reverse fault; 8, normal fault; 9, geologic section of Fig. 2.

liferi Mts, between Vaţa, Laleşinţ and Zam (Fig. 1)[1].

This basalt complex develops in this region both to the east and west of the outcropping area of the sheeted dyke complex from the ophiolitic megaslab, which represents the vestiges of an oceanic ridge. The basaltic area to the east of the sheeted dyke complex is the largest from the entire Mureş ophiolite suture, covering approximately 450 km^2. In the region west of the sheeted dyke complex, which is smaller than the former region, the basalts crop out along a narrow belt of approximately 15 km in width between this complex and the Tithonian–Neocomian flysch. They make up the Baia Scale (Savu et al., 1987), which overthrust the flysch deposits and was thrust from south-east by the sheeted dyke complex from the Căpălnaş–Techereu unit. In the same tectonic unit the rocks of the basalt complex crop out also from under the Late Kimmerian island arc volcanics (J_3–Cr_1) in the Visca and Luncoi erosion windows (Savu et al., 1985) as well as in that at Almaşu Mare–Buneşti (Savu et al., 1978), all of them situated east of the Vaţa–Zam meridian.

The ocean floor basalt complex consists mainly of submarine basic lava flows, usually in pillow lava facies (Pl. I, Figs. 1 and 2). In the zone east of the sheeted dyke complex, which probably formed on the ocean floor an abyssal plain (Figs. 1 and 2), these flows are almost horizontal, being to a little extent tectonically disturbed. In the zone of the Baia Scale, west of the sheeted dyke complex area, the basalt flows are strongly disturbed by the Alpine tectonic movements, being situated between the Căpălnaş–Techereu Nappe and the Tithonian–Neocomian flysch which they overthrust. In places the basalt flows are even reversed north-westwards, in the sense of the vergence of the Alpine structures on the north-western part of the Mureş Zone (Savu, 1988; Savu et al., 1988 b).

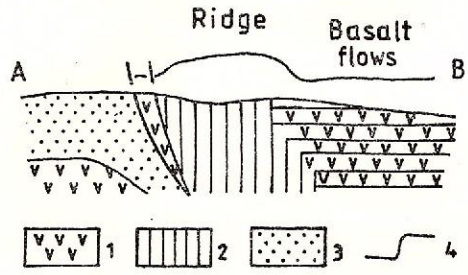

Fig. 2 — Geologic section on the AB alignment from Figure 1 through the ophiolite complexes from the Drocea Mts. 1, basalt complex; 2, sheeted dyke complex; 3, flysch; 4, magnetometric profile according to data by Andrei et al. (1984).

Thin basic pyroclastic levels (Pl. II, Fig. 1), containing sometimes larger blocks of basalts or hyalobasalts (Pl. II, Fig. 2) are intercalated among the basalt flows. At some localities, such as north of Troaş, Cuiaş, Visca and especially west of the Corbeşti levels of tachylites, glassy rocks, mostly devitrified and

[1] In order to get carthographic details, the maps of the papers published by Savu et al. (1970; 1987; 1988 a) and the geologic maps at 1:50.000, Săvârşin and Roşia Nouă sheets (1979) were used.

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in places chloritized, are found among the basalt flows. They are usually present as small spheroidal or ovoidal separations which peel off concentrically. It is also worth noting the fact that among the basalt flows from the Căpâlnaş–Glodghileşti–Sălişţioara region, situated in the area to the south-east of the sheeted dyke complex, intercalations of chemical precipitation and detrital limestones with thin Liassic quartzitic sandstone bands occur, which indicates that this area was closer to the southern continent in the oceanic zone (Savu 1984 a; Savu et al., 1988 b). Such deposits do not exist among the basalt flows from the Baia Scale; on the other hand, there occur small abyssal red argillites devoid of jaspers, indicating an area of the Mureş Ocean of a much greater depth (Savu et al., 1988 a).

The basalt lava flows consist of rocks exhibiting intersertal to subophitic structure such as basalts, hyalobasalts, amygdaloidal basalts and anamesites which are seldom associated with dolerite sills. The rocks consist of plagioclase (An 50–60) laths, augite, sometimes slightly titaniferous crystals, with which magnetite and more rarely pyrite are associated, besides variable amounts of glass. The crystallization order of minerals – the one specific for the tholeiitic basalts – is: plagioclase–clinopyroxene–magnetite–pyrite. The spilitic rocks are also frequent, in which plagioclase is an albite (An 8–10), while clinopyroxene as well as the interstitial glass were chloritized.

In the basalt complex gabbro bodies and small peridotite intrusions were emplaced, which were searched on other occasions (Cioflica, 1962; Savu, 1962 a; 1962 b; Savu et al., 1970; Savu et al., 1990 b). As it has already been shown by Savu and Andrei (1993), the small gabbro bodies at Căzăneşti and the larger gabbroic bodies at Almăşel and Cuiaş as well as the small peridotite intrusion north–east of Corbeşti are located along an important gravimetric anomaly. This indicates the presence at depth of a large dyke of gabbros and ultramafic rocks, that crossed the lower plutonic complexes of the Mureş Ocean crust and the lower part of the ocean floor basalt complex. The mentioned gabbro bodies represent apophyses of this major intrusion. During the formation process of the Mureş Ocean, whereby the ophiolitic rocks under discussion were formed, a subsea hydrothermal metamorphism manifested, that affected the basic rocks in the conditions of the zeolitic facies (Savu, 1967).

Geochemistry of Basalts

Basaltic rocks from the area situated to the east of the sheeted dyke complex zone, between Căzăneşti, Săvârşin and Zam, which represents the greatest extension of the basalts from the ophiolite zone (Fig. 1), were selected for the geochemical study of the igneous rocks from the ocean floor basalt complex in the Mureş Zone. For this purpose 44 new chemical and spectral analyses were carried out on basaltic and rarely doleritic rocks, which are presented in Table. Owing to the subsea metamorphism, tectonization and diagenesis, the Mesozoic basalts were affected by alteration processes, which are reflected by CO_2, H_2O and S contents of these rocks. By these aspects they differ from the younger volcanics.

The basaltic character of these rocks clearly results from the SiO_2 content ranging between 39 and 52 %. The latter value represents, according to Gill (1981) and Ewart (1982), the very limit between the basalts and the basalt andesites from the calc-alkalic series. The SiO_2 mean is of 47 %. The rocks containing less than 40 % SiO_2 were affected by alteration (calcitization) processes, a reason why they contain greater than normal CO_2 amounts (5–10 %) and over 16 % CaO amounts respectively.

The mean TiO_2 contents of 1.13 % range between 0.30 and 1.96 %, values characteristic of the tholeiitic basalts (Pearce, Cann, 1973). According to the TiO_2 content and the value of the FeO/MgO ratio, the rocks plot in the lower part of the abyssal tholeiitic basalt field, as is shown in the diagram (Fig. 3) drawn after Miyashiro (1975). The separated rocks plot round this position, confirming thus the ocean floor character of the basalts.

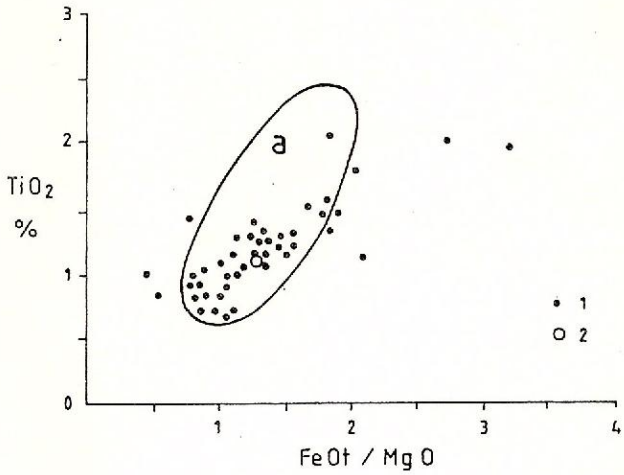

Fig. 3 – The TiO_2-FeOt/MgO diagram (according to Miyashiro, 1975). 1, Basalts; 2, mean of the basalts in the region; a, field of abyssal tholeiites.

The FeOt content ranges between 6.30 and 12.30 %, indicating the tendency of the tholeiitic magma to concentrate the iron (Fenner, 1929; Savu et al., 1984); the mean value of this oxide is of 8.64 %. At the same time the MgO content varies between 3.77 and 9.74 %, its mean content being of 6.25 %. These values plot in the field established by Miyashiro (1975) for the

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Table — Chemical composition

No	1	2	3	4	5	6	7	8	9	10	11
SiO_2	39.31	39.80	41.60	42.46	43.18	43.63	43.74	44.25	45.20	45.20	45.22
TiO_2	0.30	0.76	1.16	1.12	0.84	1.04	1.24	0.72	0.80	1.28	1.96
Al_2O_3	13.59	15.36	14.84	13.84	13.49	15.19	13.96	14.91	14.70	14.26	13.71
Fe_2O_3	5.49	5.09	6.53	6.69	4.35	6.19	6.43	5.64	3.70	5.83	5.65
FeO	1.87	2.10	3.06	2.08	3.19	3.47	2.54	2.23	2.64	3.84	7.22
MnO	0.12	0.14	0.16	0.11	0.16	0.13	0.13	0.15	0.16	0.16	0.24
MgO	5.55	5.28	6.85	5.27	6.96	7.26	5.54	6.48	5.51	5.76	6.13
CaO	16.24	16.12	12.28	16.11	14.40	10.98	12.63	13.26	12.89	9.00	8.84
Na_2O	2.65	3.35	3.95	2.11	3.54	3.89	3.76	2.83	3.50	4.41	2.72
K_2O	0.31	0.40	0.31	0.28	0.43	0.15	0.10	0.22	0.76	0.05	0.20
P_2O_5	0.10	0.09	0.16	0.16	0.15	0.15	0.19	0.12	0.08	0.15	0.23
S	0.28	0.28	0.13	0.16	0.18	0.19	0.20	0.19	0.18	0.29	0.28
CO_2	10.17	5.95	3.27	3.95	3.18	2.60	5.36	5.68	6.61	5.13	1.73
H_2O^+	3.77	4.72	5.58	5.96	5.54	5.53	3.96	2.78	3.63	4.98	5.36
Total	99.75	99.44	99.88	100.30	99.59	100.40	99.78	99.46	100.36	100.34	99.49
Ni ppm	100	75	135	160	160	130	105	115	32	85	40
Co	26	21	38	24	31	28	33	25	19	50	42
Cr	370	210	340	500	330	430	310	460	15	46	43
V	240	125	230	170	210	230	230	260	200	330	420
Sc	31	27	30	28	32	32	35	28	36	32	35
Y	18	17	21	25	18	21	28	15	30	36	42
Yb	1.4	1.4	2.2	1.8	1.3	2.1	2.8	2.4	2.4	3.0	3.6
La	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30
Zr	60	40	75	68	62	75	90	70	85	80	160
Ba	45	26	22	32	92	21	38	65	130	20	30
Sr	220	300	65	150	160	300	600	470	125	120	120
Pb	18	2.5	<2	<2	<2	<2	<2	2	2	<2	< 2
Cu	2	35	30	17	38	43	52	34	30	20	44
Ga	12	10	12	18	14	12	13	13	15	14	16
Zn	32	42	42	36	46	40	40	36	42	37	70
Nb	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10

The rocks analysed in this table are the following: basalts in pillow lava facies – 43,39,38, Zam quarry; 1, Pietriş quarry; 11, Almăşel Valley: 3,6,29, Obârşia: 41, 40, Pârâul Morii; 4, Ilteu; 9,2,27,35,DN7; 33, 16, 36,30, Hănuleasa Valley; basalts – 20, Zam quarry; 7,18, Almăşel Valley; 15,28,10, Cireşu Valley; 42, Codru Brook; 25,44, Pârâul Fugii; 12,32,19, 14, Sinteasca Valley; 37, Pârâul Morii; 23,24, Temeşoaia Valley; hyalobasalts – 5, Corbeasca Valley; tachylites – 22, 21, 26, 17, Corbeasca Valley; basaltic agglomerates – 34, Ilteu; 8, Temeşoaia Valley; dolerites – 13,31, Zam quarry; 14, Sinteasca Valley.

abyssal tholeiitic basalts. The mean value of the FeOt/MgO ratio is of 1.45, being also situated in the lower part of the differentiation line of the abyssal tholeiites.

The Na_2O contents range between 1.79 and 6.20 %, its mean value being of 3.64 %. The higher values (>4 %) indicate the presence of spilites. The latter may be represented both by spilites resulted from the reaction of the basic lava with the oceanic water, namely with the hot water vapours formed under these conditions, and by albite rocks formed through the differentiation of tholeiitic magma (Savu et al., 1984). Owing to the above-mentioned submarine alteration processes and especially to the partial sericitization of plagioclase, some tholeiitic basalts in the region contain K_2O amounts above or below 0.30 %. This component ranges between 0.02 and 2.74 %. The very low K_2O contents are probably due to the spilitization process, that is the removal of the initial K_2O content from the rocks. The same submarine alteration processes controlled also the CaO behaviour in these rocks. This component ranges between 4.52 and 16.12 %, its mean being of 10 %. In this case an increase of the CaO content in the amygdaloidal and fissured basalts and its decrease in the spilitized rocks respectively are found.

The diagram in Figure 4 after Irvine and Baragar (1971) shows the spilitization degree of the Mesozoic rocks which have a Na_2O content generally exceeding 2.50 %, a mean value established by Miyashiro (1975) for the tholeiitic basalts. Depending on this compo-

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of the basaltic rocks

12	13	14	15	16	17	18	19	20	21	22
45.70	45.86	46.23	46.30	46.43	46.44	46.53	46.70	46.91	47.02	47.03
0.84	0.96	0.60	1.16	1.80	0.64	1.40	1.20	1.60	0.92	0.88
16.04	16.01	15.35	14.59	14.10	14.95	16.56	14.04	15.71	14.10	15.36
3.87	4.21	1.65	3.93	5.94	5.26	4.56	3.45	5.70	3.60	4.22
4.69	4.80	4.82	5.06	5.90	3.23	5.75	6.07	5.24	4.89	4.43
0.36	0.24	0.14	0.16	0.26	0.11	0.19	0.17	0.23	0.14	0.14
9.74	7.16	9.73	7.18	5.54	7.28	6.99	8.17	5.31	8.38	8.20
7.93	8.85	6.80	9.69	7.89	11.46	10.79	11.59	9.01	11.56	11.23
2.15	3.54	3.41	3.62	3.98	1.63	2.49	3.07	4.72	1.79	2.12
1.40	1.59	0.48	0.12	0.33	0.60	0.25	0.24	0.28	0.42	0.40
0.09	0.10	0.12	0.13	0.24	0.15	0.17	0.12	0.21	0.15	0.15
0.64	0.29	0.20	0.23	0.24	0.18	0.23	0.17	0.33	0.25	0.24
2.60	2.57	4.54	1.72	2.13	0.89	0.77	2.59	1.23	1.68	1.78
4.71	3.26	6.30	6.08	4.73	6.68	3.31	2.64	3.03	5.58	3.95
100.76	99.44	100.37	99.97	99.51	99.50	99.99	100.22	99.51	100.52	100.13
70	57	100	57	52	185	95	48	37	210	120
28	30	29	34	40	33	38	40	30	42	30
270	155	220	155	96	410	290	135	75	500	330
200	200	160	230	460	150	370	260	400	310	220
32	29	32	34	30	18	34	37	34	38	34
20	18	20	22	32	16	27	24	29	27	19
1.6	2.1	1.4	1.8	3.8	1.2	3.4	2.9	3.6	2.0	1.6
<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	< 30
56	46	40	62	165	42	120	48	100	85	72
56	180	25	32	30	150	44	10	26	85	67
120	520	85	62	180	260	95	110	200	310	250
< 2	<2	<2	<2	2.5	<2	<2	<2	<2	<2	< 2
11	57	65	58	80	27	72	65	10	55	52
15	17	15	15	14	14	15	16	22	15	16
46	52	45	46	52	32	60	54	50	40	36
<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	< 10

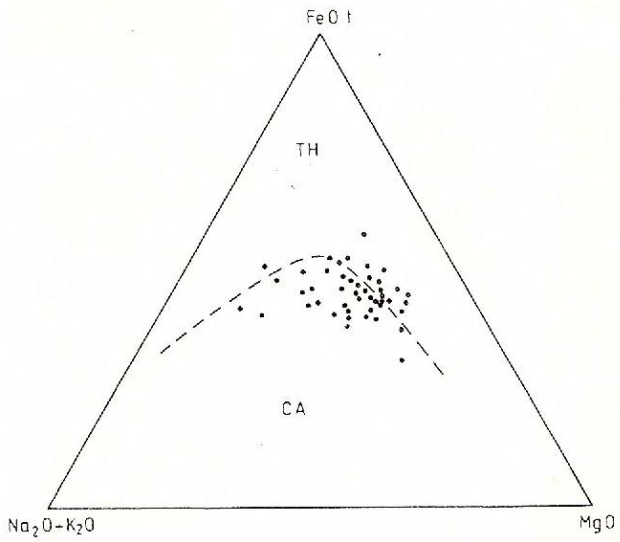

Fig. 4 — The FeOt–MgO–$Na_2O + K_2O$ diagram (according to Irvine and Baragar, 1971). TH, field of tholeiitic rocks; CA, field of calc-alkali rocks.

nent the igneous rocks from the ocean floor basalts with Na_2O content above or below 2.50 % plot in the tholeiitic field. That is why the spilitized rocks with higher Na_2O content shift from the tholeiite field to the alkali corner, being plotted in the calc-alkali field, although they do not belong to such a petrographic series.

The contents of minor elements (Table) are also characteristic of the ocean floor tholeiites. Among these elements, V, Co and Sc show lower values of the contents; the mean values of these elements reach 249 ppm in the case of V and 30 ppm in the case of Co and Sc. According to the V and Ti contents, the rocks in the diagram of Figure 5 plot between the straight lines corresponding to values of the Ti/V ratio of 20 and 50 respectively, thus underlining their character of ocean floor rocks (OFB).

Zr is a very characteristic element for the magmatic differentiation. In the present case most its values are below 100 ppm, which indicates a weak differentiation of the initial tholeiitic magma. Only a few rocks present higher Zr contents, ranging between 105 and

Institutul Geologic al României

No	23	24	25	26	27	28	29	30	31	32	33
SiO_2	47.08	47.93	47.16	47.20	47.29	47.34	47.40	47.60	47.92	47.94	48.19
TiO_2	0.92	0.68	1.40	0.72	1.04	1.28	1.16	1.28	1.56	1.36	1.24
Al_2O_3	13.24	13.81	14.09	15.04	15.71	14.00	14.10	15.41	14.30	13.80	15.66
Fe_2O_3	5.04	4.26	5.83	4.76	3.91	5.71	4.91	5.35	6.60	4.62	2.61
FeO	3.33	2.76	4.97	3.79	4.13	4.63	3.97	3.74	3.92	4.90	6.96
MnO	0.18	0.11	0.13	0.13	0.17	0.20	0.12	0.15	0.16	0.22	0.17
MgO	7.00	5.74	7.55	8.28	6.60	6.94	5.42	6.69	4.27	4.76	7.19
CaO	12.00	11.84	7.60	10.50	10.30	10.31	10.42	9.20	9.41	8.87	10.61
Na_2O	3.32	4.25	3.30	2.12	3.13	3.01	4.18	3.93	5.20	5.65	3.14
K_2O	0.60	0.02	0.15	0.36	0.30	0.02	0.06	0.06	0.13	0.25	0.07
P_2O_5	0.11	0.14	0.15	0.15	0.15	0.13	0.25	0.20	0.20	0.16	0.18
S	0.23	0.17	0.18	0.17	0.14	0.18	0.19	0.15	0.18	0.29	0.14
CO_2	3.03	3.16	1.96	1.04	2.96	4.42	4.42	1.21	3.90	4.40	0.70
H_2O	3.93	4.57	6.06	5.87	3.68	2.53	2.87	5.14	2.53	3.10	3.77
Total	100.01	99.44	100.53	100.14	99.51	100.70	99.52	100.11	100.37	99.72	100.65
Ni	75	145	26	160	85	40	140	120	33	29	60
Co	40	30	32	23	23	34	34	32	23	23	25
Cr	250	570	34	410	250	100	390	340	22	65	205
V	260	250	260	170	135	250	200	340	300	270	200
Sc	42	37	32	24	28	38	30	32	33	27	32
Y	26	22	29	17	19	30	26	25	34	25	22
Yb	2.9	2.0	2.1	1.2	1.8	3.6	2.9	3.2	5.7	3.0	2.0
La	<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	< 30
Zr	62	68	90	44	48	100	75	95	85	75	58
Ba	92	38	17	120	160	30	19	20	52	28	85
Sr	290	360	170	220	360	170	250	95	140	125	320
Pb	< 2	6	<2	<2	<2	<2	<2	<2	<2	<2	< 2
Cu	68	48	2	19	35	50	31	38	100	2.5	29
Ga	11	11	17	11	9	14	11	14	17	16	11
Zu	55	32	30	30	36	62	47	38	60	34	48
Nb	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	< 10

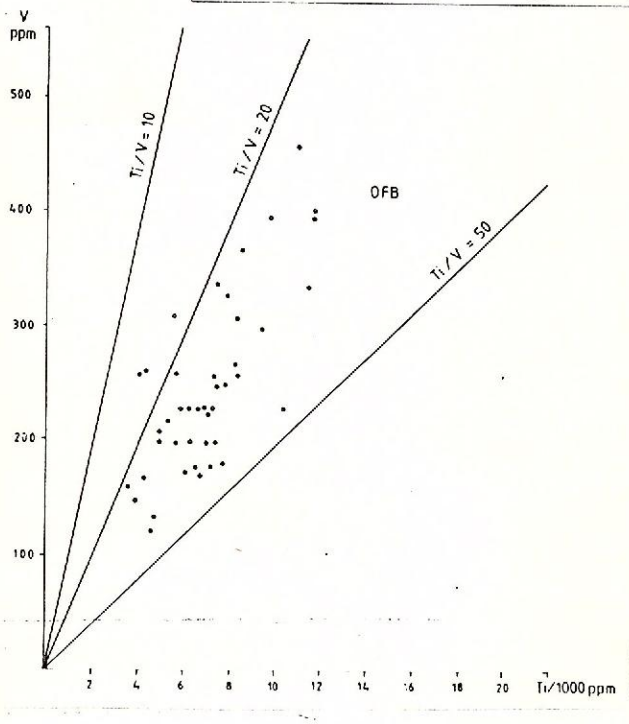

Fig. 5 – The V-Ti/1000 diagram (according to Shervais, 1982) OBF-field to ocean floor basalts.

165 ppm. These last samples have also the highest Y (30–42 ppm) and Yb (3.2–4.4 ppm) contents. Most rocks show lower Y and Yb values, reflecting also the ocean floor tholeiite character of the rocks from the basalt complex (O_1). The same characteristics of the basaltic rocks are revealed also by the contents below the detection limit of Nb (<10 ppm) and La (< 30ppm).

With a few exceptions the Ba contents from most of the basaltic rocks are below 100 ppm, values that, according to Miyashiro (1975), are characteristic of the abyssal tholeiites.

The Sr contents were partly influenced by submarine alteration processes. Thus 80 % of its values range between 100–360 ppm, the mean value being of 218 ppm. Low Sr values, ranging between 65 and 95 ppm (Table) were noticed in the spilitized rocks. A few samples present higher Sr contents that vary between 470–720 ppm.

Institutul Geologic al României

Table(continued)

34	35	36	37	38	39	40	41	42	43	44
48.60	48.90	49.40	49.68	49.74	49.97	50.30	50.70	51.79	51.97	52.22
1.04	1.20	1.00	1.08	1.90	1.28	1.24	1.12	1.36	1.92	1.72
14.44	14.69	14.24	15.31	16.56	16.30	15.50	16.20	14.56	14.84	14.30
3.56	4.74	6.01	6.49	8.27	6.36	6.64	5.33	5.40	6.41	5.66
4.76	5.16	2.53	1.48	4.26	2.50	2.61	2.66	5.06	4.50	4.98
0.16	0.17	0.13	0.17	0.17	0.13	0.16	0.13	0.15	0.17	0.21
7.27	7.08	4.00	5.58	4.00	3.77	5.29	4.79	5.06	3.01	4.81
12.29	8.93	12.20	7.53	2.25	5.20	7.03	7.61	8.04	5.04	4.52
3.64	4.14	4.69	4.00	5.36	5.15	4.94	4.24	3.04	6.20	3.89
0.36	0.16	0.25	0.93	2.74	3.40	1.27	0.82	0.37	0.95	0.07
0.13	0.15	0.15	0.15	0.27	0.29	0.24	0.22	0.29	0.22	0.24
0.30	0.24	0.33	0.16	0.28	0.22	0.16	0.15	0.33	0.28	0.21
1.62	0.76	2.31	4.71	0.96	1.46	0.30	2.01	1.62	1.37	2.93
2.35	4.36	3.82	2.52	2.97	3.49	4.06	3.49	3.31	2.86	4.14
100.52	100.67	100.26	99.79	99.73	99.52	99.74	99.47	100.38	99.74	99.90
70	32	63	2	38	50	34	110	52	33	14
27	25	19	3.5	25	26	17	30	42	30	16
250	18	250	1	55	140	17	240	110	50	4
175	180	230	30	340	180	230	180	310	400	230
40	30	29	15	24	22	30	30	32	30	20
26	25	21	65	30	30	26	28	33	37	34
2,2	2.8	3.2	6.5	3.6	4.4	1.8	1.6	3.2	4.4	2.6
<30	<30	<30	<30	<30	<30	<30	<30	<30	<30	< 30
62	75	87	220	125	120	60	85	105	145	100
40	30	29	240	205	245	75	170	37	105	10
175	270	195	135	230	230	300	340	160	220	50
< 2	<2	2.5	<2	<2	<2	<2	<2	3.5	<2	< 2
55	46	38	2	19	18	35	32	4	15	19
12	12	18	20	20	17	12	12.5	18	18	13
43	52	38	56	50	43	42	33	55	46	60
<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	< 10

Origin and Evolution of the Tholeiitic Magma

The magma which gave rise to the basalt complex as well as to the other Liassic ophiolites from the megaslab situated in the Mureş ophiolitic suture was a tholeiitic magma formed in the mantle. As results from the plotting of the Sr mean content from basalts on the diagram of Hart et al. (1970) this magma formed at the depth of 150 km under the Mureş Ocean. This conclusion corresponds to the present views on the depth at which the tholeiitic magmas form in the oceanic area. As a result of the conditions under which it formed, this magma was a tholeiitic magma (Figs. 3, 4) of plate margin, as results from Figure 6 (after Pearce and Gale, 1977). As indicated by the diagram in Figure 5, and especially that in Figure 7 (after Pearce, 1980), this magma was an ocean floor tholeiitic magma.

The tholeiitic magma evolved in a magmatic chamber at depth, which had a considerable length, considering the fact that the present length of the sheeted dyke complex, which represented a median ridge of normal type (N–type MORB) (Savu, Stoian, 1988) on the Mureş Ocean floor and which can be correlated with a gravimetric and magnetic anomaly (Fig. 2), is of over 35 km. From this magmatic chamber the parallel dykes of normal (common) basalts and dolerites started raising in the spreading zone. As these magmas reached the ocean floor, they flowed on the paleoridge slope, thus giving rise to the ocean floor basalt complex, besides the sheeted dyke complex.

Comparing the differentiation diagram of the tholeiitic magma in the sheeted dyke complex (Savu et al., 1984; Savu, Stoian, 1988) with that of the ocean floor basalt complex (Fig. 4), some differences are found. First, on the basalt diagram the curve indicating the iron concentration trend is lacking, being quite obvious on the other diagram. This trend is only slightly indicated by some basaltic rocks, in which FeOt reaches over 12 %.

The Na_2O concentration trend is obvious also on the diagram for the basaltic complex (Fig. 4), although

Institutul Geologic al României

this concentration reaches only exceptionally 6.20 % Na_2O. As already shown, this Na_2O accumulation is due to both the spilitization through the reaction of the basaltic lavas with the oceanic water and the partial differentiation of the tholeiitic magma. The rocks with the greatest sodium concentration belong to the latter case. The fact that this Na_2O concentration is due to both phenomena is indicated by the rocks shifting on the diagram from the TH field towards the alkali corner through the CA field and not through that of the tholeiitic rocks, as in the diagram of the sheeted dyke complex. The position of this trend was determined by the weak iron concentration.

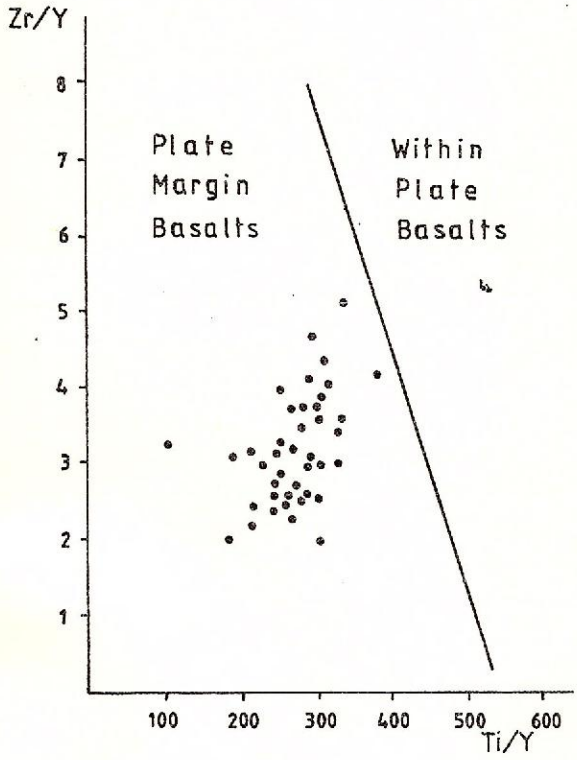

Fig. 6 — The Zr/Y - Ti/Y diagram (according to Pearce and Gale, 1977).

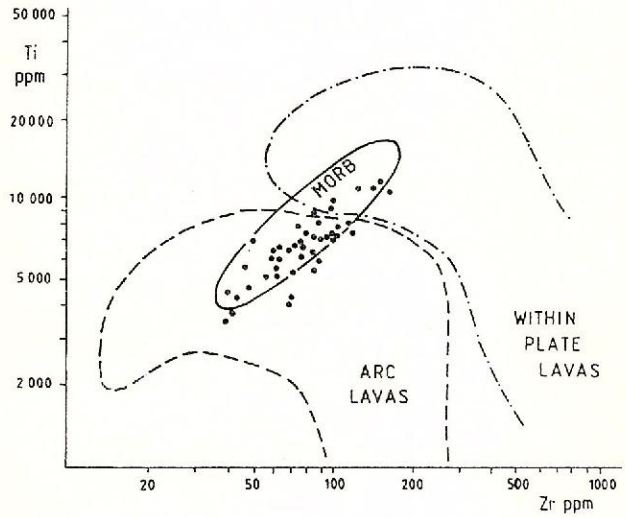

Fig. 7 — The Ti-Zr diagram (according to Pearce, 1980).

All these observations lead to the idea that the ocean floor basalt complex (O_1), as well as most of the intergranular basalt and dolerite dykes from the sheeted dyke complex (O_2) originated in a normal (common) tholeiitic magma. These rocks were emplaced in the first part of the differentiation process of the tholeiitic magma in the magmatic chamber.

The basic differentiated rocks (ferrogabbros, ferrodolerites, ferrohyalobasalts) and the acid differentiated ones (leucocrates), such as albite trondhjemitic diorites, granophyres and the albite ocean ridge granitoids from the sheeted dyke complex (Savu, Stoian, 1988) originated in the residual tholeiitic magmas, during the final stage of the evolution of the magmatic chamber at depth. That is why most of the dykes consisting of such differentiated rocks were emplaced among the other dykes of common basalts and dolerites.

Conclusions

The main ideas resulting from this paper are:

The ocean floor basalt complex (O_1) from the megaslab in the ophiolitic suture formed a field of ocean floor basalt flows in the structure of the Mureş Ocean crust on either side of the oceanic ridge (sheeted dyke complex).

This complex as well as the other associated ophiolitic complexes resulted from a tholeiitic magma formed in the mantle at the depth of ca. 150 km, developing in a magmatic chamber at depth.

The lavas from which the rocks of the basalt complex originated flowed on the floor of the Mureş Ocean, deriving from the normal tholeiitic magmas that gave rise to the common basalt and dolerite sheeted dykes at the beginning of the initial tholeiitic magma differentiation.

Consequently the basaltic rocks have the character of normal (common) tholeiitic basalts. Except some spilites rich in Na_2O, the basalt complex is devoid of the differentiated (basic and acid) rocks from the sheeted dyke complex, which formed in the final stage of the evolution of the magmatic chamber.

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Institutul Geologic al României



Fig. 1 – Basalt flows in pillow lava facies on the road (DN7) between Toc and Cuiaş (Drocea Mts).



Fig. 2 – Basalt flows in pillow lava facies in the erosion window at Visca.

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Fig. 1 – Lapilli pyroclasts level intercalated among the basalt flows in pillow lava facies on the road (DN7), between Toc and Cuiaş.

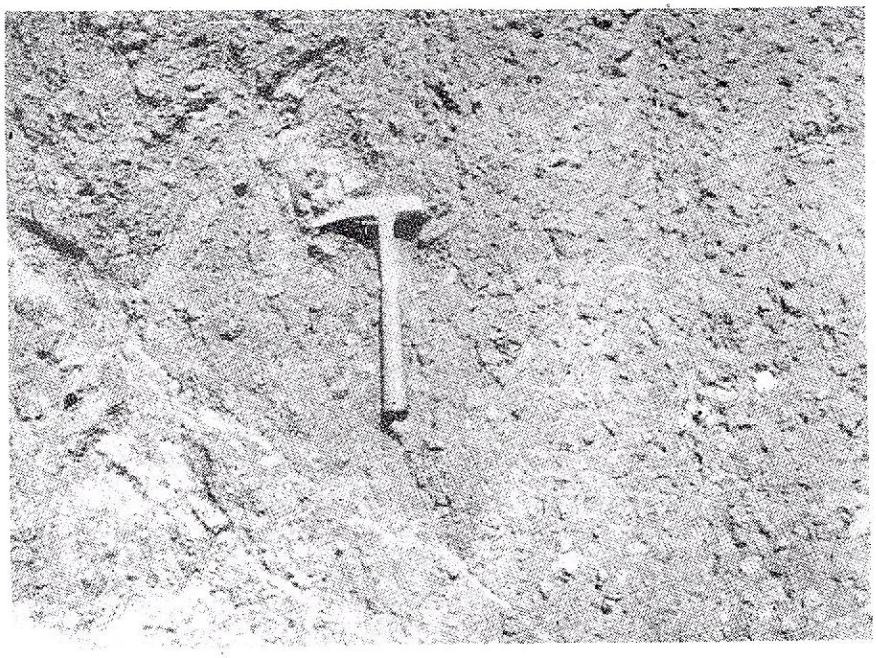

Fig. 2 – Pyroclastic rocks with larger blocks of basalts and hyalobasalts that are crossed by calcite veins and zeolites on the road (DN7) between Toc and Cuiaş.

Institutul Geologic al României

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PETROLOGICAL SIGNIFICANCE OF REE, U, Th, K, Rb, Sr, Hf AND OF THE $^{87}Sr/^{86}Sr$ IN THE BASALTIC COMPLEX OF THE MUREŞ ZONE

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Key words: Basalts. Rare earths. Trace elements. $^{87}Sr/^{86}Sr$. Mureş Zone. Apuseni Mountains.

Abstract: The Mureş Zone is an Alpine ophiolitic suture with a complex evolution, laying in the south and southeast of the Apuseni Mountains and in the north of the Banat. The ophiolitic rocks obducted from the Mureş Ocean in the axial zone of the suture belong to the first two complexes of an ocean crust: the pillow lava basaltic complex (O_1) and the sheeted dyke complex (O_2). This paper presents the distribution of REE, U, Th, K, Rb, Sr, Hf and the values of the Rb/Sr and $^{87}Sr/^{86}Sr$ ratios in the basaltic rocks of the first ophiolitic complex. These rocks have been sampled in the Cuiaş-Ciungani-Zam area. They formed an abyssal plain on the Mureş Ocean floor, situated south-east of the median ridge of the ocean, whose area is nowadays marked by the sheeted dyke complex. The values of the $^{87}Sr/^{86}Sr$ ratio (average value 0.702), the low contents of Ti, Y, Ba and Rb, the chondrite normalized REE distribution and the values of the ratios between these elements point out clearly that the basaltic complex (O_1) developed, as the sheeted dyke complex, in a normal-type segment (N-type MORB) of the Mureş Ocean median ridge. In this respect there are noteworthy the values of the Ba/La, (Ce/Yb)N, (Ce/Sm)N, and especially (La/Ce) N ratios, and the diagrams in Figures 2, 3 and 4. The rocks of the basaltic complex and those of the sheeted dyke complex are consanguineous. They resulted from the tholeiitic magma coming from the mantle, as shown by the REE patterns chondrite normalized and by the common Tb negative anomaly. The basaltic complex originates in the lavas erupted from the basalts and normal dolerites dykes, which are prevailing in the sheeted dyke complex.

Introduction

The distribution of REE, radioactive elements and of other trace elements in the Alpine ophiolites from the ocean crust slab from the Mureş Zone, which represents an Alpine ophiolitic suture with a complex evolution, has been dealt with by several researchers in their papers. Special attention has been given to the REE distribution in the sheeted-dykes complex (O_2) (Savu, Stoian, 1988). Lately, these elements have been also analysed in the Trascău Mts ophiolites, rocks coming from the two ophiolitic complexes known in the Mureş Zone (Savu, Stoian, 1991).

For a global image of the geochemistry of the Mureş Zone ophiolites we considered necessary a systematic study on the distribution of the above-mentioned elements also in the ocean floor basalt complex (O_1) from this zone of ophiolitic suture.

Distribution of the basaltic complex–petrographical and petrochemical aspects

The ocean floor basalt complex (O_1), a constituent of the ocean crust slab obducted in the Mureş Zone, is of Liassic age (180 Ma - Herz et al., 1974), as the sheeted dyke complex (O_2). It occurs in the Drocea Mts to the southeast and northeast of the sheeted dyke complex, which corresponds to the median ridge area of the Mureş Ocean. The structures of the ocean crust slab are diagonal versus the trend of the Mureş ophiolitic suture (Fig. 1). In the southeastern part of

Institutul Geologic al României

the sheeted-dyke complex, basalts probably formed and abyssal plain on the Mureş floor. The area lying between Cuiaş, Ciungani and Zam represents the widest area of basalts in the Mureş Zone (Fig. 1).

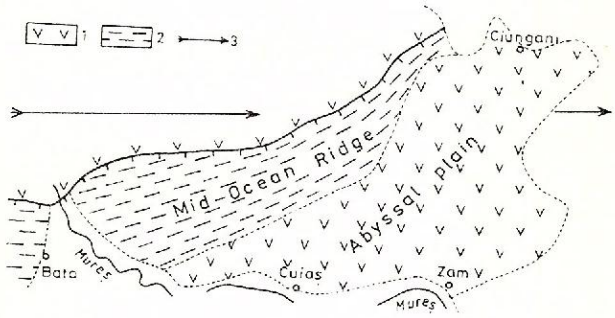

Fig.1 Structure of the ophiolitic rocks megaslab obducted in the Mureş ophiolitic suture area (Mureş Zone) and its location versus the suture trend. 1, area of the sheeted-dyke complex (median ridge of the Mureş Ocean); 2, basaltic complex area (abyssal plain) with pillow lava basaltic flows; 3, approximate trend of the ophiolitic suture in this segment.

East of the Vaţa-Zam lineament, which marks the boundary between the Liassic ophiolitic rocks, the basaltic complex respectively, and the Late Kimmerian island arc volcanics (J_3–K_1), basaltic ophiolites also occur from under volcanics in the erosion inliers at Visca, Luncoi and Almaşu Mare in the Metaliferi Mts, as well as in the Trascău Mts (Savu, 1983; Savu, 1991).

The rocks of the basaltic complex are represented by common basalts, amygdaloidal basalts, anamesites and spilites, in association with rare intercalations of tachylites, volcanic breccias and agglomerates, as well as dolerite sills. Bodies of gabbroic and ultramafic rocks are also emplaced in the complex.

The petrography of the basaltic rocks has thoroughly been studied by Savu (1962), Savu et al. (1970, 1989). These papers are accompanied by detailed maps which could be consulted for the geological structure. The basaltic rocks show an intersertal structure specific to the tholeiitic basalts. They consist of a network formed of basic plagioclase laths (An 50-70); its meshes include crystals of augite, magnetite, and possibly pyrites , as well as volcanic glass devitrified and partly chloritized under conditions of the hydrothermal metamorphism on the Mureş Ocean floor, in the epidote amphibolite, greenschist and zeolite facies (Savu, 1967; Coleman, 1977). In spilites, the plagioclase laths are albitic (An1) and augite is chloritized, concomitantly with the interstitial glass, a process which yielded leucoxene and iron hydrated oxides. Tachyllites are glassy rocks devitrified and partly chloritized within which plagioclase crystals are very rarely found. Dolerites display and ophitic or subophitic structure and consist of basic plagioclase laths (An 50–60), crystals of augite, often poikilitic, and magnetite and/or pyrite.

Table 1

Average chemical composition (44 analyses) of the rocks of the basaltic complex (according to Savu et al., 1994)

Oxides/Elements	Limits	Average
SiO_2 (%)	39.31 – 52.22	47.00
TiO_2	0.02 – 2.74	0.54
Al_2O_3	13.24 – 16.56	14.79
Fe_2O_3	1.65 – 8.24	5.15
FeO	1.48 – 12.79	4.96
MgO	3.77 – 9.74	6.25
CaO	4.52 – 16.12	10.07
Na_2O	1.79 – 6.20	3.64
K_2O	0.02 – 2.74	0.54
P_2O_5	0.08 – 0.29	0.17
FeOt/MgO	0.50 – 3.41	1.43
Ni (ppm)	2.00 – 210.00	84.11
Co	3.50 – 50.00	29.33
Cr	1.00 – 570.00	216.00
V	30.00 – 460.00	242.00
Sc	15.00 – 42.00	30.79
Y	15.00 – 65.00	26.25
Yb	1.20 – 6.50	2.65
Zr	40.00 – 220.00	83.75
Ba	10.00 – 215.00	72.52
Sr	65.00 – 600.00	221.25
Pb	2.00 – 18.00	2.52
Cu	2.00 – 100.00	36.42
Ga	9.00 – 22.00	14.32
Zn	30.00 – 70.00	44.22

According to Savu et al. (1994), the rocks of the basaltic complex constitute, beside those of the sheeted-dyke complex, a tholeiitic magmatic series. The chemical composition (Tab. 1) indicates common basalts, on the basis of the variation limits and the SiO_2 average values. The Na_2O contents higher than 2.50 per cent – a value typical of the tholeiitic basalts (Miyashiro, 1975) – point out the spilite presence. The rocks with higher TiO_2 and V contents emphasize the tendency of differentiation of the parental magma to the separation of a ferrogabbroic (ferrobasaltic) magma (Savu et al., 1984; Savu, Stoian, 1988), therefore the iron concentration (Fenner, 1929), and implicitly the concentration of the geochemically related elements, such as Ti and V. However, the differentiation process was quite weak, as indicated on the diagram in Figure 2, where the rocks of the basaltic complex plot in a field along the line representing the

Institutul Geologic al României

evolution of the basalts from the mid-ocean ridge segments of normal or "N-type MORB". The tendency of the tholeiitic magma to be enriched in Na_2O is quite obvious, as proved by the Na_2O contents of about 6 per cent, a value quite close to that of some sheeted dykes.

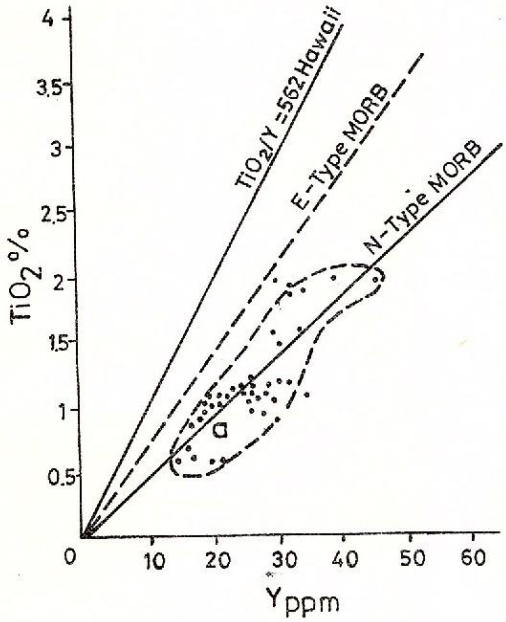

Fig. 2 TiO_2-Y diagram (after Perfit et al., 1980). Ti/Y=562, evolution line of the ocean island-type volcanism in Hawaii; E-type MORB, anomalous-type basalts line; N-type MORB, normal-type tholeiitic magma evolution line; a, basalt fields in the Mureş Zone.

The contents in other trace elements (Tab. 1) are also specific to the ocean floor tholeiitic basalts, as indicated by the low Ba values. The value lower than 2 of the FeOt/MgO ratio points out this feature.

Rb, Sr distribution and value of the $^{87}Sr/^{86}Sr$ ratio

The samples of basaltic rocks analysed for Rb, Sr and other trace elements in the present paper (Tabs. 2 and 3) have been taken off mostly from the widest area of the basaltic complex, between Cuiaş, Ciungani and Zam. Some basalts originate in the Visca inlier.

Rb and Sr contents have been determined by Energy Dispersive X-ray Analysis. In basaltic rocks Rb varies between 5.86 and 46.4 ppm the lowest values being obtained from the basalts in pillow lava facies at Toc, where Herz et al. (1974) also determined low Rb values. Sr presents low contents, too, specific to the ocean-floor basalts. This character is also given by the low value of the Rb/Sr ratio, which shows that the rocks of the basaltic complex, as well as those of the sheeted-dyke complex, originate in a magma formed in the mantle. The Sr average value (187 ppm) points out, according to Hart's et al. diagram (1970, Fig. 4), that magma was formed at a depth of 125 km.

The conclusion is clearly evidenced by the quite low values of the $^{87}Sr/^{86}Sr$ ratio, average value 0.702 ppm. Most of the values obtained are identical with those determined by Herz et al. (1974) on basaltic rocks from the same complex; a similar situation is that of the rocks from Toc, sampled from the same large outcrop of pillow lava basalts on the DN 7 Highway. Hart (1971), O'Nions et al. (1977) and Cohen et al. (1980) pointed out that the low values of the $^{87}Sr/^{86}Sr$ ratio are typical of the ocean floor basalts. Wood et al. (1979) mentioned that the value 0.702 is specific to N-type MORB. Most of the data obtained by us are situated below the value 0.703, considered as typical MORB value. One of the values is equal to it and only two of them are higher (Tab. 2).

REE distribution and petrological significance

The REE contents of the basaltic complex have been determined by neutron activation, except for Yb which has been analysed by C.Udrescu (in Savu et al., 1994), using the emission spectrography method. The data obtained are rendered in Table 3, in which the rocks analysed are separated into 3 groups after their SiO_2 content, which varies between 39 and 52 per cent.

The REE contents determined are similar with those from normal basalts of the sheeted-dyke complex in the Mureş Zone (Savu, Stoian, 1988), formed in the same median ridge of the Mureş Ocean. In the basalt and sheeted-dyke complex, the HREE content is lower in comparison with the LREE content, which represents a characteristics of the ocean floor basalts series. The smaller contents of HREE versus LREE are also emphasized by the ratio of the total of the two groups of elements, which in the three groups of basalts vary between 4.06 and 5.79 ppm (general average value 4.98). The difference between LREE and HREE is higher in case of the least differentiated basalts from the first group, in which SiO_2 is lower than 48 per cent.

REE averages in basalts are close to those presented by Werner (1984) for the ocean ridge tholeiites. As in case of the sheeted-dyke complex, there is a difference of the Tb content in basalts, too (Tab. 3). The values are lower and, consequently, this element shows a quite obvious negative anomaly (Fig. 3). In our opinion the anomaly is the result of the partial melting of the source from the mantle, as it was also suggested by Bougault et al. (1979) who considered that the fractional crystallization of the parental tholeiitic magma

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also participated in the process. As the differentiation process was weak in the basaltic complex rocks, as shown by the SiO_2 variation, all the yielded rocks correspond to basalts. Consequently, the possibility that the strong Tb anomaly should have been determined by the differentiation process is out of question. A slight tendency of Tb to form negative anomalies is also observed in case of the Pacific Ocean basalts (site 482 and 428) mentioned by Saunders (1984), as examples of the N-type MORB tholeiites.

Ocean median ridge. In the field they occur only in the sheeted-dyke complex area or in association with the gabbro bodies.

As already mentioned, there is a general trend of increase of the REE content, from basalts with a SiO_2 content lower than 48 per cent, to those with a SiO_2 content higher than 49.5 per cent; the only exception is that within the intermediary basalt group there is a decrease of the Ce and Sm contents as against the first group, although in the third group (SiO_2>49.5 per cent) the behaviour of the two elements is again normal.

Table 2

Rb, Sr (ppm) contents and the values of the Rb/Sr and $^{87}Sr/^{86}Sr$ ratios in the basaltic rocks

No.	Rock	Locality	Rb	Sr	Rb/Sr	$^{87}Sr/^{86}Sr$
1	Basalt	Toc (DN7)+	8.79	272	0.032	0.701
2	Hyalobasalt	Toc (DN7)	5.86	162	0.036	0.702
3	Basalt	Zam (quarry)	46.4	173	0.268	0.703
4	Basalt	Zam (quarry)	13.4	138	0.097	0.704
5	Basalt	Roşia Nouă	23.3	258	0.090	0.702
6	Tachylite	Corbeşti	14.5	241	0.060	0.702
7	Basalt	Zam (quarry)	23.6	131	0.180	0.704
8	Basalt	Almăşel	14.1	120	0.117	0.700

+ DN7 indicates National Highway no. 7, namely Bucureşti–Arad highway

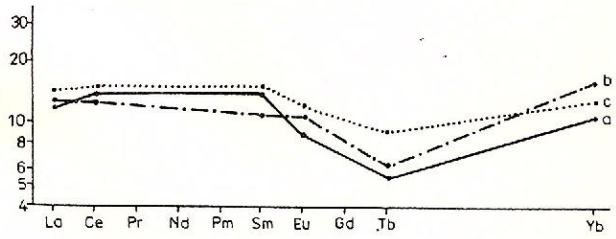

Fig. 3 Chondrite normalized REE patterns in basaltic rocks. a, basalts with SiO_2<48 %; b, basalts with SiO_2 between 48 and 49.5 %; c, basalts with SiO_2>49.5 %.

Due to the slightly differentiation of the basaltic magma, the REE differences between the basalts of the three groups are quite small (Tab. 3). Nevertheless, a tendency of increase of the REE contents concomitantly with the SiO_2 increase is clearly observed on the diagram in Figure 3, which presents the chondrite normalized REE patterns (Nakamura, 1974), of the basalts from the three petrographic groups. All the patterns range within the normalized values of 10 and 20, except for Tb which presents much more lower normalized values. In this field plot the N-type MORB basalt patterns of the sheeted-dyke complex, which shows that the basalt complex includes only lavas corresponding to basalt dykes. It is likely that the basic and acid-albitic differentiates of the tholeiitic magma remain at depth in the axial zone of the Mureş

According to their outline, the patterns of the three groups of basalts – with the common Tb anomaly which proves their consanguinity – are almost identical with the rock patterns from the sheeted-dyke complex in the Mureş Zone (Savu, Stoian, 1988, Fig. 3). Such aspects point to the common origin of the two ophiolitic complexes – their formation in the median ridge of the Mureş Ocean, of Jurassic age.

The $(Ce/Yb)N - (Ce)N^2$ diagram in Figure 4, simplified according to Saunders (1984), shows that more than 70 per cent of the rocks of the basaltic complex in the Mureş Zone plot in the basalts field from the median oceanic ridge segments of N-type MORB, except for a few ones which fall outside of this field. This position of the basalts on the diagram shows that, as on the diagrams in Figures 2 and 3, the median ridge segment of the Mureş Ocean, in which the two ophiolitic complexes were formed, is of N-type similar to that emphasized by Schilling (1975). This conclusion is based on the average values of the ratios presented in Table 3. It is notable the smaller values of the Ba/La ratio (2.99–4.11), (Ce/Yb)N ratio (0.90–1.14) and (Ce/Sm)N ratio (0.98–1.20). According to Saun-

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Table 3
REE contents in basaltic rocks

No.	Rock	Locality	La	Ce	Sm	Eu	Tb	Yb
		a) Rocks with $SiO_2 < 48$ %						
1	Basalt	Visca	3.0	8	1.8	0.87	0.37	4.0
2	Basalt	Visca	3.0	8	2.4	0.71	0.54	3.0
3	Basalt	Roşia Nouă	6.0	13	3.0	0.90	0.38	–
4	Dolerite	Zam (quarry)	2.7	15	2.6	0.55	0.52	2.1
5	Basalt	Zam (quarry)	3.5	14	2.7	0.76	–	3.6
6	Basalt	Pietriş	3.1	14	1.9	0.35	–	1.4
7	Basalt	Almăşel	5.2	12	5.8	0.72	0.86	3.6
8	Basalt	Pietriş	4.6	14	4.8	0.80	0.37	2.1
9	Basalt	Pietriş	3.6	10	1.8	0.68	0.39	3.0
10	Dolerite	Pietriş	4.6	14	2.9	0.67	–	1.4
11	Hyalobasalt	Corbeşti	4.6	10	2.2	0.61	–	1.3
12	Tachylite	Corbeşti	3.8	10	4.9	0.46	–	1.2
13	Basalt	Toc DN7	1.3	10	2.1	0.66	–	1.4
14	Basalt	Roşia Nouă	2.3	10	2.2	0.46	0.40	2.9
	Averages		3.66	11.57	2.94	0.66	0.42	2.21
LREE/HREE = 5.79; La/Yb = 1.66; (La/Sm)N = 0.87; (Ce/Yb)N = 1.12; Ba/La = 4.11; (La/Ce)N = 0.89; (Ce/Sm)N = 0.98; La/Th = 3.02								
		b) Rocks with SiO_2 between 48 and 49.5 %						
15	Hyalobasalt	Toc DN7	5.0	8	2.8	0.70	0.30	2.2
16	Basalt	Toc DN7	4.9	10	2.3	0.74	–	2.8
17	Basalt	Roşia Nouă	4.0	12	1.6	0.75	0.44	–
18	Basalt	Visca	4.0	9	3.8	1.18	0.42	7.0
19	Anamesite	Vaţa Valley	2.0	12	1.7	0.69	–	2.8
20	Spilite	Vaţa Valley	5.0	15	1.9	0.90	0.42	3.5
21	Hyalobasalt	Căzăneşti	3.0	11	2.2	0.94	0.48	3.4
22	Basalt	Căzăneşti	4.0	12	2.1	0.71	0.43	3.5
	Averages		3.99	11.12	2.30	0.83	0.41	3.60
LREE/HREE = 4.06; La/Yb = 1.27; (La/Sm)N = 1.11; (Ce/Yb)N = 0.90; Ba/La = 3.41; (La/Ce)N = 1.00; (Ce/Sm)N = 1.20; La/Th = 8.2								
		c) Rocks with $SiO_2 > 49.5$ %						
23	Dolerite	Căzăneşti	2.0	12	1.3	0.70	0.40	2.9
24	Basalt	Vaţa Valley	3.0	9	2.6	0.85	0.34	3.5
25	Dolerite	Vaţa Valley	3.0	8	2.0	0.80	0.32	2.5
26	Basalt	Vaţa Valley	4.0	10	2.9	0.93	0.43	4.7
27	Basalt	Roşia Nouă	5.0	6	4.6	1.20	0.58	–
28	Basalt	Roşia Nouă	8.0	18	4.4	1.05	0.48	–
29	Basalt	Zam (quarry)	4.5	20	2.7	0.86	0.58	4.4
30	Basalt	Micăneşti	5.3	19	5.0	0.62	0.48	3.2
	Averages		4.35	12.75	3.19	0.88	0.45	3.02
LREE/HREE = 5.10; La/Yb = 1.64; (La/Sm)N = 0.95; (Ce/Yb)N = 1.14; Ba/La = 2.99; (La/Ce)N = 0.95; (Ce/Sm)N = 1.00; La/Th = 2.6								

ders (1984), the values of the (La/Sm)N ratio, smaller or equal to one, point out the N-type MORB basalt characters of the rocks. Therefore the rocks from the basaltic and sheeted-dyke complexes are N-type tholeiites. The general average value of this ratio is 0.97. It also differentiates the N-type MORB basaltic rocks in the Mureş Zone from the basalts formed in a "plume ridge" (P-type MORB), or in other words basalts of anomalous type (E-type MORB). This delimitation is clearly shown on the diagram in Figure 2.

Quite characteristic in this respect is the value of the (La/Ce)N ratio which varies between 0.90 and 1 for all the three basalt groups in the Mureş Zone, values typical of the N-type MORB basalts.

Institutul Geologic al României

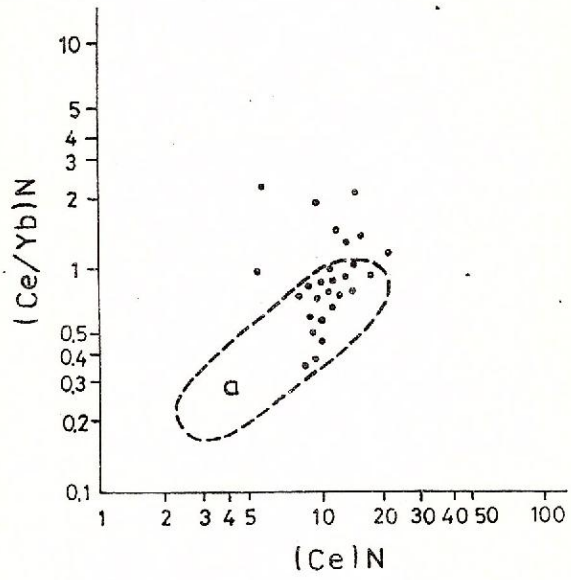

Fig. 4 - (Ce/Yb)N - (Ce)N diagram (after Saunders, 1984). a, normal-type basalt field ((N-type MORB) delimited according to the mentioned author's data.

Distribution of radioactive elements and of Hf

The U, Th and K contents in the rocks of the basaltic complex (O_1) (Tab. 4) have been determined by natural gamma spectrometry. They present low values characteristic of the ocean floor basalts, as shown by Savu et al. (1989) in the Vaţa and Visca areas. Thus, U contents vary between 0.3 and 1.5 ppm (average value 0.85 ppm); Th contents range from 0.2 to 3.2 ppm (average value 1.21 ppm). The values of the Th/U ratio are also low (0.17–3.5), the average value being 1.44 ppm. All the data are specific to the basalts from the median ocean ridge basalts (MORB).

Potassium displays very low values, ranging from 0.03 to 0.75 per cent. The lowest contents occur in spilites, rocks from which this element was removed during the spilitization process. The average value of the K content in basaltic rocks is 0.33 per cent, specific of the abyssal tholeiites (Miyashiro, 1975) or ocean floor tholeiites.

Hafnium varies in the study basalts between 0.6 and 2.5 ppm, with an average value of 1.35 ppm Hf. The Zr/Hf ratio, with an average value of 78, is also a good indicator of the origin of the ocean floor basalts of the complex O_1.

Conclusions

The rocks of the basaltic complex of the Mureş Zone occurred in the Mureş Ocean probably in an abyssal plain, southeast of the median ridge, marked nowadays by the sheeted-dyke complex. The values of the $^{87}Sr/^{86}Sr$ ratio (average value 0.702), the low contents of Ti, Y,. Ba, as well as the REE distribution and the values of the ratios of these elements, point out clearly that the basaltic complex O_1 developed, as the sheeted-dyke complex O_2, in a N-type MORB segment of the median ridge of the Mureş Ocean, Liassic in age. The values of the ratios Ba/La, (Ce/Yb)N, (Ce/Sm)N and especially (La/Ce)N are quite obvious in this respect. This tectonic setting of the basalts is rendered clearly on the diagrams in Figures 2, 3 and 4. The

Table 4

Contents (ppm) of radioelements and of Hf in basaltic rocks +

No.	Rock	Locality	U	Th	Th/U	K %	Hf	Zr/Hf
1	Basalt	Zam (quarry)	0.4	1.0	2.5	0.75	2.5	58.0
2	Dolerite	Zam (quarry)	0.4	0.8	2.0	0.46	–	–
3	Basalt (spilite)	Zam (quarry)	0.5	0.7	1.4	0.03	1.0	100.0
4	Basalt	Pietriş	1.0	1.2	1.2	0.19	–	–
5	Basalt	Almăşel	1.2	0.2	0.17	0.06	1.1	145.0
6	Basalt	Micăneşti	1.2	1.9	1.6	0.37	1.3	80.0
7	Basalt (spilite)	Pietriş	0.5	0.2	0.4	0.03	1.7	52.9
8	Basalt	Pietriş	1.3	1.0	0.8	0.55	0.6	125.0
9	Hyalobasalt	Toc	0.9	1.2	1.3	0.36	0.9	68.8
10	Dolerite with albite	Pietriş	0.9	3.2	3.5	0.03	1.4	28.6
11	Hyalobasalt	Corbeşti	1.1	1.4	1.3	0.55	–	–
12	Tachylite	Corbeşti	1.0	2.8	2.8	0.45	1.6	26.2
13	Basalt	Toc DN7	1.5	2.0	1.3	0.34	0.8	50.0
14	Basalt	Toc DN7	0.6	0.4	0.7	0.12	0.6	125.0
15	Basalt	Roşia Nouă	0.3	0.2	0.7	0.62	–	–
	Averages		0.85	1.21	1.44	0.33	1.22	78.2

+ Zr contents used for the calculation of the Zr/Hf ratio were determined by C. Udrescu (in Savu et al., in press)

Institutul Geologic al României

rocks of the basaltic complex are consanguineous with those of the sheeted-dyke complex. They have resulted from the tholeiitic magma coming from the mantle, a fact proved by the Tb common negative anomaly and by the similarity of the chondrite normalized REE patterns. The obtained data show that the basaltic complex is the result of the lavas erupted from the dykes of basalts and common dolerites of the sheeted-dyke complex.

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Institutul Geologic al României

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INTRUSIVE BRECCIAS RELATED TO MICROGRANITIC RHYOLITES IN THE DUMBRAVA–CĂPUŞU MIC AREA (NW TRANSYLVANIA)

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Key words: Breccia. Volcanic breccia. Rhyolite Brecciation. Major elements. Minor elements. NW Transylvania zone.

Abstract: On the northern border of Gilău Mts, between Dumbrava and Căpuşu Mic, an area which is limited by two NW-SE oriented fractures stands out, in which the basement consisting of crystalline schists is transgressively overlain by Senonian conglomerates, sandstones and siltstones, pierced or covered by banatitic calc-alkaline rocks. The latter ones are represented both by quartz andesitic volcanics with pyroxene and amphiboles (Maastrichtian-Paleocene, of the first cycle) and intrusive rocks (Paleocene, second cycle): andesites with amphiboles and biotite, microgranitic rhyolites. Here basalt andesites with pyroxene ± olivine crop out too, coming from another magmatic source. Microgranitic rhyolites representing final differentiation products of granitic-granodioritic magma, are associated with intrusive breccias, in whose generation fluidization processes have taken place with different intensities from one place to another, being more substantial near the magmatic body. Intrusive breccias with mixed composition always contain lithic or monomineral fragments which have been broken from the crystalline basement or from Senonian deposits. Eruptive rock fragments from the breccia are represented both by quartz andesites with pyroxene and amphiboles (at Căpuşu Mic), quartz andesites with amphiboles and biotite (SE of Dumbrava) and especially by abundant and omnipresent microgranitic rhyolites. At least two generations of breccias with the same composition are pointed out, the latest one being fine-grained (tuffaceous) with several trends and pierces or desintegrates the first one. On the whole the brecciation process developed similar to those from the breccia pipe, diatreme and pebble dyke systems, but less evolved in the case we are referring to. The evolution of the brecciation process was stopped by a rapidly consolidation of the microgranitic rhyolite body.

Introduction

In the region of the Dumbrava and Căpuşu Mic localities (Cluj county) on the northern border of the Gilău Mts, in the area framed by two NW-SE oriented fractures, there are banatitic magmatic rocks, piercing or covering the Senonian sedimentary deposits (Fig. 1). In this area, associated to rhyolites-microgranitic rhyolites which represent the latest differentiation products of banatitic calc-alkaline magma, mixed intrusive breccias have been pointed out; these breccias contain fragments from both Senonian sedimentary depostits and from crystalline schists (basement of region), as well as from banatitic magmatites (from rhyolites to microgranitic rhyolites, inclusively) but as it seems, not from pyroxenic ±olivine andesite. Senonian sedimentary deposits and banatitic magmatic rocks are overlain by Paleogene formations and Quaternary deposits.

The presence of intrusive breccias, yielded by failed explosions and, more rarely, piercing to the surface, usually accompanied by fluidization and injection processes, has been stated by one of the authors (Ştefan, 1980; Ştefan, in Mantea et al., unpubl. report) for the Vlădeasa eruptive massif. In every distribution area of the banatitic magmatites, breccias are associated to the latest magmatic (acid) differentiation products, enriched both in SiO_2 and K_2O, and in fluids. Up to the present days it has been considered that these breccias accompany only rhyolites in the main body of Vlădeasa (Cycle I, essentially a volcanic one) or microgranitic rhyolites from the irregular shaped bodies.

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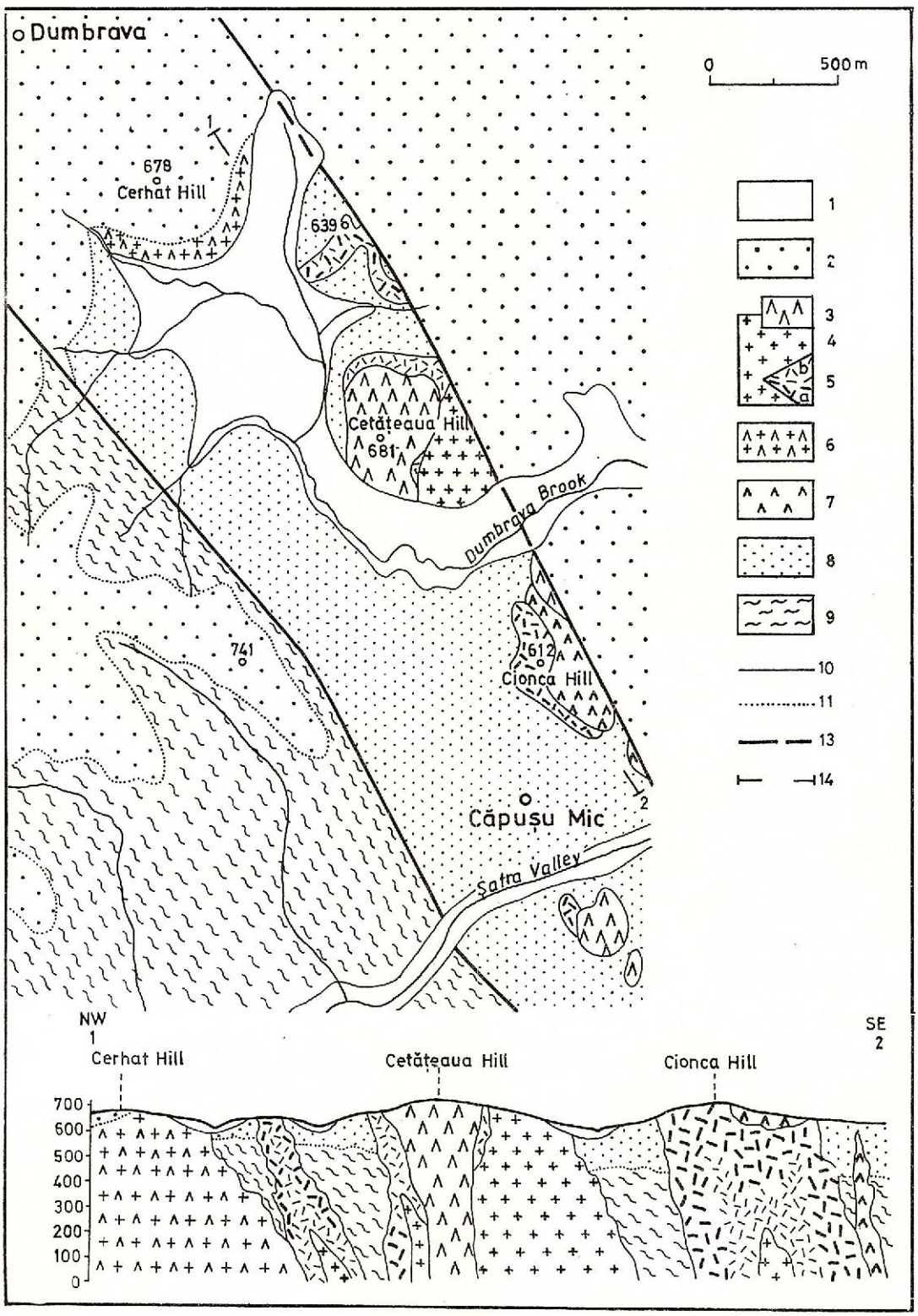

Fig. 1 — Geological sketch of the Dumbrava-Căpușu Mic area. 1, Quaternary: alluvial, proluvial, colluvial and deluvial deposits; Paleogene: 2, Jibou Beds-Cluj Limestone; 3, basalt andesites with pyroxene ± olivine; 4, micro-granitic rhyolites; 5, intrusive breccia; coarse (a) and tuffaceous (b); 6, quartz andesites with amphiboles and biotite; Paleocene-Maastrichtian; 7, quartz andesites with pyroxene and amphiboles; 8, Campanian-Santonian: sandstones, conglomerates, siltstones, 9, Cambrian-Upper Precambrian: Biharia Series: sericitic-chlorite schists, crystalline dolomites and limestones, graphite schists, amphibolites, granitoids. Conventional signs: 10, geological limit in general; 11, unconformity limit; 12, fault; 13, position of the geological section; 14, geological section.

Institutul Geologic al României

probably representing marginal facies of greater granite volumes from the depth (belonging to the 2nd cycle, only intrusive).

Detailed observations in the field, corroborated with microscopic studies, first pointed out breccias which accompany final, generally vein magmatites, with rhyolite-granitic composition and texture, enriched in alkalies, resulting from granodioritic-granitic banatitic magma differentiation.

Geological Setting

In the western part of the area, which is limited by the two fractures, crystalline formations develop, being represented by: chlorite schists ± epidote, albite; sericitic-chlorite schists; crystalline limestones and dolomites, calcareous schists; graphitic schists and amphibolites. These rocks belong to the Biharia Series from the Lupşa Nappe (Dimitrescu, Dimitrescu, Săbău; in Rusu et al., 1991, unpubl. report) which thrusts the formations of the Someş Series, developed westwards.

The crystalline formations from both series are pierced by several dykes of banatitic magmatic rocks with a NW-SE trend, represented by porphyric quartz micromonzodiorites, dacite-porphyric microgranodiorites, microgranitic rhyolites and lamprophyres.

Within the main zone under study, limited by the two fractures, the first banatitic magmatic rocks emplaced over the Senonian sedimentary formation, represented by sandstones, conglomerates and grey siltstones, are quartz andesites with pyroxene and hornblende. These magmatites can be correlated with those from the Vlădeasa eruptive massif by which the banatitic volcanism from the first cycle begins. It must be remarked that in the Dumbrava-Căpuşu Mic area dacitic and rhyolitic volcanics from the first cycle of banatitic magmatism do not occur.

Within the 2nd banatitic magmatic cycle two petrotypes are encountered in this area: 1) quartz andesites with hornblende and biotite in the Cerhat Hill at Dumbrava, which includes several fragments of crystalline schists and blocks of porphyric microdiorites and 2) microgranitic rhyolites in the Cetăţeaua Hill (SE of Dumbrava), which close this evolution cycle of the banatitic magmatism.

The intrusive breccias with changing granulation and composition have been outlined both near Dumbrava, north of the microgranitic rhyolite at the Cetăţeaua Hill, and to the south, at the Cionca Hill (north of Căpuşu Mic) and at Căpuşu Mic. The extension of this breccia can be larger as it is figured on the map (Fig. 1), as Paleogene and Quaternary deposits hinder observations.

Basalt andesites with pyroxene ± olivine, with different origin from proper banatitic rocks, in this case seem to be emplaced at least partly penecontemporaneously with microgranitic rhyolites (Ştefan et al., 1985).

The Cretaceous sedimentary formations and the eruptive rocks entirely, including intrusive breccias situated between the two fractures, are discordantly covered by red continental deposits of the Jibou Beds (Paleocene-Lower Lutetian). West of this area, Paleogene deposits transgressively cover a relief moulded on crystalline schists pierced by banatitic bodies (dacite-porphyric microgranodiorites type). East of the area which is limited by the two faults, only Paleogene and Quaternary deposits develop.

In the Dumbrava-Căpuşu Mic area Quaternary deposits occupy large spaces.

Physiography, composition and emplacement of breccias

Breccias are massive, grey, reddish, white or green rocks with variable granulation (from few centimetres to submilimetres). Lithic and mononmineral fragments are usually angular and rarely a little rounded.

The lithologic composition of breccias is variable according to the nature of the formation that builds it up. Thus, in the Căpuşu Mic area one can notice that the transition from Senonian deposits (practically undisturbed) to mixed composition breccias is accomplished through a breccia area consisting only of angular or subrounded sandstone and microconglomerate fragments. Solitary banatitic rhyolite fragments occur gradually and in the end turn into breccias rich in eruptive rocks (hornblende, pyroxene andesites and especially rhyolitic fragments). In all cases crystaloclasts and lithoclasts from prebanatitic rocks are obviously and can proceed either from Senonian sedimentary rocks or from the crystalline schist basement. Breccias were also encountered which were almost entirely constituted of banatitic magmatic (andesitic and rhyolitic) rocks, in which one cannot see but mica flakes and rare small quartz grains which had been yielded from older formations.

The breccias at Dumbrava come into contact both with Senonian micaceous fine-grained sandstones or siltstones and with microgranitic rhyolites from Cetăţeaua Hill. As at Căpuşu Mic, basaltoid andesites from this area do not seem to be involved in breccia formation.

Lithologically, the breccias from Dumbrava area are constituted both of fragments taken over from older formations (quartz, feldspar, sericite, biotite, quartzites, chlorite-sericitic schists, gneisses, etc.) and of banatitic magmatic rocks, represented by quartz andesites with hornblende and biotite and especially by microgranitic rhyolites. The crystalline schist fragments with centimetric sizes are probably directly

Institutul Geologic al României

taken over from the basement of the metamorphic rocks of the Biharia and Someş Series. On the other hand, fine mica flakes and several fine quartz grains, probably come from Senonian sedimentary formations.

Microgranitic rhyolites, sometimes eutaxitic with variable textures of the groundmass (granophyric, microgranitic, cryptocrystalline), locally as blocks of a few centimetres, are to be found as elements in the east of Dumbrava breccias, beside crystalline and quartz fragments (Pl. I, Fig. 1; Pl. I, Fig. 5; Pl. II, Fig. 1; Pl. II, Fig. 5). In other cases, andesite fragments occur beside microgranitic rhyolite fragments, being included in a fine-grained matrix (Pl. I, Fig. 6).

The granulation of the breccias east of Dumbrava and Cetăţeaua Hill is variable, as in case of those at Căpuşu Mic. In the first case, the tufaccous fraction with the same composition with the corse breccia becomes more important, especially near the microgranitic rhyolite body.

As we have already stated, the microgranitic rhyolites from the Cetăţeaua Hill body represent the final differentiation product of banatitic calc-alkaline magma. Previous data on these rocks from the viewpoint of the general banatitic magmatism evolution in the Apuseni Mts have been presented in recent papers (Ştefan et al., 1985; Ştefan et al., 1988; Ştefan et al., 1992).

Rhyolite fragments from the breccias at Căpuşu Mic are similar to those which crop out in the Cetăţeaua Hill body, being often intensively altered. Feldspars are totally replaced, both with clay minerals and silica (Pl. II, Fig. 2).

Microgranitic rhyolites from the breccia fragments at Dumbrava are similar to the ones from the body at Cetăţeaua Hill, but some differences are also noticed. Thus, in the rhyolite fragments mass, sometimes with a banded or massive structure and often with granophyric texture of the groundmass, biotite and a greater fresh oligoclase ratio can be noticed.

The comparison of the Cetăţeaua Hill rhyolites with those within the intrusive breccias was based on previously published data (Ştefan et al., 1985; Ştefan et al., 1988), as well as on the analyses of a rhyolite block from the breccia east of Dumbrava (Table). From the Table which shows the contents in major and minor elements for microgranitic rhyolites at Cetăţeaua Hill (1–4) and the values for the rhyolite block samples (5) from the breccia, comparable values result as for the analysed elements. The higher contents in CaO, Na_2O and Sr in rhyolite from breccia are probably due to the higher ratio of fresh oligoclase.

In the distribution diagram of the contents in rare earth, chondrite normalized (Henderson, 1984), for microgranitic rhyolites from the body at Cetăţeaua Hill (sample 9281) and the sample from the breccia east of Dumbrava (sample 9407), the similarity of the two curves can be observed, excepting a smaller Eu anomaly for the microgranitic rhyolite from the breccia (Fig. 2). This suggests that blocks of rhyolites from the breccia come from a rhyolitic-microgranitic magma fraction which has earlier crystallized, containing a larger amount of plagioclase.

Table

Major (%) and minor (ppm) elements in the Cetăţeaua Hill rhyolites (1–4) and breccias (5)

	1	2	3	4	5
	1408 B	573	7655	9281	9407
SiO_2	76.07	76.70	78.72	78.76	71.80
TiO_2	0.15	0.12	0.10	0.16	0.12
Al_2O_3	12.85	12.69	9.30	11.25	14.75
Fe_2O_3	1.17		0.62	0.43	0.63
FeO	0.21	1.14	0.08	0.03	0.26
MnO	0.04	0.02			0.04
MgO	0.54	0.04	0.65	0.10	0.19
CaO	0.43	0.35	1.19	0.50	1.07
Na_2O	2.86	3.08	2.80	2.67	3.59
K_2O	4.14	4.55	4.40	4.47	5.01
P_2O_5	0.05	0.01	0.66	0.04	0.12
H_2O^+	1.48	0.96	1.05	1.07	1.79
Pb	36.00	30.00	28.00	17.0	34.00
Cu	8.00	4.50	5.00	5.0	3.60
Ga	13.00	17.00	13.00	14.0	15.00
Cr	2.00	3.50	<1	8.0	2.00
V	4.00	4.00	<4	6.0	5.00
Sc	3.00	5.00	2.00	3.0	<2
Nb				20.5	16.00
Zr				145.0	100.00
Yb	4.20	1.00	2.90	2.4	<1
Y	20.00	17.00	21.00	19.0	13.00
Sr	22.00	<10	12.00	40.0	155.00
Ba	950.00	1250.00	820.00	1400.0	1100.00
Rb				167.0	163.00
U				4.1	4.60
Th				21.0	21.30
La	37.00	50.00		26.0	33.00
Ce				71.0	81.00
Sm				3.5	4.80
Eu				0.5	1.02
Tb				1.4	1.00
Hf				5.5	7.10

The lithology/granulation relationship between diverses breccia types, and implicitly the generation process of the latter, are very much similar to those which have been described in several cases from the breccia pipe, diatremes and pebble dyke systems (McCallum, 1985). The action of some fluidization processes within the above-mentioned systems has been invoked by several researchers. The experimental proofs have been

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provided by McCallum (1985). In the region we refer to, although the products of a fluidization process are obviously, the importance of this phenomenon is relatively smaller, as well as the evolution degree of breccias.

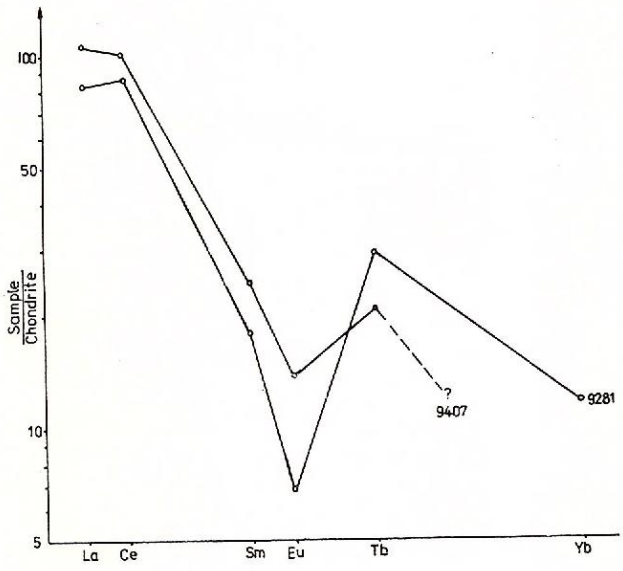

Fig. 2 - Distribution of rare earth.

The fluid pressure which has resulted due to granitic magma crystallization determined both the fracture of the crystallized envelope of the magmatic body and of the surrounding rocks, especially from its roof. Explosion processes have taken place subsequently, but could only rarely pierce to the surface, being followed by fluidization phenomena. Coarse breccias, especially at Cetăţeaua Hill and Dumbrava, are pierced by fine grained tuffaceous sequences which have resulted by desintegration of previous formed breccias in fine (litho-crystalloclastic) grains, which have been driven by a gas stream, characteristic of fluidization phenomena. Products of these processes can be noticed both at macroscopic (Pl. I, Fig. 1) and microscopic scale. In this case, either piercing phenomena of some microgranitic rhyolite fragments by fine-grained tuffaceous breccias (Pl. II, Fig. 3), or only dismembering trends of larger elements from the same magmatites (Pl. II, Fig. 4), can be noticed.

Fine-grained tuffaceous breccias with the same composition and much more uniform from the viewpoint of the granulation, undoubtly products of fluidization processes, intrude in the stratification planes of grey-blackish siltstones north of Cetăţeaua Hill. Such fine-grained rocks can be easily mistaken for common tufaceous rocks, volcanoclastics (Hedenquist and Henley, 1985), or even for epiclastics (Keihaln, fide Torske, 1983).

Within tuffaceous breccias at Cetăţeaua Hill, in the context of the presence of a fine grained material with a high clay ratio, spherical formations resembling the accretionary lapillis have been pointed out; these formations are characteristic for volcanic products, but are very likely to occur under intrusive conditions, as it has been experimentally demonstrated (McCallum, 1985).

On the whole, breccias from the investigated area are less evolved and not accompanied by important hydrometasomatic solutions. Brecciation, including fluidization processes, is followed by the intrusion of the Cetăţeaua Hill microgranitic rhyolite, which stopped the brecciation processes. The latest rhyolite related solutions contributed to the already consolidated rock silicification.

In the area of Căpuşu Mic village there is also a deep-seated rhyolitic body, next to which the participation ratio of the fine breccias resulted in fluidization processes would increase. In this case the presence of a more important hydrometasomatic solution circulation cannot be ruled out.

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A. ŞTEFAN et al. – INTRUSIVE BRECCIAS MICROGRANITIC RHYOLITES

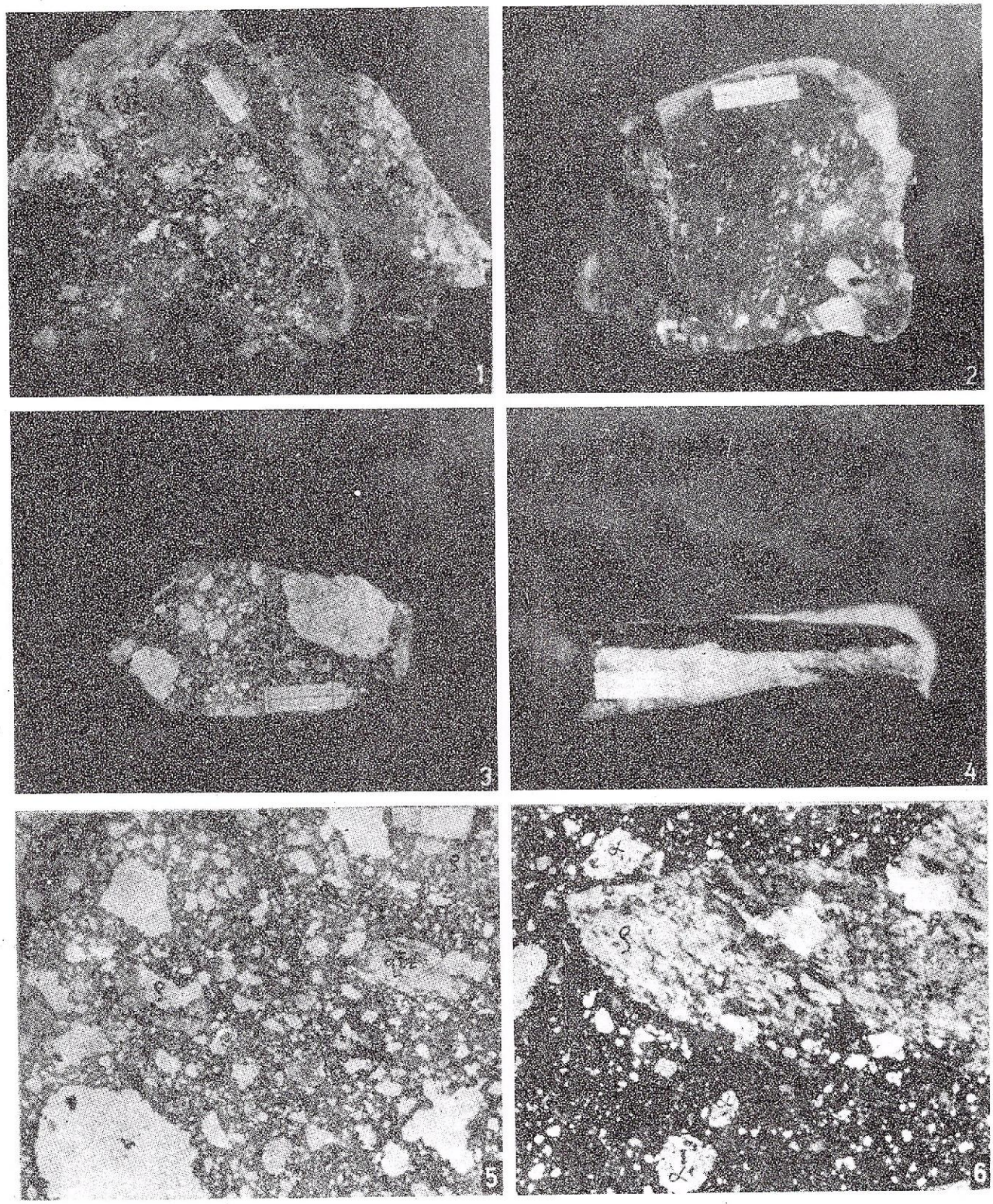

Fig. 1 – Coarse breccia pierced by tuffaceous breccia, SE of Cerhat Hill.
Fig. 2 – Coarse breccia with rhyolite fragments injected with tuffaceous breccia, Căpuşu Mic.
Fig. 3 – Breccia with altered rhyolite fragments, Căpuşu Mic.
Fig. 4 – Tuffaceous breccia in contact with black siltites. Cetăţeaua Hill.
Fig. 5 – Breccia with micaceous quartzite (qtz) and rhyolites (ρ) elements, Cetăţeaua Hill, N II, x 20.
Fig. 6 – Tuffaceous breccia with andesite (α) and rhyolite (ρ) fragments, SE of Cerhat Hill, NII, x 20.

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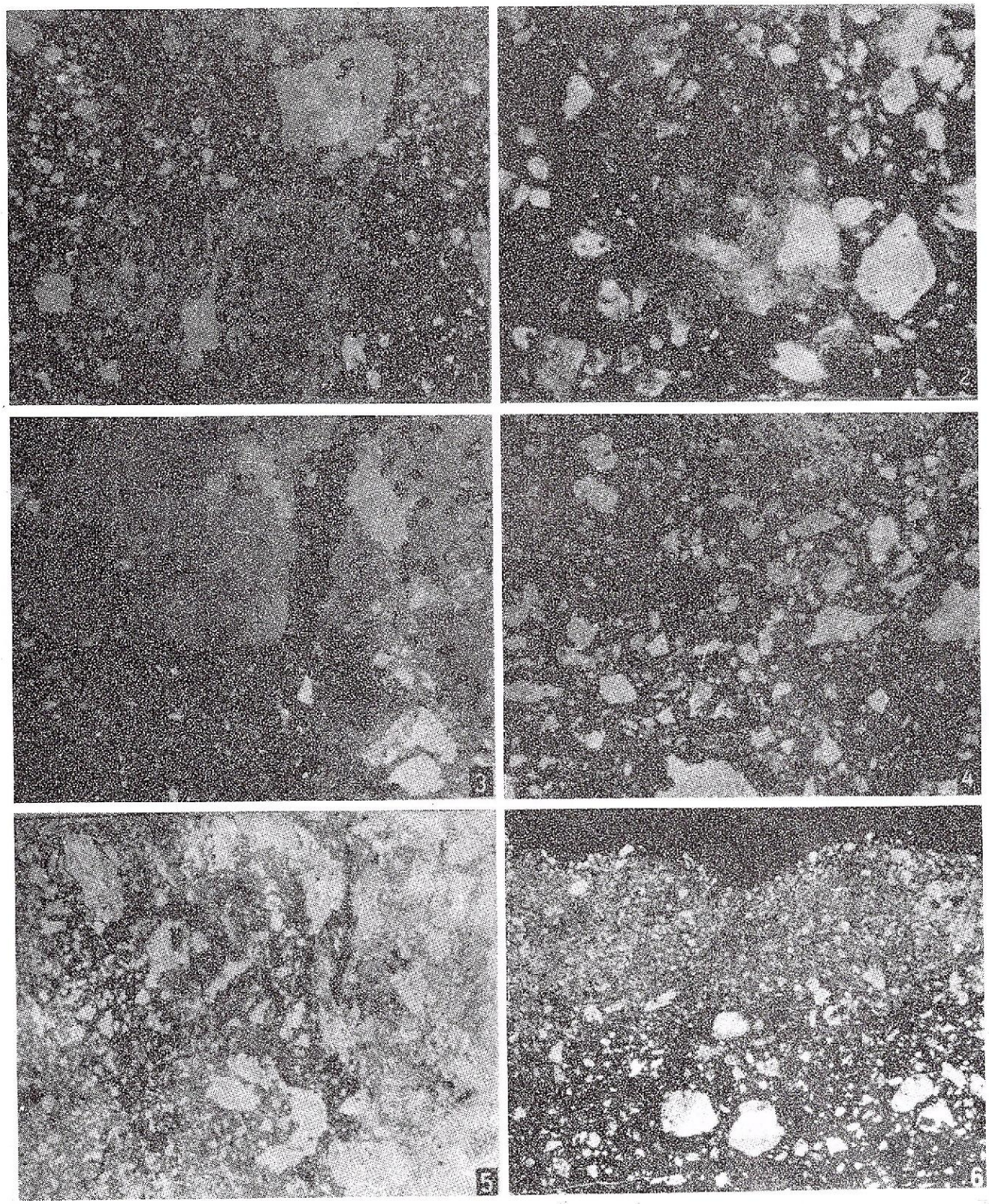

Fig. 1 – Tuffaceous breccia with fluidal rhyolite (ρ) fragments, SE of Cerhat Hill, N II, x 20.

Fig. 2 – Hydrothermally altered rhyolite fragment with feldspar relicts, Căpușu Mic, N II, x 20.

Fig. 3 – Tuffaceous breccia which pierces rhyolite fragments (ρ), Căpușu Mic, N II, x 20.

Fig. 4 – Tuffaceous breccia desintegrating rhyolite (ρ) fragments, Cetățeaua Hill, N II, x 20.

Fig. 5 – Fine-grained (tuffaceous) breccia piercing and including a more coarse breccia generation, SE of Cerhat Hill, N II, x 20.

Fig. 6 – Contact between black siltstones and tuffaceous breccia, Cetățeaua Hill, N II, x 20.

Institutul Geologic al României

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A COMPARATIVE STUDY OF THE DISTRIBUTION OF SOME MAJOR, TRACE AND RADIOACTIVE ELEMENTS IN SEVERAL PALEOGENE AND UPPER PLIOCENE-QUATERNARY BASALTS AND BASALT-ANDESITES FROM ROMANIA

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Key words: Basalts. Basaltic andesite. Xenoliths. Major elements. Trace elements. Carpathians.

Abstract: The Paleogene and Upper Pliocene-Quaternary volcanic rocks are represented by basalts, basalt-andesites and alkaline rocks. At Lucareţ and Racoş basalt flows and necks occur, but in the Apuseni Mountains and Poiana Ruscă Mountains, where the erosion was very active, dykes and necks are mostly preserved. Most of these rocks were erupted along an E-W crustal system of faults which were active since Paleogene during two periods: a Paleogene one and an Upper Pliocene-Quaternary one. Two petrologic rock series resulted, which are very characteristic. The distribution of major and trace elements discriminate the basalts from the basalt-andesites and alkaline rocks. The close contents of radioactive elements in the Racoş basalts and their ultramafic xenoliths (nodules) show that the parental magma of these basalts did not differentiate, or this process was very weak. The differences between the volcanic occurrences of each period were determined by the mantle heterogeneity and other processes.

1. Introduction

On the territory of Romania there are several occurrences of post-Laramian basalts and basalt-andesites. These rocks occur both in zones in which the Banatitic and Neogene arc volcanism has been previously active and in zones deprived of such a volcanic activity. The most characteristic volcanic occurrences are in Banat, South Apuseni Mountains, Poiana Ruscă Mountains and Perşani Mountains (Fig. 1).

According to the K/Ar and Rb/Sr datings (Herz et al., 1974; Mihăilă, Kreuzer, 1981; Ghenea et al., 1981; Lupu et al., 1986; Lemne et al., 1982, 1983, unpubl. rep.; I. Seghedi, A. Szakacs, oral com.) the volcanic rocks are of two ages: Paleogene (56–29 Ma) and Upper Pliocene-Quaternary (2.5–0.35 Ma).

Data concerning the distribution of some major, trace and radioactive elements in Paleogene and Upper Pliocene-Quaternary (Young) rocks have been obtained in different occurrences by the following geologists: Savul, Nichita (1944) and Savu et al. (1944) for Lucareţ; Kräutner (1969) for Poiana Ruscă Mts; Peltz, Bratosin (1986) and Peltz et al. (1971; 1985) for Racoş and other occurrences; Boştinescu (1982, unpubl. rep.) for the Brănişca Hill.

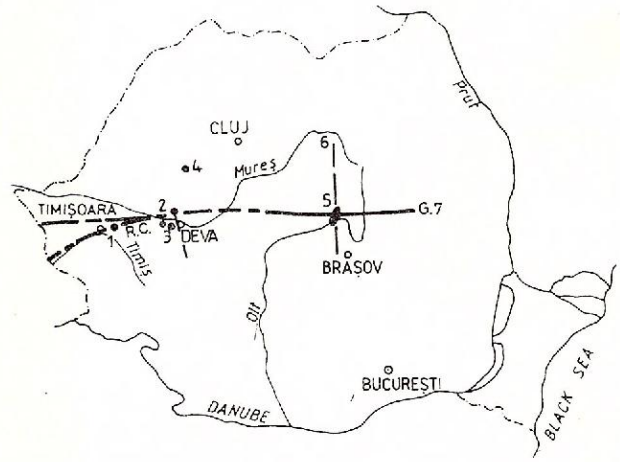

Fig. 1 Paleogene and Upper Pliocene-Quaternary volcanic rocks occurrences in Romania. 1, Lucareţ; 2, Măgura Sârbi and Brănişca Hill; 3, Herepeia; 4, Detunata; 5, Racoş; 6, cross fault; RC, the Recaş-Cuieş crustal fracture; G 7, the G 7 (Alba Iulia-Băile Tuşnad) crustal fracture.

The basaltic rocks from the Apuseni Mountains and Perşani Mountains (Racoş) were attributed to the intra-plate volcanism by Rădulescu et al. (1981). A similar characteristic was demonstrated also for the Racoş basalts by Peltz and Bratosin (1986) and for the Lucareţ basalts by Savu et al. (1994).

In the present comparative study we used data obtained in six occurrences of volcanic rocks in the men-

Institutul Geologic al României

tioned areas, which are included in the Table 1. It was elaborated with a view to reveal the peculiarities of the distribution of the major and trace elements in the volcanic occurrences and to compare them. For this purpose we had 83 new analyses.

2. Structural and petrographic aspects of the volcanic occurrences

Most occurrences presented in this paper are situated along a system of deep fractures trending almost E-W, which extend from Banat through the Mureş Couloir to the Perşani Mts (Fig. 1). Among them two crustal fractures are the most important: Recaş-Cuieş Fault evidenced by Andrei et al. (1975) and the G 7 (Alba Iulia-Băile Tuşnad) Fault evidenced by Gavăt et al. (1973), which could represent together a single great fault (Airinei, 1980). In the south of the Apuseni Mountains this fracture determined zones of tectonic breccia and crush conglomerates, green in colour, cropping out between the neck of Măgura Sârbi and Chiscădaga. West and east of this region the fault is covered by sedimentary deposits. It is northworthy that the Paleogene neck of Măgura Sârbi is situated on this crustal fault (see Gurasada map–Lupu et al., 1986), a fact which shows that it was active as far back as the Paleogene.

According to their ages, the six volcanic rock occurrences investigated by us belong to the following volcanic series: the Paleogene and the Upper Pliocene-Quaternary series. The association of basalts and basalt-andesites, including their alkaline correspondents, are characteristic in both series. Such associations occur also in other places in the world, as for instance in the Pliocene-Quaternary basaltic rock series of the West Carpathians (Hovorka, 1978) and in the Lower Cretaceous basaltic rocks of Paraná – Brasil (Mantovani et al., 1985).

The Paleogene volcanic rocks occur in the north of the Poiana Ruscă Mts and in the Mureş Couloir, but they do not exceed toward north, the great crustal fault, while the Upper Pliocene-Quaternary volcanics occur along this fracture and a little north of it, as there are the Detunata basalt-andesites. If at Lucareţ and Racoş basaltic lavas and pyroclastics occur beside the feeder necks, in the Apuseni Mountains and Poiana Ruscă Mts mostly dykes and necks and some remnants of lava flows (Brănişca Hill, Detunata Goală) are preserved, for these regions have been affected by a recent erosion which have removed the upper part of the volcanic structures.

2.1. Paleogene volcanic rocks of the Mureş Couloir

Studying the Paleogene volcanics from the Poiana Ruscă Mts and the south of the Apuseni Mountains, Kräutner (1969) separated them in two groups: a group of calc-alkaline rocks analysed in our paper, too, and the basanites younger group, which was not included in our paper because we have no analyses.

The so-called calc-alkaline group consists of basalt-andesites (Herepeia, Brănişca Hill) and alkaline rocks (Măgura Sârbi – Savu et al., unpublished data). At Herepeia (Poiana Ruscă Mts) a dyke of basic rocks (56 Ma – mean of 5 datings – Lemne et al., 1982, unpubl. rep.) occurs, intruding the crystalline schists. According to the chemical analyses (Kräutner, 1969), the rock composition indicates basalts and basalt-andesites (SiO_2 = 51.93–52.20 per cent).

The basalt-andesites (SiO_2 = 53.70–58.80 per cent, Savu et al., unpubl. data) of Brănişca Hill and the trachybasalt-andesites and trachyandesites (SiO_2 = 52.64–54.71 per cent, Savu et al., unpubl. data) of Măgura Sârbi (Apuseni Mountains) are situated on a lineament trending NNW-SSE, at about 3 km one from another. The neck of Măgura Sârbi (48 Ma, mean of 2 datings, Lemne et al., 1983, unpubl. rep.), which is exploited by the Bretea Mureşană quarry, is elongated E-W, parallel to the Recaş-Cuieş crustal fault on which it is located. Its rocks are separated in four-sided prisms in an almost horizontal position. The Brănişca Hill neck (29.6 Ma, Lemne et al., 1983, unpubl. rep.) was exploited by the Brănişca Quarry down near the level of the Mureş River, on the right riverside of which it is situated. At present only the neck walls, formed of Turronian deposits can be seen, which show that the neck has an irregular shape. North of it there occurs a remnant of the lava flows. The rock texture in these two occurrences is usually porphyritic, determined by augite and olivine phenocrysts, often substituted by secondary minerals. In the lava flows and on the neck margin the rocks have a fluidal texture and contain numerous xenoliths broken from the country rocks.

2.2. Upper Pliocene-Quaternary (Young) volcanic rocks

In this volcanic series the basalts of Lucareţ and Racoş and the basalt-andesites of Detunata are included. At Lucareţ, in Banat, there is the most characteristic occurrence of basalts. It consits of a small plateau of about 40 sq. km formed of lavas and rare pyroclastics, which have been feeded by a volcanic vent, at present a volcanic neck, situated under the lavas from the Piatra Roşia Hill, on the Recaş-Cuieş crustal fracture (see

Institutul Geologic al României

Andrei et al., 1975). These lavas, often separated in six-sided vertical columns, are almost horizontal and lie on the Upper Pliocene deposits (Mateescu, 1936), a position which correlates with the K/Ar dating of 2.5 Ma (I. Seghedi, A. Szakács, oral com.). The rocks are "transitional" alkali basalts with normative nepheline, containing SiO_2<52 per cent. They are basalts with a weak hawaiitic character, in which the relation Na_2O–$2>K_2O$ is the most frequent one (Savu et al., 1994).

The Racoş basalts (Perşani Mts) probably formed at the beginning of Quaternary an unique surface of lavas and rare pyroclastics extended north and south of the present Olt Valley, later on eroded. Rădulescu (1969) showed that these basalts were erupted along a N-S deep fracture. The Figure 1 indicates that these basalts are situated just in the intersection area of this fault and the G 7 crustal fracture, a tectonic relation which favoured the eruption through the Hegheş neck. The basaltic lavas (1.39–0.35 Ma, Herz et al., 1974; Ghenea et al., 1981, Mihăilă, Kreuzer, 1981), separated in six-sided columns, lie on the Pleistocene deposits associated with basaltic pyroclastics (Seghedi, Szakács, 1994 a,b). They contain peridotitic xenoliths (nodules) proceeding from the mantle (Măldărescu et al., 1982; 1983) and exotic quartz granules. The first have been broken by the basaltic magma from the mantle above the magmatic chamber. They have a various granulation. The SiO_2 contents are lower than 52 per cent and the relation Na_2O-2 ≠ K_2O shows that these basalts have a weak character of hawaiitic and trachytic basalts (acc. to the classification made by Bass et al., 1985).

The two rock occurrences of Detunata (Detunata Goală and Detunata Flocoasă) from the Apuseni Mountains are formed of basalt-andesites in which the SiO_2 vary between 53.38 and 55.47 per cent (Savu et al., unpubl. data). They are situated at 1.5 km one from another on a N-S hill of Maastrichtian deposits (Bordea et al., 1979), which represents a remnant of a Miocene platform (Maxim, 1944). In the northern part of Detunata Goală (0.9 Ma, E.Vâjdea, oral com.) lava flows, very characteristic six-sided columns, like the ones in the Lucareţ and Racoş basalts, can be seen. Owing to this aspect the occurrence was awarded as a natural monument. Usually the rock structure in these two occurrences is vesicular or amygdaloidal. The texture is porphyritic or glomeroporphyritic, determined by isolated or clustered augite phenocrysts. On the Detunata Flocoasă neck border the texture is fluidal.

3. Major and trace element geochemistry

The major and trace elements were determined by the Energy Dispersive X Ray Analysis on powder samples. The data are presented in Table 1 in which both the Paleogene and Upper Pliocene-Quaternary rocks have been separated in two groups: basalt and basalt-andesites, including the alkaline rocks. The mean contents of CaO are higher than 8 per cent in young basalts and in the Detunata Goală basalt-andesites (Tab. 2). In the Paleogene and the Detunata Flocoasă rocks this component is lower, its mean values varying between 5 and 7 per cent. In the young basalts the TiO_2 content is higher than 1, but it decreases in the Paleogene and Detunata basalt-andesites (Tab. 2). According to the relations between Ti, Y and Zr contents the Upper Pliocene-Quaternary basalts plot on Figure 2 (Pearce, Cann, 1973) in the intra-plate field, some of them falling out of it.

The Paleogene and young basalt-andesites and alkaline rocks plot out of this field, on its extension towards the Zr corner, which indicates rocks rich in this element. A few basalts of Racoş and some basalt-andesites reach the calc-alkaline rock field. All the Paleogene and Upper Pliocene-Quaternary basalts and basalt-andesites plot on the diagram in Figure 3 (Pearce, Cann, 1973) in the calc-alkali basalt field. Even the 'transitional' alkali young basalts of Lucareţ, which are clear intra-plate rocks, have a calc-alkaline characteristic, an aspect observed by Savu et al., (1994, Fig. 2), when they plotted them on the combined Irvine-Baragar and Hutchinson diagram. The tholeiitic continental intra-plate basaltic rocks are characteristic of the Triassic-Lower Jurassic volcanics (Savu, 1988), the parental magma of which formed in the mantle at a lesser depth than the post-Laramian one.

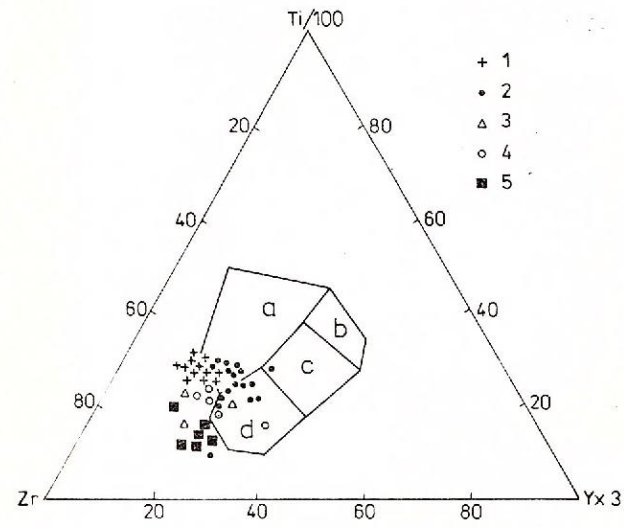

Fig. 2 – Plot of Paleogene and Upper Pliocene-Quaternary volcanic rocks on the Ti/100 Yx3–Zr diagram. a, intra-plate basalts; b, low-K tholeiites; c, ocean floor basalts; d, calc-alkali basalts (rock fields acc. to Pearce and Cann, 1973); 1, Lucareţ; 2, Racoş; 3, Detunata; 4, Herepeia; 5, Măgura Sârbi.

Institutul Geologic al României

Table 1
Distribution of the major and trace elements in the volcanic rocks

Sample	CaO %	TiO_2 %	Fe_2O_3 %	Rb ppm	Sr ppm	Y ppm	Zr ppm	Nb ppm	U ppm	Th ppm	K %
1	2	3	4	5	6	7	8	9	10	11	12
1. Lucareţ basalts (Banat), Upper Pliocene (2.5 Ma)											
a) Zăbrani quarry											
1	8.49	1.97	9.49	42	714	24	208	58	1.4	4.7	1.41
2	7.97	1.69	9.15	55	702	27	218	54	1.6	6.3	1.28
3									1.8	4.1	1.56
4	8.96	1.71	9.75	43	717	25	209	56	0.4	8.0	1.93
5	8.47	1.98	9.54	46	755	29	212	67	2.2	5.2	1.56
6									1.8	6.1	1.62
7									1.7	5.4	1.74
8									1.7	5.9	1.83
9	8.58	1.66	9.19	49	667	25	198	65	1.9	8.1	1.61
10	9.01	1.88	9.74	37	777	21	212	66	1.8	6.1	1.74
b) Zeiman quarry											
11	8.35	1.56	9.33	39	727	25	198	65	1.5	5.5	1.59
12	8.56	1.71	9.51	47	711	27	203	55	1.5	4.8	1.78
13	8.60	1.97	9.75	45	739	28	210	56	1.7	6.1	1.69
14	8.05	1.96	9.39	35	737	20	203	62	1.7	6.5	1.72
15	8.43	1.74	9.50	35	752	31	215	55	1.9	5.0	1.67
16									1.5	5.6	1.61
17	8.32	1.93	9.61	47	711	24	200	56	1.9	4.6	1.72
18									1.5	6.1	1.50
c) Şanoviţa quarry											
19									1.7	6.6	1.96
20	8.42	1.72	9.34	44	709	28	203	52	1.8	5.1	1.69
21	8.12	1.63	9.43	36	639	29	194	53	1.6	5.2	1.56
22	8.73	1.82	9.44	48	582	28	179	50	0.9	4.8	1.36
23									1.3	5.6	1.47
24	8.75	1.89	9.85	32	583	24	171	54	1.5	5.0	1.36
25									1.6	5.2	1.54
26	8.73	1.82	9.68	39	637	33	181	48	1.9	2.7	1.80
27									1.3	6.2	1.77
28									1.4	5.8	1.59
29	8.57	2.05	9.82	37	617	33	199	51	1.6	5.6	1.60
30	8.43	1.96	9.63	39	637	20	186	53	1.4	5.8	1.74
31	8.69	1.97	9.78	37	647	30	178	48	0.9	5.6	1.46
d) Ghizela quarry											
32	8.95	1.91	9.77	51	732	26	200	52			
33	8.71	2.00	9.83	34	724	21	198	56			
34	8.47	1.75	9.37	45	680	28	198	50			
2. Racoş basalts (Perşani Mountains), Quaternary (1.39–0.35 Ma)											
35	9.18	1.26	8.38	42	690	28	152	44	1.9	5.8	1.61
36	9.17	1.28	7.65	31	575	28	127	24	2.5	7.4	0.73
37	9.11	1.16	8.52	50	754	22	155	43	1.3	5.5	1.16
38	9.42	1.13	7.80	22	493	25	153	37	1.8	8.0	0.91
39	6.75	0.54	5.00	53	362	29	155	<20			
40	5.48	0.42	4.11	81	301	25	151	<20			
41	9.00	1.16	8.27	47	904	27	156	47	1.6	8.4	1.91
42	8.32	1.09	8.21	27	612	33	135	<20	1.6	6.4	0.88
43	8.09	1.20	8.94	29	561	29	129	<20	1.2	6.8	1.06
44	8.63	1.28	8.21	24	669	21	134	29	2.0	7.0	1.22
45	9.36	1.11	8.36	36	651	27	122	29	1.0	6.8	0.91

Institutul Geologic al României

Table 1(continued)

1	2	3	4	5	6	7	8	9	10	11	12
46	9.17	1.33	8.14	31	670	21	130	31	1.1	6.4	1.21
47	8.97	1.13	8.22	31	652	28	130	24	1.3	6.1	1.36
48	8.80	1.23	8.29	32	752	24	122	26	1.4	6.7	1.30
49	8.87	0.96	8.34	20	689	32	122	27	1.4	6.1	1.05
50	8.08	1.31	8.14	31	677	24	133	<20	1.5	6.1	1.26
51	7.96	1.34	8.31	38	753	23	134	22	1.4	6.1	1.73
52	8.04	1.44	8.41	36	854	27	144	25	1.2	5.9	1.62
3. Detunata Flocoasă basalt-andesites (Apuseni Mountains) Quaternary (0.9 Ma)											
53	5.12	0.61	5.72	59	520	15	148	<20	1.6	6.6	1.05
54									1.3	5.0	1.28
55									1.7	3.5	1.07
56									1.8	4.5	1.18
4. Detunata Goală basalt-andesites (Apuseni Mountains), Quaternary (0.9 Ma)											
57									1.6	4.4	1.37
58									1.8	5.4	1.21
59									2.3	4.2	1.41
60									1.9	4.2	1.00
61									1.3	5.8	1.35
62	7.61	0.71	6.46	28	535	18	116	<20	1.7	5.7	1.24
63									1.3	4.6	1.00
64	8.46	0.73	6.52	37	538	21	110	<20	1.7	4.0	1.15
65									1.3	5.5	1.03
5. Herepeia basalt and basalt-andesites (Poiana Ruscă Mts), Paleogene (56 Ma)											
66	7.80	0.94	7.92	29	464	17	139	<20	0.9	3.7	0.94
67	6.82	0.69	7.55	20	486	30	121	<20	0.8	3.9	0.55
68	6.73	0.88	7.85	21	494	22	139	<20	1.3	4.6	0.77
69									0.9	4.9	0.79
70	7.23	0.82	7.92	25	469	24	145	20	0.9	5.1	0.81
71	6.90	0.92	8.22	15	402	23	136	20	1.1	4.4	1.14
6.Brănişca Hill and Măgura Sârbi basalt-andesites and alkaline rocks (Apuseni Mountains), Paleogene (48–29.6 Ma)											
72	0.94	1.11	14.30	39	266	11	182	24	2.1	7.8	0.98
73	5.82	0.60	6.49	34	527	24	152	<20	1.9	8.0	2.63
74	5.72	0.68	6.11	46	530	18	162	<20	2.4	9.0	2.08
75	6.53	0.44	6.70	39	536	18	162	<20	3.1	7.5	1.51
76									2.5	7.8	1.82
77	6.03	0.52	6.40	48	517	21	167	<20	1.4	6.1	2.83
78	5.22	0.46	5.83	48	524	14	156	<20			
79									1.8	6.6	1.42
80									2.3	6.3	1.37
81									1.9	6.8	1.42
82									1.5	7.2	1.50
83									1.9	8.6	1.08

The highest mean contents of Fe_2O_3 were determined in the Lucareţ basalts (9.56 per cent) and the lowest (5.7 per cent) in the Detunata Flocoasă basalt-andesites (Tab. 2). The mean values of this component in the Paleogene and Detunata Goală occurrences varies between 7 and 8 per cent. The highest contents of Sr were also determined in basalts in which the values exceed 700 ppm, while in the Paleogene and Quaternary basalt-andesites and trachytic rocks these contents decrease, standing between 536 and 463 ppm.

The same discrimination between basalts and all the basalt-andesites and associated alkaline rocks can be obtained by the content of Nb. The higheast mean values (26–56 ppm Nb) were determined in the first rocks, while in all the basalt-andesites and alkaline rocks the contents maintain under 20 ppm (Tab. 2). The mean value of Zr/Nb ratio vary in the young basalts between 3.5 and 5.2, while the mean value of Y/Nb ratio (Tab. 2) varies in the same basalts between 0.5 and 1. According to Erlank and Kalbe (1976) the higher values

Institutul Geologic al României

of the last ratio, as there are most values of the Racoş basalts, indicate parental magmas derived from mantle sources depleted by extraction of previous meltings.

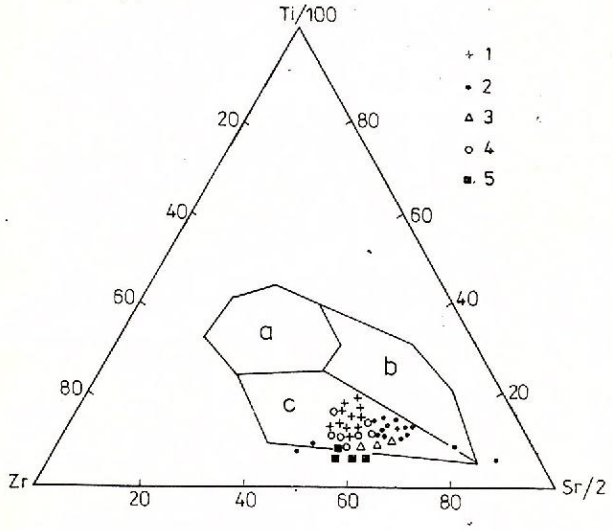

Fig. 3 – Plot of Paleogene and Upper Pliocene-Quaternary volcanic rocks on the Ti/100–Sr/2–Zr diagram. a, ocean floor basalts; b, low-K tholeiites; c, calc-alkali basalts. The same legend like in Fig. 2.

of the mean values of the six occurrences plot in the intra-plate rock field (WPB). The alkaline rocks from the Paleogene neck of the Măgura Sârbi and the basalt-andesites from the Detunata Flocoasă plot very near this field at high values of Zr/Y ratio, but not in any other field of the diagram. The mean values situated in and out of this field occupy on the diagram the same position like the individual rocks of the Racoş basalts (Peltz, Bratosin, 1986) and the Triassic-Lower Jurassic continental tholeiitic basalts of the East Carpathians, Făgăraş and North Dobrogea (Savu, 1988). On this diagram as in the other ones, the plots of individual analyses or of their mean values are clustered in small swarms which indicate close geochemical characteristics.

The mean contents of Y (Tab. 2) very clear discriminate the basaltic rocks with 26 ppm Y from the Paleogene and the Detunate basalt-andesites, the mean values of which vary between 15 and 23 ppm Y.

Table 2

Mean values of the major and trace elements from the volcanic occurrences

No.	CaO %	TiO_2 %	Fe_2O_3 %	Rb ppm	Sr ppm	Y ppm	Zr ppm	Nb ppm	Zr/Nb	Y/Nb	U ppm	Th ppm	K %	Th/U
1	8.53	1.83	9.56	41.82	692.30	26.34	198.82	55.73	3.56	0.47	1.56	5.59	1.72	3.58
2	8.46	1.13	7.85	36.72	777.86	26.27	138.00	26.33	5.24	0.99	1.51	7.21	1.24	4.77
3	5.12	0.61	5.72	59.00	520.00	15.00	148.00	–	–	–	1.60	4.90	1.14	3.06
4	8.03	0.72	6.49	32.50	536.50	19.50	113.00	–	–	–	1.65	4.86	1.19	2.94
5	7.09	0.85	7.89	22.00	463.00	23.20	136.00	–	–	–	0.98	4.43	0.83	4.52
6	5.04	0.63	7.63	42.33	483.33	17.66	163.50	–	–	–	2.07	7.70	1.69	3.71

The analysed rocks are from: 1, Lucareţ; 2, Racoş; 3, Detunata Flocoasă; 4, Detunata Goală; 5, Herepeia; 6, Brănişca Hill and Măgura Sârbi volcanic occurrences.

In case of Rb the values are very variable and, as it is shown in Table 2, they do not indicate a clear separation between the Paleogene and the Upper Pliocene-Quaternary volcanics, or between basalts and basalt-andesites. As a rule, the Rb/Sr ratio is very low, its mean values varying between 0.4 and 0.11. Herz et al. (1974) showed that such low values indicate parental magmas formed in the mantle. The high contents of Sr suggest, according to Hart et al. (1971), that the parental magma of the Paleogene and young volcanics formed in the mantle at a great depth.

The mean contents of Zr (Tab. 2) are also variable in the rocks of the six occurrences. The highest contents (mean 198 ppm Zr) were determined in the basalts of Lucareţ and the lowest (mean 115 ppm Zr) in the Detunata Goală basalt-andesites. On the Zr/Y–Zr diagram (Fig. 4 – Pearce, Norry, 1979) the majority

4. Radioactive elements

The radioactive elements U, Th and K (Tab. 1) were determined by Natural Gamma Spectrometry. The contents of U are low and vary within limited ranges, their mean values (Tab. 2) maintaining between 0.58 and 2 ppm. These ranges were determined in the Paleogene basalt-andesites and alkaline rocks. On the diagram in Figure 5 (Kravchenko, 1974) all the Paleogene and Upper Pliocene-Quaternary volcanics plot at higher values of U and K than the mean value of continental tholeiites (CT) and even that of the alkali basalts (AB). In the young basalts there are high contents of Th (5.6–7.2 ppm). In all basalt-andesites the mean value of Th vary between 4.4 and 4.9 ppm. The Paleogene alkaline rocks of Măgura Sârbi have the

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highest contents of Th (7.7 ppm Th) from the all analysed volcanic rocks.

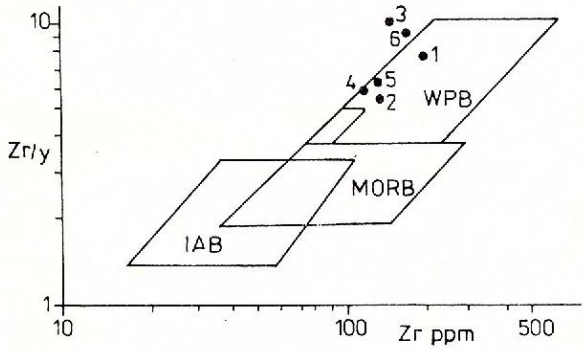

Fig. 4 – Plot of Paleogene and Upper Pliocene-Quaternary volcanic rocks on the Zr/Y–Zr diagram (mean values of Table 2). WPB, intra-plate basalts; MORB, middle ocean ridge basalts; IAB, island arc basalts (rock fields acc. to Pearce and Norry, 1979). 1, Lucareţ; 2, Racoş; 3, Detunata Flocoasă; 4, Detunata Goală; 5, Herepeia; 6, Măgura Sârbi.

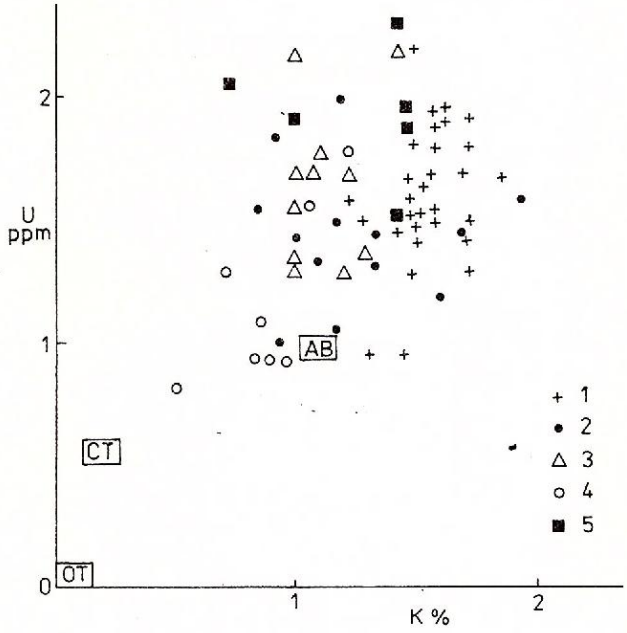

Fig. 5 – Plot of Paleogene and Upper Pliocene-Quaternary volcanic rocks on the U-K diagram. 1, Lucareţ; 2, Racoş; 3, Detunate; 4, Herepeia; 5, Măgura Sârbi. (OT), ocean tholeiites; (CT), continental tholeiites; (AB), alkali basalts.

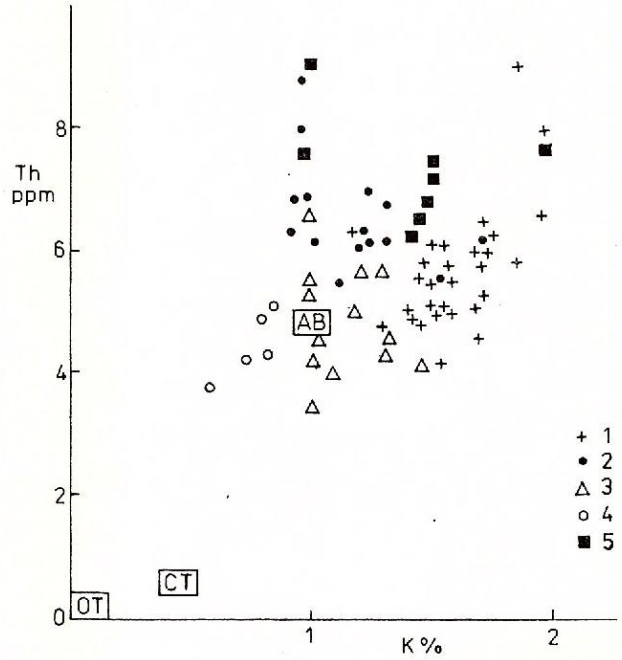

Fig. 6 – Plot of Paleogene and Upper-Pliocene-Quaternary volcanic rocks on the Th–K diagram. The same legend like in Fig. 5.

On the diagram in Figure 6 (Kravchenko, 1974) a few rocks plot around the mean value of the alkali basalts (AB), but most rocks are distributed at little higher values of Th and K. The Th/U ratio usually is higher than in other basalt types.

The K has contents characteristic of the alkali basalts of hawaiitic type (Tab. 1 and 2). The mean values are higher than in other types of basaltic rocks.

The patterns of the main elements analysed in this paper (Fig. 7), normalized by their mean contents in the mantle according to Wood et al. (1979), are similar to those of the intra-plate basaltic rocks described by these authors in different occurrences in the world. All these volcanic series are very characterised by a strong decrease of the values, or a stressed slope of the patterns respectively, from Rb to Y. All the patterns are similar, indicating the origin of their parental magmas in mantle sources with a similar chemical composition. A peculiarity occurs in case of Lucareţ basalts, the pattern of which shown a small positive anomaly for Nb.

The mean contents of radioactive elements in the peridotitic xenoliths (nodules) from the Racoş basalts (Tab. 3) are close to those from the host rock (Tab. 2).

It leads to the conclusion that the source in the mantle, from which the basaltic magma derived by partial melting, consists of the same peridotite as the xeno-

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liths. The close mean values of the radioactive elements in the two types of rocks show that the basaltic magma did not differentiate, or this process was very weak in comparison with the initial pyrolith.

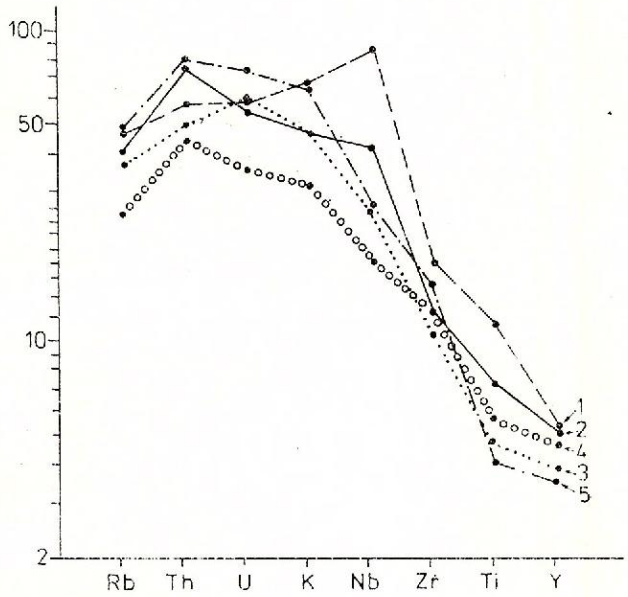

Fig. 7 — Pattern variation of some selected elements normalized by their mean contents in the mantle (according to Wood et al., 1979). 1, Lucareţ; 2, Racoş; 3, Detunata Goală; 4, Herepeia; 5, Măgura Sârbi.

Table 3
Distribution of U, Th and K in peridotitic xenoliths from the Racoş basalts (Perşani Mts)

No.	U ppm	Th ppm	K %	Th/U
1	2.40	8.60	1.74	3.58
2	2.10	8.40	1.83	4.00
3	1.50	5.90	1.12	3.93
4	1.70	6.20	0.82	3.64
5	1.40	4.60	1.34	3.28
6	1.80	9.40	1.02	5.22
Mean	1.81	7.18	1.31	3.94

5. Conclusions

The different types of rocks from the six occurrences analysed in this paper formed during two geological periods without any important tectonic movements, excepting some faults.

The first period settled in Paleogene after the Laramian movements ceased (Rădulescu, Săndulescu, 1980). During this period two Paleogene rock groups formed: first the calc-alkaline rocks consisting of basalts and basalt-andesites associated with alkaline (ultrabasitic) volcanics, situated in the Mureş Couloir, which have ambiguous geochenfical characteristics; then formed the 'basanites' (Kräutner, 1969) constituted of alkali tholeiites and other alkali basalts, which are intraplate rocks, situated in the Poiana Ruscă Mts.

By their age (56–29.6 Ma) and geochemical characteristics the basalt-andesites and the associated alkaline rocks appear as a transitional rock group between the arc Banatitic volcanics (70–60 Ma) developed in the north of the Poiana Ruscă Mts and in the Mureş Couloir (Savu, 1962; Savu et al., 1992) and the Paleogene alkali basalts (basanites) of 29 Ma (mean of 2 datings, Lenne et al., 1982, unpubl. rep.). On the Gurasada map (Lupu et al., 1986) the Paleogene basalt-andesites are also situated in an intermediate position between the Banatitic and basanitic rocks. The second period settled at the end of the Pliocene after the Late Alpine movements ceased (Rădulescu, Săndulescu, 1980). During this period the subalkali basalts with normative nepheline of Lucareţ (2.5 Ma) and Racoş (1.39–0.35 Ma) and the basalt-andesites of Detunata (0.9 Ma) were erupted. The subalkali basalts are intraplate rocks, a characteristic evidenced by their contents of Ti, Zr, Y and Sr (Pearce, Cann, 1973) and by their position on the Ti Zr and Zr/Y Zr tectonic setting diagram (Peltz, Bratosin, 1986; Savu et al., 1994). The Detunata basalt-andesites display ambiguous geochemical characteristics close to those of the Paleogene basalt-andesites.

The geochemical differences between the rock occurrences of each period or its epochs have been determined by the depth of the magma melting, the melting rate and particularly by the mantle heterogeneity. This heterogeneity was the result of a previous irregular metasomatic process (Wilson, 1989) that affected the mantle during the ocean crust subduction.

The contamination and the differentiation of the parental magmas from which derived the different analysed rocks were almost absent or very weak, for each volcanic structure from both periods had a rapid evolution.

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Institutul Geologic al României

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UPPER PLIOCENE TO QUATERNARY BASALTIC VOLCANISM IN THE PERŞANI MOUNTAINS

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Institutul Geologic al României. Str. Caransebeş nr. 1, 78344 Bucureşti 32



Key words: Volcanic features. Trachybasalts. Xenoliths. Lava flows. Base surges. Fallout. Perşani Mountains. East Carpathians.

Abstract: Upper Pliocene-Pleistocene (2.25–0.5 Ma) small-sized volcanic edifices are aligned on a NE-SW trend over 22 km in the western part of the Perşani Mts (East Carpathians). From north to south, three isolated volcanic centers occur (Sărata, Racoş and Mateiaş), then a larger volcanic area develops - probably initially caldera type - with a lot of eruptive centers (Hoghiz-Bogata Valley) and some isolated occurrences at Comana. The volcanics consist of trachybasalts with abundant lherzolite xenoliths. Volcanic structures are complex, witnessing several superimposed eruptive phases: an initial phreatomagmatic phase caused ejection of juvenile material and great quantities of fragmented bed-rock material, with formation of initial maars; in the second phase lava flows and possible ephemeral lava lakes were generated; the third phase – again phreatomagmatic – with mostly volcanian base surge (with low angle cross bedding) deposits and plinian fallout deposits (proximal with bomb sags), rich in mantle xenoliths; the fourth phase produced strombolian scoria and spatter cones, consisting of agglutinates, agglomerates and lapilli deposits.

Introduction

In the Perşani Mts (Southeastern Transylvania), an intracontinental basaltic volcanic area, representing the youngest volcanic activity in the Eastern Europe, occurs. This north-northeast to south-southeast trending area, including several simple and complex volcanic edifices, is about 22 km long and 8 km wide. It lies between the Sărata Valley to the north and Comana to the south.

Due to its unique spectacular features, the basaltic volcanics were intensely studied by many workers in the past. The first descriptions belong to Hauer and Stache (1863), Herbich (1878), Koch (1900), Wachner (1915) and the earliest petrographic study was done by Laţiu (1928).

As the earliest volcanological approach, Preda (1940) firstly established the succession of volcanic products, the following four sequences being distinguished: (1) lower tuffs, (2) lava flows, (3) upper tuffs and (4) scoria deposits. The volcanic structure and the volcanic history of the Racoş-Hoghiz area have more recently been presented by Mihăilă et al.(1972) and Mihăilă , Peltz (1977). Rădulescu (1969), Rădulescu et al. (1981) and Rădulescu, Săndulescu (1980) suggested an intraplate geotectonic setting for the basaltic volcanism in the Perşani Mts. Other papers were focused on age (Tövissi, 1972; Tövissi, Chintăuan, 1972; Casta, 1980; Mihăilă, Kreutzer, 1981; Ghenea et al., 1981), mantle xenoliths (Măldărescu et al., 1982, 1983), petrology (Peltz, Bratosin, 1971, 1986; Peltz et al., 1972; Peltz, Peltz, 1983; Peltz, Stoian, 1985; Peltz et al., 1985).

Following recent geological investigation based on lithological and petrographical observations, a more detailed picture of the eruptive history in the Perşani Mts has been obtained.

The petrographical and petrochemical investigations highlight the mineralogical and chemical homogeneity of the alkali-basaltic volcanics, disregarding their position in the lithological-columns and occurrence areas. According to these data, they are mainly of trachybasaltic composition (Peltz, Bratosin, 1986).

Volcanic sequences

Explosive and effusive basaltic volcanic products occur in four areas: Sărata Valley, Racoş-Mateiaş, Hoghiz-Bogata-Trestia Valley and Comana de Sus (Fig. 1). The northernmost area is located along the

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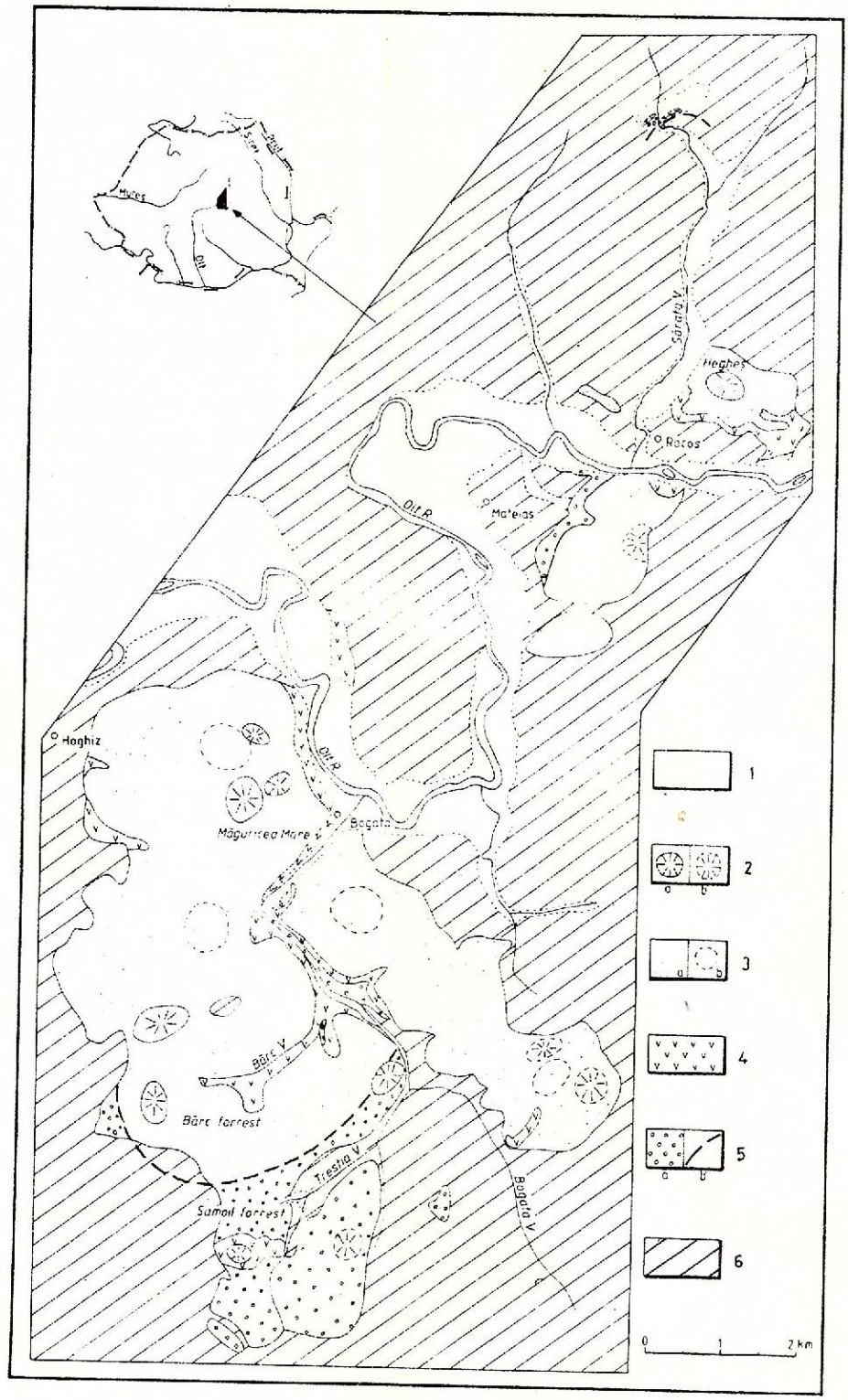

Fig. 1 — Geological sketch of the area with Quaternary basaltic volcanism in the Persani Mts (Romania). 1, Holocene alluvia; 2, Fourth stage: scoria cones: a)mapped; b) presumed; 3, Third stage pyroclastics: a) fallout and base surge deposits; b) presumed tuff-rings features; 4, Second stage: lava flows; 5, First stage pyroclastics: a) base surge, fallout and flow deposits; b) presumed outline of tuff-ring border; 6, Prebasaltic Mesozoic and Cainozoic sedimentary and volcanic formations.

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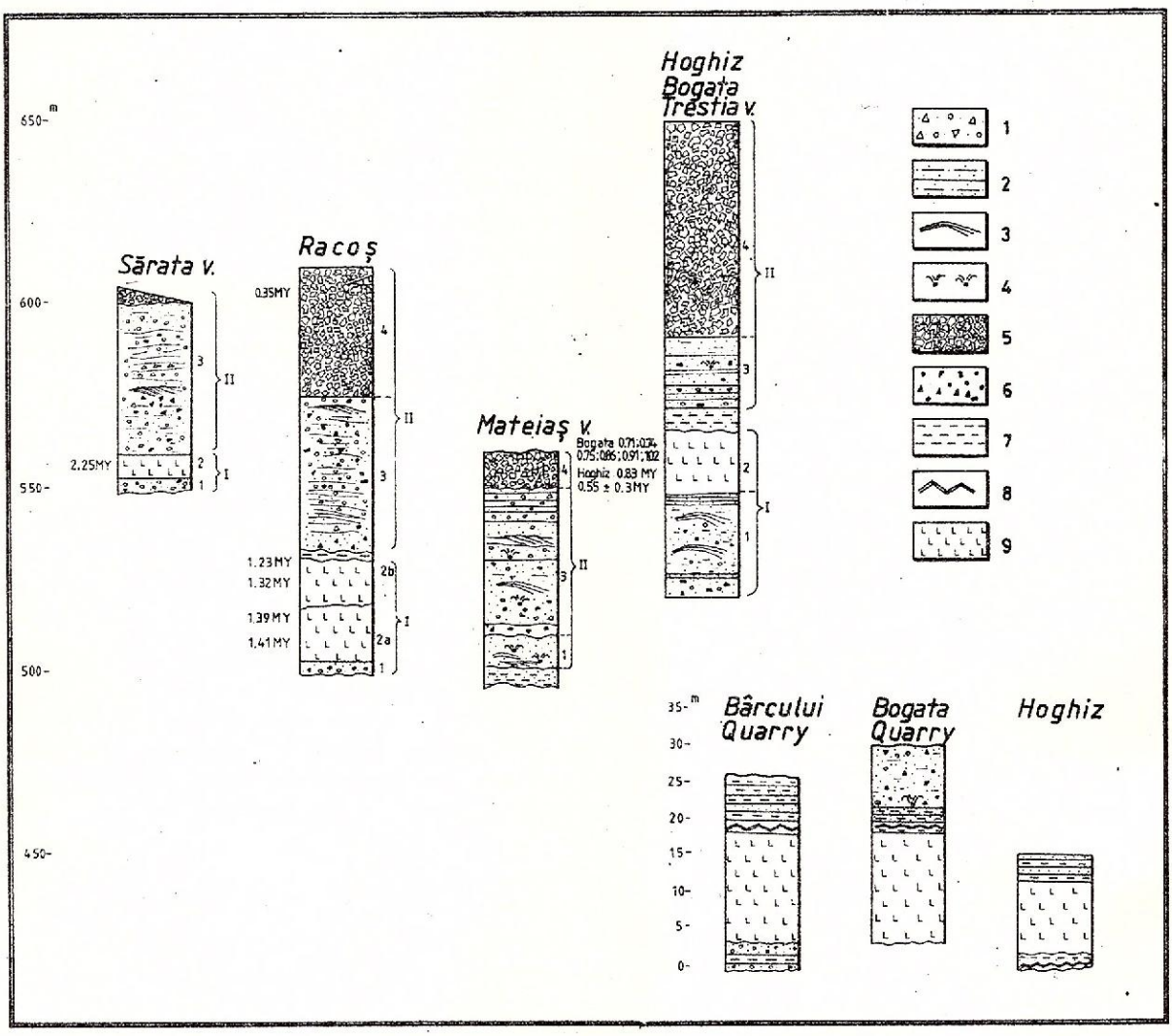

Fig. 2 – Synthetic lithological columns in basaltic sequence in Perşani Mts. 1, lapillistone, lapilli tuffs; 2, tuffs; 3, dunes; 4, ballistic blocks; 5, agglutinates and scoria tuffs; 6, accidental fragments; 7, clays; 8, paleosoils; 9, lava flows.

Sărata Valley, 5 km north of Racoşul de Jos village, covering about ca. 0.3 sqkm. The second area, of about 3.5 sqkm, occurs at both sides of Olt River near Racoşul de Jos and Mateiaş villages. The third and most extensive occurrence covers almost 40 sqkm, between Hoghiz and Bogata to the north and the waterheads of the Trestia Brook to the south. The southernmost area is situated at the Comana Valley covering 1.5 sqkm; however, it is not included in our following considerations. All these occurrence areas are distributed along NNE-SSW trending tectonic line.

The succession of volcanic sequences is presented in four generalised lithological columns, corresponding to the mentioned occurrence areas. K-Ar ages are displayed too, according to Casta (1980), Mihăilă, Kreutzer, 1981, Ghenea et al. (1981) (Fig. 2).

1. In the east-side bank of the Sărata Valley, an about 60 m long outcrop occurs. In its lowermost part a 2.5 m thick sequence of pyroclastic deposits, consisting of alternating poorly stratified and poorly sorted lapillistones and lapilli tuffs, can be seen. They display in places crossbedding laminae and lenticular features. Essential pyroclasts include porous or scoriaceous lapilli embedded in a poorly welded ashy matrix. Typical lapillistones tend to occur as lenticular beds. Occasionally larger fragments appear as scoria and as cauliflower-type bombs. The pyroclastic deposit is rich in accidental material represented by lithic fragments disrupted from the prevolcanic basement (Myocene rhyolitic tuffs and quartz pebbles).

At the upper part of the sequence, well-bedded lapillistones, containing fewer accidental fragments, prevail. The observed features suggest phreatomagmatic eruptions with pyroclastic surge and fallout emplace-

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ment mechanism.

The described deposits are topped by a lava flow with variable thickness (7–10 m) and with a blocky-jointed basal part. At the top of the lava flow, a transition to lapilli tuffs and lapillistones showing distinct bedding planes can be observed. At the upper part of this 5–10 m thick sequence, well-stratified pyroclastics, almost lacking essential components, crop out. Fine and coarse accessory basaltic lapilli, with a compact structure and accidental material (Mesozoic limestones, diatomites, quartz pebbles and Miocene tuffs) are the major components. Lherzolitic mantle xenoliths are also present. This sequence displays base surge depositional features originating in a phreatic eruption.

Some isolated basaltic scoria blocks have been found near the outcrop area, suggesting the possible presence of the strombolian fallout deposits, too.

2. In the Racoş-Mateiaş area, two lithological columns corresponding to two different volcanic structures will be described. At Racoş, plan-parallel layered lappilli tuffs occur in the lowermost part. Their essential components consist of dense or porous basalt fragments. The sequence is 2–2.5 m thick, poorly sorted and normally graded. It contains abundant accidental fragments, represented by quartz pebbles, Miocene tuffs and older limestones from the basement (Pl. I, Fig. 1).

In the main quarry of Racoş, at the base of the outcrop, the transition from first-phase pyroclastics to lava flow can be well noticed. Various developments of the cooling-generated joint pattern across the flow can also be studied. The thin (0.3–1 m) vesicular blocky-jointed basal part turns upwards into a thick (5–10 m) collonade in the middle of the flow and then into platty-jointing (1–2 m) at its uppermost part. In the western corner of the quarry, a vertical contact between lava and former pyroclastics marks the margin of a lava pond. The lava itself locally displays curved columns and two superposed collonades, separated by a thin platty-jointed zone suggesting two superposed flow units in the pond. Locally, some interesting structural features, mainly in the upper flow units, can be observed here in some blocky-jointed rocks. Chilled borders and grey or pinkish-brown core, showing granulated structures are obvious. Hit with a hammer, these rocks are disaggregate in small irregular, polyhedral, more or less equant, fragments (Pl. I, Fig. 2). These features were previously described as "cucuruz basalts" by quarry workers. Their average grain-size is similar to that of lapilli from almost all lapilli tuff deposits.

In the eastern part of the main quarry a lava tunnel with elliptical section has been found in the lower part of the lava flow. It is now partially filled by with an argillaceous material containing basalt blocks. Peridotitic xenoliths showing a rounded shape and 15–35 cm in diameter are frequently enclosed in the lava flows. Joint-plane fissures and larger vesicles contain in places incrustations and druses of well-crystallized calcite, generated from post-volcanic solutions.

In the main quarry a well-bedded lacustrine sedimentary sequence (argillaceous sand) with a thin discontinuous limestone level at the base is exposed on top of the lava flows.

In the eastern quarry of Racoş, thick sequences (tens of meters) of well-bedded pyroclastics overlie the lava flows, as a second pyroclastic level in the lithological column. A succession of finely bedded (cm, dm) poorly sorted lapilli tuffs, enclosing scoria and abundant accidental material, is well exposed. At the lava flow-pyroclastics boundary, a phreatic deposit made exclusively of accidental fragments is present. The patchy appearance of this sequence is a consequence of different abundance of light-coloured accidental fragments contrasting with dark essential material (Pl.II, Fig. 1). Locally, the deposits are updomed and cut by numerous normal faults, likely due to shallow intrusions accompanying the last phase activity (Pl. II, Fig. 1).

In the upper part of the quarry, unconformably overlying reddish scoria lapilli, layers of the neighbouring Hegyes scoria cone can be seen.

The Hegyes scoria cone is well exposed, due to its long-time excavation, now displaying only its remnants - a central pillar and surrounding lowered border. Repeated unconformities are visible in the walls of the central pillar, suggesting a complicated inner structure and significant changes of the summit area during the building up of the cone. Cone-building material mainly consists of coarse scoria fragments, bombs and spatter including tiny lava tongues in the central part and of finer unconsolidated scoria ash and lapilli layers at the periphery. Grain-size variation in the latter reflects pulsating intensity during the eruption.

At Mateiaş, near the Mateiaş Brook mouth, fine sand deposits with few coarse sand interlayers, several meters thick, occur. The deposit consists of basaltic component with frequent admixture of Miocene Perşani tuff fragments. Several thin mica-sand interbeds also occur. It displays plane-parallel and in places sandwave laminations with bedding set thicknesses of 0.05–0.5 cm. The bedding sets locally show some grading and sorting. Backset lamination in dune-like structures dips 15–30^0 towards SW, suggesting an origin in a NE situated volcanic source.

Upstream, along the Mateiaş Valley, a sequence of pyroclastic deposits is exposed. They consist of plane-parallel bedded tuffs (centimeter to decimeter thick) displaying multiple normal grading (Pl. II, Fig. 2), rarely reverse grading and in places dune-like cross-

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bedding (Pl. II, Fig. 2), and fine tuffs showing mantle bedding and a network of fine polygonal cracks. Compositionally, they are characterized by the variable abundance of essential and accidental fragments.

Another important feature of this sequence is the presence of numerous bomb-sags formed by the impact of accidental ballistic blocks that fell in the soft moistened fine ash layers (Pl. III, Fig. 1). Bomb sags are obviously symmetrical and only a few assymetrical ones have been found, suggesting their source towards east. Impacting blocks are lacking from certain sags. These small impact depressions favoured coarse-fraction accumulation from the base of the next deposit (Pl. III, Fig. 1). The features of the sequence are characteristic of subaerial fallout and surge mechanisms.

At the bottom of the sequence a 0.5 m thick bed of clay is exposed.

The next sequence is 2–3 m thick and consists of massive unsorted lapilli tuff, showing local concentrations of accessory basaltic blocks (0.5–1 m in diameter) or Miocene Perşani tuff accidental fragments (0.3–0.7 m in diameter). Its features are consistent with the pyroclastic flow origin of the sequence.

Another bedded sequence (10 meters thick), showing a slightly coarser grain-size, follows. It is represented by lapillistones and some fine and coarser tuff interlayers, within which the essential component prevails. Accidental Perşani tuff fragments enclosed in this sequence – and in the previously described one as well – are well rounded and display desiccation cracks (Pl. III, Fig. 2). Planar and wavy bedding with internal cross-lamination in places, normal grading and the presence of bomb sags are the main structural features of the sequence, suggesting subaerial fallout and base surge emplacement processes.

In the Mateiaş Valley area, the uppermost sequence is a 10-15 m thick massive to poorly bedded deposit, consisting of more or less welded ash and basaltic scoria, agglutinated spatter in places, typical of strombolian explosive activity.

In the Mateiaş quarry, on the left side of the Olt River, a lava flow sequence more than 50 m thick crops out. A giant inclusion of bedded tuffs (10/7 m) engulfed in the lava flow can be seen in the quarry, suggesting former explosive activity. Another sequence of bedded tuffs is present at its top. It is not very clear whether the lava flow is confined to a local vent, or its eruption center is situated somewhere within the Racoş area, where the effusive activity was important.

The generalised lithologic column for the Hoghiz-Bogata-Trestia Valley area is shown in Figure 2.

Like in all other areas, the first-phase volcanic activity displays an explosive character, during which thick sequences (tens of meters) of bedded pyroclastics accumulated. They are well exposed on the left side of the Trestia Brook (Pl. IV, Fig. 1) and in La Gruiu area. These deposits consist of various grain-sized pyroclastic rocks, ranging from blocks to fine ash-sized fragments. They are typically composed of coarse accidental clasts (Mesozoic limestone, conglomerate pebble and gravel, Miocene rhyodacitic tuffs, disrupted from the basement) of more or less porous essential basaltic lapilli, and of an abundant basaltic ash matrix. Locally, abundant mantle xenoliths and loose amphybole and pyroxene megacrysts are included.

In the northern and eastern part of the area, the sequence is thinner (1–5 m) and in places it overlies a clay deposit displaying paleosol features at its top (Pl. IV, Fig. 2). In the Bogata area, this sequence has not been found. Depositional features suggest subaerial fallout and surge deposits.

Initial explosive activity was followed by massive lava outpourings. Lava flow sections are well exposed in several quarries along the Bârc and Bogata valleys and in natural outcrops along the Hoghiz and Trestia valleys, and in the left bank of the Olt River, near Bogata. The lava flow thickness is relatively constant (10–15 m). The basal part is vesicular, displaying horizontal platy jointing, which turns upwards into vertical columns (Pl. V, Fig. 1). The upper part occasionally presents blocky fragmentation similar to that of Racoş lava flows, showing inhomogeneous and granulated structure. At Hoghiz this kind of structures seems to be generalised for the entire lava pond.

On top of the lava flow, an argillaceous sequence has accumulated, as it was observed in the Bârc and Bogata quarries. This lacustrine deposit contains several tiny intercalations of fine airfall tuff.

The next sequence is well exposed in the Bogata quarry, being represented by well-bedded pyroclastic deposits (7–10 m thick). In their lower part accidental fragments prevail, represented by limestone, Perşani tuff (most of them rounded with desiccation cracks), quartz pebble and gravel. Some bomb sags have been observed too. In the upper part of the sequence, cross-bedding lamination typical of base surge depositional processes can be seen, whereas in the lower part of the sequence fallout deposition prevails.

As in the whole Perşani basaltic province, volcanic activity finished with scoria cone building strombolian eruptions (Pl. V, Fig. 2). Seven cones displaying well preserved morphology have been recognised; among them Măguricea Mare, Măguricea Mică, Gruiu and Trestia are best exposed. Local presence of scoria blocks and bombs (Pl. VI, Fig. 1) in some other areas, without associated relevant morphology, has also been found. Strombolian fallout deposits exposed in the quarry north of Măguricea Mare are similar or identical to those described at Racoş (Hegyes Hill) (Pl. VI, Fig. 2).

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Volcanological interpretations

Despite of some differences in their K-Ar ages (Casta, 1980; Mihăilă, Kreutzer, 1981), the various occurrence display similar sequence successions of the explosive and effusive products. First-phase activity was explosive everywhere. Large-scale phreatic and phreatomagmatic explosive activity occurred, during which a complex sequence of subaerial fallout and base surge deposits accumulated. Phreatic and phreatomagmatic eruptive centers were likely of maar and/or tuff-ring type. Their initial features were not preserved, due to the subsequent volcanic activity. Based on bedding plane position measurements and using overall deposit geometry and some morphological criteria, some of them have tentatively been identified, the largest one in the Barc-Trestia valleys area.

Initial explosive activity was followed by massive lava outpourings, probably in the craterial areas of the former maar structures. The observed lava flow sections suggest an unique flow unit, except for Racoş area, covering land surfaces around effusive vents. The source areas were located at Sărata, Racoş, Bogata-Hoghiz and Barc-Trestia valleys. The source of the Mateiaş lava flow could be either from a local vent, or from the Racoş center.

The two initial – explosive and then effusive – phases could be interpreted as a first-stage volcanic activity all over the area. Geological observations suggest a longer repose time following this first stage, during which local lacustrine environments developed and argillaceous sequences accumulated at the top of the lava flows. Thin intercalations in some clay deposits (Barc Valley) suggest intermittent small-scale explosive activity occurring at this time.

The second-stage activity started again with phreatic and phreatomagmatic explosive eruptions. The resulted deposits show similar features to those of their first-stage counterparts, including prevailing fallout and surge deposition. Volcanic activity was not synchronous at different eruptive centers, since overlapping spatial relationships between the same-stage pyroclastics belonging to different vents have been encountered, such as along Mateiaş Valley, where phreatomagmatic deposits from a local vent overlie distal phreatomagmatic tuffs belonging to an eruptive vent in the Racoş area. Vent locations during this stage were partly the same as for the first stage activity, but new vents has likely been added. They were recognised as circular depressions, well expressed morphologically in the Hoghiz-Bogata-Trestia Valley area. Their initial features likely were of tuff-ring type.

Volcanic activity ended in the Perşani Mts with scoria cone building strombolian eruptions, as a consequence of the decreasing magmatic volatile content and /or external water availability to trigger explosive magma-water interaction.

The occurrence of the two main eruptive stages is interpreted as a result of two magma pulses, each of them starting with violent explosive eruptions, followed by less energetic eruptive events.

The best preserved volcanic features are the scoria cones. They are superimposed upon first-stage volcanic edifices partially destroyed by second-stage phreatic and phreatomagmatic eruptions. Of all, the Hoghiz-Bogata-Trestia Valley occurrence area displays the most complex structure, including several adjoining and/or coalescing tuff-ring type edifices, partially invaded by ponded lava flows during the first stage, and superimposed second-stage scoria cones or their remnants.

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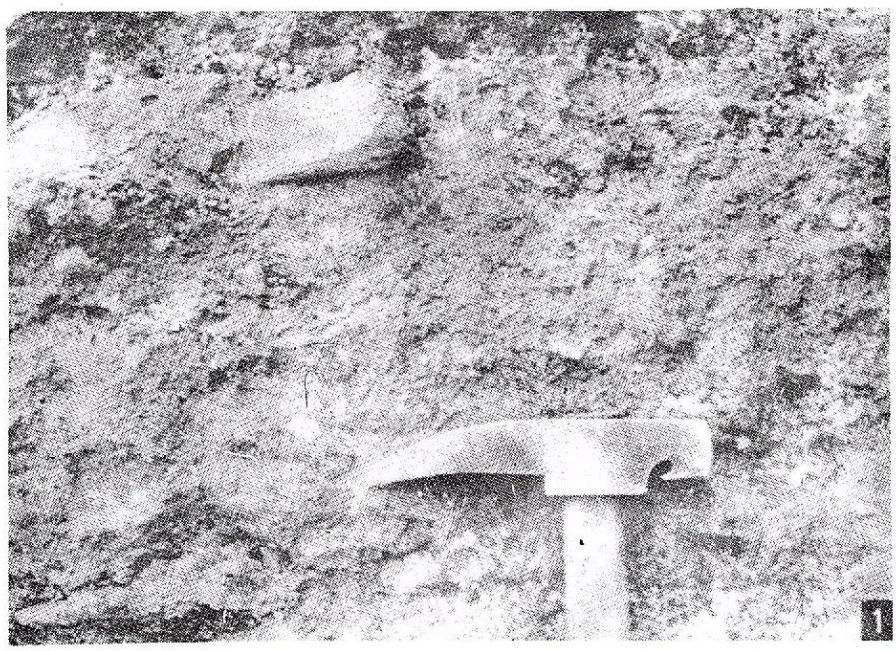

Fig. 1 – Accidental fragments of Miocene rhyolitic Perşani tuff (1) and Mesozoic limestone (smaller, light-coloured) in first-stage phreatomagmatic deposit; Racoş quarry.

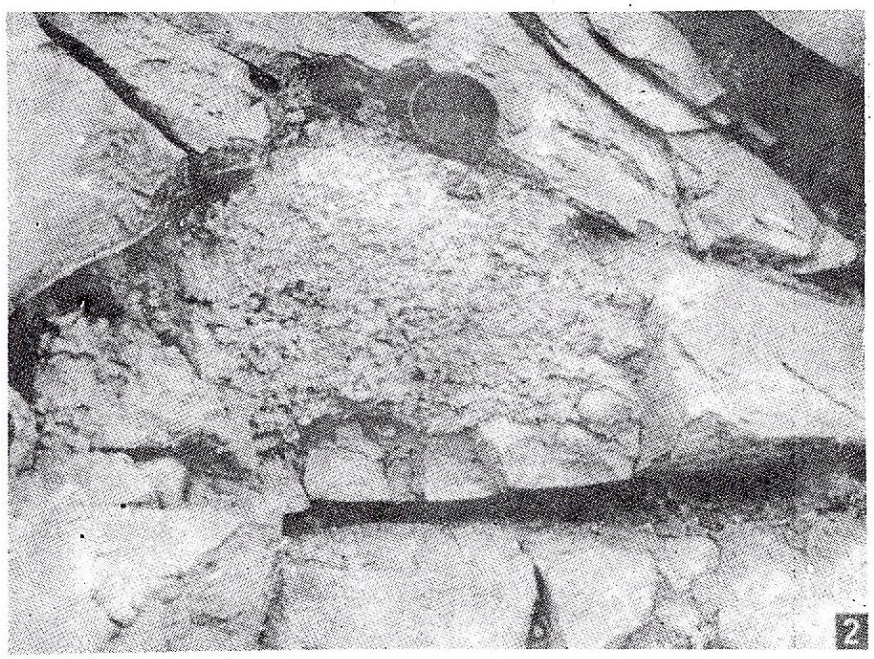

Fig. 2 – Detail in blocky jointed lava flow with granulated core ("cucuruz basalt") and massive crust within a polyhedral block; Racoş quarry.

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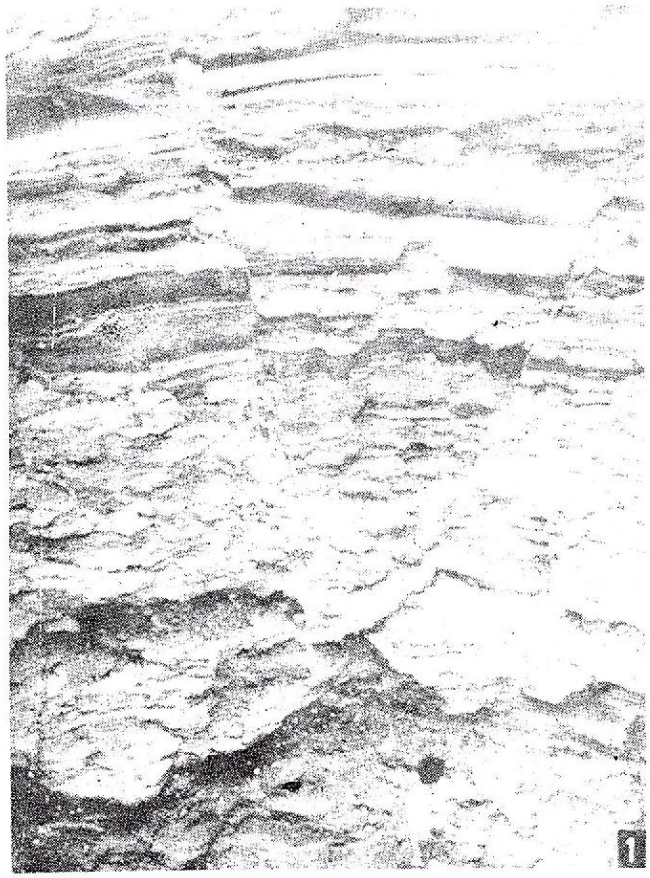

Fig. 1 – Layered third-stage phreatomagmatic deposits showing postdepositional uplift and normal faulting in the north-eastern quarry at Racoş. Small light-coloured fragments represent accidental material.

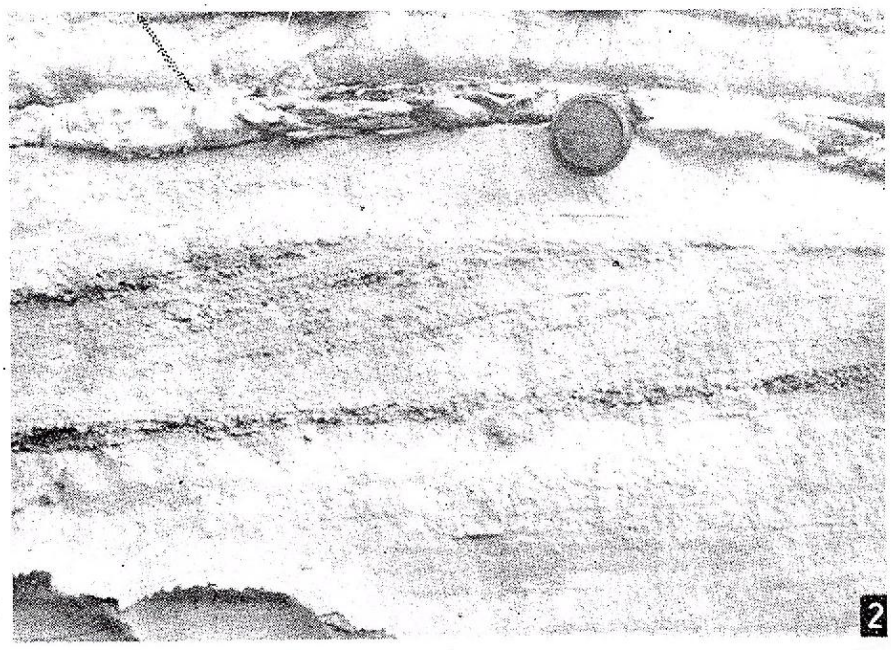

Fig. 2 – Horizontally layered phreatomagmatic ashfall and cross-bedded base surge deposits; first-stage pyroclastics; Mateiaş Valley.

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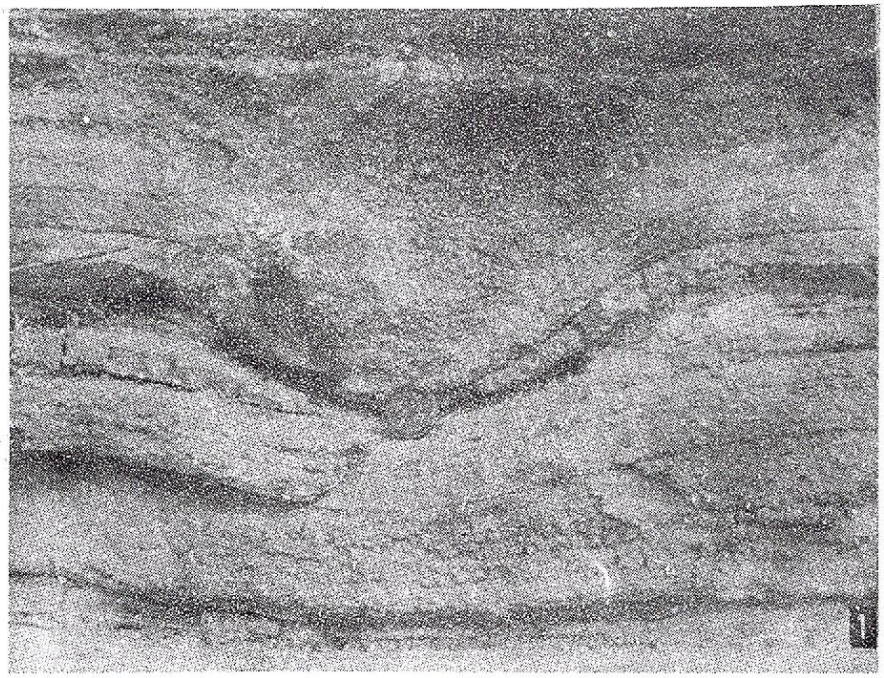

Fig. 1 – Bomb-sag in horizontally layered phreatomagmatic ashfall tuffs; Mateiaş Valley. Note the coarse-grained filling in the sag from the overlying bed. Polygonal desiccation crack network in the light-coloured fine fallout ash at the bottom of the picture is an evidence for wet-ash deposition.

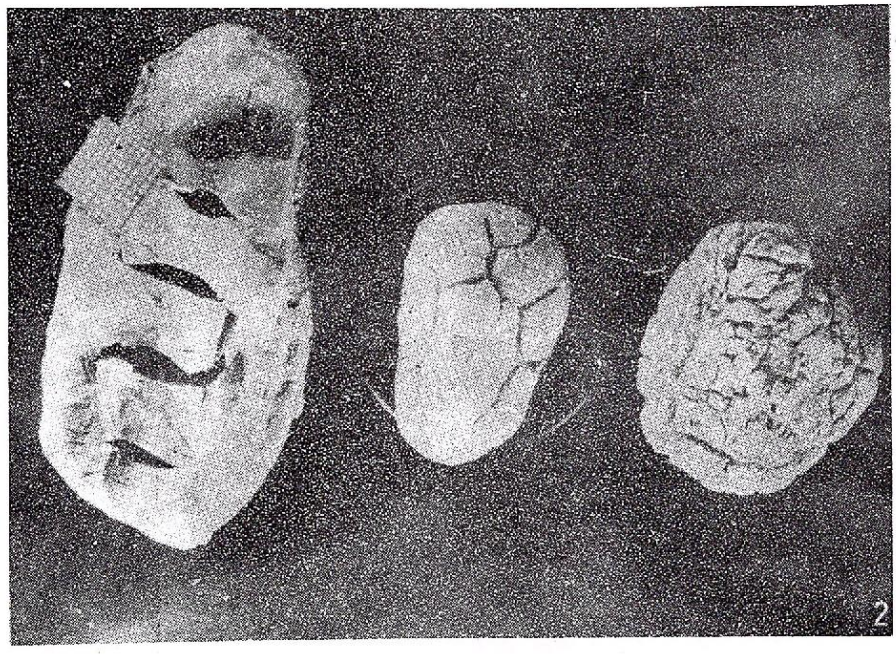

Fig. 2 – Rounded accidental Miocene tuff fragments with desiccation cracks in base surge deposits; Mateiaş Valley.

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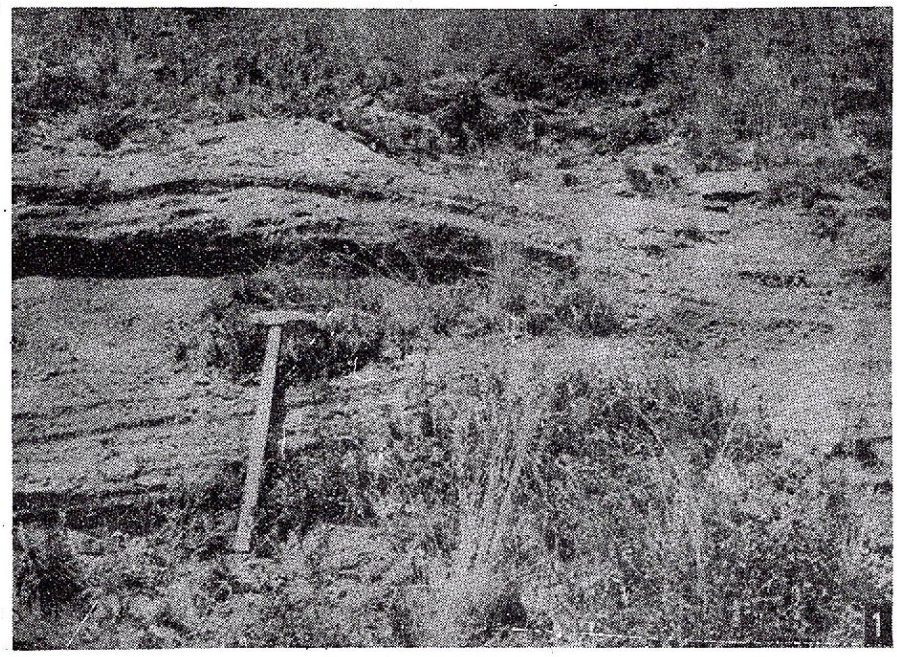

Fig. 1 – Cross-bedded base surge deposits within first-stage pyroclastic sequence croping out in the northern side of the Trestia Valley.



Fig. 2 – Unconformable boundary between Quaternary argillaceous deposits – solidificated in its upper part – and layered pyroclastic fall deposits. Note mantle bedding at the base of pyroclastics Bârc Valley.

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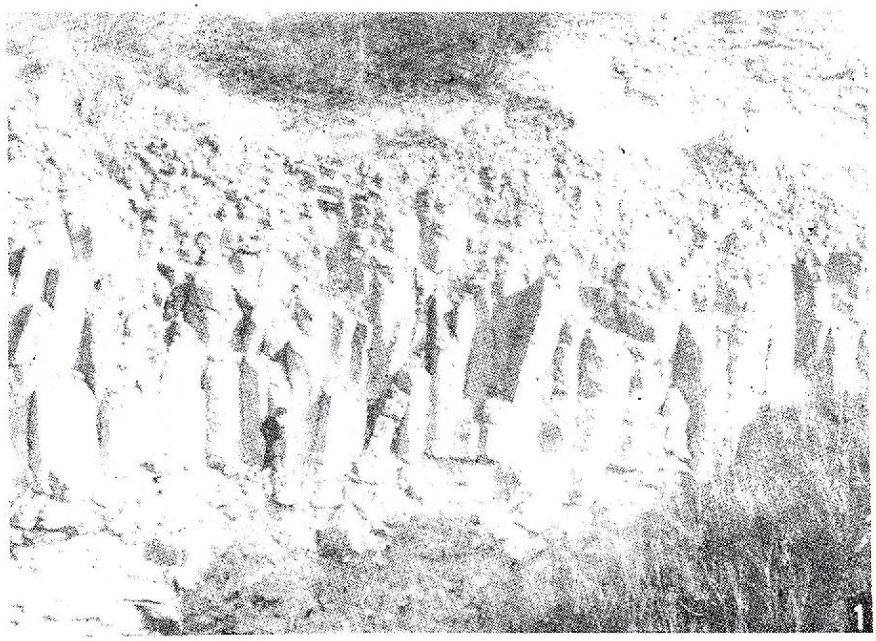

Fig. 1 – Lava flow displaying basal platty jointing (a), vertical columns in the middle part (b) and blocky fragmentation at the top (c). It is covered by third-stage pyroclastics. Bogata quarry.

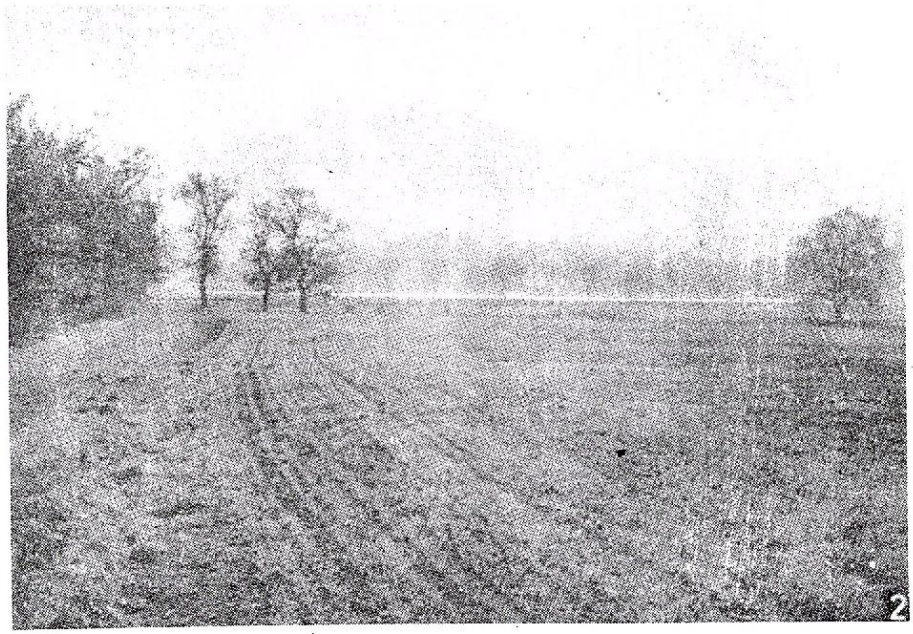

Fig. 2 – Typical scoria cone morphology slightly modified by subsequent erosion. Gruiu cone.

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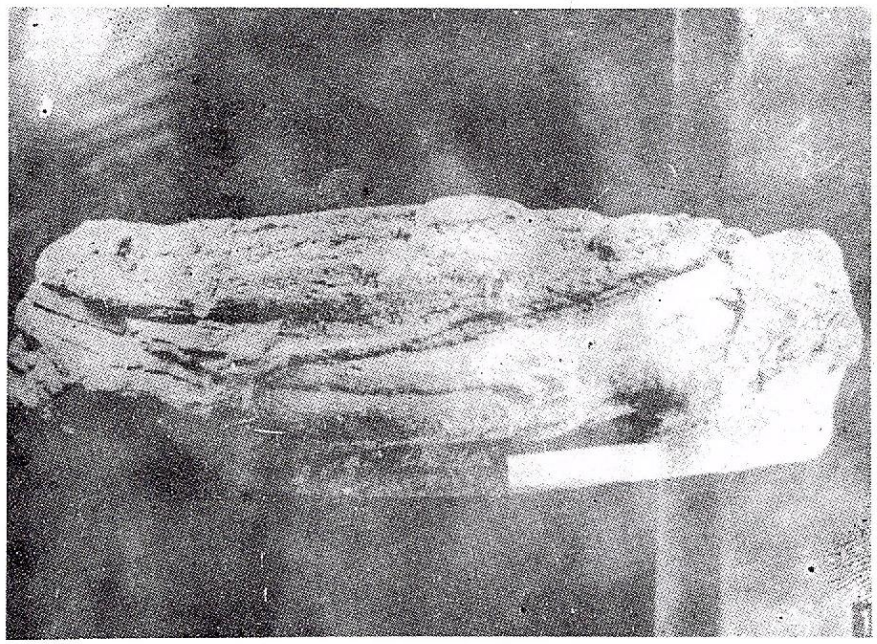

Fig. 1 – Slightly contorted ribbon bomb from one of the scoria cones west of Trestia Valley.

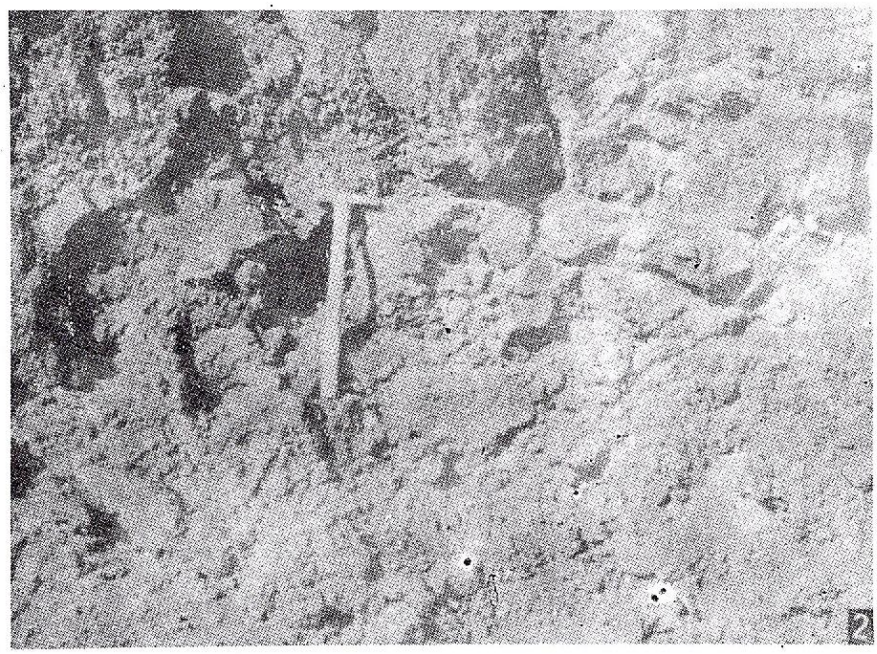

Fig. 2 Agglutinate (spatter deposit) in the central part of the Măgiricea III scoria cone. Note bread-crust cracking on the surface of some scoria fragments (right bottom) and highly vesicular interiors of many scoria.

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PETROLOGY, GEOCHEMISTRY AND ORIGIN OF THE PLEISTOCENE WITHINPLATE BASALTS OF LUCAREŢ (BANAT – ROMANIA)

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Key words: Basalts. Alkali Basalts. Pleistocene. Partial melting. Major elements. Minor elements. Rare earth. Banat. Romania.

Abstract: The plateau of the olivine alkali basalts is situated in Banat (Romania), at about 35 km east of Timişoara. It consists of lava flows and more rarely of pyroclastics. The eruption of basalts took place on the Recaş-Cuieş crustal fracture, during a distension period at the beginning of the Pleistocene. The volcanic activity manifested itself by quite fluid basaltic lava flows and yielded alkali basalts with olivine and virtual nepheline. The rocks came from an essexitegabbroic magma formed in the mantle at a quite great depth by the melting (0.5–5%) of the garnet peridotitic substratum. They display obvious features of continental withinplate basalts, as shown by the contents of major and minor elements and their ratios as well as by the diagrams presented in the paper.

Introduction

The olivine basalts of Lucareţ constitute a plateau lying between the mentioned locality, Şanoviţa and Topolovăţu Mare, at about 35 km east of Timişoara, on the northern side of the Bega channel and of the Timiş Valley. The plateau has been eroded by the southern and eastern valleys, so that its surface dips slightly towards the confluence of the Chizdia Brook with the Bega channel (Fig. 1). Nowadays it is overlain by soil.

From the geological and petrographical point of view the Lucareţ basalts were known since the last century. The first observations belongs to Loczy (1882), Schafarzik (1882), Gáal (1891) and Adda (1898) who studied particularly the rocks from the Lucareţ quarry. Adda pointed out that the basalts erupted subaerially during the Pontian through a central channel situated under Piatra Roşie Hill – the uppermost point of the basaltic plateau, starting with strong explosions. Şanoviţa basalts were studied later by Mauritz (1930). Mateescu (1936) considered that the Lucareţ basalts were products of an epiorogenic volcanism which manifested at the beginning of the Quaternary on a N-S trending fracture, as the Gătaia basalts and the Lipova and Buziaş mineral waters. The Lucareţ basalts were also attributed to the Quaternary by Drăghici (1953, unpubl. rep.), who emphasized that the basaltic lavas overlie the Upper Pliocene sands.

A detailed petrographic study on basalts, beside petrochemical data, was ellaborated by Savul and Nichita (1944), who described the succession of the basaltic eruptions in the Lucareţ quarries. The mentioned authors remarked that the rocks display an Atlantic character and originate in a teralitgabbroic magma, as shown by their slightly alkaline character. Peltz et al. (1971) and Peltz and Bratosin (1988) also studied the Lucareţ basalts and inferrred that they originate in a gabbroic magma.

In 1966 several boreholes were drilled in the basaltic plateau by ISEM (IFLGS) for building stones (Tănăsuică, 1966, unpubl. rep.). The technical characteristics of the basalts were presented by Pârvu et al. (1977).

As no modern petrological and geochemical study has been elaborated up till now, we hope that the present paper will meet the long-felt want.

Structural types of basalts and their extension

The basaltic plateau consists of two obvious nappes or lava flows, with different extension, separated by a level of basaltic scoria (Fig. 1), which have been found in the Lucareţ quarries (Savul, Nichita, 1944) and in the boreholes south of it (Tănăsuică, 1966, unpubl. rep.). The drilling data have been used for the extrapolation of the limits of the upper basaltic flow. In the Lucareţ quarries the lava flow is about 10 m thick, whereas in the boreholes in the Zăbrani Hill its thickness is more than 16 m. That flow overlies tuffs and basaltic scoria and it determined the caustic metamorphism which caused the burning of the Upper Pliocene

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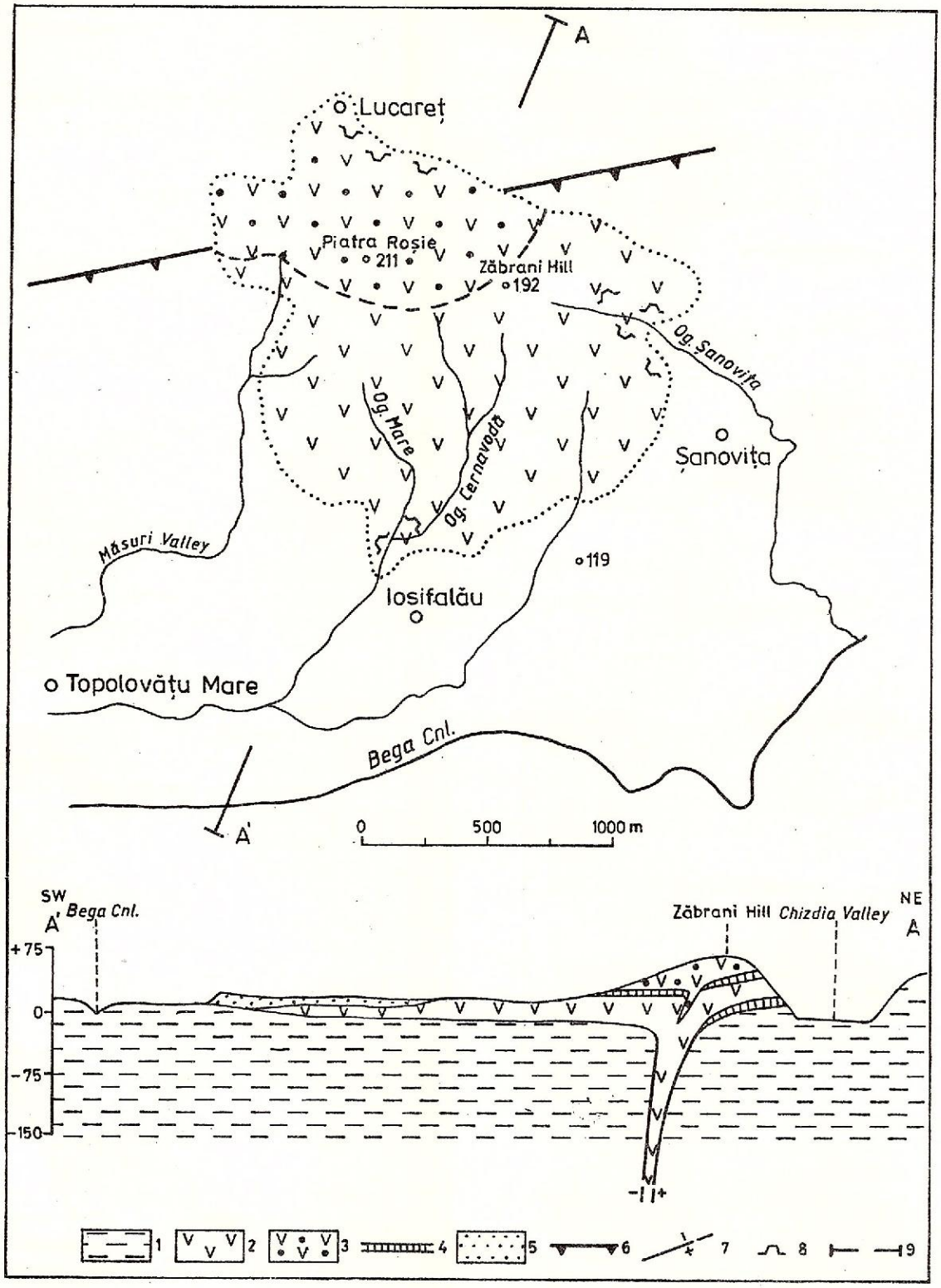

Fig. 1 – Geological map of the Lucareț basalts. 1, Pliocene sedimentary deposits; 2, lower basalts; 3, upper basalts; 4, basaltic scoria level; 5, terrace deposits; 6, trend of Recaș-Cuieș crustal fracture; 7, location in section of the Recaș-Cuieș crustal fracture; 8, quarries; 9, line of the geological section.

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sands. The lower flow occupies about 25 sq km of the basaltic area. Drăghici (1953, unpubl. rep.) stated that the wells from the Bega alluvial plain intercepted the basalts, which points out that the flow extends more southwards where it was eroded (see the geological section in Fig. 1). It is an argument in favour of Adda's opinion (1898) who considered that the basaltic flows occupied an area of 40 sq km.

At Lucareţ the lower flow is overlain by a thin level of basaltic scoria which becomes thicker (about 7 m) southwards, at the expense of the upper flow basalts, that disappear (Fig. 1). This flow represents the last basaltic lava eruption. Its thickness varies between 1 and 4m and occupies a reduced surface round the Piatra Roşie Hill, which marks the volcanic channel and the basaltic neck, respectively.

The basaltic neck is over-shaped, elongated towards ENE-WSW, as shown by the geophysical evidence. It is enrooted in the plane of the Recaş-Cuieş crustal fracture (Andrei, Cristescu, 1966) having the same trend as the elongation of the basaltic neck (Fig. 1). Although older, the mentioned fracture moved again at the end of Neogene. The elongated shape of the basaltic neck shows that towards the depth it tended to become a feeder dyke.

At about 3.5 km WSW of the Lucareţ basalts plateau, on the extension of the crustal fracture towards Recaş, one can observe another olivine basalt occurence much smaller, figured on the Geological Map of Romania sc. 1:500000 and mentioned by Pârvu et al. (1977) north of Topolovăţu Mare. It might also be enrooted on the plane of the Recaş-Cuieş crustal fracture, as in case of the linear eruptions; it is also possible it might represent a vestige of the Lucareţ basalts plateau which extended there, but was eroded.

The basalt flows, especially the lower one, which are being worked in the Şanoviţa and Zăbrani quarries, include three basalt types: massive basalts, vacuolar basalts (locally "schweizer") and nodular basalts (locally "cucuruz" or "Sonnenbrenner"). In the Zăbrani Hill quarry (Fig. 1) the lower lava nappe consists clearly of two levels (flows ?) of basalts with different structures. The lower level is repesented by vacuolar basalts, separated as plates after a system of horizontal fissures, parallel with the lava flow plane. As in case of the basalts of this type, the vacuoles have usually a diameter of 0.2-1 cm; nevertheless, in the Lucareţ quarry vacuoles of 18/13 cm with lava outgrowths on their walls were observed in 1969. The upper level consists of nodular basalts separated, like those in the Şanoviţa and Lucareţ quarries, in the characteristic columns after three vertical fissure systems, a phenomenon specific to the lower basalt flow.

In the Lucareţ plateau, basalts display an uniform petrographic and chemical composition (Table 1). They show, a porphyritic structure determined by the olivine phenocrysts and augite microphenocrysts, small crystals of augite, apatite, Ti-magnetite, ilmenite, pyrite and glass. The texture is massive, with a fluidal aspect round the phenocrysts. Basalts also include in places xenoliths of recrystallized metamorphic rocks, as well as small nests and agglomerations of augite crystals, probably formed by magma reaction with some quartz xenoliths.

Olivine (forsterite) phenocrysts occur either isolated or grouped, which gives the glomeroporphyritic structure of the rock. The largest phenocrysts are corroded by magma. Some nonco.roded crystals show an obvious fayalitic zone on the margins, a component also occurring in the CIPW normative composition (Table 2). In some rocks the olivine crystals – small or large – are altered on the margins, where a thin aureole of secondary iron oxides is formed. Savul and Nichita (1944) supposed that the oxidation aureole was formed on account of a fayalitic zone found also on the margin of the corroded olivine phenocrysts. That being the situation, there are two hypotheses on the origin of the phenocrysts: (1) they represent magma corroded phenocrysts; (2) they represent partly melted xenocrysts carried by the magma from the mantle, around which the fayalitic zone was deposited later on. In the second case the phenocrysts without oxidation aureole should be surrounded by the fayalite zone, but it has not been observed up till now.

Augite represents a variety with a higher content of Ti and Fe (Tables 1 and 2), with $c \wedge Ng = 45–47^0$. Plagioclase microlites possess an average andesine composition (An_{40}) (see Table 2); they are more basic towards the interior ($An_{65–60}$) and on the margin they display a zone which can reach the oligoclase composition (An_{20}). All this shows that the interstitial glass has a high-Na_2O content pointing out the hawaiitic character of the basalts.

Geochemistry and tectonic setting

Table 1 shows the chemical composition of the basalts, from which it results that SiO_2 varies between 45.75 and 49.20 per cent, with an average of 47 per cent, values that situate the rocks much below the upper limit of 52 per cent SiO_2 of the basalts (Gill, 1981; Ewart, 1982). The ratios between the main major chemical constituents of the rocks plot them on the diagram in Figure 2 (Irvine, Baragar, 1971; Hutchinson, 1982) in the calc-alkaline field. It is noteworthy that the crystallization order of the minerals in the Lucareţ basalts follows the trends of the calc-alkaline or basic alkaline magmatic series: olivine-clinopyroxene-plagioclase.

Institutul Geologic al României

Table 1
Chemical composition of the Lucareţ Basalts

No*	1	2	3	4	5	6	7	8	9	10	11
SiO_2	45.75	46.45	46.51	47.27	47.41	47.61	47.97	48.03	48.13	48.57	49.20
TiO_2	2.24	2.25	2.44	2.36	2.36	2.28	2.32	2.39	2.44	2.28	2.20
Al_2O_3	14.85	14.40	15.50	15.40	13.84	14.55	14.65	14.85	14.04	15.25	14.59
Fe_2O_3	3.89	4.68	9.38	4.22	5.54	2.44	2.59	6.08	2.65	3.79	2.97
FeO	6.57	5.83	2.16	6.48	5.50	7.95	7.73	4.96	7.70	6.79	7.66
MnO	0.16	0.17	0.17	0.17	0.18	0.01	0.17	0.16	0.16	0.17	0.17
MgO	7.91	7.64	6.90	7.47	7.67	8.57	8.07	7.50	8.39	8.33	8.44
CaO	9.53	9.78	8.48	8.71	9.35	9.05	8.59	8.54	8.39	8.33	8.44
K_2O	1.37	1.65	1.50	1.64	1.80	1.60	1.61	1.59	1.50	1.59	1.49
Na_2O	4.32	4.14	3.05	3.58	4.26	3.83	4.15	3.50	3.68	3.91	3.61
P_2O_5	0.63	0.46	0.60	0.53	0.45	0.40	0.39	0.42	0.41	0.43	0.40
CO_2	1.53	1.99	–	–	1.12	1.73	–	–	–	–	0.50
S	0.10	0.02	0.12	0.11	0.03	0.04	0.05	0.05	0.34	0.02	0.06
H_2O^+	0.60	0.61	2.69	1.54	0.91	0.57	1.53	2.42	1.45	1.10	1.01
Total	99.45	100.06	99.50	99.48	100.42	100.63	99.82	100.49	99.61	100.77	100.36
Ni**	150	150	210	185	160	140	185	200	140	100	90
Co	21	17	33	24	23	17	27	22	18	31	26
Cr	140	135	280	245	170	150	230	200	170	150	125
V	220	135	220	220	190	160	240	210	160	135	115
Sc	20	17	24	20	22	16	22	22	18	15	13
Nb	75	46	78	60	85	36	70	50	43	40	38
Zr	270	210	300	250	280	190	280	250	200	230	200
Y	35	26	40	36	35	28	36	34	30	19	18
Yb	1.4	1.3	1.9	1.3	1.4	1.2	1.4	1.5	1.3	2.2	2.4
La	46	47	44	40	47	35	42	47	33	30	36
Ba	620	520	670	650	600	450	700	730	570	600	500
Sr	1200	950	1000	1050	1000	850	1150	1200	920	1000	860
Cu	23	22	37	23	23	19	37	25	17	36	20
Zn	34	44	<30	50	45	44	500	40	44	34	38
Ga	16	13	17	12	14	15	14	11	13	13	15
Pb	4	13	3	5.5	2	3	460	9	6	<2	<2
U		1.8	2.1		2.4		1.6	1.3	2.0		
Th		5.2	7.1		5.8		5.3	5.2	4.6		

*The rocks analysed in this table are: 1 and 3, olivine basalts from the Zăbrani quarry; 2 and 4-11, olivine basalts from the Şanoviţa quarries.

**The trace elements have been determined by emission spectrography. Sn<2.

The high contents of TiO_2 from these rocks, with average values of 2.32 per cent, and of Na_2O and K_2O, with average values of 3.82 per cent and 1.58 per cent, respectively, as well as of P_2O_5, with average values of 0.46 per cent, in comparison with the contents of the above constituents from other types of basalts are also worth mentioning. All the components emphasize, by their contents, the alkaline character of the basalts, as rendered on the diagram in Figure 3 (Bass et al., 1973). The slightly alkaline character of the basalts is also proved by the CIPW normative composition of the rocks (Table 2), in which the normative nepheline values vary between 0 and 5.89 per cent (average value 2.41 per cent).

On the diagram in Figure 4 (Bass et al., 1973), with one exception, all the rocks plot at the bottom of hawaiite field, pointing out the slightly alkaline character of the basalts, rocks in which the relation $Na_2O > K_2O$ is the most frequent. The average values of the Niggli parameters (Burri, 1959) - si–109.7; al–20.4; fm–47.2; c–21 9; alk–10.8; k–0.21; mg–0.57; ti–4.0 - show that the basalts came from an "essexitgabbroic"-type magma from the series of sodic gabbroic magma. This character is also emphasized by the relation $Na_2O > K_2O$, the average value of the total alkalis - 5.48 per cent - and the average value of the ratio K_2O/Na_2O - 0.41 per cent. These values are quite close to those established by Cullers and Graf (1984) for the alkali basalts.

Institutul Geologic al României

Table 2
Normative (CIPW) composition of the Lucareţ basalts

No	1	2	3	4	5	6	7	8	9	10	11
or	8.15	9.81	9.11	9.85	10.73	9.51	9.66	9.63	9.00	9.50	8.90
ab	27.58	29.45	26.53	27.63	28.72	30.15	24.80	30.35	27.50	27.81	30.86
an	17.19	15.93	24.85	21.44	13.45	17.89	16.85	20.61	17.63	19.58	19.40
ne	4.99	3.14	–	1.70	4.15	1.33	5.89	–	2.23	3.05	–
di	10.26	11.46	10.67	12.02	16.48	7.85	13.81	14.99	14.14	12.78	8.85
he	2.38	1.79	–	2.60	1.43	2.81	5.06	0.24	4.44	3.14	3.00
en	–	–	10.74	–	–	–	–	3.57	–	–	3.46
fs	–	–	–	–	–	–	–	0.07	–	–	1.35
fo	10.56	9.69	1.39	9.34	8.16	12.49	9.82	6.04	10.28	10.55	9.58
fa	3.10	1.91	–	2.55	0.90	5.65	4.54	0.12	4.08	3.28	4.11
il	4.28	4.30	4.76	4.55	4.52	4.36	4.47	4.65	4.71	4.38	4.22
mt	5.67	6.82	0.01	6.21	8.10	3.56	3.81	9.03	3.90	5.55	4.35
hm	–	–	9.63	–	–	–	–	–	–	–	–
cc	3.50	4.55	–	–	2.57	3.95	–	–	–	–	1.15
ap	1.61	1.17	1.57	1.37	1.15	1.02	1.01	1.09	1.06	1.10	1.03
py	0.19	0.04	0.23	0.21	0.06	0.08	0.09	0.10	0.65	0.04	0.11

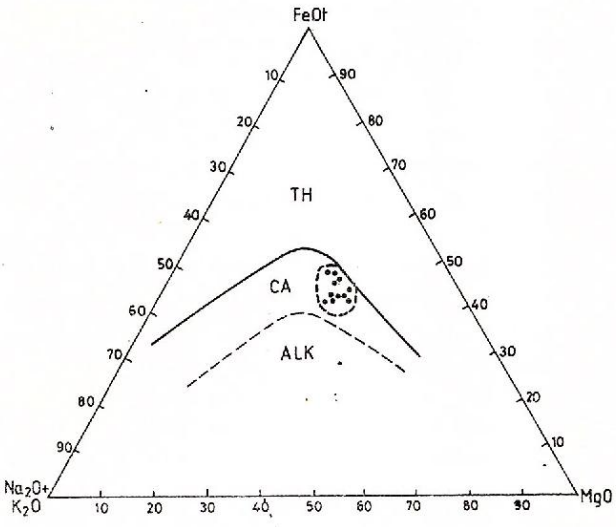

Fig. 2 - Plot of the basalts on the FeOt - MgO - $Na_2O + K_2O$ diagram. Th, tholeiitic rocks field; Ca, calc-alakaline rocks field; Alk, alkaline rocks field.

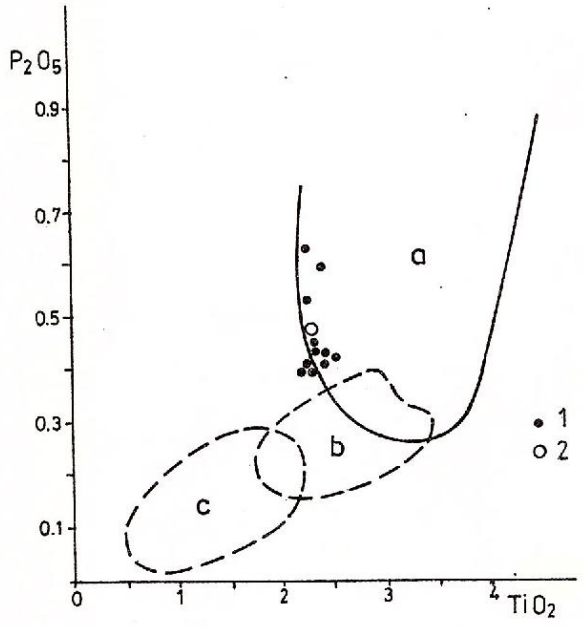

Fig. 3 - Plot of the basalts on the P_2O_5 - TiO_2 diagram. a, alkali basaltic field; b, ocean island basalt field; c, ocean ridge basalt field. 1, analysed rocks; 2, average value of the analysed rocks.

Although TiO_2 shows high contents in all the rocks, the values are similar with those of the normal basalts, varying between 115 and 240 ppm. It shows that the high TiO_2 contents do not represent the result of the basaltic magma differentiation, but the ratio between the two elements was determined by the conditions of partial melting of the ultramafic substratum from the mantle at a very great depth, thus resulting a typical continental withinplate magma, as shown in Figure 5 (Pearce, 1980). We enjoy this conclusion the more so as the basaltic rocks show that the parental magma differentiation was almost inexistent, if we take into account the quite reduced SiO_2 variation.

Higher contents than in other basaltic types are also presented by Zr (average value 242 ppm). These high contents make the average value of the ratio Ti/Zr = 57, lower than in other basaltic types. The Zr/Y ratio

Institutul Geologic al României

plot the basalts in the slightly alkaline basic rocks of continental withinplate type (Fig. 6, Pearce, 1980).

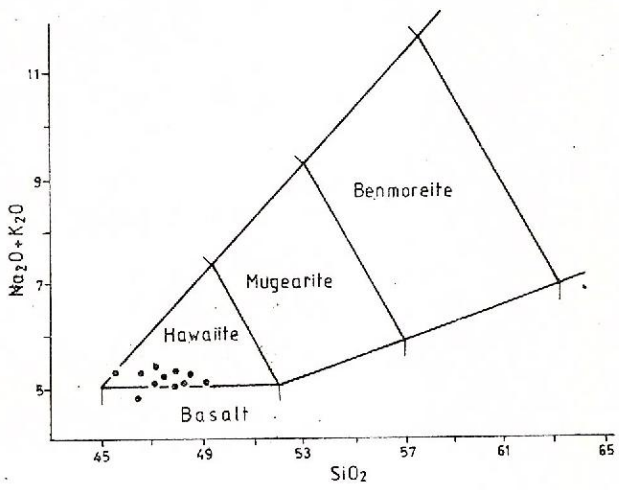

Fig. 4 - Plot of the basalts on the $Na_2O + K_2O$ - SiO_2 diagram.

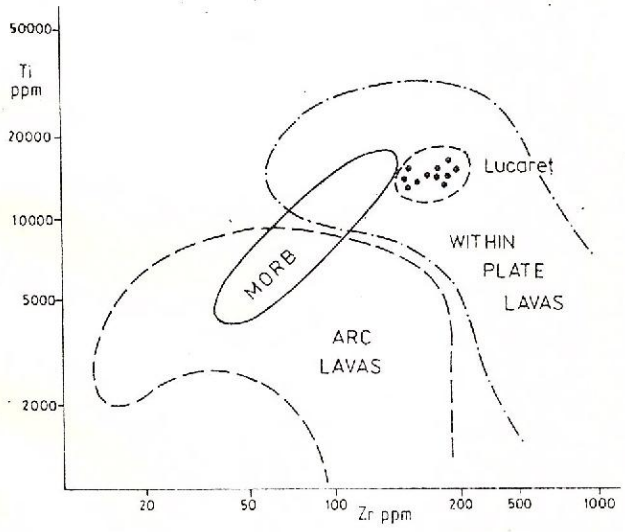

Fig. 5 - Plot of the basalts on the Ti - Zr diagrams. MORB, field of the median oceanic ridge basalts.

In order to confirm this tectonic setting, Table 3 presents – by comparison – the contents of some major and trace elements from the plateau basalts, withinplate-type rocks from the different continents and from the Lucareţ olivine basalts. The higher average contents of Ti (>10.000 ppm) occur in all the basaltic series. Higher value are also observed in case of Zr, Y and Ca/Al_2O_3 ratio. The limits of the values of the Ca/Al_2O_3 ratio vary within the limits established by Werner (1984) for the continental withinplate basalts. The average content of Sr is quite high (1019 ppm) in the Lucareţ olivine basalts, as against the other basaltic series. An average value of 859 ppm Sr occurs in the olivine basalts of Nuanetsi (Africa).

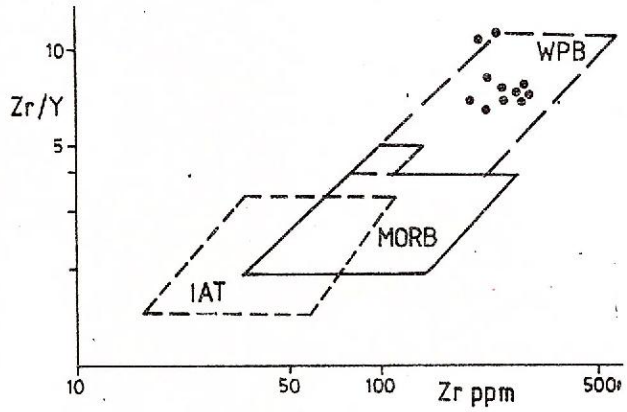

Fig. 6 - Plot of the basalts on the Zr/Y - Zr diagram. WPB, withinplate basalt field; MORB, field of the median oceanic ridge basalts; IAT, island arc tholeiitic basalt field.

All the study rocks (Table 1) present very high contents of Ba (average value 598 ppm), which also point to the continental withinplate basaltic character of the Lucareţ rocks, a character which is also emphasized by the high contents of Sr. These contents also indicate that the slightly alkaline parental magma was formed in the mantle, at a very great depth.

The high contents of Nb (average value 56 ppm) in all the Lucareţ basalts point out the character of continental withinplate alkali basalts of these rocks. The relations between the Nb/Y and Zr/P_2O_5 ratios plot the rocks in the field of alkali basalts on the diagram in Figure 7 (Floyd, Winchester, 1975).

The high and constant contents of the Cu and Zn in all the basalts (Table 1), with average values of 26 ppm Cu and 82 ppm Zn, also point to the parental magma formation in the mantle.

The U contents in the Lucareţ basalts are more than twice higher than those from the ocean floor basalts, and the Th contents (Table 1) are almost four times higher than those from the same basalts. The contents of U and Th are close to those from the medium-K volcanic series from the island arcs. The average value of Th/U ratio – 2.5 – is also higher than that from the ocean floor basalts and close to the value from the medium-K volcanic series from the island arcs.

Institutul Geologic al României

Table 3
Average contents of some chemical elements in basic rocks of continental withinplate type

No	Location	Number of analyses	Ti ppm	Zr ppm	Y ppm	Sr ppm	CaO/ Al_2O_3	Authors
1	Continental basalts	35	15150	215	29	460	–	Pearce and Cann(1973)
2	Decan traps-India	9	11850	132	28	187	–	Pearce and Cann(1973)
3	Nuanetsi basalts – Africa	14	15600	192	–	614	–	Cox et al. (1967)
4	Nuanetsi olivine basalts – Africa	47	16200	300	–	859	–	Jamieson (1969, in Cox, 1972)
5	Basalts from Tuli syncline – Kenya	12	16800	328	–	683	–	Pearce and Cann (1973)
6	Paraná basalts-Brasil	23	13500	–	–	403	–	Mantovani et al. (1985)
7	Basalts and dolerites from North Dobrogea	42	14400	254	41	341	0.51	Savu (1988)
8	Lucareţ olivine basalts	11	13910	242	21	1019	0.60	In this paper

Table 4
Distribution of REE, Hf and Ta in the basaltic rocks

No	Sample	La	Ce	Sm	Eu	Tb	Yb	Hf	Ta
1	(1)	30	68	4.7	1.50	0.87	1.4	4.7	1.2
2	(2)	25	37	4.8	1.01	0.90	1.3	3.7	4.0
3	(5)	26	29	5.8	1.25	0.60	1.4	3.4	1.5
4	(7)	27	30	4.8	1.20	0.64	1.4	5.5	0.93
5	(9)	26	39	5.1	0.97	0.70	1.3	5.2	2.6
Average		26.80	40.60	5.04	1.19	0.74	1.36	4.50	2.01
ΣREE = 77.71; LREE/HREE = 21.89: Eu/Sm = 0.24; (La/Yb)N = 13.33									

The numbers in brackets indicate the sample numbers in Table 1.

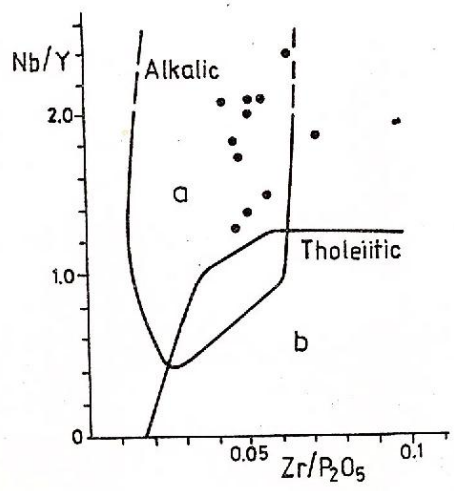

Fig. 7 - Plot of the basalts on the Nb/Y - Zr/P_2O_5 diagram. a) alkali basalt field; b) tholeiitic basalt field.

REE, Hf and Ta distribution in basaltic rocks

Table 4 presents the contents of REE, Hf and Ta from the Lucareţ basalts as well as their average values and those of some characteristic ratios. The values are typical of the group of alkaline basic rocks, separated by Cullers and Graf (1984) into three subgroups. The Lucareţ basalts present values closer to those characteristic of the "transitional" alkali basalt subgroup. Thus, the total REE = 77.71 ppm is close to the lower limit of 90 ppm of the variation of this value in the alkaline basic rocks. Due to the high-LREE content of these rocks, the average value of the LREE/HREE ratio (22) is higher than in other basaltic types. Likewise, the average value of the Eu/Sm ratio (0.24) is identical with the value of the lower variation limit of this ratio in the alkali basaltic rocks. The value of the (La/Yb)N ratio (13.33) is fairly close to the upper limit (12) of

Institutul Geologic al României

the variation of this ratio established by Coish et al. (1986) for withinplate basic rocks.

According to their outlines, the patterns of the Lucareţ basalts (Fig. 8) are similar with those presented by Werner (1984) for the continental plateau basalts from Columbia and Keweenawan. They are also similar with the patterns of the alkali basalts from Iceland and Azore Islands presented by Saunders (1984) for the basalts from the anomalous ridge segments, considered, as a matter of fact, products of the oceanic withinplate volcanism (OWPB). These patterns point to a continuous decrease of the REE contents from the light ones (LREE) to the heavy ones (HREE), as already shown. Only two out of the five REE analyses indicate a quite low negative Eu anomaly, which shows that the fractional crystallization, the differentiation of the parental alkali basaltic magma, respectively, was almost inexistent.

On the basis of the value of the Ce/Yb ratio - chondrite normalized (Nakamura, 1974) - and of the (Ce)N contents, the Lucareţ basalts plot on the diagram in Figure 9 (Saunders, 1984) in the alkali basalts field from the Hawaii, Azore, Reunion and Iceland, along the curve A. It represents the equilibrium batch partial melting of a garnet lherzolite consisting of 55 per cent olivine, 25 per cent orthopyroxene, 15 per cent clinopyroxene and 5 per cent garnet. The position of the rocks on the diagram shows that for the formation of the parental alkali basaltic magma the peridotitic source from the mantle was melted about 0.5-5 per cent, as well as that of the alkali basaltic magma from the mentioned oceanic islands. Kay and Gast (1973) pointed out that, theoretically, for the formation of the alkali basaltic magma about 0.8-2.9 per cent of the peridotitic source has to be melted, a field in which plots also the average value of the Lucareţ basalts on the diagram in Figure 9.

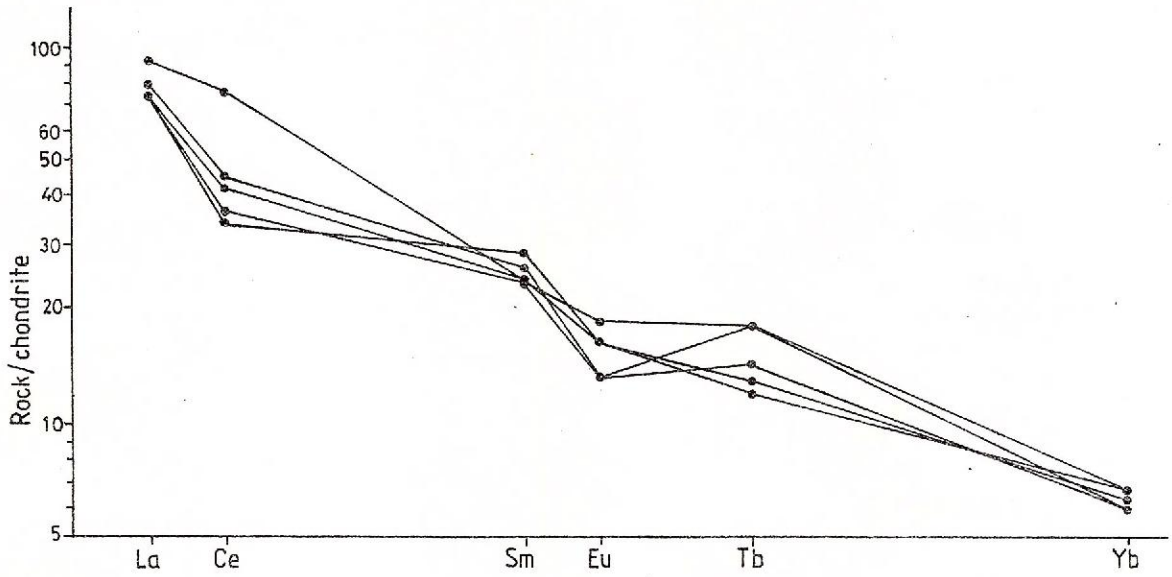

Fig. 8 - Chondrite normalized REE patterns of the basaltic rocks.

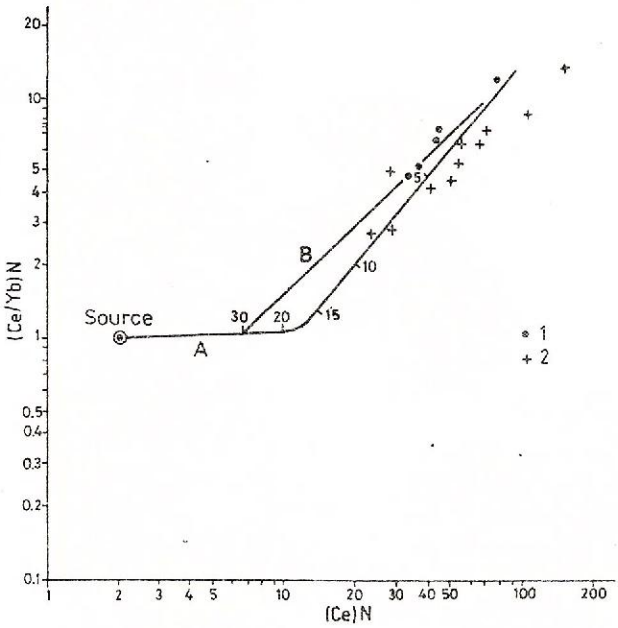

Fig. 9 - Plot of the basalts on the (Ce/Yb)N - (Ce)N diagram. A) equilibrium batch partial melting of a garnet lherzolite source; B) fractional crystallization line of an eclogite within an open system; the numbers of curve A indicate the percentage of the partial melting: 1, Lucareţ basalts; 2, alkali basalts or olivine alkali basalts of ocean withinplate type from Hawaii, Azore, Reunion and Iceland.

Institutul Geologic al României

Conclusions

The Lucareţ plateau consists of lava flows and more rarely of basaltic pyroclastics. According to several forceresearchers, the N-S trending fracture system played a major part in the starting of the Quaternary volcanic activity in the Banat, but the eruption of the Lucareţ basalts occurred on the Recaş-Cuieş crustal fracture plane, with a WSW-ENE trending, approximately in the zone of intersection with fractures from the previous system.

Taking into account that the rocks do not form a volcanic cone and that no crater was found in the eruption zone although the volcanics are quite young, it results that the volcanic activity did not developed by central-type eruptions, as presumed, but by large flows of quite fluid basaltic lava, typical of the linear eruptions.

The volcanism developed during a distension (decompression) period in the Alpine areas at the beginning of the Pleistocene, as a reaction to the compression phases which accompanied the last Neogene tectonic movements. The two above-mentioned fracture systems were generated or reactivated.

The distension process was accompanied by a thermo-tectonic phenomena which yielded in the mantle a slightly alkaline basaltic magma by the melting (0.5-5 %) of the garnet peridotitic substratum, thus giving rise to olivine alkali basalts found in the Banat as well as in Transylvania (Detunata, Racoş), the Pannonian Basin, the Moesian Platform etc.

The basaltic rocks display obvious continental withinplate basalts (CWPB) characteristics, as shown by the contents of major and trace elements. The REE contents and the ratios between them point to transitional alkali basalts of continental withinplate type.

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Institutul Geologic al României

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LITHOSTRATIGRAPHIC AND METAMORPHIC CORRELATIONS: A TENTATIVE WAY OF EXPLORING THE EARLY HISTORY OF THE GETIC CRYSTALLINE

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Key words: Metamorphic rocks. Metamorphism. Stratigraphic units. Nappes. Stratigraphic controls. Crustal blocks. Collision. Getic Nappe. South Carpathians.

Abstract: The current models regarding the Getic Crystalline are based on the idea of a single metamorphic event that affected an undisturbed depositional sequence. A great deal of more recent information claims for a polymetamorphic history and even for cryptic discontinuities within the pile. In the Western outcropping area, a well-defined sequence exists that can be successfully extrapolated to all other zones. Each and every lithostratigraphic unit contains, besides a characteristic lithology, two superposed mineral assemblages surprisingly constant within the limits of one unit. The succession consists of three entities, from bottom to top: the Armeniş Formation (AF), the Vaideeni Complex (VC), and the Negovanu Mare Formation (NMF). The limits between the lithostratigraphic entities are conformable to the prevailing S_2 foliation and banding, and bear no special sign of dynamic action, though metamorphic conditions and lithology change on either side. This advocates for a succession of Precambrian metamorphic nappes, namely from top to bottom: the Semenic Nappe (SN), built of the NMF, the Voineasa Nappe (VN), containing the VC, and the Armeniş Unit, formed by the AF rocks. The pile is "intruded" by the Ursu Formation (UF). A presumable mechanism is invoked, according to which the units belonged to three different blocks that came into collision, making possible the thrust of the SN over the AU, squeezing between them eclogite-bearing rocks belonging to the VN. The proposed piezothermic arrays are based on mineral assemblages and reasonable estimates of heat flow and conductivity. Subsequent emplacement of the UF is explained bz diapiric raise, initiated by lithospheric convection.

1. The state of the art vs. the outline of the issue

The large extension, the accidented morphology of the outcropping areas, the difficulties related to highlighting the lithological entities and interpreting their mutual relationships were major hindrances in making up an integrated image of the Getic Crystalline.

Defined as the Lotru Series (Ghika-Budeşti, 1932), the Getic Crystalline failed to be the subject of a lithostratigraphical horizontation until the sixties, except for the separations operated by Pavelescu (1959), that the author himself decided to abandon afterwards.

During the period 1970–1975, two basic studies, at the same time spatialiy extended and detailed, were carried out, that tackled the lithostratigraphy and metamorphism of the Sebeş Lotru Series in two regions of the western South Carpathians. The first one dealt with the Semenic Mts (Savu, Micu, 1964), and it was subsequently improved and detailed by Savu (1966, 1970), so that in its final version it stated five lithostratigraphic entities (complexes). At the same time the author correlated the sequences in this zone with those constructed for the Lotru Mts (Savu, 1968; Savu, Schuster, 1977), that in their turn were progressively subject to some alterations.

The second detailed scheme was drafted by Bercia (1975) for the Godeanu Outlier. The author distinguished three rock complexes made up of five horizons, which fail to directly correlate with those in the Semenic Mts.

Out of the two detailed studies, however, a general lithostratigraphic succession results, consisting of biotite-gneisses (paragneisses) at the bottom, followed by quartz-feldspar gneisses and amphibolites, and micaschists and gneisses (paragneisses) at the top. The terms of this sequence were considered as a single

Institutul Geologic al României

premetamorphic pile in depositional continuity, affected by a sole major metamorphic event characterized by a Barrovian-type zonality. Metamorphic zones are claimed to either intersect (Savu, 1970) or parallel (Bercia, 1975) lithologic boundaries. In the Godeanu Mts, Bercia (1975) also identified a Pyrineean-type zonality, grading into Barrovian metamorphics by a transition zone crossed by isograde that pass undeflected from one baric type to the other.

Mainly because marker beds are not highlighted, the limits of the lithostratigraphic entities are not precisely defined, this affecting the quality of the lithostratigraphic separations and the correlation possibilities among different zones.

Knowledge related to the Lotru Series obviously increased with progress in the completion of the national geological map, scale 1:50 000; nevertheless, the storage of information did not lead to an essential clearing up in the lithostratigraphic correlations. This was mainly due to heterogeneous approaches based on different outlooks, the absence of a coherent vision and the proliferation of local names, hard to inventorize, and especially to correlate.

The need to systematize the existing data led to the elaboration of some attempts dealing with the correlation of the whole Getic Crystalline (Savu, 1978; Kräutner, 1980), studies which are mostly affected by the approximation degree of the available materials. Thus, it was assumed that the pile had resulted from a single depositional cycle and, therefore, it was named the Sebeş-Lotru Group (Kräutner, 1980), accepting that initial superposition and continuity relationships remained unaltered during the metamorphic evolution. The lack of marker at the boundaries between the lithostratigraphic units implied correlations on successions of units with ill-defined limits, resulting a vacillating and sometimes a contradictory character for the boundaries. A sequence of five formations was proposed, out of which the more evident marker unit (SL^{1a}) is an amphibolite and quartz-feldspar-gneiss formation, interposed between two piles of micaschists and gneisses.

The recognition of polymetamorphism on mineralogical-textural (Hârtopanu, 1978) and structural (Iancu, Hârtopanu, 1979; Iancu, 1983) grounds induced a more thorough investigation of the metamorphic parageneses, rocks bearing contrasting assemblages in contact pointing out concealed tectonic relationships. After a review of eclogites and ultramafics occurrences in the Getic Crystalline, Iancu et al. (1987) assert their emplacement along a Proterozoic cryptic paleosuture. This statement implies a tectonic empilement of the present sequence.

Old discontinuities and masked tectonic units constituted the topic of recent investigations throughout the outcropping area of the Getic Crystalline (Iancu et al., 1989, 1990). On the basis of lithological contrasts and inferred blastomylonite levels, a succession of Precambrian metamorphic nappes was deduced.

However, a stronger support should be needed to motivate the existence of such discontinuities and to identify them. In this respect the study of the metamorphic history within the pile may be a valuable tool in tracing out the contrasts. Some facts previously recorded offer a very promising starting point: Bercia (1973) noticed a quasi-coincidence of the lithostratigraphic boundaries with the isograde, while Balintoni (1986) perceived a fair correlation between each superposed event and its corresponding mineral assemblages. These features, oposed to Savu's scheme (1970), fit in with a great deal of situations where metamorphic assemblages are extremely constant over large areas and display a strong "lithostratigraphic" control.

The aim of this paper is an attempt of correlations at the scale of the Getic Crystalline, a tug into account both the lithologic sequences and the metamorphic history, as recorded in mineral assemblages, in order to provide an image of the edification of the present succession. Discontinuities should be pointed out mainly on juxtaposition of rock volumes with contrasting metamorphic history, and less on lithological criteria or lineaments of shear zones supposed to represent blastomylonitic planes. The matching of the three criteria, however, may warrant reliable arguments for identifying and tracing old deep-seated tectonic contacts.

2. Lithostratigraphic sequence

The starting point of the first lithostratigraphic image was offered by the central-eastern Semenic Mts (Savu, Micu, 1964), where the lowermost outcropping term of the Lotru Series was separated, and subsequently named the "sillimanite bearing paragneisses and carbonatic rocks complex (C_1)" (Savu, 1970).

During the survey of the Armeniş sheet of the national geological map 1:50000 (Săbău in Năstăseanu et al., 1990, unpubl. report) this "complex" was recovered as the Armeniş Formation. Mapping revealed that the biotite gneisses of this formation support a thin amphibolite and quartz-feldspar gneiss level (the Piatra Scrisă amphibolite). This is overlain by porphyroblastic almandine, kyanite and staurolite bearing mica gneisses and schists (Pl. I), named the Valea Mare Formation (Ibid.).

Despite the fact that the Armeniş Formation gneisses and the Valea Mare Formation are separated by no more than a few tens of meters, the metamorphic conditions, as indicated by the corresponding assemblages, obviously contrast. Within the Ar-

Institutul Geologic al României

meniş Formation an assemblage devoid of muscovite exists, consisting mainly of biotite, plagioclase and sillimanite. Neither muscovite was found to coexist with quartz, nor sillimanite with K-feldspar, whereas in the Valea Mare Formation the prevailing assemblage is almandine-staurolite-kyanite-rutile (a higher pressure and a lower temperature one). This state implies a tectonic contact postdating the making up of the above-mentioned assemblages. By the other hand, no conspicuous shear zone exists at the boundaries between the entities, the transition being structurally congruent. As a consequence, the boundaries must be recrystallized under medium grade conditions.

Tracing the extension of these relationships, one can notice that they also persist in the Godeanu Outlier (Pl. I) and continue eastwards. The Armeniş Formation maintains the same lithology, containing also lowermost terms in the northern Godeanu Outlier. The other two formations change their composition, so that the names under which they were defined must be abandoned, as having only a local relevance. The equivalent of the Piatra Scrisă level in the western Godeanu Outlier consists of thicker amphibolites containing intercalations of almandine and sillimanite bearing muscovite schists, grading into quartz-feldspar gneisses with increased lithological variability eastwards. The equivalent of the Valea Mare Formation appears on extended areas, having more characteristic outcrops and more perceptible features elsewhere.

In keeping with these amendments, the entities that can be traced throughout the Getic Crystalline are, from bottom to top, the Armeniş Formation (AF), the Vaideeni Complex (VC), and the Negovanu Mare Formation (NMF), the features of which will be detailed further on.

2.1. The Armeniş Formation (AF)

The lithology of the Armeniş Formation is composed of sillimanite bearing biotite gneisses, pearl-gneisses, amphibolites and amphibole-biotite-gneisses, quartz-feldspar gneisses, anthophyllite and/or cummingtonite bearing gneisses, marbles and calc-silicate rocks.

The lower boundary of the Armeniş Formation is unknown, the lowermost lithology forming a banded sequence with alternating amphibolites and amphibole-biotite gneisses, which appear in the core of an antiform in the northern Godeanu Outlier.

The uppermost level is frequently built up of sillimanite nodules bearing gneisses that contain numerous marble intercalations, bordered upon Vaideeni Complex amphibolites. The relationship can be followed along the eastern verge of the Semenic Massif, in the Poiana Ruscă Mts and on the borders of the Godeanu Outlier, except for the eastern one. The contact and its adjacent structural elements are congruent to the prevailing S_2 metamorphic foliation.

In the northern Godeanu Mts, a succession of two terms is evident, of which only the uppermost one also appears in other areas. The bottom term, the Gugu Member (GM), consists of a banded alternation of amphibolites, biotite and/or Mg-amphibole gneisses, pearl gneisses, subordinate quartz-feldspar gneisses, separated by Conovici and Săbău (1993) as the banded gneisses and amphibolite subformation. The top term, named the Micuşa Member (MM), contains chiefly sillimanite bearing biotite-gneisses and marble intercalations. The boundary between the two subformations is tightly to isoclinally folded, the tectonic competence of the Gugu Subformation favouring the accomplishment of structural patterns built up during two superposed synmetamorphic folding episodes, F_1 and F_2, that one can notice also inside the Gugu Member. Within the Micuşa Member, the prevalent structural feature is the S_2 foliation, that parallels the upper boundary of the Armeniş Formation and is slightly corrugated.

As to the thickness, it is worth stressing that several meanings of the term exist, i.e. the initial thickness, the thickness of the pile and the thickness perpendicular on the structural elements. Out of this only the last one is relevant, provided that both boundaries of the considered entity are apparent, the other significances being misleading in a polydeformed polymetamorphic sequence.

The maximum apparent thickness of the Armeniş Formation, as the bottom limit is not exposed, is ca 3000 m in the Pârâul Mâţului-Gugu summit-Moraru summit zone (Godeanu Mts).

The outcropping areas of the Armeniş Formation range between the eastern margin of the Semenic Mts and the Cimpa basin in the Parâng Mts. In the Semenic Mts, a narrow strip is exposed along the Timiş Valley between the localities of Teregova and Sadova Veche (C_1 complex, Savu, 1970). In the Poiana Ruscă Mts, it partly correlates with the Băuţar and Bucova formations (Maier, Lupu, 1979), whereas in the Godeanu Mts it borders the outlier, except for the eastern part, between the Iauna and Naiba valleys, being coincident to most of the G_{11} Complex (Bercia, 1975). The easternmost outcrops exist in the Parâng Mts in two tectonic windows along the Cimpa Valley, that exposes an alpine rabotage outlier from under the bottom of the Getic Nappe. The Armeniş Formation is coincident with SL^{gn_a} separated by Kräutner (1980) on the grounds of correlating the existing information, except for some erroneous assignments. It is the case of the bottom of the succession drafted by Savu (1974), included here, although the mentioned author did not assign it to C_1, the equivalent of the Armeniş Forma-

Institutul Geologic al României
156529

tion; also of the sequences described by Hann, Gheucă (1978), Savu, Schuster (1977), Pavelescu (1970) and Focşa, Hurduzeu (1967). All these belong to higher levels in the pile.

Metamorphic conditions are extremely constant inside the Armeniş Formation, a feature fully shared by all the other formations of the Lotru Series, the metamorphic conditions changing abruptly at the boundaries between the formations. The ressemblance goes even further, as in each formation one can notice parageneses formed during two metamorphic events, of which those belonging to the first event are relict.

The M_1 event is defined in the Armeniş Formation by the almandine-rutile-kyanite-staurolite assemblage. The corresponding metamorphic conditions are estimated by Conovici, Săbău (1993) to 550–650^0 C and over 9.7 Kbar. The M_2 event, resulting in an almost complete reorganization of the rocks, led to a particular assemblage, devoided of muscovite, but with no coexisting sillimanite-muscovite pair, and the biotite-sillimanite-quartz assemblage, rather than the cordierite-K feldspar one. The presumable metamorphic conditions are probably close to 714^0 and 6Kbar (Ibid), the invariant point involving the thermal breakdown of muscovite with quartz and the anatexis curve; it would make possible the visible assemblages and imply strong silica and alkali depletion, leading to a strongly Mg-rich composition, fairly conspicuous in the Armeniş Formation.

2.2. The Vaideeni Complex (VC)

The name of the Vaideeni Complex is first mentioned as such in Balintoni et al. (1989), modified from Săbău et al. (1987). Its lithology consists of a sequence of quartz-feldspar gneisses, amphibole gneisses, mica-gneisses and schists, containing ultramafics, eclogites anatectic granitoids, quartzites and microblastic biotite gneisses. Eclogites and ultramafics exhibit high-grade assemblages, that underwent retrograde alterations during the medium-grade events affecting the pile. Iancu et al. (1987, 1989) claim for the existence of two high-grade superposed assemblages, an "eclogitic" and a "granulitic" one, but do not provide satisfactory arguments regarding the latter. The above-mentioned authors, however, point out a "Precambrian cryptic suture" outlined by these exotic bodies.

The microblastic gneisses, initially separated by Conovici (oral presentation, 1980) as the Jidoştiţa Series, were assumed by Iancu, Hârtopanu (1982) to represent the rocks of a Precambrian tectonic unit subjacent to the Lotru Series. Nevertheless, the fact that these gneisses are confined only to the Vaideeni Complex, where they appear as intercalations in mapping (southern Sebeş Mts) and in drill cores (Porţile de Fier Outlier), and the lack of any metamorphic contrast between them and the Vaideeni Complex mica gneisses, rather support the idea that they represent an intraformational tectonic facies, accomplished by rising the density of mineral nuclei along intense shear surfaces paralleling S_2.

The bottom boundary of the Vaideeni Complex is well exposed in the eastern Semenic Massif and in the Godeanu Outlier, where marble intercalations bearing biotite gneisses belonging to the Armeniş Formation support the Vaideeni Complex amphibolites, including mainly muscovite mica-gneisses and schists intercalations, that occasionally contain almandine and sillimanite.

The upper limit of the Vaideeni Complex brings into contact its different lithologies with the Negovanu Mare Formation micaschists. The boundary parallels the S_2 schistosity and the array of adjacent intercalations. The most frequent Vaideeni Complex lithologies adjacent to the contact are quartz feldspar gneisses and amphibolites. In case the Vaideeni Complex ends with feldspar-rich gneisses, a Mn-rich level appears (the Delineşti level), making an excellent marker horizon consisting of spessartine rocks, Mn-bearing quartzites and schists (Delineşti-Semenic Mts; Dâlmele, Dealul Negru-Lotru Mts; Sadu Valley; Ursani-Căpăţânii Mts). No obvious superposition relationships can be reported inside the Vaideeni Complex; furthermore, the lithons geometry and the presence of exotic inclusions strongly support its construing as a mélange.

The thickness of the Vaideeni Complex is highly variable, ranging from a few tens of meters (E Semenic Mts) to kilometers, in the Sebeş, Lotru and Căpăţâna Mts. Anyhow, in this latter case no more than the magnitude order can be inferred, since only the upper limit is exposed.

The prevalent structural features involves variable-sized tight folds, sometimes extending over kilometers (Dinică , MS 1990), which grade into tabular aspects towards the boundaries, paralleling them.

The Vaideeni Complex extends over most of the outcropping area of the Getic Crystalline (Pl. II), approximately correlating with the SL^{gn} formation defined for the western massifs by Kräutner (1980). The tentative separation, due to the same author, of three formations inside this level (SL^{gni}, SL^{la}, SL^{gns}) throughout the Getic Crystalline west of the Olt River is obviously affected by a principle error, since it claims for a higher detaliation in areas where the correlation is accepted to be more doubtful.

In the Godeanu Outlier, the Vaideeni Complex includes a part of the amphibolites formerly assigned to the G_{11} level, G_{12} level and G_2 complex (Bercia, 1975), in the Căpăţana Mts the entire sequence described

Institutul Geologic al României

by Hann, Gheucă (1980), the Voislova Formation in the Poiana Ruscă Mts (Maier, Lupu, 1979), whereas in the Lotru Mts it contains the paragneiss complex (Savu, Schuster, 1977), equivalent to the Steaja Gneiss (Arion, 1988, unpubl. report), as well as the Vidra and Frumoasa gneisses (Ibid), equivalent to the amphibolites and quartz feldspar gneisses formation (Savu, Schuster, 1977).

In the Semenic Mts, Savu (1970) defined the orthoamphibolites and manganese-bearing rocks complex (C_5), allegedly located in a synform which preserved the uppermost part of the Lotru pile. However, the rock assemblage is definitely typical of the Vaideeni Complex: amphibolites, quartz-feldspar and mica-gneisses that support spessartine rocks at the transition toward kyanite-bearing micaschists, that obviously appear in a higher position (e.g. Ogaşu Urât, Delineşti). Moreover, eclogites are found in mica-gneisses (Bârzava Valley) and frequent Jidoştiţa-type microblastic gneisses appear. The structural array is given by isoclinal folds grouped in an open anticlinorium gently plunging to SW. The flanks of this anticlinorium are very slightly deviating from the horizontal. The Vaideeni Complex extends eastwards in the N Semenic Mts, containing quartz-feldspar gneisses exposed beyond the locality of Rugi.

The metamorphism in the Vaideeni Complex is characterized by the almandine-muscovite-staurolite-kyanite-rutile assemblage formed during M_1. Its preservation degree is generally low and inhomogeneous. The conditions are identical with those of the M_1 in the Armeniş Formation, containing the same assemblage. A very pervasive M_2 superposed event resulted in the sillimanite-almandine-muscovite ± staurolite assemblage. No systematic sampling of the parageneses is available, but the numerous investigated points indicate constant conditions. These are estimated at 5–6 Kbar around 680^{0}C (Conovici, Săbău, 1993) for the Vaideeni Complex rocks (not defined as such in the mentioned paper.

2.3. The Negovanu Mare Formation (NMF)

The Negovanu Mare Formation makes up the uppermost lithostratigraphic term of the Lotru Series. Its name comes from the homonymous peak in the Lotru Mts. The lithology is constantly rich in mica bearing rocks, garnet, kyanite, staurolite and rutile bearing micaschists being extensive and typical.

The bottom limit is discontinuously marked by the spessartine rocks of the Delineşti level, frequent pegmatite lenses, or the lithologic contact towards the Vaideeni Complex amphibolites. A peculiar aspect is recorded in the Lotru Mts, and at a less extent in the Sebeş Mts, where the Negovanu Mare Formation micaschists overlie mica-gneisses of the Vaideeni Complex. Here, the boundary is materialized by a thin discontinuous quartzite level with regional extension (Arion, oral, com., 1988).

The Negovanu Mare Formation begins with micaschists containing large almandine porphyroblasts, also frequently kyanite and/or staurolite, then follow micaschists lacking porphyroblastic minerals that contain (Li-) pegmatites or phlebitic muscovite gneisses. The pelitic character of this entity continues upward with a remarkable constancy to the highermost levels exposed.

Structurally, the Negovanu Mare Formation appears as a horizontal tabular level, generally keeping up to the highermost ranges (Hârtopanu, unpublished doctoral thesis, 1982), as most of the authors agree. The internal structural boundaries parallel the inferior limit. The maximum preserved thickness of the lithologies ascribed to the Negovanu Mare Formation was stated by Savu (1970) to be almost 20 000 m, that looks highly overestimated and obviously at variance with the recorded mineral assemblages. The thickness invoked by Bercia (1975) seems more realistic (1500 m); however, in many areas this figure cannot be documented to exceed some 100–200 m.

At any rate, this thickness gives no idea of the initial one, as it merely represents a part of the Negovanu Mare Formation that escaped erosion.

The Negovanu Mare Formation correlates with the G_3 complex of Bercia (1975) and with C_2–C_4 complexes defined by Savu (1970), being equivalent with SL^{gnms}, and partly SL^{gns}, as separated by Kräutner (1980). Difficulties in correlation mainly stem from the poorly argued separation of the SL^{gns} that contains in some cases Negovanu Formation micaschists, in others Vaideeni Complex ones, or even both.

The Negovanu Mare Formation metamorphism bears the imprint of two superposed parageneses constant throughout the formation. The M_1 event took place under the same conditions as in the other two formations, at ca 10 Kbar and 550–650^0 C. The kyanite-staurolite-almandine-rutile-muscovite assemblage is notably well preserved, mainly due to the M_2 features. This event is illustrated by a similar assemblage, formed, however, at lower pressure, as ilmenite neoformation is recorded and also rutile is partly altered into the same mineral.

The uppermost part of the Negovanu Mare Formation may contain in places neoformation sillimanite, that may indicate a temperature rise concurent to the pressure decrease (Pârvova, Smogotin brook, Conţu, Frăsincea).

Institutul Geologic al României

3. The status of the Ursu Formation

The Ursu Formation (UF) represents a domain of the Getic Crystalline having a peculiar history, separated on grounds of lithology, tectonic setting and metamorphic evolution in the Căpăţâna Mts and Petrimanu Zone (Săbău et al., 1987; Hann et al., 1988), and also in the Godeanu Outlier (Conovici, Săbău, 1993).

Its lithology is mainly composed of ortoclase-biotite-sillimanite- cordierite gneisses. Amphibolites, quartz-feldspar gneisses and unusual lithologic types (leptynites, khondalites, kinzigites) are also frequent.

Its boundaries toward the Lotru Series are hardly manifest, as they are welded under low-pressure metamorphism conditions (450^0 C, 2 Kbar in the Căpăţâna Mts; 600^0 C, 3.5 kbar in the N Godeanu Mts). The exposures of the Ursu Formation generally have the aspect of diapiric domes and plumes that impose a thermal halo upon the host rocks, represented by all three divisions of the Lotru. Series (Armeniş Formation, Vaideeni Complex, Negovanu Mare Formation).

The structural style corresponds to a plastic flow, the gneisses having blastomylonitic features and being involved in frequently convolute folding. No lithologic sequence is seizable.

Apart from the already mentioned areas, the Ursu Formation appears in the left side of the Mănăilcasa brook (Berza et al., 1986, unpubl. report), previously ascribed to the Danubian Realm, in the Godeanu and Bahna outliers, where it has been included in the Lotru Series despite the metamorphism history contrast (Bercia, 1975; Hărtopanu, 1975).

The metamorphic evolution of the Ursu Formation accomplished through successive mineralogical adjustments to the conditions imposed during the diapir ascension, under an intense steady-state strain. Diachronic assemblages join and follow each other on a retrograde path from the former khondalitic paragenesis (garnet±cordierite+silimanite+feldspar) to the muscovite+andalusite+biotite±staurolite assemblage (Săbău et al., 1987). The cessation of the diapiric emplacement is recorded by the same late-stage assemblage both in the thermal halo and (subordinately) inside the domes, corresponding to ca. 450^0 C at 2 kbar (Ibid).

4. The construction of the Precambrian structure

The existence of exotic blocks (inside the Vaideeni Complex) and metamorphic contrasts within the pile clearly prove its present tectogenous character. The heterogeneous metamorphic history of the four component entities claim for a tectonic stacking of formations or complexes with separate previous evolution. The Armeniş Formation rocks constitute the lowermost Armeniş Unit (AU), which supports the Voineasa Nappe (VN), made up of the Vaideeni Complex rocks. The top of the sequence is represented by the Semenic Nappe (SN) formed of the Negovanu Mare Formation rocks. As the three units evolved separately prior to the thrust, they must have belonged to different blocks, namely the Armeniş (AB), Voineasa (VB) and Semenic (SB) blocks. The Ursu Formation diapirically pierces the pile resulted from the superposition of the mentioned units. As the boundaries between the entities are welded by medium-grade metamorphic assemblages, the stecking of the pile apparently took place in deep-seated conditions. A realistic model of the achievement of the present structure must comply with the constraints resulting from the metamorphic features mentioned above. The facts that need a coherent scenario can summarized as follows:

– eclogite emplacement within the Vaideeni Complex;

– making up of M_1 assemblages in different rock volumes under identical conditions;

– overprinting of M_2 assemblages, as to be generally prevalent and to differ from the M_1 ones in the Vaideeni Complex and Armeniş Formation, whereas in the Negovanu Mare Formation they should by and large keep constant with them.

– adjoining of the three entities during M_2 along boundaries that parallel the S_2 metamorphic foliation;

– the status of preservation of the M_1 and M_2 assemblages;

– piercement of the pile by the Ursu Formation, its retrograde path and the emplacement of the thermal aureole.

For observing these constraints, the entities in the pile have to move along the piezothermic arrays in Figure 1, where the segments determined by recorded mineral assemblages are marked in solid line and those inferred in dashed line. The arrays were traced following England, Thompson (1984), for a suitable choice of physical parameters.

Eclogite emplacement (only) within the Voineasa Block may be achieved by driving them from the bottom of the crust during tectonic stacking on a subfluence zone, inferred by Hann (1983) (Fig. 2A). The depth eclogites come from was arbitrarily chosen to be 70 km, which seems a realistic estimate. A thermal maximum of ca 600–650° C at 10 Kbar (M_1 conditions) can be reached on the piezothermic array, for an estimated thermal conductivity of 1.5 W/m.K, if the erosion of a crust thicknened by thrusting was immediately onset under a heat flow of 45 m W/m^2, a value typical of old orogens. The erosional rate must have been high in an exceedingly thick crust resulted from doubling an already thickened crust. Assuming

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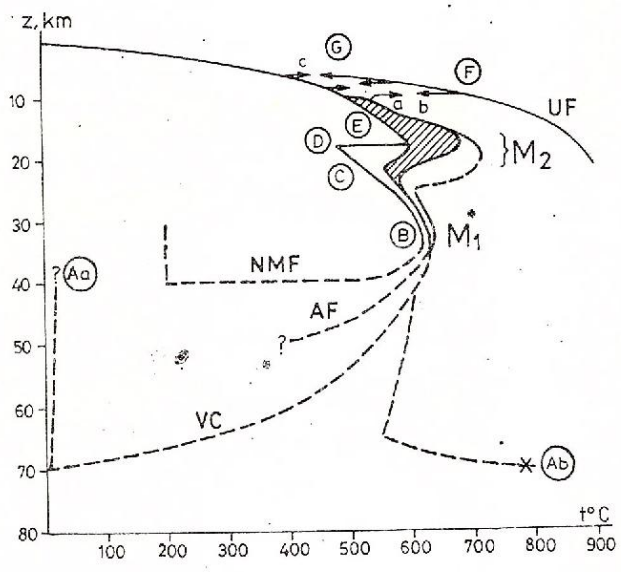

Fig. 1 – Piezothermic arrays of the Getic Crystalline entities. For explanation and abbreviations see text. a, b, c – metamorphic conditions corresponding to recorded assemblages; M_1, M_2 – peak metamorphism conditions; B – stage illustrated in Fig. 2B and similar; dashed line – possible evolution; solid line – evolution indicated by mineral assemblages; * – eclogites.

an erosional rate of 0.7 mm per year, the Aa-B and Ab-B segments describe the erosion and thermal relaxation of the Voineasa Body until the reaching of M_1 peak metamorphic conditions, followed by subsequent uplift and cooling.

The other two blocks attain the same conditions after an evolution for which information is lacking. In order to illustrate this, one of the possible scenarios can be chosen, according to which the Negovanu Mare Formation evolved as a result of doubling of a 35 km thick crust by thrusting, onset of the uplift after 20.10^6 years at an erosion rate of 0.35 mm/year. The choice for the condition was made according to England, Thompson (1984), assuming a heat flow of 45 mW/m^2, in a pile with a thermal conductivity of 1.5 W/m.K. Figure 2B illustrates the M_1 event for the three blocks. The figure may have a diachronic character, as there is no need for M_1 to take place simultaneously in the three blocks, providing identical peak conditions be achieved. After M_1, the AB evolution bears the imprint of a thermotectonic M_2 event, the condition of which could have been reached in different ways. For simplicity sake, the dashed line represents an evolution through AB crust shortening by a factor f=1.3, whereas the heat flow rises to 60 m W/m^2, an average value for the continental crust.

At the time corresponding to M_2, the three blocks came into collision (Fig. 2 C, C in Fig. 1). The upper part of the AB crust was probably chopped off along a horizontal shear zone imposed on the lowresistence zone confined to the invariant conditions mentioned above, whereas the top of the SB (SN) is overthrust, driving on its bottom intensely strained material belonging to the median block that constitutes the VN, containing eclogite relics (Fig. 2 D, point D in Fig. 1). The isotherm normalization follows (Fig. 2E, zone E in Fig. 1). This goes with a metamorphic peak M_2 in the VN, and a temperature pulse in the SN. Mineral assemblages readjust in the V, N, whereas in the SN a textural readaptation concurrent with neoformation of an assemblage containing kyanite, staurolite and possibly subordinate sillimanite (close to the bottom) are registered. The new assemblages overprint mainly on strongly sheared zones.

Squeezing the VN between two blocks in thrust relationships very likely could generate highly strained levels, expressed as Jidoştiţa type intraformational blastomylonites. They usually contain sillimanite, but kyanite is also found close to the village of Petros (S Sebeş Mts). Spessartine accumulation at the SN/VN boundary possibily resulted from mineral reactions between Mn-bearing almandine from the Negovanu Mare Formation micaschists and K-feldspar coming from the Vaideeni Complex, that produced biotite, (Mn-bearing) muscovite and spessartine.

From this stage on, the pile registered a unitary evolution, following an array fixed by the physical parameters of the AB. The onset of a convective regime in the deep zones of the pile, under compressional stress, resulted in diapir domes and plumes made up of the Ursu Formation rocks that pierce the pile. The mineral alterations that occurred illustrate conditions corresponding to points a, b, c in Figure 1, prevailing during the stages represented in Figure 2 F, G. The Ursu Formation ascension concludes in the stage 2F, which is postdated by uniform uplift and cooling. The present status of the Getic Crystalline is illustrated in Figure 2 H.

Subsequently, only events having a minor significance occur, e.g. younger (Hercynian?) slip planes marked by chloritized blastomylonite lineaments. The continuous compressive regime is also argued by the achievement of a low-amplitude Hercynian nappe system described by Iancu et al. (1989), evident in the southern Banat. The lowermost Bozovici Unit has a VC basement and a Paleozoic cover. It supports the Miniş Unit, consisting of low-grade rocks (the Miniş Series), whereas the uppermost units (Oreviţa and Nera respectively) are formed only of the Lotru Series rocks. Alpine tectonics is represented by a system of thrusts and faults (Pl. II).

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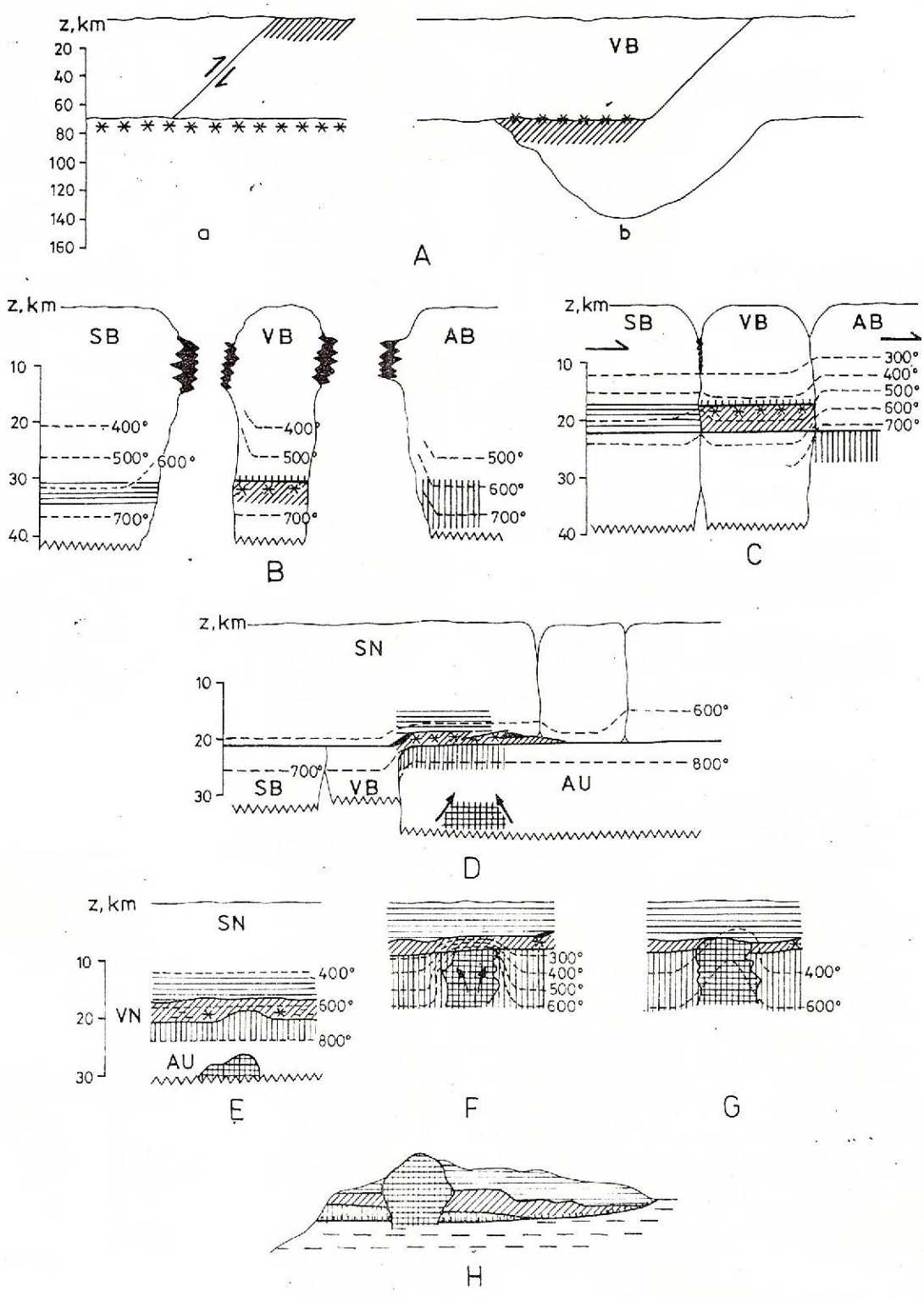

Fig. 2 – Schematic diagrams for the LS and VF evolution. Explanations in the text. Aa – subfluence onset in the VB, Ab – tectonic stacking of the VB, B – position of the blocks at the M_1 diachronic moment; C – position of the blocks at collision; D – achievement of the metamorphic nappe structure; E – thermal relaxation; F, G – diapir doming of the VF; H - present status. Arbitrarily reduced horizontal scale, for H arbitrary scale; thin dashed line – isotherms.

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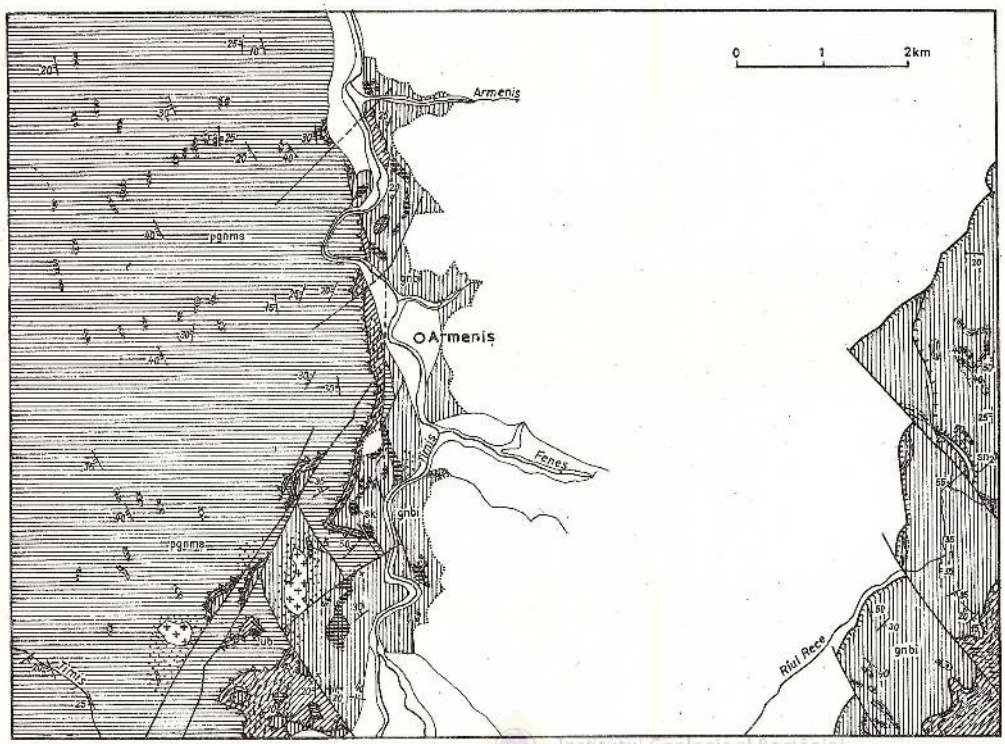
LITHOSTRATIGRAPHICAL CORRELATION BETWEEN THE EAST SEMENIC MTS AND THE WESTERN RIM OF THE GODEANU OUTLIER



LEGEND

LOTRU SERIES
SEMENIC NAPPE
Negovanu Mare Formation
- Pegmatites
- Mica-gneisses and schists

VOINEASA NAPPE
Vaideeni Complex
- Amphibolites, amphibole gneisses
- Micaschists
- Ultrabasites
- Quartz-feldspar gneisses

ARMENIȘ UNIT
Armeniș Formation
- Marbles
- Calc-silicate rocks
- Amphibolites
- Pegmatites
- Biotite gneisses

- Intrusives
- Thermal contact haloes
- Intrusion boundary
- Lithologic boundary
- Discordant boundary
- Alpine thrust
- Precambrian thrust
- Scale limit
- Fault
- S_2 Foliation

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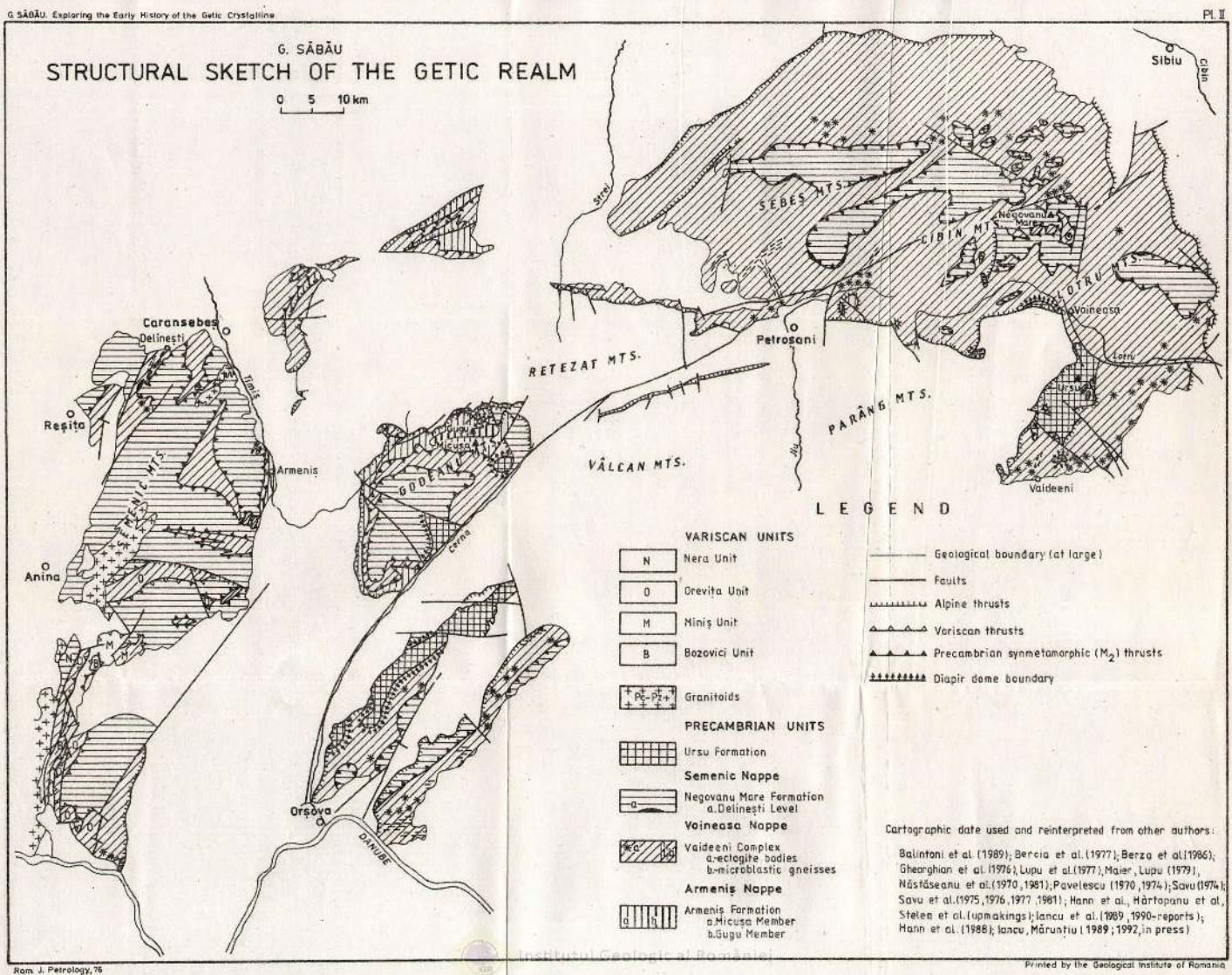
G. SĂBĂU

STRUCTURAL SKETCH OF THE GETIC REALM

0 5 10 km



LEGEND

VARISCAN UNITS

- N — Nera Unit
- O — Orevita Unit
- M — Miniş Unit
- B — Bozovici Unit
- Pc-Pz+ — Granitoids

PRECAMBRIAN UNITS

- Ursu Formation

Semenic Nappe

- Negovanu Mare Formation
  a. Delineşti Level

Voineasa Nappe

- Valdeeni Complex
  a-ectogite bodies
  b-microblastic gneisses

Armeniş Nappe

- Armeniş Formation
  a. Micusa Member
  b. Gugu Member

- Geological boundary (at large)
- Faults
- Alpine thrusts
- Variscan thrusts
- Precambrian synmetamorphic (M_2) thrusts
- Diapir deme boundary

Cartographic date used and reinterpreted from other authors:

Balintoni et al. (1989); Bercia et al. (1977); Berza et al. (1986); Gheorghian et al. (1976); Lupu et al. (1977); Maier, Lupu (1979); Năstăseanu et al. (1970, 1981); Pavelescu (1970, 1974); Savu (1974); Savu et al. (1975, 1976, 1977, 1981); Hann et al., Hârtopanu et al., Stelea et al. (upmakings); Iancu et al. (1989, 1990-reports); Hann et al. (1986); Iancu, Mărunţiu (1989; 1992, in press)

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Rom. J. Petrology, 1994, 76, p. 129–141

REACTIVATED METAMORPHIC COMPLEXES IN FOLD AND OVERTHRUST BELTS (e. g. SOUTH CARPATHIANS)

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Key words: Polymetamorphism. Stratigraphic units. Tectonic units. Nappes. Precambrian. Paleozoic. Shear Zones. South Carpathians.

Abstract: The Alpine overthrust nappe structures of the South Carpathians show a polyphasic character, presenting both spatial overlapping (repeated shearings) and spatial-temporal migration evidences. Inside the Alpine structural units, Paleozoic and pre-Paleozoic structures are preserved as inherited structures, represented by Variscan overthrust nappes and pre-Variscan crustal discontinuities. They delimit major lithostratigraphic (Paleozoic and Precambrian) units, generated through orogenic compression (folds, cleavages, foliations etc.) during the pre-Alpine tectogenetic events. The oldest and deepest structures show numerous evidences of dynamic and dynamo-thermal reactivation on various scales and under various physical conditions.

1. Introduction

The South Carpathians, southern branch of the Carpathians, represent a segment of the Alpine-Carpathian-Himalayan linear chain, characterized by overthrust structures on the Moesian Platform.

The paper presents several cartographic materials illustrating the particular features of the South Carpathians Alpine structures, as well as the complex pre-Alpine relationships and structures preserved in the pre-Mesozoic basement. The structural sinthesis is based on regional mapping data and the qualitative analysis of the micro and meso-structural elements, differentiated in time and space.

The Paleozoic and Precambrian formations from the Getic Realm involved in Alpine or older structures are briefly presented in Table.

2. Alpine Structures

The South Carpathians border south- and eastwards with the Moesian Platform, which is overthrust by Carpathian units of eastern vergence (Murgoci, 1905; Codarcea, 1940[1]. Westwards, the tertiary deposits of the Pannonian Basin overlie a basement consisting of complex alpine structures as follows: basement nappes, spatially correlatable with units of the Northern Apuseni Mountains, and obduction nappes with ophiolites and associated sediments, belonging to some units connecting the major suture (Vardar zone) with the Transylvanian nappes of the Mureş zone (Săndulescu, 1984).

The major alpine structures in the westermost part of the South Carpathians (Banat Zone) are illustrated in Figures 1 and 2.

The spatial-temporal differentiation of the alpine structural units has been made according to the following criteria:

- characteristics and ages of the Mesozoic sedimentary covers;
- structural-formational discontinuities (overthrust planes, post-nappe faults etc.);
- effects of alpine metamorphism.

Austrian (Midcretaceous) nappes can be distinguished, which have been previously grouped (Iancu, Mărunţiu, 1989) in: Danubian, Infragetic (Arjana, Severin), Getic and Supragetic units. The Laramian (intra-Senonian) tectogenesis resulted in new shearings and shiftings over great distances of the Getic and Supragetic units over the Danubian units, and together with the latter, over the Moesian Platform. Towards the Pannonian Basin, the Laramian shearings generated overthrust planes within the Southern Apuseni Mts.

[1]Data regarding the alpine overthrust nappes were synthetized by: Năstăseanu et al., 1981; Berza et al., 1983; Iancu, 1986; Iancu et al., 1990; Balintoni et al., 1989; Hann, Szasz, 1984.

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Table – Lithostratigraphic and Lithotectonic Entities in the Getic Crystalline

Age	Tectonostratigraphic units		Lithology	Mineral assemblages-Dynamo-thermal events
MIDDLE-LOWER PALEOZOIC	Buceava Fm.	Agriș volcanogenous Subformation	Metabasalts, metasiltites, matavolcanic rocks	prehnite, pumpellyite (chlorite, albite, calcite, sericite)
		Șopot terrigenous Subformation	Metaconglomerates, metasandstones, minor metabasalts, metadiorite-metaplagiogranite	
	Moniom Gr.	Cârșie Formation	Metaconglomerates, metasandstones, metapelites, metasiltites	chlorite, actinolite, epidote/clinozoisite, (albite, quartz calcite)
		Valea Satului volcano-sedimentary Formation	Metapelites, metasiltites, metacarbonatic rocks, basic metatuffs, minor metarhyolites, metadacites and metagabbroic rocks	
		Lescoviția Formation	Metasandstones, metagraywacke, metasiltites, basic metatuffs and metatufites	sericite, chlorite, quartz, albite, epidote, clinozoisite, stilpnomelane
LOWER PALEOZOIC (Cambrian-Ordovician-Silurian?)	Caraș Group	Mînuș Formation	Sericite-chlorite-graphite-quartz schists, feldspar quartzites, minor metarhyolites and basic metatuffs	chlorite, epidote, actinolite, albite, muscovite, ± biotite
		Tălva Mare Formation	Mica-feldspar quartzites, minor quartz-sericite schists	chlorite, epidote/clinozoisite, actinolite, albite, muscovite, stilpnomelane± biotite±garnet
		Dognecea terrigenous Formation	Quartz-muscovite-chlorite±graphite schists, minor greenschists and metacarbonatic rocks	
		Ralnic volcano-sedimentary Formation	Metabasites, minor muscovite-chlorite schists, quartzites and calc-schists; metagabbro-dioritic and metaultramafic rocks	
		Naidaș Formation	Metabasites, metarhyolites, metadacites, quartzites (partly feldspathic), mica-quartz schists, muscovite-chlorite schists, quartz-feldspar schists	
		Călușu Formation	Basic metatuffs and metabasalts; metarhyolites and rhyolitic metatuffs; quartz-muscovite-chlorite schists	albite, chlorite, epidote, green biotite, muscovite
UPPER PROTEROZOIC	FĂGĂRAȘ GROUP		Plagiogneisses, graphite quartzites, phibolites, carbonate rocks	M_1, M_2 - biotite, muscovite, almandine, plagioclase, sillimanite (fibrolite); hornblende; M_3 - (Variscan retrograde event): chlorite actinolite, albite, muscovite
MIDDLE PROTEROZOIC	SEBEȘ GROUP		Micaschists, plagiogneisses, amphibolites, amphibole gneisses, quartz-feldspar gneisses crystalline limestones, manganiferous rocks, pegmatites, metatectic migmatites	M_1 - garnet, staurolite, kyanite, rutile, quartz; M_2-M.P.-garnet, staurolite, kyanite, muscovite, plagioclase, biotite; L.P.-sillimanite±staurolite; sillimanite, K-feldspar, cordierite; andalusite.
	LOTRU GROUP		Massive and banded amphibolites, amphibole gneisses, metagabbros, metapyroxenites, metaperidotites (ophiolite-like complex); mica or quartz plagiogneisses, feldspar quartzites, biotite-quartz-feldspar gneisses, pegmatites, migmatites	M_1, M_2 (isofacial) sillimanite ±kyanite, garnet, K-feldspar, plagioclase
MIDDLE LOWER (?) PROTEROZOIC	CUMPĂNA GROUP		Augen gneisses, linear gneisses, massive and banded quartz-feldspar gneisses, migmatised mica- and quartz plagiogneisses, minor amphibolites	M_1, M_2 (isofacial) metamorphic events: kyanite, staurolite, garnet biotite, muscovite, K-feldspar
	LUCHIȚA GROUP		Quartz plagiogneisses, mica plagiogneisses, minor amphibolites, quartz-feldspar gneisses (generally microblastic rocks) and lens-like metamafic-ultramafic rocks	M_1 - poor relics in ultramafics (pyroxene); M_2 - hornblende, plagioclase, biotite, muscovite, garnet; M_3 (static areal over print): andalusite, sillimanite (fibrolite)
	URSU GROUP		Biotite-cordierite gneisses, amphibole gneisses, metagabbros, metaultramafites, minor quartz-feldspar gneisses, mica plagiogneisses and metacarbonatic rocks	M_1 (granulite event) garnet, Mg-cordierite, plagioclase, orthoclase, sillimanite (?); pyroxene; M_2 - sillimanite, biotite, cordierite, spinel, plagioclase, orthoclase; M_3 -Fe-cordierite, biotite, garnet, muscovite, staurolite; M_4 (static overprint) andalusite, sillimanite (fibrolite); muscovite, cordierite

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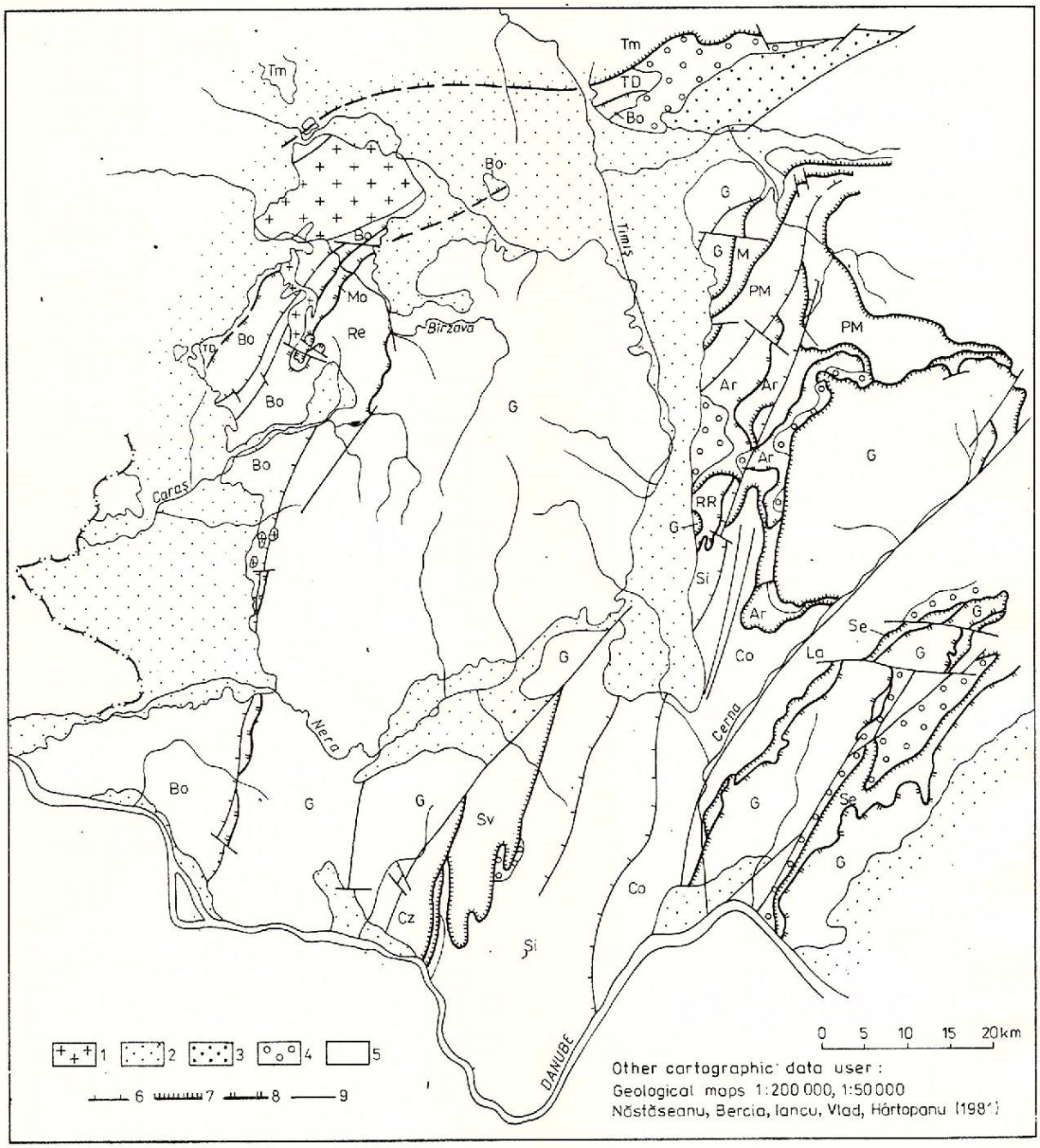

Fig. 1 - Alpine Structural units of the Banat-Mehedinţi zone. 1, Banatitic magmatites (Upper Cretaceous-Paleocene); 2, Neogene-Quaternary post-tectonic formations; 3, post-Laramian cover (Maastrichtian-Paleocene); 4, post-Austrian cover (Vraconian-Senonian); 5, pre-Austrian formations (Mesozoic pre-Austrian sedimentary cover, Variscan molasse, Paleozoic and Precambrian metamorphic rocks); 6, reverse fault; 7, Laramian overthrust plane; 8, Austrian overthrust plane; 9, fault. The tectonic units are: Tm – Timiş; TD – Tâlva Drenii, Bo – Bocşa; Mo – Moniom; Re – Reşiţa; G – Getic; Cz – Cozla; Sv – Svinecea, M – Măru; PM – Poiana Mărului; Ar – Arjana; RR – Râul Rece; Si – Sirinia; Co – Cornereva; La – Lainici; Se – Severin.

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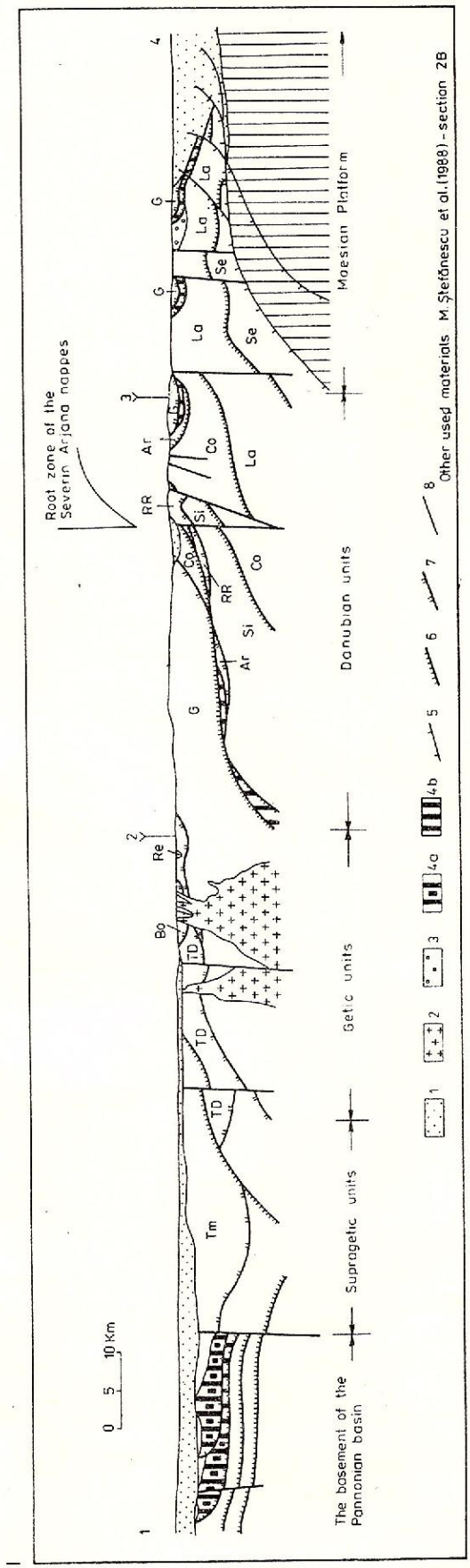

Fig.2 - Geological cross-section on the Banat-Mehedinți zone (Carpathian orogen reinterpreted after Iancu in Ștefănescu et al., 1988, unpubl. report). 1, Neogene-Quaternary sedimentary cover; 2, Banatitic magmatites (Upper Cretaceous-Paleocene); 3, Post-Austrian sedimentary cover (Vraconian-Senonian); 4, Ophiolites (a) Mureș zone; (b) Severin zone; 5, Reverse fault and Miocene thrusts; 6, Laramian overthrust plane; 7, Austrian overthrust plane; 8, Fault. The labelling of the tectonic units are the same as in Fig.1, and Sc - Schela; MoPl – Moesian Platform.

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The intra-Miocene overthrusts reactivated the pre-existing units and the post-nappe covers, as a consequence of the compressional and underthrusting impact caused by Moesian Platform, component part of the Euro-Asian craton.

The data available point to some characteristic features of the Alpine structures:

a) the shearing (basement and/or cover nappes), varying in age (Austrian and Laramian), are spatially overlapped polyphasic structures which explain the crust thickening by repeated shearings, alternatively or simultaneously with those induced through orogenic compression; the relation of banatitic magmatites with the overthrust nappes permits their characterization as subsequent magmatites with respect to the Austrian and Laramian phases, the polyphasic character of the intrusions being demonstrated based on petrology and age by Russo-Săndulescu Doina et al. (1984);

b) the alignments of the post-Austrian and post-Laramian banatites may be tectonically controlled by vertical crustal discontinuities correlatable with subduction processes (Rădulescu, Săndulescu, 1973);

c) the crust thickening by recurring shearings was accompanied by a spatial-temporal migration of the overthrusts from the median zones (median massifs) towards the margins of the orogenic zone;

d) the bilateral vergence of the structures is accompanied by a strong eastern asymmetry, the external part of the chain being strongly overthrust on the platform (Fig. 2);

e) the drastic shortening in the orogenic compression phases took place over the area of the mobile zones with oceanic crust or thinned crust. This is proved by the indirect evidence of the oceanic crust subduction, the existence of the obduction nappes with ophiolites in tectonic melange (Mărunţiu, 1987) and the tight folding, refolding and manifestation of Alpine metamorphism effects (cleavages, mineral neoformations) in cover units of Arjana-Severin Realm and Danubian Units;

f) the Alpine reactivation of the pre-Alpine basement took place especially in the shear zones, where it manifested as dynamic metamorphism of limited extension. This permitted the preservation of the pre-Alpine structures inside the Alpine basement units;

g) the crustal fractures represented either by distentional faults or by strike-slip faults (especially in the intra-Miocene ones) as well as the curved aspect of the chain are due to a great extent to the pushings exerted by Moesian Platform.

3. Paleozoic and Pre-Paleozoic Structures

Within the Alpine nappes, pre-Alpine structural units have been pointed out, generated during some Paleozoic and pre-Paleozoic deformational and metamorphic events. We further give several such examples from the Getic Realm (Fig. 3a, b). Thus, structures of this kind are represented by regional structural discontinuities of the overthrust plane type (Figs. 3, 4, 6), meso- and micro- structures generated through orogenic compression. These structures involve either Paleozoic and Precambrian formations (Fig. 5) or only Precambrian formations (Fig. 6).

In most cases, the structural discontinuities represent low angle crustal shearing effects taking place under different physical conditions, either independent of regional dynamo-thermal events (generally at upper crustal levels), or simultaneously with the latter, often at deep crustal levels (deep-seated shear zones and metamorphic nappes).

The effects of the Paleozoic deformations during some "Variscan" and/or "Caledonian" events, well preserved in the Danubian Realm (Iancu, Mărunţiu, 1989), manifest by partial reactivations of the Precambrian basement through dynamic metamorphism and regional retromorphism (diaphtoresis). There are great volumes in which the Precambrian basement is not yet reactivated , preserving the pre-Paleozoic structures.

3.1. Paleozoic Structures

The examples presented in the paper provide the morphological image of some Variscan nappes (transected by Carboniferous granitoids and covered by the Upper Carboniferous-Permian molasse) in the Bozovici zone (Fig. 4) and in the Leaota Massif (Fig. 3b). These crustal low-grade discontinuities, marked by mylonites (blastomylonites) are preserved within the same Alpine unit (Getic Nappe). They affect the retromorphosed or not Precambrian basement, as well as formations varying in age - Cambrian-Ordovician and Devonian-Lower Carboniferous (Minis, Căluşu and Buceava Formation, respectively) (Table).

The outline of the nappes and the discontinuous occurrence of the Paleozoic formations in tectonic windows, under the Precambrian metamorphics, represent the effects of the post-nappe folding at the end of the Variscan orogenesis.

Some of the discontinuities noticed in the exposed areas of the Precambrian formations may be of Paleozoic age (Fig. 3a, b), but the present state of knowledge does not allow us to make any specifications in this respect.

In western Banat, the palynologically dated Paleozoic formations crop out over relatively large areas, enabling the deciphering of their structures (Fig. 5). They belong to the Moniom Group (Upper Devonian-Lower Carboniferous) from the Moniom Alpine Nappe

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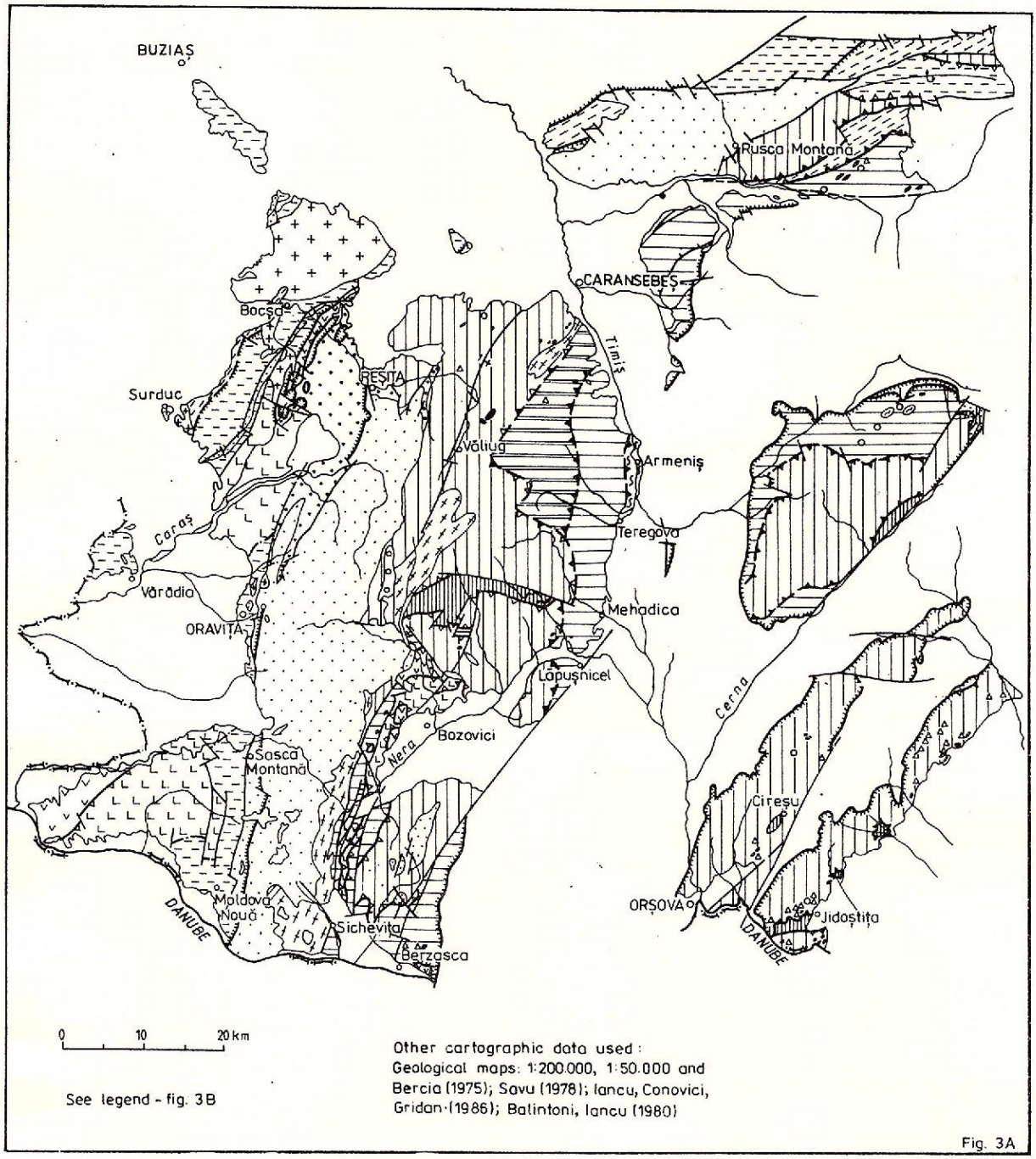

Fig. 3 Structural sketch of the Getic Realm (3A - western part; 3B - eastern part) (Iancu, Mărunţiu). 1, Cenozoic post-tectonic sedimentary deposits; 2, Banatitic magmatites (Upper Cretaceous-Paleocene); 3, Mesozoic sedimentary formations and associated magmatites; 4, Variscan molasse; Paleozoic formations; 5, Devonian-Lower Carboniferous (Buceava Fm., Moniom Group, Lescoviţa Fm.); 6, Variscan granitoid rocks; 7, Cambrian-Silurian (Miniş Fm., Caraş Group, Căluşu Fm.); Precambrian Formations; 8, Făgăraş Group; 9, Sebeş Group; 10, Lotru Group; 11, Cumpăna Group (a, Cumpăna s. str.; b, Brebu gneisses); 12, Luchiţa Group; 13, Ursu Group; 14a, peridotite; 14b, eclogite; 14c, granulite; 15, Alpine reverse fault; 16, Alpine overthrust plane; 17, Paleozoic (Variscan) overthrust plane; 18, pre-Paleozoic tectonic discontinuity (metamorphic nappe); 19, tectonic discontinuity of uncertain age; 20, green schist facies mylonitic zone; 21, chloritoid-bearing blastomylonites; 22, fault; 23, reversed antiform; 24, reversed synform.

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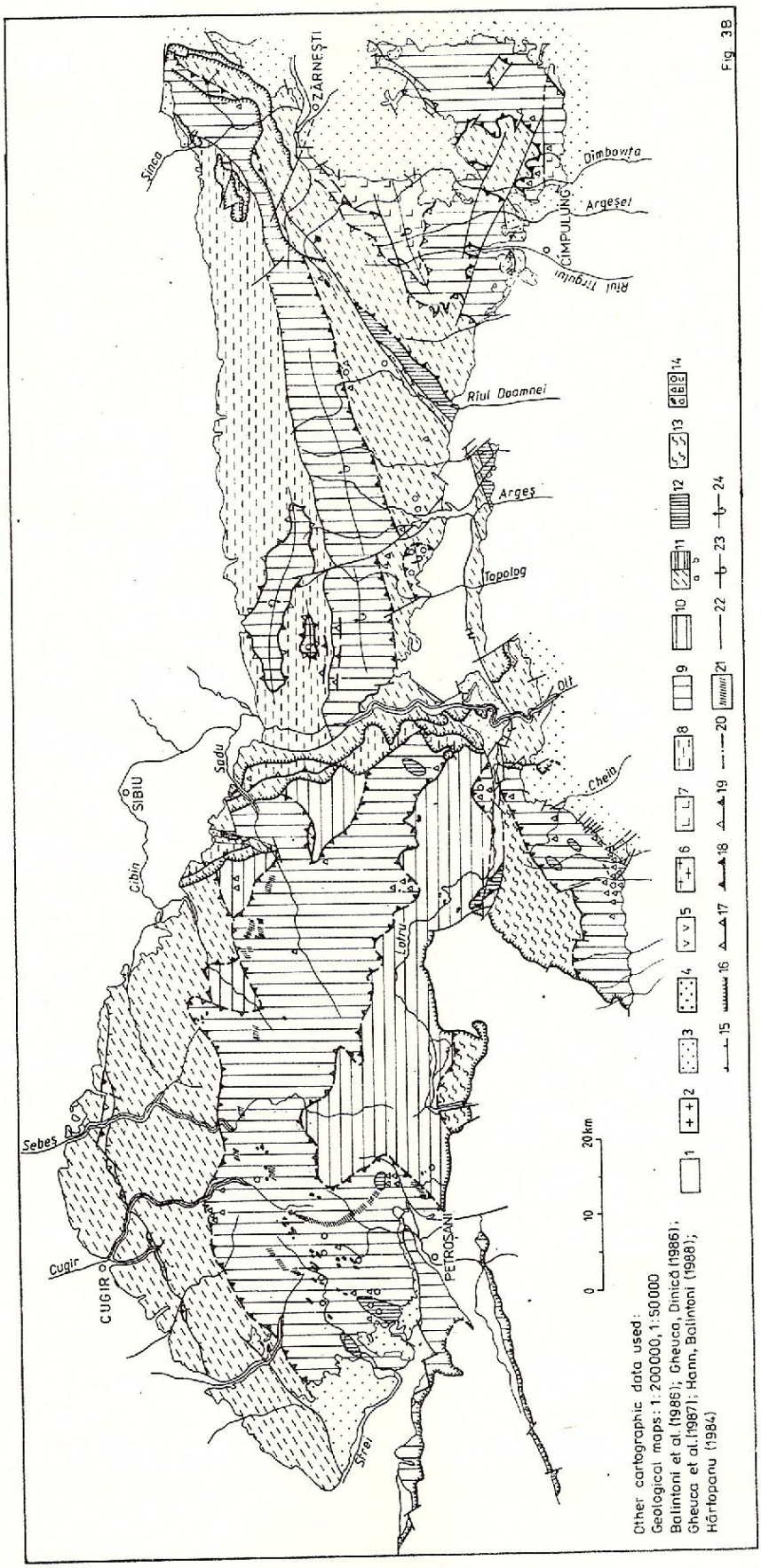

Fig. 3B

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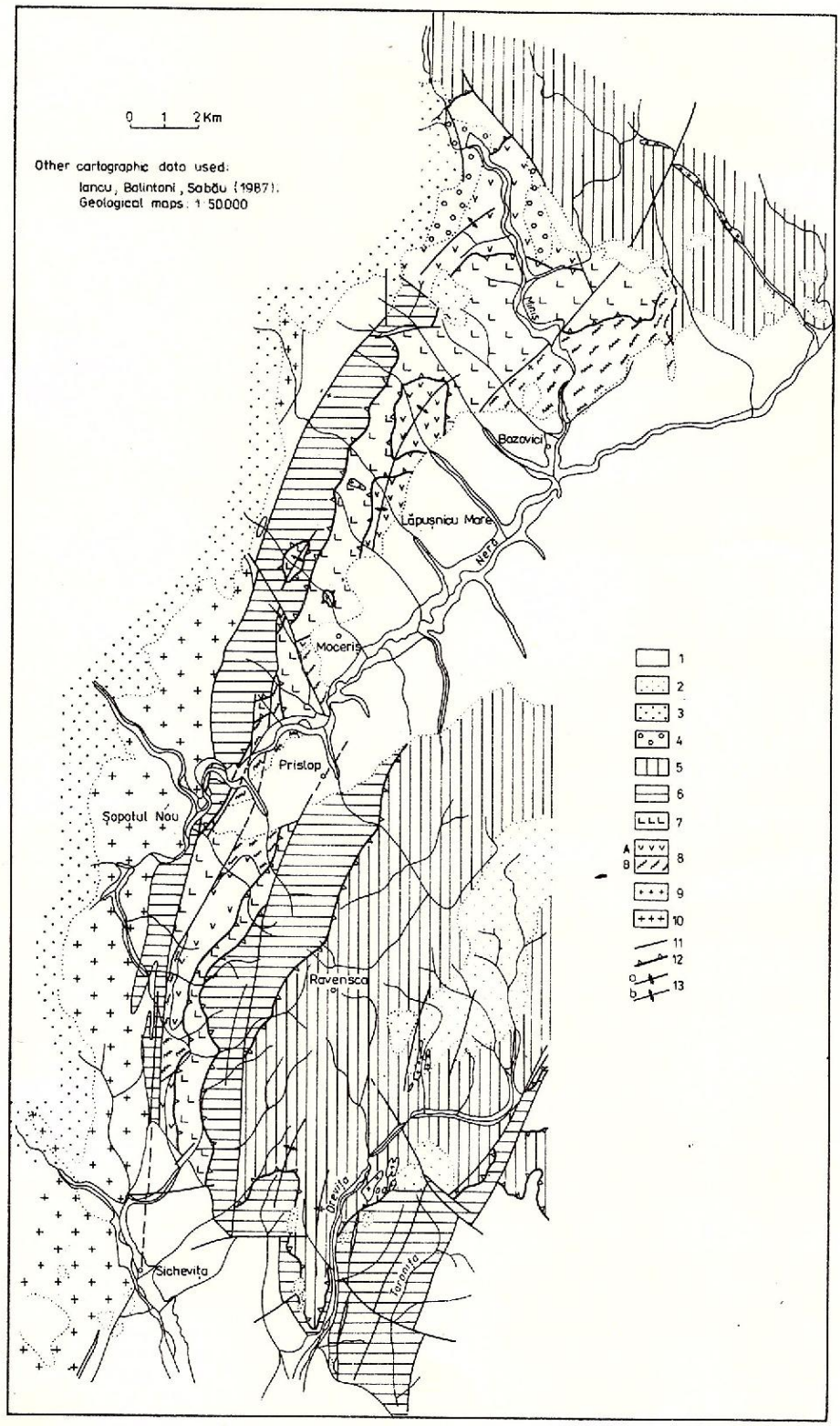

Fig. 1 Paleozoic nappe structure in the Alpine Getic Nappe (Banat-Bozovici zone). (Iancu, Mărunțiu). 1, post-Laramian cover; 2, post-Austrian cover; 3, pre-Austrian cover; 4, Variscan molasse; Variscan structural units: 5, Nera Unit: Sebeș Group (Proterozoic); 6, Oreviţa Unit: Lotru Group (Ravensca retromorphosed Fm.); 7, Miniș-Unit: Miniș Fm. (Lower Paleozoic); 8, Bozovici Unit: A, Buceava Fm. (Paleozoic); B, Lotru Group (Proterozoic); Igneous rocks: 9, Banatitic magmatites (Upper Cretaceous-Paleocene); 10, Sichevița granitoid rocks (Carboniferous); 11, fault; 12, Variscan overthrust plane; 13, post-nappe folds: a, synform; b, antiform.

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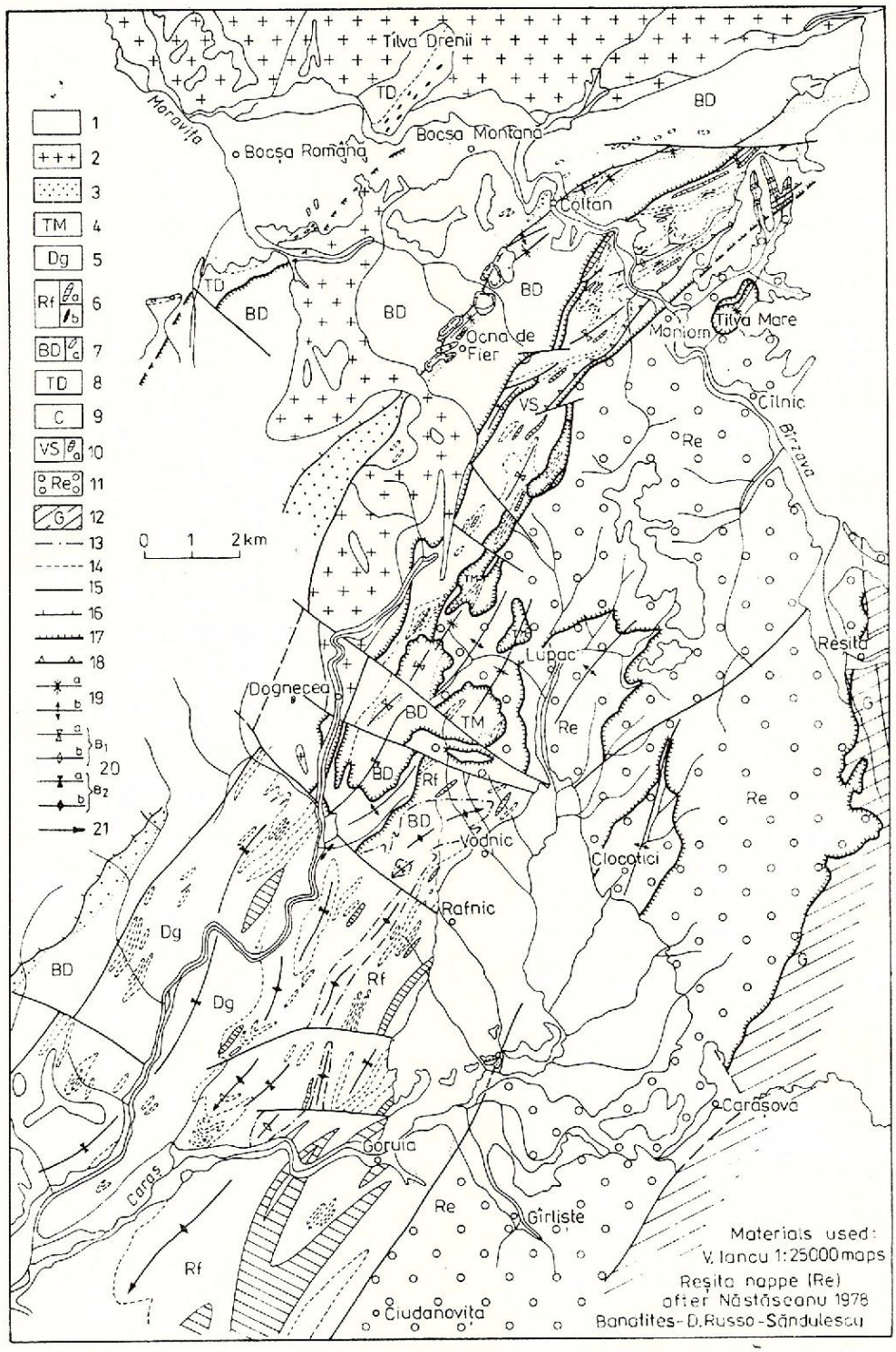

Fig. 5. Structural sketch of the Bocșa-Caraș zone (Banat) (Iancu). 1, post-Laramian cover; 2, Banatitic magmatites (Upper Cretaceous-Paleocene); Bocșa Nappe: 3, Upper Carboniferous-Mesozoic sedimentary cover; Caraș *Group* (Lower Paleozoic): 4, Tâlva Mare Fm.; 5, Dognecea terrigenous Fm.; 6, Rafnic volcano-sedimentary Fm.; a, metagabbro-diorites; b, metaultramafics; *Făgăraș Group*: 7, Bocșița-Drimoxa Fm. (Proterozoic); a, orthoamphibolites; Tâlva Drenii Nappe: *Cumpăna Group* (Proterozoic); 8, Tâlva Drenii Fm.; Moniom Nappe: *Moniom Group* (Middle Paleozoic); 9, Cârșie Fm.; 10, Valea Satului volcano-sedimentary Fm.; a, metagabbrodiorites; Reșița Nappe: 11, Upper Carboniferous-Lower Permian and Mesozoic sedimentary cover; Getic Nappe: 12, Proterozoic basement and Upper Carboniferous-Mesozoic cover; 13, pre-Alpine structural discontinuity; 14, lithologic limit; 15, fault; 16, reverse fault; 17, Austrian overthrust plane; 18, Variscan thrust plane; 19, Alpine folds: a, synform; b, antiform; 20, Paleozoic B1 and B2 folds: a, synform; b, antiform; 21, pericline.

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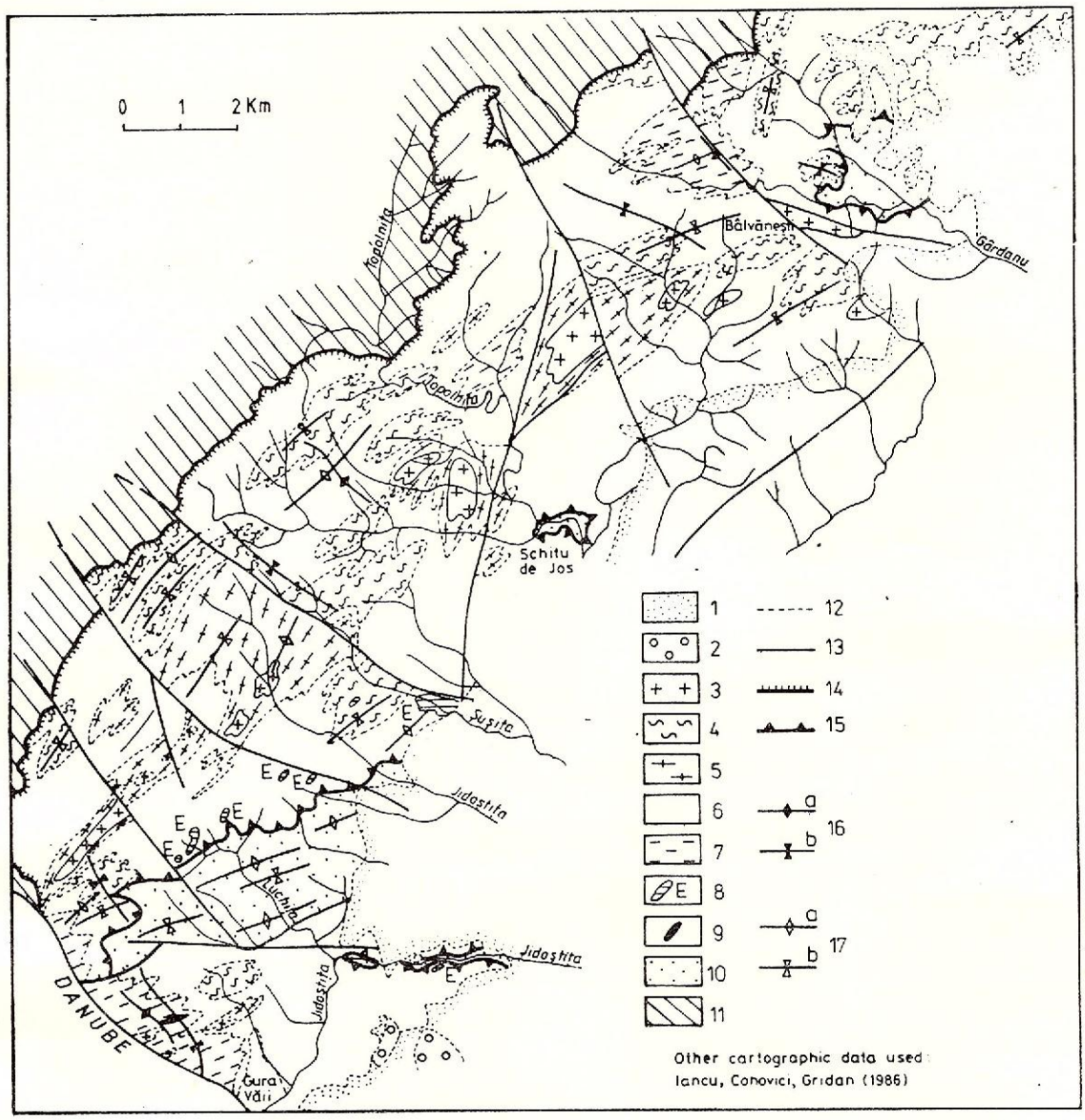

Fig. 6 – Structural Map of the pre-Paleozoic metamorphics in the southern part of the Porțile de Fier Outlier (after Iancu et al., 1986). 1, Cenozoic post-tectonic cover; Getic Nappe Porțile de Fier Outlier: 2, Mesozoic sedimentary cover; Sebeș Group, Lower plagiogneisses Fm.: 3, granitoides (diatexites); 4, Micaschist Sfm. (micaschists, plagiogneisses); 5, Quartz-feldspar gneisses Sfm. (quartz-feldspar gneisses, amphibolites, plagiogneisses); 6, Plagiogneisses Sfm., (mica plagiogneisses, amphibolites, quartz-feldspar gneisses); 7, Quartz plagiogneisses Sfm. (quartz plagiogneisses, amphibolites); 8, eclogitic and granulitic rocks; 9, metaperidotites; 10, Luchița Group, Quartzite Fm. (quartzites, plagiogneisses); 11, Severin Nappe, Lower Cretaceous limy-shaly flysch; 12, lithologic limit; 13, fault; 14, Alpine overthrust plane; 15, pre-Paleozoic-tectonic discontinuity (blastomylonitic zone); 16, B1 folds: a) antiform; b) synform; 17, B2 folds: a) antiform; b) synform.

and to the Caraș Group (Cambrian-Ordovician-Silurian) from the Bocșa Alpine Nappe. Inside the Alpine Units, some Paleozoic structural discontinuities may be regarded as overthrust planes. Other contacts, such as that between the Caraș Group formations and the Precambrian basement, may represent either normal contacts affected by progressive deformations or subsequently folded overthrust type contacts.

The Moniom Group, showing Variscan deformations and metamorphism (Sudetic phase) is characterised by:

– regional deformation (F1 folds and axial plane S1 foliations);

– preservation of both sedimentary and magmatic pre-metamorphic structures;

– differentiated blastesis of metamorphic minerals

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(stronger in carbonatic rocks and in green rocks and limited in pelite-psammitic rocks);

– existence of the transposition effects of some lithologies after the directions imposed by folding and cleavage (prevailingly mechanical transposition).

The Caraş Group shows the following structural and metamorphic characteristics:

– regional deformations (F1 folds, S1 foliations) taking place under physical conditions specific for the greenschist facies, the stilpnomelane (chlorite) and, locally, biotite or garnet zones;

– recrystallization and generalized mineral neoformation in metapelite-psammites and basic rocks (pre- and synkinematic blastesis of chlorite, epidote and actinolite and post-kynematic blastesis of albite);

– metamorphic mineral blastesis in the S1 foliation plane or, in the case of concentric folds, parallel to the bedding;

– partial preservation of some magmatic minerals and structures in metarhyolites, metadacites, metagabbros;

– partial mineral neoformation (S2) in the greenschists facies and effects of refolding, generally after coaxial directions with the B1 fold hinges.

The polyphasic character of the Caraş Group deformations, the metamorphic neoformation in a new dynamo-thermal event, as well as the lithological and structural-deformational contrast with the formations of the Moniom Group, indicate a longer history of the lower Paleozoic sequences, which are involved also in "Caledonian" events.

At present it is dificult to specify whether:

– the formations of the Caraş Group represent remnants from a Lower Paleozoic chain with independent evolution and tectonically joined to a Variscan chain, or

– the Paleozoic formations belong on the whole to the same mobile zone with a longer evolution, similar to the Variscan of western Europe (Matte, 1986).

In this context, the refolding and the partial metamorphic reorganization of the Caraş Group formations may be polyphasic, or may have a polycyclic character.

3.2. Pre-Paleozoic Structures

In the order to illustrate the regional structural relations between the main Precambrian formations of the Getic Realm from the South Carpathians, we present the synthetic sketch from Figure 3A, B. The pre-Paleozoic structural discontinuities delimit the main lithofacial rock associations, which are proved by petrologic evidences to originate in different initial, pre-orogenic, geotectonic environments.

The table shows the principal Precambrian (medium to high) litho-tectonic or structural-formational units which are represented as groups of formations (grade rocks); a brief metamorphic-deformational evolution of these groups is also presented.

The deepest and the oldest (pre-M1) crustal discontinuity within the Getic Realm is marked by a regional alignment of: eclogite, granulite and metaperidotite anisofacial exotic bodies (Fig. 3 a, b). The effects of the deep shearings at the crustal-mantle level are represented by the microstructures preserved in the nuclei ofthe metaultramafic rocks (plastic flow texture) and eclogite bodies (high pyrope garnet, omphacitic pyroxene and kyanite porphyroclasts included in a granulitic matrix with second garnet and clinopyroxene generation, sometimes with flattened simplectitic structures). The successive decompressions and reequilibrations (M1M2) of these rocks at the level of the amphibolite facies are obviously subsequent to some previous metamorphic-deformational processes taking place under conditions specific for the eclogite and granulite facies.

The dispersion of the exotic rock lenticular bodies over areas reaching tens of kilometers in length and hundreds of metres in thickness represents the effect of the strong crustal shearings. The recurring transpositions accompanied by dismembering and stretching as well as the retrograde adaptations at the level of the host rocks, make difficult the location of the crustal discontinuity marked by the exotic rocks (Iancu et al., 1988).

The other pre-Paleozoic structural discontinuities represent deep crustal shearings and may be considered to have resulted through various processes and in different moments.

The criteria according to which the discontinuites can be established and spatially followed are complex:

– changes of major lithological rock associations;

– metamorphic-deformational evolution of the various rock assemblages (independent evolution of a group or common evolution for several groups);

– presence of dynamic metamorphism effects with areal or linear development on a regional scale (microblastic, mylonitic rocks);

– presence of the overlapped reactivation effects (remetamorphosed or refolded deep-seated shear-zones).

On the whole, two main types of deformational evolution, precollisional and collisional, can be distinguished. The precollisional evolution is illustrated by effects of the extensional deformations of the thinned crust characterized by high geothermal gradients, simultaneously with the development of some isotropic (granulitic) structures or of plastic deformation accompanied by stretching, boudinages and dismemberings of some rock bodies originating from an older crust, e.g. fragments of granulitic terrains of Ursu type or of Cumpăna type augen gneisses and Brebu (plagio-

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garnet) gneisses from the Semenic Mts.

The collisional evolution, and subduction-obduction type processes, are illustrated by:

– occurrence of dismembered oceanic crust accompanied by eclogites;

– ubiguity of the effects of compressive type regional deformations (folds, foliations), accompanied by medium to low-pressure metamorphism and migmatization;

– isofacial reactivation or re-equilibrations under varied baric conditions (overlappings of metamorphic zones or of various facies series);

– generation of anatectic granitoids associated with the metamorphic or post-metamorphic events.

Crustal discontinuities are represented either by mylonitic zones (with blastesis under amphibolitic to eclogite facies conditions) or by apparent transitional zones, marked by common metamorphic planes (metamorphic nappes).

The internal structures of the pre-Paleozoic units are rather diversified and need a special treatment. All the lithological entities show evidence of a polymetamorphic-polydeformational evolution. This is illustrated in Figura 6, where the Precambrian structures are pointed out, which resulted by the overlapping of the efffects of the deformations specific for the M1 and M2 events (Grenvillian or Assynthic tectogeneses), undergone by sequences of Sebeş Group. The transposition in various stages of the D1 structures after D2 structural directions generated "bodies" of complicated morphology, making very difficult the identification of the lithological markers.

3.3. Structural Control of Some Metamorphogenetic Mineralization Occurrences

The deep crustal shear zones are of special interest for the structural control they provide in the mobilization and concentration of some metallic elements and the generation of ore bodies.

The products of the deep-seated dynamic metamorphism (blastomylonite belts, metamorphic transition zones) are locally accompanied by ore occurrences, concentrated either in metamorphic quartz bodies (e.g. Jidoştiţa-Porţile de Fier Outlier - Iancu et al., 1980, unpublished report) (Fig. 3A) or in various mylonites (e.g. Nergăniţa Valley-Semenic Mountains, Râmeşti Valley - Căpăţâna Mountains). In the former case, the quartz bodies with gold-bearing sulphides (arsenopyrite, pyrite, galena) belong to the S2 foliation of the sequences situated above the tectonic discontinuity. In the later case, the pyrite and pyrrhotite concentrations plot in the blstomylonitic foliation planes, in the maximal shear zones.

From the structural-metallogenetic point of view, the following aspects can be underlined:

– generation of some metalliferous concentrations in the deep crustal shear zones, through dynamic metamorphism (e.g. Jidoştiţa zone, acc. Iancu , 1994, in press);

– reactivation through dynamic metamorphism of some preexisting metalliferous concentrations: pre-metamorphic (e..g. Vâlsan Valley, acc. Udubaşa et al., 1988) or metamorphic (e.g. Romanilor Valley, Căpăţâna Mountains);

– reactivation of Precambrian metalliferous concentrations in more recent cycles (Variscan, Alpine) through their mobilization in mobile solutions with metallogenetic content (e.g. tectogenetic occurrences of Valea lui Stan type, acc. Udubaşa, Hann, 1988).

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GNEISS DOMES IN THE SEBEŞ-LOTRU SERIES, SEBEŞ MOUNTAINS

Ion STELEA

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Key words: Granitoids. Granitoid gneiss. Gneisses. Orthogneiss. Paragenesis. Textures. Domes. Sebeş Mountains. South Carpathians.

Abstract:The gneissic rocks in the Sebeş Mountains contain two mineral parageneses which characterize different textural varieties. The old paragenesis (oligoclase I + biotite + garnet) is well preserved in the granular textures. The younger paragenesis (oligoclase II + muscovite + microcline) characterizes the ophtalmitic and stromatic textures. The gneiss bodies with cores of granular textures (orthogneisses) outcrop in domal structures. In the Sebeş Mountains they form the lower Complex of the Sebeş-Lotru Series.

Introduction

The northern slope of the Sebeş Mts (South Carpathians) is mostly constitued of gneissic rocks. A gneissic rock is defined, like any other one, by its mineralogical composition and texture. Different composition and textures characterize different type of gneisses. Their correct cartography is very important to understand metamorphic evolution of the gneissic fields.

As the mineralogical composition variations are not easily observed in the field (e.g. one cannot differentiate oligoclase from albite or microcline), the gneisses cartography according to composition is quite difficult, even impossible. However, textural variations can be easily recognised. Moreover, the textural changes correspond to mineralogical ones.

I. Textural types of gneisses. Texture-paragenesis relationships

The high mobility of the quartzofeldspathic phase during metamorphism lead to varied textures in the gneissic rocks. The gneissic textures widespread in the north of the Sebeş Mts are granolepidoplastic, ophtalmitic (augen), stromatic (banded) and granular. The ophtalmitic and stromatic textures are later and they replace older granular and granolepidoblastic textures.

Gneissic rocks contain two medium-grade parageneses belonging to two Prealpine regional metamorphic events (M_1 and M_2). The M_1 paragenesis, prekinematic to S_2 foliation, consists of oligoclase I, biotite and garnet. It occurs in all gneisses, but is well preserved in the granular and granolepidoblastic ones. The M_2 paragenesis, synkinematic to S_2 foliation, consists of oligoclase II, muscovite and microcline. In the replacement of M_1 by M_2 paragenesis we observed two stages:

a) without essential change of the old textures (granular or granolepidoblastic), when the second of paragenesis is quantitatively insignificant and contains only oligoclase II and muscovite;

b) with the change of the old textures into new ones (ophtalmitic and stromatic), when the second paragenesis is well developed and contains oligoclase II, muscovite and microcline.

The relationships between parageneses and textures show that the granular and granolepidoblastic textures are previous to the M_2 metamorphism, whereas the ophtalmitic and stromatic ones are synkinematic to this metamorphic event. Except for the granular textures, that are of magmatic origin, all the other textures are metamorphic, their development being controlled by deformation on S_2 foliation.

The two oligoclase generations differ by physiographic characters (size and shape). The oligoclase I grains, having smooth grain-boundaries, are albite-law twinned and marginally show myrmekites. Usually they are replaced by clinozoizite, sericite or calcite. The oligoclase II is coarse-grained, untwinned and have sinuous grain-boundaries. Sometimes is replaced by calcite. Both generations show close anorthite content (about 20 per cent). Greater differences can be observed in the granular gneisses (from basic to acide oligoclase).

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Important amounts of microcline crystallize in the second paragenesis only in the case of the total change of the rock texture that involves a strong synmetamorphic deformation. The deformation also explains the increase of retrograde reactions in the retextured gneisses: biotite breakdown into muscovite and clinozoizite neoformations on plagioclase I partially replaced by oligoclase II and microcline. Is not clear if microcline replaces or not even oligoclase II.

Microcline is scarce in the gneiss body from Luncani Valley (Fig. 3). Here, oligoclase I recrystallizes into oligoclase II during M_2 event, but microcline does not crystallize. Post-M_2, albite and calcite replaced oligoclase II during a brittle deformation episode related to the Alpine Cioclovina strike-slip fault.

The plagioclase synkinematic recrystallisation, as well as the biotite breakdown into muscovite, occur in many metamorphic rocks from South Carpathians. These processes belong to the M_2 regional metamorphism. Microcline represents a total new mineral phase. A pre-M_2 K-feldspar generation in these rocks is not known. Its existence can be presumed only if we accept that it has been completely recrystallised. But the microcline physiographic features, determined by the filling of the voids generated by cataclasis and by the oligoclase substitution in percentages varying from one grain to another, are in favour of its metasomatic origin.

The equivelence between the textural, mineralogical and genetical terms for the Sebeş Mts gneisses is rendered in the table below:

Textural terms	Mineralogical terms	Genetical terms
granular gneisses	plagioclase gneisses granitoid gneisses tonalitic gneisses	orthogneisses
granolepidoblastic gneisses	plagioclase gneisses plagioclase and microcline gneisses	paragneisses (?)
ophtalmitic gneisses stromatic gneisses	plagioclase and microcline gneisses	tectonogneisses

On the map, the orthogneisses include both granular and associate ophtalmitic and stromatic gneisses. Granular gneisses are in fact granitoids with tonalitic mineralogical composition and coarse, non-oriented texture. Such rocks have been described in the Romanian geological literature as granitic gneisses. The para-origin of mica gneisses is not proved so that the term paragneisses must be carefully used. However, for the Sebeş-Lotru Series it describes the mica-rich gneisses. The ophtalmitic and stromatic gneisses have been generated by deformation, especially on the expense of granitoid gneisses. Because sometimes is difficult to recognize the pre-deformation gneissic rocks, we suggest the term tectonogneisses for their origin.

II. Geometry of the orthogneisses bodies. Litostratigraphic implications

The orthogneisses bodies occur as "islands" in the paragneisses "sea" which covers the northern slope of the Sebeş Mts. Their actual configuration is the result of the interaction of several factors: internal texture, relationships with the S_2 foliation and erosion level. Their size vary as follows: 2–15 km long axis and 1–5 km short axis. The smaller bodies cannot be identified as such because they are distroyed by metamorphic transposition after S_2 foliation.

Most of the orthogneisses bodies preserve in the central part one or several cores of granitoids enveloped in ophtalmitic gneisses, in their turn enveloped in stromatic ones. In accordance with their morphology, two structural situations can be defined: (1) gneissic domes and (2) gneissic layers.

1. Gneissic domes

They are ellipsoidal bodies of orthogneisses which always contain granitoid cores in the central part. Usually S_2 foliation dips outwards (Figs. 1 and 2), indicating a wideness at depth, below paragneisses (an antiform) and only sometimes dips inwards suggesting a narrowing at depth (a synform) (Fig. 3). However, the foliation surrounds the orthogneisses, so that any folds are excluded.

Taking into account their morphology these bodies are similar to the domes described by Eskola (1949). Initially they consisted of granitoid rocks (tonalites), which were subsequently metamorphosed into ophtalmitic and stromatic gneisses during the regional M_2 metamorphic event. The shape of the domes becomes lenticular by deformation and this shape explains the variation of the foliation dipping around them.

The orthogneisses domes show a textural zonality parallel with their contour: granular gneisses (granitoids) in the central part, ophtalmitic gneisses in the intermediary zone and stromatic gneisses on the border. Stromatic gneisses can also occur inside the dome,

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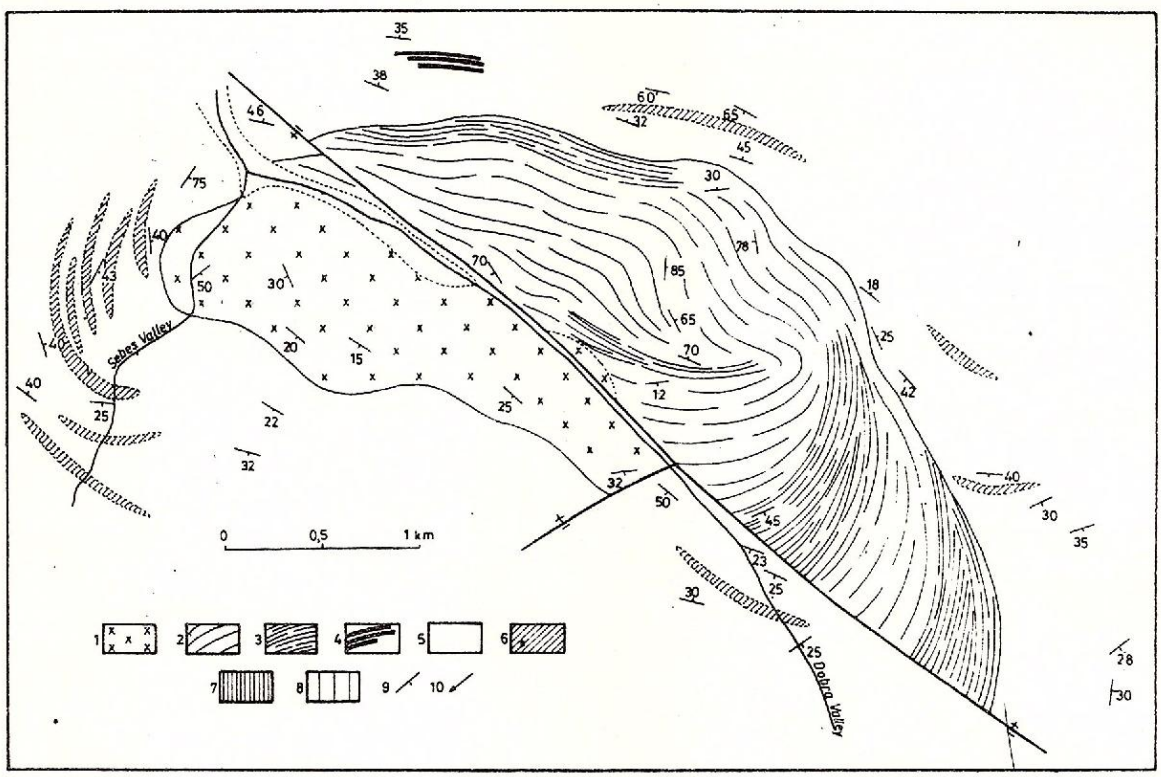

Fig. 1 — Structural sketch of the orthogneisses body in Dobra Valley. 1, granitoids; 2, ophtalmitic gneisses; 3, stromatic gneisses; 4, blastomylonites; 5, paragneisses; 6, amphibolites; 7, crystalline limestones; 8, sedimentary formations; 9, S_2 foliation; 10, aggregate lineation (l_2).

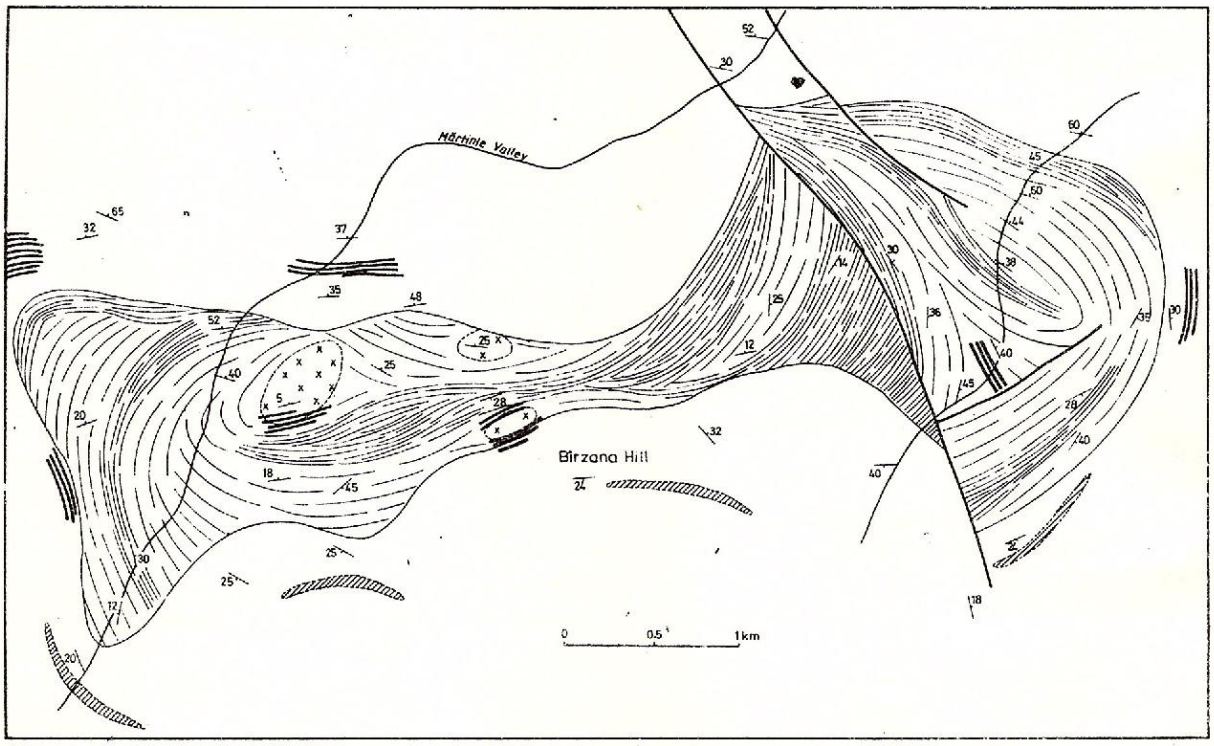

Fig. 2 — Structural sketch of the orthogneisses body in the Bârzana Hill. Same legend as Figure 1.

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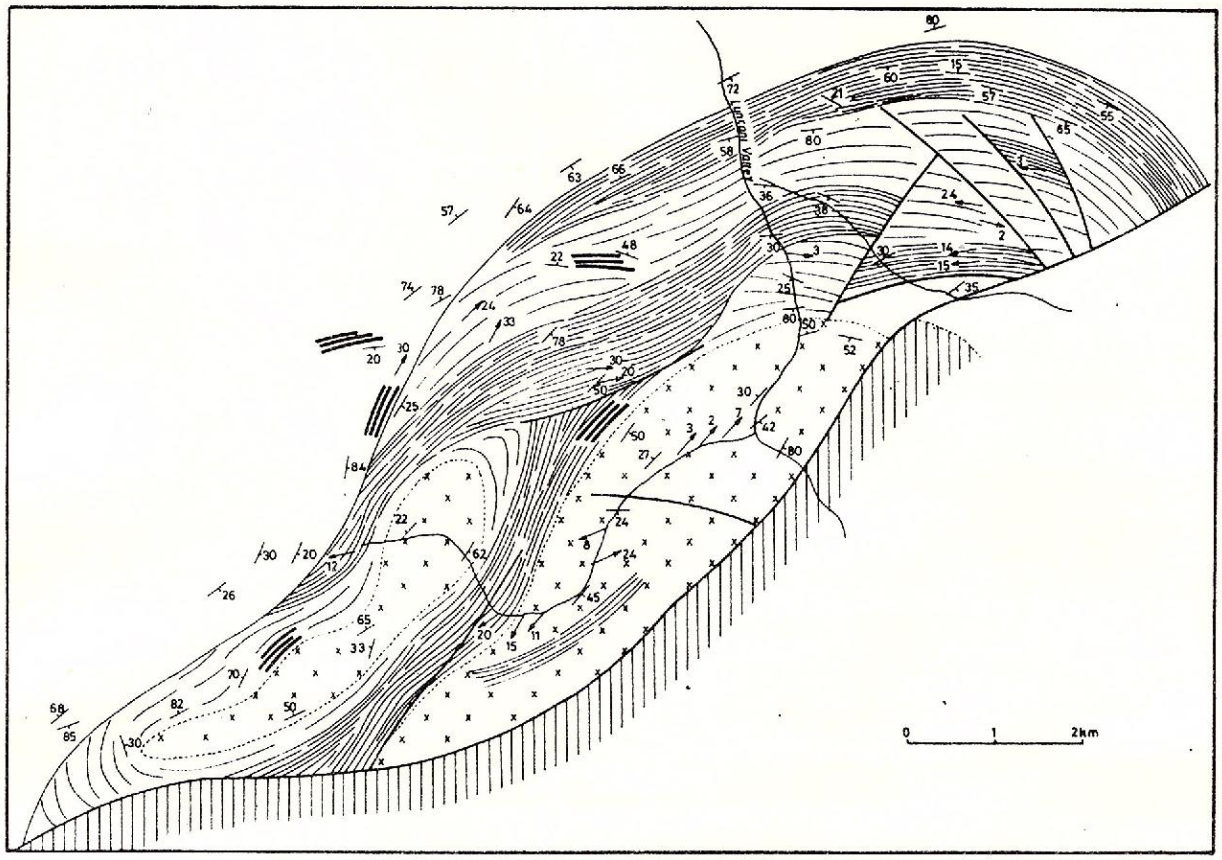

Fig. 3 – Structural sketch of the orthogneisses body in the Luncani Valley. Same legend as Figure 1.

usually between two neighbouring granitoid cores (Fig. 3).

Around the domes frequently occur biotite blastomylonites formed at the expense of the surrounding paragneisses. Fine-grained biotite blastomylonites also occur in the interior, especially at the contact of granular gneisses with the other ones. This textural zonality reflects a zonality of the stress distribution within the domes during the M_2 metamorphism. Certainly, a significant part in the blastomylonites generation was played by the competence contrast between granitoids and paragneisses.

Similar dome structures were described by Iancu et al. (1990, unpubl. report) in the Semenic Mts. They consist of nodular plagioclase gneisses with relict granoblastic textures, which grade at the exterior into mica plagiogneisses by means of the blastomylonitized gneisses.

In the Sebeş Mts. most of the gneiss domes outcrop in the eastern part of the northern slope: Sebeş and Dobra valleys, Barzana Hill. The biggest dome (15 km long axis and 5 km short axis) crops out in the westernmost part, in the Luncani Valley basin.

2. Gneissic layers

They are orthogneisses bodies that outcrop only in the thalweg of the main valleys: Dobra, Mărtinie, Pianu. The presence of the granitoid gneisses(in association with amphibolites) along the valleys from the southern and central part of the Sebeş Mts was commented by Hârtopanu (1982, unpubl.). They are overlain by a horizontal pile of paragneisses and micaschists.

In the northern part of Sebeş Mts. the gneissic layers mostly consist of gneisses with ophtalmitic texture. When they exist, the granitoid cores are randomly disposed as against the cartographic bodies contour (Fig. 4). The textural zonality around them is not parallel with their outline. The lack of blastomylonite shows that the present cartographic contour is only the result of the erosion.

The S_2 foliation penetrates the bodies after wide curves with thecentres (i.e. possibly granitoid cores) frequently outside the present outline. One can infer that the orthogneiss layers horizontally extend under the paragneiss to form a continuous level in the

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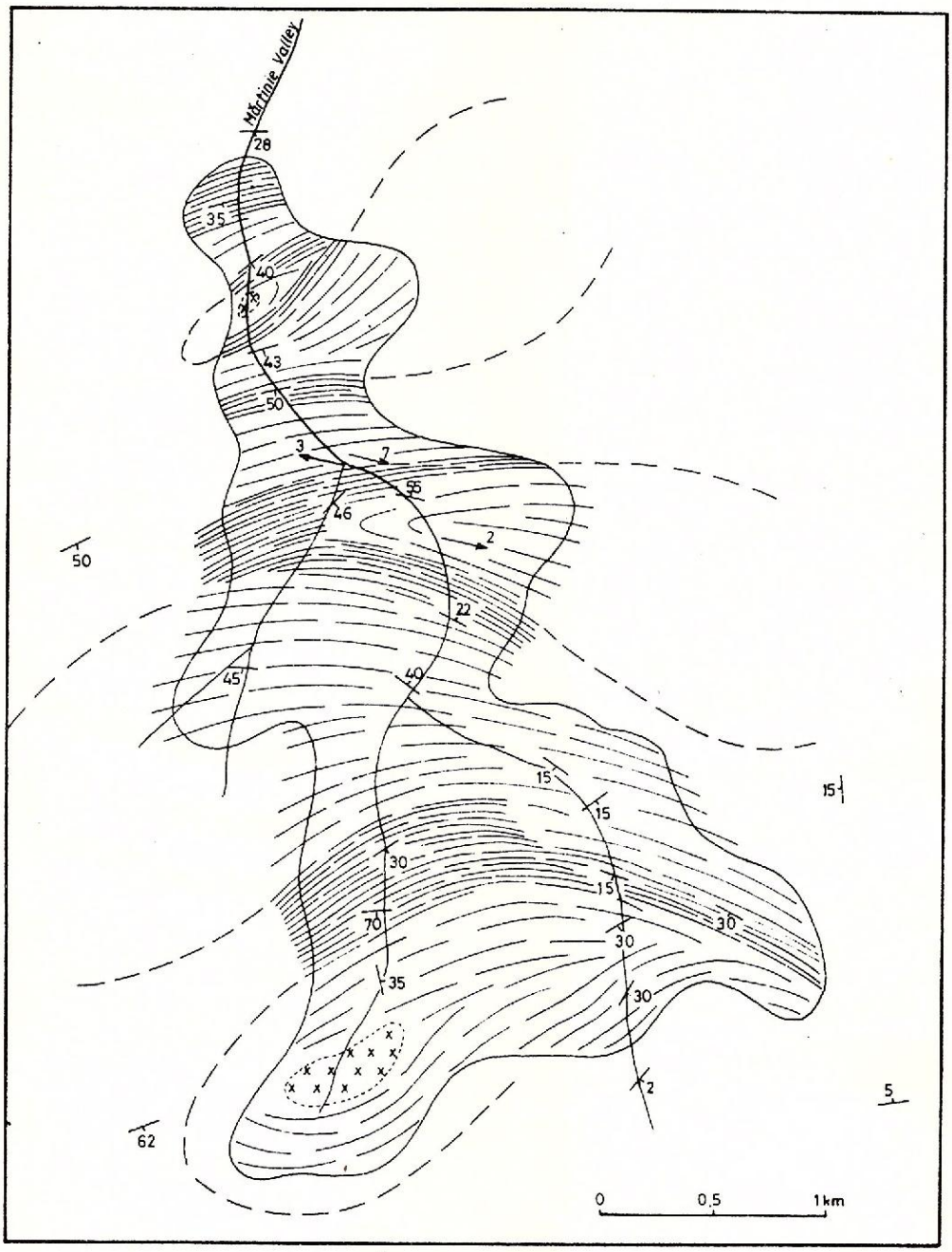

Fig. 4 - Structural sketch of the orthogneisses body in the Mărtinie Valley. Same legend as Figure 1.

basement of the Sebeş-Lotru Series.

3. *Lithostratigraphic implications*

At the beginning of the century, the gneissic rocks in the Sebeş Mts were of a great interest for the researchers. They recognize isometric bodies of granites and gneisses with mineralogical compositions and textures, different from the mica gneisses ones, so that many orthogneisses bodies mapped by us have been previously mentioned under different interpretations.

Laufer (1924) mentioned a "granites and orthogneisses" body in the Luncani Valley. Vendl (1932) described several gneisses varieties in the central and southern part of the massif: "biotite-bearing granitic gneisses" in the Frumoasa, Galbina, Taia, Auşelu, Pietrosu, Sipcea and Bistra valleys, "hornblende-bearing granitic gneisses" in the Frumoasa Valley,

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"augen gneisses" in the Sebeş Valley, "biotite and muscovite-bearing gneisses" in the Dobra Valley. Ghika-Budeşti (1940) separated "augen gneisses" bodies in the Sebeş, Cugir, Cibin valleys and "granitic gneisses" in the Bistra, Dobra and Frumoasa valleys.

All these gneisses varieties were included by Savu et al. (1977, unpubl. report) in the quartzofeldspathic gneiss complex (C_3), with its maximum extent on the northern side. Hartopanu (1982, unpubl.) noted the preferential disposition of the granitic gneisses in the valleys thalweg and hence inferred the tabular structure of the Sebeş Lotru Series, with gneisses in the bottom, paragneisses at the middle part and micaschists at the top.

On the geological map of Romania, scale 1:50000, Şugag sheet (Stelea et al., 1991), as well as on the map annexed to the present paper, the orthogneisses bodies have been included in the orthogneisses complex, as the lower complex of the Sebeş-Lotru Series. It is overlain by the paragneiss and micaschists complex.

The lower complex outcrops either as domes or as tabular bodies. With a remarkable lithologic homogeneity, it consists of granular, ophtalmitic and stromatic gneisses. Small bodies of amphibolites and rhyodacites dykes are rarely found.

The middle complex covers most of the northern slope and consists of mica gneisses with granolepidoblastic texture, locally replaced by ophtalmitic and stromatic ones. Interlayered micaschists amphibolites, microcline gneisses with microgranular texture (not discussed in this paper), eclogite, metaperidotite and granodiorite bodies, rhyodacites dykes and pegmatite lenses are found as well. At the upper part there is a characteristic lithological assemblage of amphibolites, microclines gneisses, pegmatites and eclogites.

The upper complex overlies the central zone of the Sebeş Mts and extends in the northern part only along interfluves. It consists of micaschists with garnet, staurolite and kyanite, locally replaced by sillimanite. Interlayered mica gneisses, pegmatites, amphibolites, metaperidotites, eclogites, manganese sillicate schists occur too. The quartzofeldspatic phase in micaschists also recrystallize during the M_2 event, but with minor textural changes.

The lithostratigraphic limits of the Sebeş-Lotru Series are pre-existent to M_2 metamorphic event (S_1 surfaces). Since the three complexes show different, rheological properties, the limits between them behaved as shear-zones during the M_2 event. The presence of blastomylonites along them proves this.

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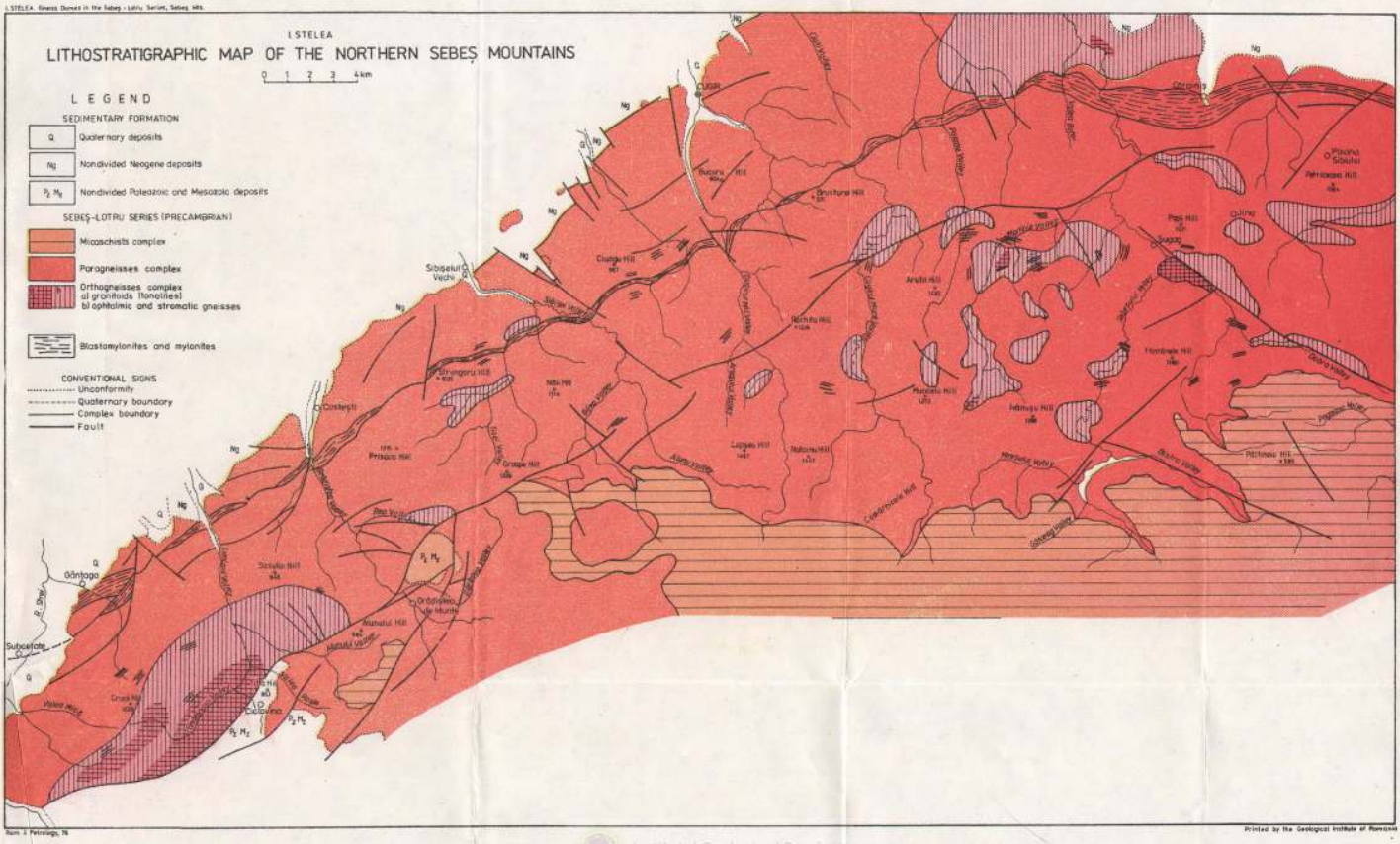
I. STELEA. Gneiss Domes in the Sebeş - Lotru Series, Sebeş Mts.



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INSULA DE CRISTALIN DE LA BREBU-ZORLENCIOR, O VERIGĂ DE LEGĂTURĂ ÎNTRE PÂNZELE SUPRAGETICE DIN MUNŢII BANATULUI ŞI MASIVUL POIANA RUSCĂ (CARPAŢII MERIDIONALI VESTICI)

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Key words: Metamorphic rocks. Precambrian. Paleozoic. Tectonic units. Nappes. South Carpathians.

Abstract: Brebu-Zorlencior Crystalline Island, a Link between the Supragetic Nappes in the Banat Mountains and the Poiana Ruscă Massif (Western Southern Carpathians). The Brebu-Zorlencior Crystalline Island presents geological information necessary for the structural correlation between Supragetic Nappes of the Banat Mountains and the Poiana Ruscă Massif. This area consists of three Austrian basement nappes: the Bocşa Nappe overlies the Tâlva Drenii Nappe, overlying the Moniom Nappe. The nappe boundary is modified by subsequent faults. The overthrust plane between the Bocşa Nappe and the Tâlva Drenii Nappe is marked by quartz lenses of tectonic origin. The Bocşa Nappe consists of medium-grade metamorphics showing retrogressive features (Bocşiţa-Dricmocsa lithogroup). The crystalline basement is overlain by a Jurassic sedimentary cover. The Tâlva Drenii Nappe consists of Precambrian ocular migmatites (Cumpăna lithogroup). The Moniom Napppe is made up of low-grade Paleozoic metamorphics with a Cenomanian-Coniacian cover. These Austrian Nappes are included in the Lotru-Bistra Laramian Nappe.

1. Introducere şi istoric

Situată la nord-est de localitatea Brebu şi la vest de oraşul Caransebeş, insula de cristalin se găseşte în imediata apropiere a satului Zorlencior, la marginea sud-estică a acestuia. Şisturile cristaline formează un deal cu altitudinea maximă de 487 m, denumit "Măgura Brebului", cu o suprafaţă de aproximativ 8 kmp, care se ridică din întinderea largă a depozitelor marine neogene înconjurătoare.

Cercetările noastre au fost întreprinse cu ocazia cartărilor pentru elaborarea hărţii geologice scara 1 : 50.000, foaia "Caransebeş" (Hann et al., 1992, date nepublicate), realizându-se o imagine cartografică inedită.

Pe baza rezultatelor obţinute şi în concordanţă cu evoluţia cunoaşterii din ultimii ani a structurilor alpine din Munţii Banatului, respectiv din masivul Poiana Ruscă, devine acum posibilă corelarea acestora şi prin intermediul diferitelor formaţiuni cristaline şi sedimentare ce aflorează pe acest teritoriu.

Având o suprafaţă redusă şi o poziţie izolată, această zonă a fost mai rar cercetată de către geologi. Prima cartare de amănunt o datorăm Adelei Arghir Drăgulescu, care prezintă în anul 1963 (date nepublicate) un raport de prospecţiuni geologice însoţit de o hartă la sc. 1 : 10.000. Ulterior, în anul 1969, autoarea foloseşte o parte din aceste date la redactarea hărţii geologice a României, sc. 1 : 200.000 foaia 25 "Deva". Încadrează tipurile petrografice descrise într-o serie gnaisică (gnaise oculare şi gnaise aplitice) şi în seria şisturilor cu porfiroblaste de albit, ambele fiind atribuite seriei de Sebeş-Lotru; mai distinge însă în sudul insulei şi o serie epimetamorfică, cu un complex al rocilor verzi şi un complex blastodetritic. Formaţiunile sedimentare mezozoice constituind cuvertura pre-laramică au fost de asemenea menţionate. O referire mai precisă, bazată pe date paleontologice asupra vârstei unora dintre formaţiuni şi pe corelarea

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lor cu cele din bazinul Rusca Montană, se face mult mai târziu de către Bucur şi Strutinski (1982).

Menţionări cu privire la insula cristalină de la Brebu-Zorlencior au fost făcute cu ocazia unor corelări litologice sau structurale din cuprinsul segmentului vestic al Carpaţilor Meridionali. In anul 1961 apare Ghidul Excursiilor al Asociaţiei Geologice Carpato-Balcanice (Codarcea et al.), în care se arată că în insula de la Măgura Brebului se găsesc două unităţi tectonice: unitatea Ocna de Fier ce cuprinde şisturile cristaline ale complexului gnaiselor micacee, inclusiv şisturile cloritice cu porfiroblaste de albit; şi unitatea de Locva ce conţine şisturi epimetamorfice (un complex filitos şi tufogen de roci verzi). Intregul ansamblu este atribuit domeniului getic.

Maier (1974) întrebuinţează noţiunea de "cristalin supragetic" în sensul lui Streckeisen (1934) şi corelează, pe baza unor date noi, seria şisturilor cu porfiroblaste de albit descrisă în insula Brebu-Zorlencior de Arghir (1963, date nepublicate) cu zona Bocşiţa-Drimoxa (Ocna de Fier) descrisă ca atare deja în anul 1931 de către Codarcea, cu complexul şisturilor cu porfiroblaste de albit din munţii Locva, respectiv cu complexul de şisturi muscovitice, cu porfiroblaste de oligoclaz de la est de localitatea Tincova (Poiana Ruscă) descris de Neacşu et al. (1961, date nepublicate).

Recent Dimitrescu (1990) corelează din punct de vedere litostratigrafic şisturile cu porfiroblaste de albit din insula cristalină Brebu, nu numai cu cele asemănătoare întâlnite în munţii Banatului şi în vestul masivului Poiana Ruscă, ci şi cu cele din masivul Serbo-Macedonean, iar spre est şi nord de Poiana Ruscă cu cele din munţii Cibinului şi Lotrului, din munţii Iezer-Păpuşa-Leaota, respectiv cu cele din Munţii Apuseni, masivele Bihor-Gilău. Autorul constituie astfel o unitate litostratigrafică prealpină, denumită "grupul Vlasina-Locva-Leaota".

2. Descrierea unităţilor tectonice

In cuprinsul insulei de cristalin afloreză formaţiuni metamorfice şi, pe zone foarte restranse, depozite sedimentare mezozoice, caracteristice pentru trei dintre pânzele supragetice occidentale (Fig. 1). Privind de sus în jos, au fost recunoscute panza de Bocşa (Iancu, 1986), pânza de Talva Drenii (Kräutner et al., 1981; Iancu, 1986) şi, cu poziţie inferioară, pânza de Moniom (Iancu în Russo-Săndulescu et al., 1982; Iancu, 1985).

2.1. Pânza de Bocşa

Această unitate tectonică este constituită dintr-o stivă de metamorfite mezometamorfice, adesea cu aspecte retromorfe, pe care stau transgresiv şi discordant depozite sedimentare calcaroase eocretacice-mezojurasice; ea şariază atât peste unitatea de Talva Drenii cât şi peste cea de Moniom.

Şisturile cristaline afloreză în partea central-nordică, respectiv sudică a regiunii, aici evident la est de planul tectonic marcat de cuarţite care le separă de gnaisele liniare şi oculare ale pânzei de Talva Drenii. Plagiognaisele sunt mai mult sau mai puţin micacee, fără să se observe însă şi micaşisturi propriu-zise. Rocile sunt intens foliate, uneori microcutate, contrastând astfel puternic faţă de gnaisele liniare învecinate. Din punct de vedere mineralogic sunt alcătuite din cuarţ, feldspat plagioclaz (oligoclaz, albit) mice, granaţi, clorit, apatit, minerale opace şi semiopace. Cuarţul apare în două generaţii. Se observă cristale de cuarţ de dimensiuni ceva mai mari prezentând diverse grade de tectonizare (extincţii ondulatorii, poligonizări), cristale ce au un caracter relict. A doua generaţie de cuarţ este microcristalină cu extincţii de obicei normale şi structuri mozaicate şi s-a format pe seama cuarţului de generaţia întâi. Dar cuarţul secundar se găseşte şi în cadrul procesului de substituire a oligoclazului printr-un agregat cuarţo-muscovitic. Frecvent cristalele de cuarţ se alungesc în lungul foliaţiei.

Plagioclazul apare, de asemenea, în două generaţii şi poate forma porfiroblaste. Plagioclazul de generaţia întâi este un oligoclaz microgranoblastic, numai uneori porfiroblastic, sericitizat, rar maclat polisintetic. De obicei coexistă cu plagioclazul din generaţia a doua, care este un albit porfiroblastic limpede, cu incluziuni de oligoclaz. Dar substituirea poate fi şi totală, rămânând numai albitul.

Micele, reprezentate printr-un biotit, parţial sau total substituit prin clorit, şi muscovit sub formă de fascicule fusiforme, ondulate sau recurbate, sau formând o reţea filiformă anastomozată cu ochiurile alungite în direcţia foliaţiei. Aceste aspecte indică faptul că şisturile au fost supuse unor deformări intense şi repetate.

Granaţii se observă numai sub microscop, sunt pe cale de a fi substituiţi prin clorit şi deseori apar sparţi. Rocile conţin şi minerale de titan (ilmenit, leucoxen), mai rar rutil sagenitic format în timpul substituirii biotitului prin clorit. Acest ultim aspect este însă foarte caracteristic pentru procesele retromorfe ce afectează metamorfitele de intensitate medie. Remarcăm astfel că întreaga stivă de şisturi a suferit un retromorfism destul de penetrativ, care s-a suprapus peste foliaţia S2.

Şisturile cristaline din panza de Bocşa se aseamănă foarte mult cu ceea ce Codarcea (1930) a descris ca fiind complexul gnaiselor micacee din zona Buchin-Ocna de Fier, la est de Bocşa, munţii Banatului, corelare

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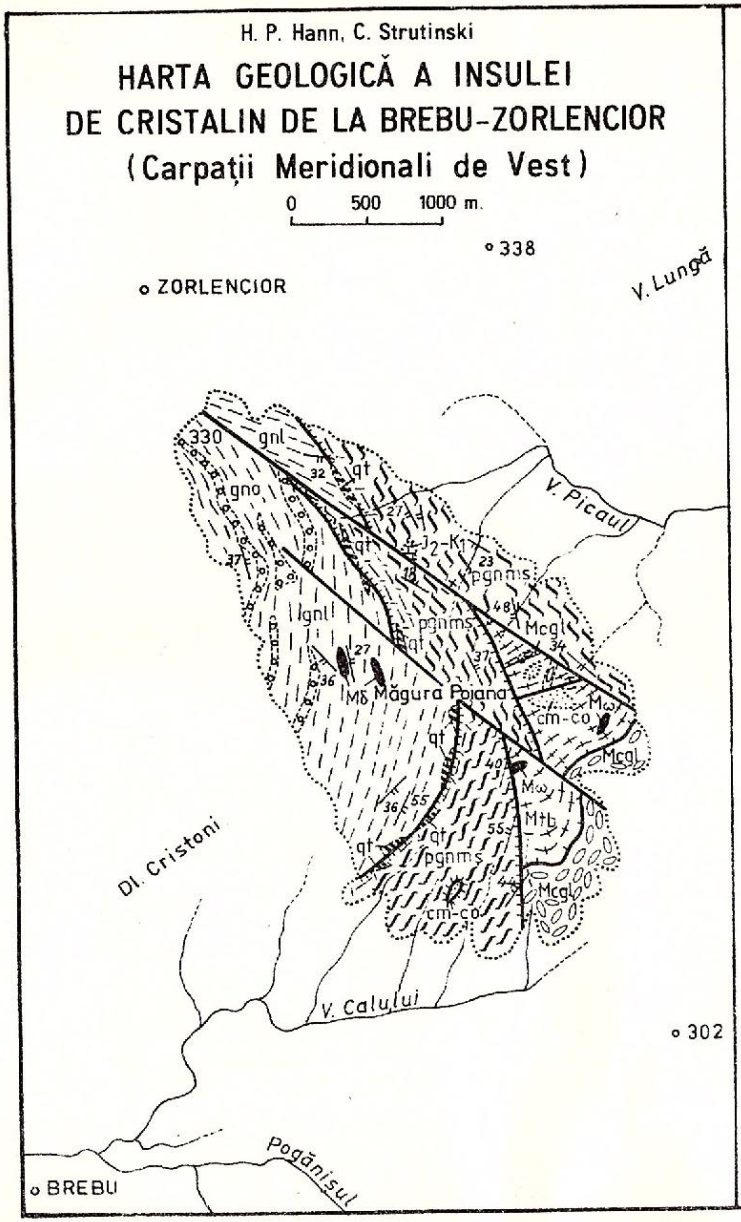
H. P. Hann, C. Strutinski

HARTA GEOLOGICĂ A INSULEI DE CRISTALIN DE LA BREBU-ZORLENCIOR (Carpații Meridionali de Vest)

Fig. 1



CUVERTURA POST-LARAMICĂ

CUATERNAR și NEOGEN — Depozite aluviale ale teraselor, sillite, argile nisipoase, nisipuri

UNITĂȚI SUPRAGETICE

Pînza de Bocșa

EOCRETACIC MEZO JURASIC ? (nedivizat) — J_2+K_1 — Calcare detritice și calcare dolomitice necristalizate

Formațiunea (litozona) Bocșița-Drimoxa - metamorfism barrovian pre-caledonian în faciesul amfibolitelor cu almandin, zona cu almandin, metamorfism regresiv în faciesul șisturilor verzi

PRECAMBRIAN SUPERIOR B — pgn mș — Plagiognaise ± micacee cu granat, porfiroblaste de oligoclaz, aspecte retromorfe cu clorit

Pînza de Tilva Drenii

Grupul (litogrupul) Cumpăna

Formațiunea (litozona) Tilva Drenii-Tincova - metamorfism barrovian precambrian policiclic în faciesul amfibolitelor cu almandin

PRECAMBRIAN SUPERIOR B — gnl — Plagiognaise cu nivele de gnaise liniare și de gnaise albe cuarto-feldspatice (leptinite)

gno — Gnaise oculare

PALEOZOIC SUPERIOR ? — Mδ — Metadacite

Pînza de Moniom

CONIACIAN-CENOMANIAN — cm-co — Conglomerate, gresii carbonatice, calcare, marnocalcare grezoase

Grupul (litogrupul) Leșcovița - metamorfism regional varistic în faciesul șisturilor verzi, zona cu clorit

Formațiunea (litozona) Cîrșie

CARBONIFER INF.-DEVONIAN — Mcgl — Metaconglomerate polimictice, metagresii

Formațiunea (litozona) Valea Satului

Mtb — Metatufuri, metatufite bazice (șisturi cu clorit, epidot, albit ± cuarț, sericit, calcit) cuarțite

Mω — Metagabbrouri

ROCI FILONIENE

CRETACIC SUPERIOR ? — qt — Cuarțite albe de origine tectonică

SEMNE CONVENȚIONALE

- Limită litologică
- Limită de discordanță
- Pînză de șariaj
- Falie inversă
- a b — Falie verticală sau subverticală cu decroșare în plan orizontal (a) sau vertical (b)
- Foliație S_2 în roci metamorfice
- Zone de laminare

Institutul Geologic al României

reluată de acelaşi autor şi în anul 1961. Ulterior aceste corelări au mai fost realizate şi de Maier (1974), Kräutner et al. (1981), Iancu (1986), Balintoni et al. (1989), incluzand însă în mod eronat şi gnaisele oculare de tip Tâlva Drenii-Cumpăna din insula de cristalin de la Brebu în stiva şisturilor cu porfiroblaste de plagioclaz. Spre est, în munţii Poiana Ruscă, şisturile cu porfiroblaste de oligoclaz-albit se reîntâlnesc la est de localitatea Căvăran în lungul văii Vălişoara şi au fost descrise de către Neacşu et al., (1961, date nepublicate) şi de către Kräutner (1972, foaia 1:50.000 "Nădrag" şi 1990, date nepublicate) sub denumirea de complexul, respectiv "formaţiunea de Vălişoara". Aderand la clasificarea litostratigrafică informală a secvenţelor proterozoice, propusă de Balintoni, Berza şi Hann (în Balintoni et al., 1989), includem stiva de şisturi, deci pânza de Bocşa, în metalitofaciesul Negoi, formaţiunea (litozona) Bocşiţa Drimoxa, corelabil la rândul său cu litogrupul Sibişel din Carpaţii Meridionali centrali.

Depozitele sedimentare de vârstă cocretacic-mezojurasică sunt reprezentate prin calcare cenuşii micritice, calcare dolomitice recristalizate, gresii calcaroase cenuşii-negricioase, şi se întâlnesc într-un mic petic situat în partea central-nordică a insulei de cristalin. Atribuirea vârstei s-a făcut prin analogie cu depozitele identice descrise de Codarcea (1930) în regiunea Bocşa, ulterior figurate şi de către Năstăseanu (în Russo-Săndulescu et al., 1982) pe foaia 1:50.000 "Bocşa". În Poiana Ruscă, în versantul stâng al văii Vălişoara, şisturile cu porfiroblaste de plagioclaz suportă de asemenea o cuvertură sedimentară jurasică (Dincă în Kräutner et al., 1972).

2.2. Pânza de Tâlva Drenii

Fondul petrografic al metamorfitelor pe care le conţine această unitate tectonică este constituit din gnaise liniare, roci dezvoltate la sud-vest de culmea principală a insulei cristaline.

În alcătuirea lor intră cuarţ, feldspaţi, mică şi granaţi foarte mărunţi, fisuraţi, parţial alteraţi, de mărimi microscopice. Roca se caracterizează printr-o aliniere riguroasă a componentelor leucocrate după direcţia şistozităţii, care este foarte bine exprimată şi unică. Cuarţul, adesea dezvoltat în "şnururi" este microgranoblastic. Feldspaţii formează granule ovoidale constituite din microclin şi oligoclaz. Micele filiforme, sau dispuse în benzi subţiri, sunt reprezentate prin muscovit şi biotit, sau numai prin una dintre cele două varietăţi.

În masa gnaiselor liniare se întâlnesc frecvent nivele metrice sau decimetrice de gnaise oculare. Aceste roci sunt constituite din benzi alternante milimetrice sau submilimetrice de cuarţ, feldspat şi mice, care mulează ochiuri lentiliforme de microclin şi cuarţ ce pot atinge mărimi de 1–2 cm. Plagioclazul din benzile fine alternante este reprezentat prin oligoclaz, uneori maclat polisintetic, niciodată prin microclin, care se întâlneşte exclusiv în ochiuri. Micele sunt reprezentate prin biotit, subordonat muscovit. Pe lângă biotitul mai larg dezvoltat şi orientat după foliaţie, apare şi un biotit scurt, orientat oblic sau perpendicular pe foliaţie, părând a străbate lamelele lungi şi concordante. S-au mai observat apatit şi minerale opace.

Alături de tipurile de roci menţionate mai sus se mai întâlnesc benzi metrice, evidente, de gnaise albe cuarţo-feldspatice constituite din cuarţ, plagioclaz şi muscovit. Uneori plagioclazul a fost parţial substituit prin albit ± feldspat potasic. Muscovitul conţine minerale opace. Rocile prezintă o foliaţie slab exprimată, şi la prima vedere par a fi cuarţite laminate.

Pe afluentul stâng al văii Roba, în cadrul stivei de gnaise liniare s-a întâlnit un corp de rocă, de 1–2 m grosime, format aproape exclusiv din albit. Aceste albitite pot conţine rare cristale de cuarţ şi, ca mineral accesoriu, zircon.

Imaginea alcătuirii petrografice a unităţii de Tâlva Drenii nu poate fi completă fără să menţionăm şi cele două corpuri de dacite microcristaline care se dezvoltă aparent concordant în interiorul gnaiselor liniare. Sub microscop se observă o structură porfirică, fenocristalele de plagioclaz sericitizat, subordonat de cuarţ corodat fiind răspândite într-o masă fundamentală cuarţo-feldspatică sericitoasă. Plagioclazul este maclat după legile Karlsbad, Albit şi Periclin şi prezintă de obicei conture idiomorfe. Unul din corpuri este slab foliat. Considerăm că aceste dacite nu sunt afiliate provinciei banatitice, ci sunt mai vechi, corespunzând probabil cu keratofirele cuarţifere descrise de Codarcea (1930) mai spre vest, în zona Ocna de Fier.

Spre nord, pânza de Tâlva Drenii este tăiată de o falie cu decroşare în lungul căreia vine în contact cu formaţiunile cristaline ale pânzei de Bocşa, situată în mod normal deasupra. Spre est, unitatea de Tâlva Drenii este separată de pânza de Bocşa printr-un plan de încălecare, care datorită unor falieri ulterioare, a fost redresat aproape de verticală. Importanţa acestui plan tectonic este marcată nu numai de deformările mecanice şi cataclazările care îl însoţesc, ci şi de prezenţa unor lentile de cuarţite de origine tectonică situate în interiorul zonelor de laminare aparţinand ambelor unităţi tectonice. Aceste roci prezintă frecvent o structură brecioasă şi culori albicioase-cenuşii. Sub microscop se observă că sunt constituite în proporţie de 90-99 % din cuarţ, restul fiind plagioclaz, muscovit, sericit, carbonat, minerale opace. Plagioclazul este foarte alterat şi tectonizat. Masa de cuarţ este heteroblastică şi formează atat cristale bine dezvoltate, prismatice, deseori cu contur

Institutul Geologic al României

cristalografic (prisme bipiramidale) dispuse neorientat, cât şi mase microcristaline care umplu spaţiile dintre cristalele dezvoltate porfiroblastic. Acestea din urmă pot apare asociate sub formă de glomerule, iar, pe de altă parte, sunt frecvente cazurile în care roca, aparent cu aspect masiv, lasă să se recunoască coexistenţa unui cuarţ granoblastic cu cel microgranoblastic. Toate aceste aspecte sugerează că lentilele de cuarţite s-au format în timpul când planul tectonic era activ, structura internă a lentilelor reflectând modul intermitent de funcţionare a fracturii.

Se ştie că cuarţul se mobilizează foarte uşor în timpul deformărilor care afectează rocile supuse proceselor tectonice plicative sau disjunctive. Asemenea mobilizări de cuarţ se cunosc şi în alte regiuni, cum ar fi de exemplu în lungul faliei Oltului din Carpaţii Meridionali Centrali.

Dacă privim din punct de vedere litologic, constatăm că metamorfitele descrise mai sus prezintă similitudini cu cele cunoscute spre vest în munţii Banatului, la Tâlva Drenii la nord-est de localitatea Bocşa, unde se dezvoltă aceleaşi gnaise oculare şi gnaise liniare, descrise încă de Codarcea (1930), figurate apoi de Iancu (în Russo-Săndulescu et al., 1982) pe foaia 1:50.000 "Bocşa" şi străbătute de intruziunile banatitice de vârstă paleocenă. Fâşii mai înguste din aceeaşi secvenţă, dar cu intercalaţii de gnaise albe cuarţo-feldspatice se întâlnesc şi la sud de Bocşa (Iancu, în Ghenea et al., 1984). Înspre est, în masivul Poiana Ruscă, la est de localitatea Tincova se întâlnesc de asemenea asociaţii litologice foarte asemănătoare cu cele din insula de cristalin gnaise liniare, gnaise oculare şi gnaise albe cuarţo-feldspatice. Aceste roci au fost descrise de către Vasilescu (1975, date nepublicate), Kräutner et al. (1972, 1975, date nepublicate).

De altfel Kräutner a arătat încă din anul 1968 (date nepublicate) că stiva de metamorfite de la Tincova, ce cuprinde gnaise oculare, aparţine unui cristalin de tip Cozia. În concordanţă cu toate aceste observaţii, rocile metamorfice ale pânzei de Tâlva Drenii aparţin, aşa cum au arătat Balintoni et al. (1969), meta-litofaciesului Sebeş-Lotru, grupul (litogrupul) Cumpăna, respectiv formaţiunii (litozonei) Tâlva Drenii-Tincova şi au suferit deci un metamorfism de tip precambrian-barrovian, policiclic în faciesul amfibolitelor cu almandin.

2.3. Pânza de Moniom

Formaţiunile cristaline care alcătuiesc corpul pânzei sunt slab metamorfozate şi suportă o cuvertură sedimentară de varstă cenomanian-coniaciană. Din punct de vedere geometric această unitate se situează în partea bazală a stivei de pânze supragetice. Planul de şariaj a fost redresat aproape de verticală, datorită unor falieri ulterioare. Înspre nord a fost decroşat de o fractură şi mai nouă, orientată nord-vest - sud-est.

Şisturile epimetamorfice afloreză în estul insulei de cristalin şi aparţin litogrupului (seriei de) Leşcoviţa, întrucât se pot recunoaşte cele două litozone (formaţiuni) constituente: "Cârşie" şi "Valea Satului", aşa cum au fost descrise şi datate înspre vest în cuprinsul foii 1:50.000 "Bocşa" (Iancu, în Russo-Săndulescu et al., 1982), sau ulterior de către Iancu (1985).

2.3.1. Formaţiunea (Litozona) Valea Satului (Devonian?-Carbonifer inferior)

Şisturile clorito-albitice ± epidot, adesea cu structură milonitică, alcătuiesc fondul petrografic al acestei litozone în care se dezvoltă apoi nivele de şisturi sericito-cloritoase, cuarţite, mici corpuri de metagabbrouri.

Şisturile clorito-albitice sunt roci de culoare verzuie-cenuşie, microcristaline, extrem de foliate, prezentând frecvent şi o microrubanare tectonică, având deci un caracter de milonite. Sub microscop se observă clorit, plagioclaz, epidot, cuarţ, carbonat, sericit, minerale opace. Cloritul este de regulă constituentul principal al rocii, alcătuind o păslă orientată, în ochiurile căreia apar celelalte minerale. Cu epidotul asocierea este deseori foarte intimă. Cuarţul formează de obicei un component accesoriu sub forma unor cristale orientate paralel cu foliaţia, mai rar litoclaste. Plagioclazul este reprezentat prin albit microgranoblastic, sau prin varietăţi calcice relicte sub formă de cristaloblaste mărunte. Carbonatul, întâlnit frecvent, formează lamine concordante cu foliaţia, sau apare depus pe fisuri oblice sau perpendiculare faţă de aceasta şi este reprezentat atât prin calcit cât şi prin ankerit.

Şisturile sericito-cloritoase formează nivele subţiri, masa rocii fiind constituită dintr-un agregat sericito - cloritos intim concrescut cu albit microporfiroblastic. Caracteristic este hidroxidul de fier, care apare ca o impregnaţie difuză în toată roca.

Metagabbrourile se întâlnesc în diverse stadii de transformare. Astfel, pot fi alcătuite dintr-un agregat microcristalin de clorit, biotit, carbonat, plagioclaz şi minerale opace (magnetit, pirită), roca având o textură masivă. Carbonatul pseudomorfozează frecvent prisme de amfibol. În alte cazuri roca constă practic dintr-un agregat paslos de actinolit, având textura laminară, orientată, cu baghete de amfiboli frecvent recurbate.

Institutul Geologic al României

2.3.2. Formaţiunea (Litozona) Cârşie (Carbonifer inferior)

Litozona Cârşie însoţeşte întotdeauna litozona Valea Satului, cu care este cutată împreună. În alcătuirea ei intră sernifite, gresii şi microconglomerate laminate.

Sernifitele sunt roci cenuşii cu textură orientată evidentă, constituite din cuarţ, feldspat, mice, material opac şi carbonatic. Într-o masă fundamentală microcristalină, având compoziţie cuarţ-albit-sericit-carbonat, apar fragmente clastice de ordinul milimetrilor de cuarţ, albit - tablă de şah, biotit, sau agregate compuse din aceste minerale. Şistozitatea rocii este imprimată de orientarea preferenţială a sericitului şi a cuarţului.

Gresiile şi microconglomeratele, de culoare negricioasă-cenuşie, prezintă un caracter sedimentar evident, dar sunt laminate. Sunt alcătuite din fragmente de şisturi cristaline mezometamorfice, aşezate într-o matrice constituită din acelaşi material, dar fin triturat.

Depozitele sedimentare de cuvertură sunt reprezentate prin marnocalcare sau gresii calcaroase care conţin o bogată microfaună de microforaminifere şi calcisferule specifice intervalului stratigrafic Turonian-Coniacian (Bucur, date inedite), formaţiuni ce capătă o largă dezvoltare mai spre est, în cadrul bazinului Rusca Montană (Bucur, Strutinski, 1982). În munţii Banatului aceste depozite au fost menţionate de Iancu (1985), la est de Ocna de Fier. Sedimentele formează două petice dispuse transgresiv peste litozonele Valea Satului şi Cârşie. Se întâlnesc de asemenea în sudul insulei cristaline Brebu, unde apar intens laminate într-o mică fereastră de sub şisturile cu porfiroblaste de plagioclaz ale pânzei de Bocşa. În vecinătatea peticelor menţionate mai sus s-au găsit şi numeroase blocuri şi fragmente de conglomerate, ce ar putea fi cenomaniene, reprezentând baza succesiunii Cretacicului superior.

O parte din rocile cristaline, care constituie pânza de Moniom - formaţiunea (litozona) Valea Satului, se întâlnesc şi în masivul Poiana Ruscă, la limita lui sudică. Acest fapt a fost scos în evidenţă pentru prima dată de către Strutinski şi Hann (1986), care au arătat că rocile caracteristice acestei litozone reapar în lungul unei falii la sud-vest de localitatea Rusca Montană. Totodată au demonstrat că şisturile pe care Kräutner et al. (1972) le-au descris sub denumirea "seria de Maciova", ce afloreză în sud-vestul masivului pe o suprafaţă foarte restrânsă, reprezintă de asemenea o secvenţă caracteristică formaţiunii (litozonei) Valea Satului din pânza de Moniom.

3. Corelarea structurală cu masivele învecinate

Nici o încercare de corelare a unităţilor suprageticе inferioare din masivul Poiana Ruscă cu cele din munţii Banatului nu poate reuşi decât ţinând cont de informaţiile geologice oferite de insula cristalină Brebu-Zorlencior. Lipsa unor date descriptive exacte a împiedicat pe diferiţii autori de hărţi structurale să traseze în mod corect liniile tectonice care traversează insula. Sperăm ca prin prezenta lucrare să eliminăm acest neajuns. Se mai poate adăuga că pentru prima dată a fost recunoscută ca aflorând în cuprinsul insulei de cristalin şi unitatea de Talva Drenii.

Aşa cum rezultă din capitolul precedent, recunoaşterea şi delimitarea unităţilor structurale din cuprinsul insulei a fost posibilă atât prin identificarea efectelor metamorfismului dinamic care însoţeşte formarea planelor tectonice cât şi datorită observaţiilor privind similitudinea litologică a diferitelor formaţiuni, cristaline sau sedimentare, întâlnite aici, cu cele care afloreză în regiunile învecinate de la vest şi de la est (Fig. 2).

Se observă că unitatea de Bocşa, cu o poziţie superioară, şariază atât peste pânza de Talva Drenii, cât şi peste pânza de Moniom, situaţie evidenţiată în munţii Banatului de către Iancu (1985). Planul de încălecare al pânzei de Bocşa peste pânza de Talva Drenii din masivul Poiana Ruscă a fost figurat de Iancu (1986), Balintoni, Iancu (1986) şi Balintoni et al. (1989). Pânza de Bocşa a fost corelată regional cu pânza de Uria din nordul munţilor Sebeş şi din regiunea văii Oltului din Carpaţii Meridionali (Balintoni et al., 1989), având poziţii spaţiale similare.

Unitatea de Talva Drenii, situată geometric sub pânza de Bocşa a fost evidenţiată de Kräutner et al. (1981) în munţii Banatului, iar de către Iancu (1986) atât în munţii Banatului cât şi în masivul Poiana Ruscă. Pe harta structurală a Carpaţilor Meridionali (Balintoni et al., 1989) ea a fost corelată cu pânza de Măgura din versantul drept al văii Oltului şi nordul masivului Sebeş, după datele lui Hann şi Balintoni (1988).

Pânza de Moniom (Iancu, 1985), bine dezvoltată în munţii Banatului, şi care după Săndulescu (1984) ar reprezenta o digitaţie frontală a pânzei de Locva, reapare spre est, aşa cum au arătat Strutinski şi Hann (1986) şi în sudul masivului Poiana Ruscă, reprezentând pânza supragetică inferioară situată geometric cel mai jos. Cuvertura sedimentară a acestei pânze (cenomanian-coniaciană) este comună în bazinul Rusca Montană atât pentru pânzele de Bocşa şi Talva Drenii cât şi pentru pânza getică, demonstrând astfel realizarea întregului edificiu în tectogeneza mezocretacică (austrică).

Aşa cum au arătat Balintoni et al. (1989), domeniul getic-supragetic este implicat în două pânze de şariaj

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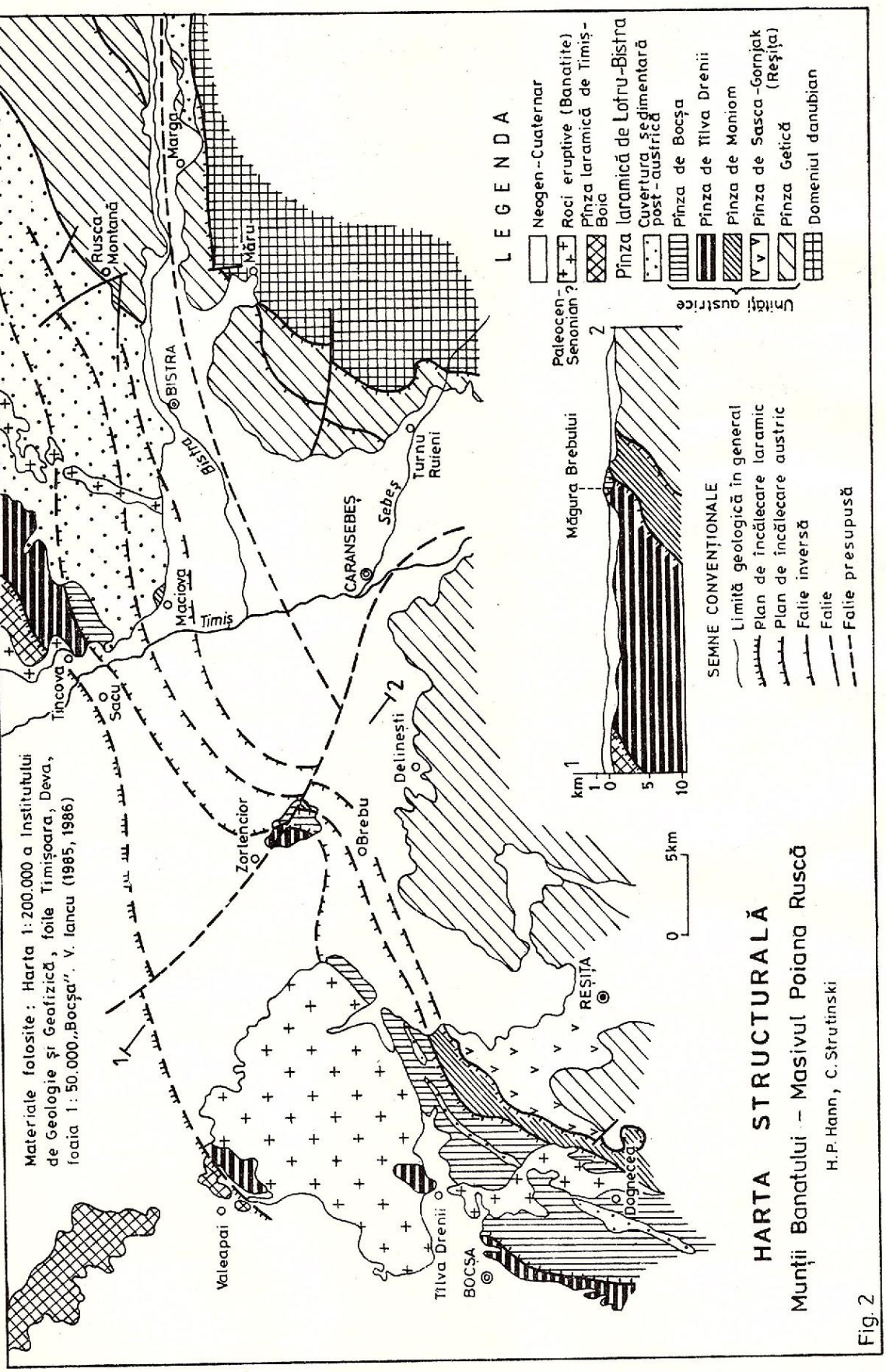

Fig. 2

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majore de vârstă laramică, fiecare din ele conţinând o stivă de pânze formate în tectogeneza paroxismală anterioară, cea austrică. Astfel s-a deosebit pânza de Lotru-Bistra, în poziţie inferioară, şi pânza de Timiş-Boia, în poziţie superioară. Unităţile de Moniom, Tălva Drenii şi Bocşa fac parte, de jos în sus, ca pânze supragetice de soclu, din pânza majoră laramică de Lotru-Bistra.

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Summary

The occurrence of the metamorphic basement beneath the Neogene sedimentary cover in the Măgura Brebului hillock (487 m), near Zorlencior (Banat), permits to better correlate and to link the tectonic units from the Banat Hills with those from the Poiana Ruscă Massif (South Carpathians). Three different units have been delimited within the study area, corresponding to the Bocşa, Talva Drenii and Moniom nappes, initially outlined in the Banat Hills.

The Bocşa Nappe consists of Precambrian medium-grade, partially retromorphosed, metamorphic rocks, unconformably overlain by a sparsely preserved sedimentary cover of mainly calcareous rocks of Eocretaceous (?Middle Jurassic) age. The metamorfic rocks crop out in the central-northern part of the hillock, as well as in the south, where they are flanked tectonically to the west by the rocks of the Talva Drenii Nappe. The rocks are intensely foliated and frequently microfolded. Petrographically they are represented almost

Institutul Geologic al României

exclusively by micaceous plagiogneisses ± garnet. The most distinctive feature of these rocks is the presence of albite porphyroblasts that partially substitute an older generation of microgranular oligoclase.

Petrographically, the Talva Drenii unit (Precambrian) is dominated by a linear gneiss that makes up the bulk of the western and south-western part of the metamorphic outcrop area. It consists of a microgranular assemblage of quartz, feldspar (oligoclase and microcline), mica (biotite and/or muscovite), garnet, apatite, and opaque minerals. Intercalated within the linear gneiss, metric or submetric levels of augengneiss occur, that are characterized by the presence of relatively large (1-2 cm) augen of microcline or quartz. Two dykes of slightly metamorphosed porphyrite (metadacite) almost concordantly crosscut the linear gneisses. Apparently, the Talva Drenii unit lacks a sedimentary cover.

The Moniom Nappe consists of two clearly different metamorphic formations, the Valea Satului, and the Cărşie Formation, respectively. Both are metamorphosed under greenschist facies conditions.

The Valea Satului Formation (?Devonian - Lower Carboniferous) is mainly represented by chlorito-albitic schists ± epidote that include small lenticular bodies of highly transformed metagabbros. Beside this dominantly basic assemblage, epiclastic quartzites and chlorito-sericitic schists are subordinately present. The Cărşie Formation (Lower Carboniferous) consists of laminar sernifites and less foliated meta-sandstones and -microconglomerates with well-preserved sedimentary structures. The sedimentary cover of the Moniom Nappe is represented by Upper Cretaceous (Turonian-Coniacian) micropalaeontologically dated marls and calcareous sandstones.

The three petrographically distinct units that have been outlined within the Măgura hillock have their equivalents both to the west, in the Banat Hills, and to the east, in the Poiana Ruscă Massif. These units are considered to represent basement nappes of Austrian age, making up a tectonic pile that acted as a unitary (Supragetic, Lotru-Bistra) nappe during the Laramian orogenic movements.

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Institutul Geologic al României

Institutul Geologic al României

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LES SCHISTES CRISTALLINS DES SÉRIES DE BIHARIA ET DE MUNCEL DANS LE BIHOR MÉRIDIONAL

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Key words: Metamorphic rocks. Tectonic units. Stratigraphic units.Nappes. Apuseni Mountains.

Abstract: *Crystalline Schists of the Biharia and Muncel Series in the South Bihor.* The Biharia Series of the Biharia Nappe is made up of an Ophiolitic Formation and a Tufogene-Carbonatic Formation. The Muncel Series of the Muncel Nappe is divided into: Necşeşti Formation (chlorite and sericite schists with albite), Sericite-schists and Gneiss Formation and Porphyroid and Graphite-quartzite Formation. The Muncel Nappe is overthrust northwards on the Biharia Nappe, the overthrust having a post-erosional character.

Les travaux géologiques entrepris entre 1983 et 1987 pour la rédaction de la feuille géologique à l'échelle de 1 : 50.000 Biharia (Bordea et al., 1988) nous ont fourni quelques données concernant les séries de Biharia et de Muncel du Bihor méridional. Les schistes cristallins de ces séries appartiennent à deux nappes, connues depuis longtemps, les Nappes de Biharia et de Muncel.

1. Nappe de Biharia

Série de Biharia

La série de Biharia, bien individualisée comme entité lithostratigraphique, n'était pas encore subdivisée il y a vingt ans (Pálfy, Rozlozsnik, 1939; Giuşcă, 1937; Dimitrescu, 1958; Ionescu, 1962; Marza, 1969). Le premier auteur (dans la littérature publiée) qui a proposé une subdivision de cette série a été Ionescu (1970, 1971), qui sépare dans sa thèse de doctorat deux "complexes". Dans le cadre du "complexe" inférieur, il distingue un premier horizon comportant des schistes albitiques à chlorite et muscovite, dans lesquels se trouvent, de manière subordonnées, des intercalations de schistes chloriteux à albite et des schistes muscovitiques, suivies par un second horizon de schistes chloriteux à porphyroblastes d'albite.

Dans le cadre du "complexe" supérieur, Ionescu distingue un troisième horizon (dans la numérotation d'ensemble) formé par des schistes albitiques à minces intercalations de quartzites feldspatiques et des schistes chloriteux à porphyroblastes d'albite et un quatrième horizon où prédominent les schistes chloriteux à al bite qui admettent des intercalations de gneiss albitiques, de calcaires dolomitiques et de chloritoschistes à calcite.

Une autre subdivision lithostratigraphique a été réalisée par Lazăr, Berbeleac et Ştefan (1980). Le complexe inférieur (600 m) de la série de Biharia serait formé par des schistes quartzeux à albite, à minces intercalations de schistes chloriteux à porphyroblastes d'albite, de schistes à chlorite, albite, épidote et actinolite et de schistes amphiboliques.

Le complexe moyen (120–150 m) serait représenté par des schistes chloriteux à porphyroblastes d'albite, à intercalations de calcaires dolomitiques et de schistes carbonatiques - quartzeux à chlorite. Le complexe supérieur (300 m) serait constitué par des schistes chloriteux à porphyroblastes d'albite admettant des intercalations de schistes muscovitiques à porphyroblastes d'albite et de schistes actinolitiques à épidote.

Nous ajoutons que dans la partie sud-est du Gilău, Solomon et al. (1984) séparent, dans la série de Biharia, un complexe inférieur à métagrauwackes, un complexe moyen à schistes tufogènes basiques et un complexe supérieur à schistes quartzeux à séricite; ce dernier représente à notre avis (Dimitrescu, 1988), la formation de Necşeşti de la série de Muncel.

En ce qui concerne les intrusions qui traversent la série de Biharia, initialement n'ont été reconnus que des corps d'orthoamphibolites (Pálfy, Rozlozsnik, 1939; Giuşcă, 1937; Dimitrescu, 1958; Ionescu, 1962, 1970, 1971), représentant des métagabbros et

Institutul Geologic al României

des métadiorites. Mârza (1969) est le premier auteur qui attribue à une partie des gneiss albitiques du sud-est du Gilău le caractère de métagranites trondhjemitiques; ultérieurement dans la même région, Solomon et al. (1984) reconnait des granitoïdes gneissiques. Il est suivi par Balintoni (1985) et Hartopanu et al. (1982). Dans la partie méridionale du Bihor Lazăr et al. (1980) montrent aussi que la série de Biharia est traversée par des corps de métagabbrodiorites et de métagranites.

Remarquons aussi qu'on a parlé dans le passé du caractère "géosyclinal" et "ophiolitique" de la série de Biharia (Ianovici et al., 1976). Lazăr et al. (1980) précisant même qu'il serait question d'une formation de spilites-kératophyres.

A la suite de mes travaux, je suis arrivé à la conclusion qu'il suffit de subdiviser la série de Biharia en deux formations, la limite entre celles-ci étant tracée, de manière conventionelle, un peu au-dessous du premier niveau de calcaires dolomitiques.

1. La formation ophiolitique comprend les trois premiers horizons de Ionescu (1970, 1971), le complexe inférieur de Lazăr et al. (1980) et, probablement, la plus grande partie du complexe inférieur de Solomon et al. (1984). Cette formation comporte des schistes chloriteux à porphyroblastes d'albite (provenant de roches spilitiques, mais peut-etre aussi de grauwackes feldspatiques). Elle est traversée par de nombreux corps d'orthoamphibolites (provenant avec certitude de gabbros, de gabbrodiorites et de dolérites), alternant avec des corps de gneiss albitiques (représentant des métatonalites-métaplagiogranites). Elle se trouverait donc dans la partie supérieure du profil ophiolitique typique, notamment dans le complexe des gabbros stratifiés, connus pour être associés localement à des corps de plagiogranites - tonalites. On pourrait aussi considérer cette formation comme un complexe leptyno-amphibolique (CLA). Il est vrai qu'une petite partie des roches albitiques représentent des mtagrauwackes (Solomon et al., 1984), le fait résultant de la comparaison microscopique avec les métaplagiogranites. Ces schistes albitiques ne sont en effet que des schistes toujours à porphyroblastes, dans lesquels ces derniers constituent presque entièrement la roche, au détriment de la partie micacée - chloriteuse et en l'absence presque totale du quartz.

2. La formation tufogène-carbonatique de la série de Biharia est constituée toujours par des schistes chloriteux à porphyroblastes d'albite admettant, surtout vers sa base, des intercalations de calcaires dolomitiques et de schistes carbonatiques à chlorite, suivies par des amphibolites, des schistes albitiques et des schistes à chlorite, albite et épidote. L'association métagabbros-métaplagiogranites manque. Cette formation provient surtout de roches tufogènes basiques, parfois intermédiaires (kératophyres) et acides (ici entreraient éventuellement les "métaplagiovolcanites" Mârza, 1969, du Gilău de sud-est), à de faibles intercalations de roches sédimentogènes carbonatiques (± pélitiques ?). La formation tufogène-carbonatique correspond au quatrième horizon de Ionescu (1970, 1971), aux complexes moyen et supérieur de Lazăr et al. (1980) et au complexe moyen de Solomon et al. (1984).

Un problème surgit à propos d'un supposé niveau de calcaires cristallins, apparaissant à la base de la formation supérieure de la série de Biharia, qui comprendrait toutes les roches carbonatiques connues, niveau dont l'existence est soutenue par Lazăr et al. (1980) et par d'autres géologues. Je ne considère pas l'existence de ce niveau constant comme démontrée, et c'est pourquoi nous n'avons pas construit une tectonique de détail en partant du postulat que, partout où apparaît une roche carbonatées celle-ci se trouverait à la base de la formation citée. C'est aussi le motif pour lequel nous avons divisé la série de Biharia en deux et non pas en trois formations.

L'age de la série de Biharia est très probablement vendéen (Protérozoïque terminal), ce qui résulte des données palynologiques sur la série elle-meme, ainsi que sur la formation suprajacente de Necşeşti (Visarion, 1970; Visarion, Dimitrescu, 1971; Solomon et al., 1984; Dimitrescu in Zoubek, 1988).

La série de Biharia est polymétamorphique (Balintoni, 1985). Son premier métamorphisme, plus intense dans la formation ophiolitique, ne paraît pas avoir toutefois dépassé le faciès des amphibolites à épidote et albite; il existe, en tout cas, des reliques de grenat et de biotite chloritisés. Le second métamorphisme, à caractère rétrograde, correspond à l'isograde de la chlorite. Dans la formation tufogène-carbonatique nous n'avons pas trouvé des minéraux reliques, ce qui pourrait suggérer soit l'intensité plus élevée de la diaphtorèse, soit l'intensité plus basse du premier métamorphisme. La série de Biharia est recouverte en discordance par la formation des métaconglomérats de la série carbonifère de Păiuşeni.

2. Nappe de Muncel

A) Série de Muncel

Il y a 20 ans la série de Muncel n'était pas encore subdivisée (Pálfy, Rozlozsnik, 1939; Giuşcă, 1937; Ionescu, 1962). La première subdivision a été réalisée toujours par Ionescu (1970, 1971), qui sépare:

- un niveau inférieur à gneiss albitiques, schistes chloriteux à albite, épidote et actinolite, schistes chloriteux á porphyroblastes d'albite, dolomies cristallines;
- un niveau moyen, dans la composition duquel entrent des schistes sériciteux, parfois à chlorite ou à graphite, à intercalations de calcaires cristallins;

Institutul Geologic al României

– un niveau supérieur, constitué par des porphyroïdes (métadacites) et des métagranites.

Dans la même région, Lazăr et al. (1980) séparent, dans la série de Muncel, deux complexes. Le complexe inférieur (500 m) est formé par des schistes quartzeux à séricite et des schistes sériciteux à intercalations de porphyroïdes, d'épigranites à lentilles de calcaires cristallins. Le complexe supérieur (2000 m) est constitué par des schistes graphiteux alternant avec des quartzites graphiteux, des schistes quartzeux à séricite ± feldspath, sa partie supérieure admettant des intercalations de porphyroïdes.

Dans une région située immédiatement à l'est, nous avons séparé, dans la même série (Dimitrescu, Ioachim, 1972; Dimitrescu et al., 1974; 1977; Ianovici et al., 1976), trois membres:

– le membre de Necşeşti (Dimitrescu, 1988): schistes séricito-chloriteux alternant avec des schistes sériciteux à albite et des schistes chloriteux (± albite), à de rares intercalations de porphyroïdes métakératophyriques;

– le membre moyen: schistes sériciteux, à intercalations de porphyroïdes métarhyolitiques et de gneiss oeillés;

– le membre supérieur: quartzites graphiteux, schistes sériciteux à biotite, paragneiss fins à biotite, schistes amphiboliques, porphyroïdes métarhyolitiques à biotite.

Le membre de Necşeşti correspond à ce que Borcoş (1962), Márza (1969) et Solomon et al. (1984) ont séparé comme le complexe supérieur de la série de Biharia dans le sud-est du Gilău, attribué aussi par Balintoni (1985) et Hartopanu et al. (1982) à la même série, mais sans être distingué du reste de celle-ci.

À la suite de nos levers de terrain dans le Bihor méridional, nous sommes arrivés à la conclusion que la série de Muncel peut être divisée en trois formations:

1) La formation de Necşeşti, apparaît comme une bande mince à la base de la Nappe de Muncel à l'ouest, au nord et à l'est de la crête de Muncel-Romana, au sud de la vallée de Plămani (zone de la mine de Brusturi), au nord et nord-est du sommet de Dolea ainsi que le long de la vallée supérieure de Răul Mic. Partout cette formation avait été attribuée à la série de Biharia, apparaissent à la base de la nappe de Muncel (Giuşcă, 1937; Ionescu, 1962; Giuşcă et al., 1973; Lazăr et al., 1980; Berbeleac in Borcoş et al., 1983); Ionescu (1970, 1971) avait attribué la formation à la série de Muncel, mais ensemble avec des roches appartenant à la série de Biharia.

2) La formation des schistes sériciteux et des gneiss comporte un niveau de schistes sériciteux et de phyllonites, toutes les intercalations de mylonites granitiques et de gneiss oeillés de la zone, tout comme quelques intercalations de porphyroïdes métarhyolitiques et métadacitiques. Cette formation correspond aux niveaux moyen et supérieur de Ionescu (1970, 1971) et au complexe inférieur de Lazăr et al. (1980). Les calcaires cristallins mentionnés par ces auteurs appartiennent, en fait, à notre avis, à la formation tufogène-carbonatique de la série de Biharia, en constituant des klippes de rabotage à la base de la nappe de Muncel; ceci résulte de leur association avec des lames minces de schistes chloriteux à albite, par éxemple à Ciungi (affluent droit de la vallée de Dolea-Brusturi).

3) La formation des porphyroïdes et des quartzites graphiteux correspond au complexe supérieur de Lazăr et al. (1980) et comporte une masse principale de porphyroïdes métarhyolitiques et des métatufs acides, à intercalations de schistes sériciteux et graphiteux.

Les données palynologiques obtenues plus à l'est indiqueraient les suivants âges pour les trois formations:

a) la formation de Necşeşti appartient au Cambrien inférieur (Visarion, 1970; Visarion, Dimitrescu, 1971; Solomon et al., 1984), pouvant, d'après les derniers auteurs cités, comprendre aussi la base du Cambrien moyen;

b) la formation des schistes sériciteux et des gneiss reviendrait au Cambrien moyen;

c) la formation des porphyroïdes et des quartzites graphiteux se place dans le Cambrien supérieur, en passant probablement aussi dans l'Ordovicien (Visarion, Dimitrescu, 1971; Dimitrescu in Zoubek, 1988).

En ce qui concerne le magmatisme acide associé à la série de Muncel, on peut distinguer deux types de roches:

a) Les porphyroïdes ont été caractérisés par Ionescu (1962, 1970, 1971) comme étant des métarhyolites et des métadacites; elles sont accompagnées par des métatufs acides (Lazăr et al., 1980). Nous les avons considérées comme ignimbrites métamorphisées (Dimitrescu, 1973). Nous disposons d'une seule analyse chimique sur les porphyroïdes (Pálfy, Rozlozsnik, 1939). Ce volcanisme acide d'âge cambrien, connu aussi plus á l'est (Dimitrescu, 1958) est corrélable avec celui de la série d'Arada (Dimitrescu, 1966) et avec celui du groupe de Tulgheş (Carpathes Orientales).

b) Dans la formation médiane de la série de Muncel s'intercalent des roches dénommées gneiss oeillées par Pálfy, Rozlozsnik (1939) et Dimitrescu (1958, 1973), épigranites par Giuşcă (1937) et Lazăr et al. (1980), métagranites ou granites gneissiques cataclasés à microclin, par Ionescu (1962, 1970, 1971) et métaporphyres granitiques, par Giuşcă et al. (1967) et Ianovici et al. (1968).

Dans le passé, nous avons soutenu (Dimitrescu, 1973) que ces roches représenteraient des métaconglomérats arkosiens à ciment surtout quartzeux, intercalés dans une formation volcano-sédimentaire.

Institutul Geologic al României

Tâchons de résumer les principaux caractères du corps de roches en discussion. Le "contrôle stratigraphique" de la manière d'apparition est claire; les gneiss en cause s'intercalent comme un corps unique entre des schistes sériciteux (à très rares porphyroïdes) dans son mur et des schistes sériciteux à porphyroïdes plus abondantes dans son toit, en constituant dans leur ensemble ce que nous avons séparé comme formation des schistes sériciteux et des gneiss. La continuité du corps de gneiss, de puissance comprise entre 30 et 300 m, est presque ininterrompue depuis Luncşoara de Sus-Brusturi de l'ouest, jusqu'à Mihoeşti (Câmpeni) à l'est, c'est-à-dire sur environ 40 km. Ce corps orienté EW est plissé de manière large, ensemble avec les formations avoisinantes. Dans la vallée de Cioroasa (Avram Iancu) nous avons trouvé, dans ces gneiss, trois intercalations centimétriques de calcaires cristallins. Une analyse chimique d'un gneiss de Avram Iancu a donné des teneurs en SiO_2 et Al_2O_3 trop élevées pour une roche éruptive, les paramètres Niggli s'encadrant dans le champ des roches détritiques.

En ce qui concerne les caractères microscopiques des gneiss, nous avons affirmé que leur structure suggère une origine blastodétritique, mais avec une déformation postcristalline évidente.

A présent nous sommes convaincus que ces gneiss, au moins ceux dépourvus de caractère oeillé du Râul Mic (Bihor méridional) sont des mylonites granitiques. On pourrait formuler avec précaution l'hypotèse d'un corps stratoïde concordant de granites et de porphyres granitiques, à caractère hypabyssal, mise en place dans une formation sédimentaire-volcanogène acide, à puissantes intercalations de tufs soudés. La supposition d'une coexistence avec des niveaux de conglomérats arkosiens pourrait conduire à l'acception d'un phénomène de convergence, voie facile de ne résoudre qu'en apparence les problèmes difficiles. Il est maintenant connu que des structures cataclastiques ont été confondues avec des structures (blasto-) détritiques. Les intercalations de calcaires cristallins pourraient avoir été entraînées, par voie tectonique, à l'intérieur des gneiss, ou pourraient avoir été formées par différenciation métamorphique.

La série de Muncel est certainement rétromorphe, à degré initial de métamorphisme qui n'a pas dépassé probablement l'isograde du grenat (nous avons observé des reliques chloritisés de ce minéral) du faciès des amphibolites à épidote et albite; dans des conditions d'intensité plus élevée, les délicates structures porphyriques reliques (blastoporphyriques) des porphyroïdes n'auraient pas pu se conserver. Nous ne pouvons pas apprécier avec certitude en ce moment si entre la mise en place des magmatites et le premier métamorphisme de la série de Muncel, d'une part, et le métamorphisme dynamique généré par le charriage de la nappe de Muncel, d'autre part, une phase de rétrogréssion serait intervenue. Ce que nous voulons souligner c'est le caractère mylonitique des anciennes roches granitiques et le caractère phyllonitique d'une grande partie des schistes de la série de Muncel du Bihor méridional, ces caractères s'atténuant beaucoup vers l'est. L'aspect des "taches" chloriteuses, observables sur les foliations des schistes et des phyllonites, suggère leur provenance de paragneiss ou de micaschistes.

Un problème qui se pose à propos de la série de Muncel est la corrélation des formations moyenne et supérieure de celle-ci avec la série de Baia de Arieş, corrélation soutenue par Balintoni (1985). En essence, l'argumentation de l'auteur (qui a tenu à affirmer son point de vue sur la feuille à 1 : 50 000 Valea Ierii, Hărtopanu et al., 1982) repose sur la présence dans la série de Baia de Arieş des gneiss oeillés de "type Mihoeşti" et des quartzites graphiteux, ainsi que sur le caractère rétromorphe des schistes sériciteux et séricito-chloriteux (à grenat et biotite relique) de la série de Muncel.

L'étude lithostratigraphique et métamorphique de la série de Baia de Arieş a été approfondie (Iancu, Balintoni, 1986; les mêmes auteurs de la feuille à 1 : 50 000 Poşaga). La série a été subdivisée, la formation inférieure de Sălciua étant suivie par la formation supérieure de Cioara. A son tour, dans la formation de Salciua, ces auteurs ont distingué le terme inférieur des plagiogneiss micacés de Stremînoasa, à intercalations d'amphibolites et de gneiss quartzo-feldspathiques nodulaires ("de Mihoeşti") et un terme supérieur des micaschistes à almandin, staurotide ± disthène, à intercalations d'amphibolites, de calcaires cristallins et de quartzites blancs et noirs (graphiteux). Dans la formation de Cioara on a séparé les plagiogneiss à biotite ± muscovite ± almandin ± microclin de Valea Cioarei et les micaschistes quartzeux supérieurs de Valea Matrii, à intercalations d'amphibolites.

De la comparaison de ces données de détail avec nos propres données concernant la série de Muncel, on peut observer qu'il n'existe que peu d'éléments communs. La subdivision lithostratigraphique est totalement différente. Pour la série de Baia de Arieş, on remarque, dans son contenu pétrographique, des intercalations abondantes et caractéristiques de roches carbonatées, d'amphibolites et de quartzites blancs et noirs dans un fond métapélitique (Balintoni, Iancu, 1986 a); par contre, dans les deux formations supérieures de la série de Muncel, les roches carbonatées et les quartzites blancs font défaut, tandis que les amphibolites n'apparaissent qu'en quantités infimes et seulement dans la zone de Bistra-Lupşa (Dimitrescu et al., 1974). En échange, les porphyroïdes

Institutul Geologic al României

(métavolcanites acides), très caractéristiques pour la série de Muncel, sont absentes ou presque de la série de Baia de Arieş. Les quartzites graphiteux se rencontrent, il est vrai, dans les deux séries, mais à eux seuls ils ne peuvent pas soutenir le parallélisme (il y en a aussi, par exemple, dans les séries de Someş, de Tulgheş, de Chiril, etc.).

Quant aux mylonites granitiques - gneiss oeillés de la série de Muncel, ils disparaissent à l'est de Mihoeşti. Sur la feuille Poşaga les "gneiss de Mihoeşti" (comme les appellent I. Balintoni et V. Iancu) ne sont figurés que dans la vallée de Poşaga et immédiatement au sud et au nord de celle-ci, sous la forme de quatre minces lentilles à la base de la nappe de Baia de Arieş (ces lentilles sont intercalées dans les micaschistes supérieurs de la formation de Sălciua et non pas dans les plagiogneiss inférieurs de Streminoasa, comme ils résulterait de la légende de la feuille de Poşaga). Une de ces lentilles avait été figurée sur la carte de Mârza (1969), mais comme porphyroïde. Le parallelisme entre ces roches et "les gneiss de Mihoeşti" ne nous paraît pas en dehors de tout soupçon et, en tout cas, il ne peut pas conduire à la corrélation des séries cristallines respectives.

D'autre part, les mêmes auteurs (Iancu, Balintoni, 1986) réalisent une étude détaillée de la succession des paragénèses, dans la série de Baia Arieş, en distinguant trois étapes: un premier métamorphisme barrovien M_1 en faciès des amphibolites à almandin, zone à staurotide ± disthène (rarement sillimanite); un second métamorphisme régional M_2, isofacial avec le premier; un réthromorphisme régional M_3, à formation de muscovite, chlorite, albite et épidote. Or entre les histoires métamorphiques des séries de Muncel et de Baia de Arieş, le seul point commun est la diaphtorèse, qui ne peut pas constituer evidemment un critère de corrélation lithostratigraphique. D'ailleurs, à propos des comparaisons des histoires métamorphiques, nous contestons le principe admis par Balintoni, Iancu (1986 b) d'après lequel "le nombre des générations minérales est tout au plus égal au nombre d'orogenèses qui ont affecté la roche". Par "orogenèses", les auteurs susmentionnés comprennent les cycles orogéniques, et si, par exemple, une paragenèse est démontrée comme hercynienne, ils attribuent automatiquement la précédente comme calédonienne ou cadomienne. Or il est bien connu et vérifié qu'un cycle orogénique comprend plusieurs phases tectoniques-métamorphiques: le cycle alpin, dans les Alpes, de deux à trois. On ne peut pas admettre à "priori" que dans les Carpathes les phénomènes se déroulent autrement, comme implicitement le postulent les auteurs cités et d'autres qui les imitent.

En conclusion, la corrélation de la partie supérieure de la série de Muncel avec la série de Baia de Arieş, soutenue par Balintoni (1985), ne repose pas sur des arguments irréfutables.

B) Tectonique

Dans la série de Muncel, dans les schistes et dans les phyllonites, la structure mésoscopique est dominée par une schistosité, numérotée provisoirement S_2, qui a un caractère mylonitique prononcé. Cette schistosité provoque la transposition totale d'une foliation plus ancienne S_1, le phénomène ne pouvant être décelé que par endroits.

D'autre part, dans les porphyroïdes et dans les corps de gneiss, on ne peut observer que la foliation S_2. Très rarement (à l'ouest du sommet de Dolea), dans les schistes sériciteux, on peut observer en plus une foliation moins pénétrative S_3.

Les linéations minérales et les crénulations sont orientées NW-SE. Les foliations S_2, ainsi que les linéations inclinent presque exclusivement vers le sud dans toute la nappe de Muncel.

La plupart des auteurs qui ont traité de la tectonique du Bihor méridional (Rozlozsnik, 1935; Giuşcă, 1937; Giuşcă et al., 1937; Ionescu, 1962, 1970, 1971; Lazăr et al., 1980) sont d'accord sur le fait que la nappe de Muncel, constituée en totalité ou en grande partie par la série de Muncel, est charriée vers le nord sur des différents termes de la nappe de Biharia, étant l'unité tectonique à la position la plus élevée de la région.

Giuşcă (1937) a considéré la nappe de Muncel comme n'étant constituée que par ce que nous appelons à présent la formation des schistes sériciteux et des gneiss, et la formation des porphyroïdes et des quartzites graphiteux. Ce qu'on appelle aujourd'hui la formation de Necşeşti a été attribuée par cete auteur et par d'autres qui lui ont succédé, à la série de Biharia. Dans ce sens, la série de Biharia apparaissait, soit en fenetre tectonique (Raul Mare, Raul Mic, Valea Plămanilor), soit comme une mince bordure rabotée à la base de la nappe de Muncel (à l'ouest, au nord et à l'est de la crête de Muncel - Romana).

Ionescu (1962) a considéré la formation de Necşeşti comme faisant partie de la série de Biharia; mais en admettant, comme nous mêmes (Dimitrescu, 1958, 1966), la continuité entre les séries de Biharia et celle de Muncel (qui n'avaient pas été encore subdivisées à cette date-là), l'auteur n'a plus reconnu des fenêtres ou des klippes de rabotage au sud de la ligne frontale de charriage. La nappe de Muncel a une extension plus large que celle donnée par Giuşcă (1937), en comprenant des termes des deux séries cristallines.

Berbeleac (in Giuşcă et al., 1973; Lazăr et al., 1980; Cioflică et al., 1982; Borcoş et al., 1983) reconnait de nouveau, sur la base d' études plus approfondies, les trois fenetres mentionnées. De rabotage il n'en est

Institutul Geologic al României

pas question; autour de la crête de Muncel-Romăna, à la base de la nappe de Muncel, une bande mince de terrains est de nouveau attribuée à la série de Biharia (en fait, la formation de Necşeşti), en position normale sous la série de Muncel.

Au cours de nos travaux de terrain, nous avons constaté que les fenêtres de Valea Plămânilor et de Râul Mare-Râul Mic aval existent exactement comme elles avait été reconnues par Giuşcă (1937), Giuşcă et al. (1973), Cioflică et al. (1982), Borcoş et al. (1983).

Dans la première, la série de Biharia de la nappe de Biharia est recouverte en discordance par une bande mince de la série carbonifère de Păiuşeni (fait qui a échappé à Ionescu, 1962, 1971, mais pas aux autres auteurs); la nappe de Muncel, recouvre celle-ci.

Dans la fenêtre de Raul Mare-Râul Mic aval, la série de Biharia de la nappe apparaît directement au-dessous de la formation des schistes sériciteux et des gneiss de la nappe de Muncel, sans l'intermédiaire de la série de Păiuşeni et en l'absence de la formation de Necşeşti de la nappe de Muncel. En échange la fenêtre de Râul Mic des anciens auteurs n'existe que dans les portions aval et amont extrême; dans la portion immédiatement en amont de la mine, apparaît la formation de Necşeşti, reposant normalement sous le reste de la série de Muncel.

La base de la nappe de Muncel est jalonnée par de nombreuses klippes de rabotage de calcaires dolomitiques cristallins de la formation tufogène-carbonatée (série de Biharia), accompagnées parfois par des lames réduites de schistes à chlorite et albite. Ces calcaires ont été attribués par les autres auteurs, soit à la série de Păiuşeni, soit à la série de Muncel.

Il n'est pas exclu que certaines lames de schistes chloriteux à albite, attribuées par nous à la formation de Necşeşti de la nappe de Muncel, autour de la crête de Muncel - Romana, soient en réalité toujours des klippes de rabotage, provenant de la série de Biharia; peut-être n'ai-je pas été toujours en mésure de les séparer.

La nappe de Muncel a ainsi un caractère post-érosif (épiglyptique); elle recouvre, soit la série de Păiuşeni, soit directement des divers termes de la série de Biharia. Grâce à la troncature basale on trouve à la base de la nappe de Muncel, soit la formation des schistes sériciteux et des gneiss, soit la formation de Necşeşti, soit des klippes de rabotage comportant des lames de la série de Biharia. Les caractères, en bonne partie mylonitiques et phyllonitiques des roches, suggèrent la présence, vers la base de la nappe de Muncel, d'une ceinture de cisaillement.

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Strand, T. (1972) The Norwegian Caledonides. p. 1–20. In: Kulling, O., Strand, T. (eds.) Scandinavian Caledonides, 560 p., Interscience Publishers.

c) books:

Bălan, M. (1976) Zăcămintele manganifere de la Iacobeni. Ed. Acad. Rom., 132 p., Bucureşti.

d) maps:

Ionescu, I., Popescu, P., Georgescu, G. (1990) Geological Map of Romania, scale 1:50,000, sheet Cîmpulung. Inst. Geol. Geofiz., Bucureşti.

e) unpublished papers or reports:

Dumitrescu, D., Ionescu, I., Moldoveanu, M. (1987) Report. Arch. Inst. Geol. Geofiz., Bucureşti.

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A Comparative Study of the Distribution of Some Major, Trace and Radioactive Elements in Several Paleogene and Upper Pliocene–Quaternary Basalts and Basalt–Andesites from Romania *Savu H., Tiepac I., Grabari G.* 91
Upper Pliocene to Quaternary Basaltic Volcanism in the Perşani Mountains *Seghedi I., Szakács Al.* 101
Petrology, Geochemistry and Origin of the Pleistocene Withinplate Basalts of Lucaret (Banat – Romania) *Savu H., Udrescu C., Stoian M., Tiepac I.* 109
Lithostratigraphic and Metamorphic Correlations: A Tentative Way of Exploring the Early History of the Getic Crystalline *Săbău G.* 119
Reactivated Metamorphic Complexes in Fold and Overthrust Belts (e.g. South Carpathians) *Iancu V., Mărunţiu M.* 129
Gneiss Domes in the Sebeş–Lotru Series, Sebeş Mountains *Stelea I.* 143
Insula de cristalin de la Brebu–Zorlencior, o verigă de legătură între pânzele supragetice din munţii Banatului şi masivul Poiana Ruscă (Carpaţii Meridionali vestici) *Hann H. P., Strutinski C.* 149
Les schistes cristallins des séries de Biharia et de Muncel dans le Bihor méridional *Dimitrescu R.* 159

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