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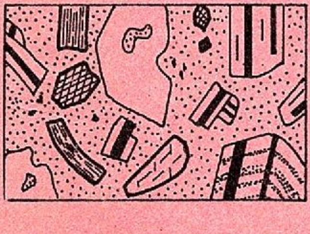

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The **Romanian Journal of Petrology** (Rom. J. Petrology) is now at its first volume in the new form. However, the publication goes back to 1910, as the first volume of the "Dări de seamă ale Şedinţelor" (D.S.) has appeared as proceedings of geologists working with the Geological Institute of Romania. The journal (D.S.) appeared initially as a single volume (till volume 54, 1969), then with five series, the present issue being a direct continuation of the D.S./series 1(Mineralogy–Petrology).

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PETROLOGY OF THE SICHEVIŢA GRANITOIDS (SOUTH CARPATHIANS)

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Key words: Granitoids. Major elements. Minor elements. Petrology. South Carpathians – Getic and Supragetic crystalline – Western Almăj Mts.

Résumé: *Pétrochimie du granodiorite de Sicheviţa (Carpathes Méridionales).* Le granitoïde de Sicheviţa comporte des granites (monzogranites, syénogranites), granodiorites, tonalites, diorites quartzifères et trondhjémites. Ces entités sont traversées par des filons de lamprophyres, diorite-porphyres, aplites, aplites granitiques et pegmatites. Il y a une bonne corrélation pour la classification pétrologique et chimique, illustrées dans les diagrammes QAP et TAS (fig. 1, 2). Le granitoïde est calco-alcalin (fig. 3). Les granites sont potassiques, les tonalites, les trondhjémites et la plupart des granodiorites ayant des caractères sodiques (fig. 4). Pendant le processus de différentiation les valeurs diminuent pour Sc, Co, Cr, Mg, Ni, Fe_{tot}, V, Ca; la valeur K augmente et les valeurs Sr et Na restent presque constantes. Les éléments radioactifs sont présents en quantités réduites (U, Th, Rb). Sur base de six mesures radiométriques (tab. 4) et des rapports réciproques granite/schistes cristallins adjacents, on peut considérer l'âge hercynien, possiblement calédonien du granitoïde de Sicheviţa.

Introduction

The Sicheviţa granitoid is located in the south-western area of the Almăj Mts (Banat, Caraş-Severin district).

The granitoid rocks trending north-southwards occur along a 25 km strip. The maximum width of about 8 km is at the Sicheviţa village. In the south, along the Danube Valley, these rocks are exposed over 6 km width. In the north, between the villages Gîrnic and Şopotul Nou, the granitoid crops out over 2-3 km width.

The geologic structure of the Sicheviţa-Şopotul Nou area belongs to the Getic tectonic unit. The metamorphic basement includes the Sebeş-Lotru Group and the Miniş and Buceava formations (Savu, 1979; Iancu, Mărunţiu, 1989).

In the eastern part, the Sicheviţa granitoid crosses the crystalline schists which are metamorphosed in the contact area; in the west, it is transgressively overlain by the Mesozoic sedimentary formations from Reşiţa-Moldova Nouă area (Năstăseanu et al., 1981).

The Miocene sedimentary formations overlie transgressively and unconformably both the granitoid rocks and the crystalline schists.

Previous Studies

The Sicheviţa granitoid and the crystalline schists occurring in the south-western Banat were cited by Böckh, Schereter and Schafarzik in the second half of the 19-th century and in the first decades of the 20-th.

According to Streckeisen (1934) the Liubcova (Sicheviţa) granitoid is younger than the crystalline schists crossed by it.

Codarcea (1940) changed the name of the Liubcova granitoid to the Sicheviţa granitoid. In his opinion, this pluton consists of granite-granodioritic rocks of different types: aplites, aplite-granites, abunding in potash feldspar, coarse-grained granites with orthose phenocrysts, granodiorites and biotite quartz diorites.

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Constantinof and Focşa (1959, unpubl. report) assigned to this pluton the following rock types: tonalites, granodiorites and granites.

Rare elements of the Sicheviţa granitoid massif were investigated by Bîrlea et al. (1963, unpubl. report), Bîrlea and Bîrlea (1965, unpubl. report).

The microtectonic studies of Bercia and Bercia (1968, unpubl. report) pointed to relevant unconformity between the trend of primary foliations of granitoid rocks and the trend of foliations of the crystalline schists crossed by granites.

A synthesis study of geological prospectings for rare elements in the Sicheviţa massif belongs to Bîrlea (1970, unpubl. report) who also elaborated a comprehensive study of accessory minerals from the same area (1975, unpubl. thesis of doctor's degree) .

The metallogenetic potential of granitoids has been recently investigated by Stan et al. (1986, unpubl. report).

Petrography of Granitoid Rocks

The Sicheviţa granitoid was minutely investigated by Bîrlea (1975, unpubl. report) who recognized the following components: granites, porphyry granites, pegmatoids, granodiorites and quartz diorites (bearing biotite and hornblende). These groups are generally valid, although according to the petrochemical studies, the quartz diorites mostly belong to tonalites. Bîrlea herself shows that according to Streckeisen's (1973) model classifications these rocks should be assigned to tonalites or trondhjemites. Finally the author adopted the classification proposed by Jung and Brousse (1959).

An accurate map division of granites, granodiorites, tonalites and trondhjemites is not possible. However, the areas in which one of the mentioned rock types prevails have been put down.

The Sicheviţa granitoids exhibit massive structure and coarse-grained, pegmatoid or normal textures. Cataclastic structures and textures occur in the Vreia Valley trending north-southwards.

The mineralogic composition of the Sicheviţa granitoid includes: quartz, plagioclase, potash feldspar± biotite±muscovite±hornblende, accessory minerals (apatite, titanite, zircon, magnetite, garnet, pyrite, monazite).

The quartz grains are more or less equigranular, corroded, elongated and crushed in places due to disjunctive tectonic stress.

The plagioclase (An 18-35 %) exhibits idiomorphic or xenomorphic, frequently hipidiomorphic contours. Plagioclase zoning occurs frequently with granitoids containing femic minerals (tonalites and quartz diorites).

K-feldspar reaches centimetric size in pegmatoid granitoids. It shows perthitic, microperthitic or cross-hatched texture. Karlsbad twins are frequent. Poikilitic texture of K-feldspar with plagioclase, biotite, muscovite and quartz inclusions appears.

Biotite constitutes well developed, highly pleochroic lamellae; chloritisation occurs frequently. Sometimes biotite turns into muscovite.

Green hornblende with typical green-yellow-bluish pleochroism shows an extinction angle $c/ng=18-25^0$. Sometimes it turns into actinote with $c/ng=10-13^0$ or into biotite.

Granites show two textural-structural facies:

a) medium-grained granites, occurring mainly in the south, near the locality Cruşoviţa. They are characterized by the equal size of component minerals (1-3 mm), the leucocrate character and the massive structure. Mineralogical composition: quartz – 25-48 %, plagioclase – 25-40 % (An 20-25), potash feldspar – 20-48 %, biotite – 0.1-5 %, muscovite – 0.4-2 % and accessory minerals < 1 %.

b) coarse-grained granites exposed mainly north-west and south-west of the village Liuborajdea. The rocks have porphyry textures with orthoclase phenocrysts ca. 2 cm long, cemented by holocrystalline matrix with uniformly grained minerals (ca. 4-6 mm). The coarse-grained granites show crystalline schists enclaves in places. Mineralogic composition: quartz – 28-50 %, plagioclase – 34-43 % (An 18-25), potash feldspar - 10-25 %, biotite – 5-9 %.

Granodiorites are closely associated with medium-grained granites, tonalites and trondhjemites in places. They show the same mineralogic composition as granites, but differ little from the latter by modal composition: quartz – 22-45 %, plagioclase – 32-60 %, (An 22-35), potash feldspar – 6-20 %, biotite – 2-13 %.

Tonalites and trondhjemites prevail between Cameniţa and Şopotul Nou. They frequently alternate with granodiorites and equigranular granites. Tonalites and trondhjemites are medium-grained (1-4 mm), exhibit massive structure and holocrystalline texture. The hornblende occurrences in tonalites and quartz diorites

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make them different from granites and granodiorites. The tonalites and the trondhjemites differ from granites and granodiorites also by reduced potash feldspar amounts. The latter is sometimes lacking. Zoned plagioclase feldspars are characteristic of tonalites and trondhjemites.

Mineralogic composition: quartz – 25-40 %, plagioclase (An 25-35 %), potash feldspar – 0-2 %, biotite – 2-20 %, hornblende 0.1-5 %, accessory minerals – 0-1.5 %.

Unlike tonalites, trondhjemites contain reduced biotite and hornblende amounts (M < 7 %). Only rarely the quartz ratio decreases below 10 %. Then they grade into quartz diorites.

Petrography of Vein-like Rocks

Vein-like rocks are represented by aplites, granitic aplites, pegmatites, granodiorite-porphyries and lamprophyres. The length of these veins ranges from meters to hundreds of meters. They are not more than 3-4 m thick. Most frequent are aplites and granodiorite-porphyries; lamprophyres occur sparsely. Aplites grade into granitic aplites and pegmatites. The mineral size of pegmatites does not exceed 1.5 cm.

Aplites, aplite-granites and pegmatites are light-coloured rocks, yellow-whitish, exhibiting equigranular textures. They are not preferentially oriented within the granitoid massif. The mineralogic composition is: potash feldspar (25-35 %) represented by cross-hatched microcline, perthites, microcline-perthites, plagioclase (30-40 %), slightly sericitized, slightly zoned in places (An 16–20) and quartz (20-25 %). Muscovite (0.5–1 %) and biotite (0.1–0.5 %) occurrences are sparse.

Aplites are allotriomorphic-equigranular, while pegmatites are holocrystalline-coarse-grained.

Granodiorite-porphyries are dark grey, exhibiting porphyritic or aphanitic structure. Mineralogic composition: sericitized plagioclase with polysynthetic or Karlsbad twins (25-30 %). These minerals occur as phenocrysts. The coarse-grained groundmass (40-55 %) consists of quartz and plagioclase.

Lamprophyres are of dark green-black colour and exhibit aphanitic textures. They are usually altered and pierced by calcite veinlets. The texture is porphyritic, with hialine or intersertal groundmass. The phenocrysts are represented by olivine, augite, hypersthene±green hornblende.

Chemistry of Igneous Rocks

The geochemical characterisation of Sicheviţa granitoids and of associated vein-like rocks is based on 48 complete silicate analyses, 35 spectral analyses and 95 analyses for estimating the U, Th and Rb contents.

Major Elements

The main oxide percentages reported for the Sicheviţa granitoids range as follows: SiO_2=63.12-75.67; Al_2O_3=13.29-18.71; Fe_2O_3=0.40-3.37; FeO=0.00-2.85; MgO=0.10-2.60; CaO=0.60-4.59; K_2O=1.69-4.60; Na_2O=3.19-4.60 (Tab. 1).

On the QAP diagram (Fig. 1, normative composition, Rittmann method) the granitoid rocks are distributed as follows: granites (syenogranites and monzogranites), granodiorites and tonalites. On the TAS diagram proposed by Andreeva et al. (1981) most of the samples are plotted within the fields of granites and granodiorites; some leucogranitic features of granitoids and quartz diorite occurrences are also pointed out (Fig. 2).

Granites are of calc-alkaline character showing very slight alkaline tendencies (Fig. 3).

Granites (syenogranites and monzogranites) are of potassic character, while tonalites and trondhjemites are of sodic nature. Granodiorites are partly sodic, partly potassic (Fig. 4).

Out of the vein-like rocks (Tab. 2) aplites yield high SiO_2 (73.40-75.96 %) and alkalis (3.78-5.95 %) contents, and very low Fe_2O_3 (0.57-1.67 %), FeO (0.00-0.55 %), MgO (0.03-0.50 %) and CaO (0.36-1.50 %) ones, similarly to pegmatites: SiO_2=71.40 %, K_2O=6.80 %, Na_2O=3.31 %, Fe_2O_3=0.58 %, FeO=0.60 %, MgO=0.85 %, CaO=0.40 %.

On the contrary, lamprophyres yield low SiO_2 (46.86-48.62 %) and alkalis (K_2O=0.80-1.04 %, Na_2O=2.47-3.00 %) contents and higher FeO (2.80-3.30 %), Fe_2O_3 (4.08-5.85 %), MgO (9.00-9.33 %) and CaO (8.80-10.92 %) ones.

Granodiorite-porphyries exhibit oxide ratios, in between aplite-pegmatites and lamprophyres, obviously tending to the former: SiO_2=71.93-74.52, Fe_2O_3=0.49-0.85, FeO=0.55-0.92, MgO=0.40-1.04, CaO=1.41-2.45; K_2O=2.27-3.32, Na_2O=3.63-4.57.

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TABLE 1

Chemical Composition of Sichevița Granitoid Rocks

No.	Oxides / Sample	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	TiO_2	H_2O^+	CO_2	Location	Rock type
1	7	75,67	13,39	0,40	0,34	0,02	0,37	1,55	4,58	3,20	0,04	0,05	0,49	0,00	Right tributary Gramensca Valley	MZG
2	13	75.46	13,29	0,76	0,00	0,03	0,28	0.91	4.60	3,19	0,06	0,08	1,03	0,00	Odăile Valley	SYG
3	45 *	74,95	14,84	0,73	0,09	0,05	0,54	1,32	3,40	3,84	0,02	0,00	0,47	0,00	Gravii ravine	MZG
4	135	74,02	14.81	1,34	0,00	0,03	0,39	1,14	2,23	3,80	0,10	0,20	0,90	0,00	Seacă Valley	MZG
5	13T	73,67	14,54	0,72	0,62	0,03	0,37	1,10	4.06	3,61	0,16	0,12	0,70	0,00	Cremenița Valley	SYG
6	0165	73.41	14,74	0,92	0,75	0,05	0,37	0,60	3,95	3,59	0,10	0,12	1,09	0,00	Răchiței Valley	SYG
7	11	73,08	14,54	1,69	0,00	0,04	0,49	1,79	3,43	3,96	0,06	0,20	0,67	0,00	Gramensca Valley	MZG
8	0103	73,10	15,41	0,86	0,00	0,03	0,30	1,57	3,39	3,80	0,10	0,16	0,90	0,00	Lăpușnic Valley	MZG
9	122	72,97	14,54	0,51	1,07	0,05	0.53	1,40	3,38	4,25	0.08	0,20	0,66	0,00	Gramensca Valley	MZG
10	14T	72,89	14,89	1,49	0,10	0,06	0,46	1,55	3,04	3,90	0,16	0,16	0,98	0,00	Porîmb Valley	MZG
11	12T	72,23	15,21	1,56	0,12	0,04	0,50	2,16	3,05	4,05	0,12	0,20	0,70	0,00	Găvojdia Roșie Valley	MZG
12	0173	71,37	14,84	0,86	1,53	0,04	0,68	1,65	3,78	4,00	0,09	0,24	0,62	0,00	Răchița Valley	
13	93 *	71,23	15,33	1,05	1,26	0,07	1,20	1,02	4,16	3,40	0,06	0,15	0,90	0,00		SYG
14	321 *	70,60	15,00	0,85	2,50	0,03	0,95	1,85	3,11	3,37	0,08	0,45	0,90	0,25		MZG
15	85 *	70,55	15,59	1,28	1,38	0,09	1,10	1,40	4,10	3,83	0,06	0,18	0,66	0,00	Isvorului ravine	MZG
16	92 *	70,20	15,37	1,46	1,59	0,10	1,50	0,86	3,86	4,25	0,06	0,20	1,03	0,00	Gramensca Valley	MZG
17	40 *	70,10	16,78	1,27	0,30	0,08	0,60	1,30	4,40	4,60	0,07	0,00	0,88	0,00	Crușovița Valley	MZG
18	57	69,69	15,14	2,00	1,06	0,05	1,09	2,66	3,36	3,65	0,12	0,32	0,65	0,00	Vreia Valley	MZG

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Table 1 (continued)

19	28 *	69,35	18,60	1.20	0.00	0.14	0.10	1.20	3.47	4.50	0.10	0.00	0.38	0,50	Moran ravine	TO-TRD
20	54	68.97	15,89	0,86	1,93	0,05	1,07	2,47	3,60	3,73	0,12	0,28	0,77	0,00	Vreia Valley	MZG
21	87	68.87	16,10	0,48	2,18	0,05	1,48	3,34	2,13	3,83	0,11	0,52	0,72	0,00	Crușovița Valley	GD
22	126	68.63	16,61	0,54	2,08	0,05	2,08	3,07	2,00	3,41	0,12	0,50	0,76	0,00	Gramensca Valley	GD
23	90	68.44	15,99	0,70	2,38	0,08	1,56	3,40	2,13	3,55	0,13	0,52	0,74	0,80	Crușovița Valley	GD
24	0180	67.94	16,26	1,45	2,03	0,06	1,40	2,40	2,35	3,88	0,16	0,46	1,19	0,00	Răchița Valley	GD
25	79	67.04	16,43	1,27	1,68	0,06	1,42	3,69	2,32	4,41	0,16	0,48	0,37	0,38	Danube Valley	GD
26	113	66.87	16,81	0,90	1,92	0,04	1,62	2,99	2,48	4,33	0,14	0,52	1,13	0,00	Gramensca Valley	GD
27	40	66.34	16,86	1,93	1,53	0,06	1,50	4.11	1,94	4,11	0,16	0,56	0,83	0,00	Nera Valley	GD
28	34	65.94	17,61	1,14	1,99	0,06	1,77	3,11	2,50	3,99	0,14	0,46	1,16	0,00	Bresnic Valley	GD
29	41	65.85	16,65	1,36	2,11	0,06	1,83	4,11	1,85	4,01	0,16	0,60	1,02	0,00	Nera Valley	TO-TRD
30	0168	65.68	16,56	0,74	2,04	0,05	1,95	3,18	3.30	4,15	0,16	0,56	1,53	0,00	Răchița Valley	GD
31	39	65.22	17,21	1,50	2,35	0,07	1,88	4,45	1,72	4,07	0,13	0,56	0,69	0,00	Minelor ravine	TO-TRD
32	74	65.15	17,46	1,61	1,94	0,06	1,95	4,59	1,69	4,29	0,18	0,64	0,77	0,00	Danube Valley	TO-TRD
33	36	65.05	17,01	3,37	0,57	0,07	1,88	4,30	2,03	4,30	0,18	0,60	0,80	0,00	Bresnic Valley	TO-TRD
34	27	63.52	17,56	2,85	1,36	0,08	1,79	3,57	2,41	4,02	0,20	0,76	1,21	0,50	Danube Valley	GD
35	90 *	63.27	18,39	1,69	2,65	0,11	2,45	3,56	2,20	4,25	0,09	0,57	0,60	0,00	Gramensca Valley	GD
36	17 *	63.12	18,71	1,23	2,85	0,13	2,60	3.50	2,50	3.90	0,13	0,57	0,19	0,64	Gramensca Valley	GD

MZG: Monzogranites, *TO-TRD*: Tonalite-Trondhjemites, *GD*: Granodiorites
Analyses according to Lidia Bîrlea (1975)

Analyst: Elena Colios

TABLE 2

Chemical Composition of Vein-like Rocks Associated with the Sichevița Granitoids

No.	Oxides / Sample	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	TiO_2	H_2O^+	CO_2	Location	Rock type
1	12	75,96	13,59	0,72	0,17	0,03	0,03	0,36	3.97	4.03	0,04	0,08	0,73	0,00	Odăile Valley	AP
2	15	75,26	13,74	0,57	0,00	0,03	0,07	0,38	5.95	5,09	0,06	0,08	0,53	0,00	Right tributary Vreia Valley	AP
3	118	75,04	13,69	0,62	0,11	0,02	0,08	0,64	4,92	3,96	0,04	0,06	0,50	0,00	Left tributary Cheilor Valley	AP
4	53	74,52	13.92	0,85	0,55	0,05	0,40	1,41	3,32	3,63	0,06	0,20	0,75	0,00	Vreia Valley	GP
5	61 *	74,40	14,92	0,63	0,09	0,02	0,50	1.06	5,45	2,85	0,03	0,00	0,88	0,00	Vreia Valley	AP
6	124	73,89	14,96	0,67	0,10	0,04	0,17	1,38	3,78	4,08	0,06	0,08	0,62	0,00	Gramensca Valley	AP
7	311 *	73,40	14,30	1,67	0,30	0,02	0,50	1,50	4,40	3,50	0,00	0.00	1,30	0,30		AP
8	94	72,60	11,99	0,49	0,92	0,03	0,78	2.54	2,21	4,57	0,08	0,18	0,45	0.00	Crușovita Valley	GP
9	111	71,93	15,64	0,50	0,83	0,03	1,04	2,06	2,54	4.16	0,09	0,22	0,69	0,00	Left tributary Gramensca Valley	GP
10	124 *	71,40	15,57	0,58	0,60	0,04	0,85	0,40	6,80	3,31	0,04	0,07	0,63	0,00		PG
11	374 *	48,62	16,00	5,85	2,80	0,14	9.00	8,80	0,80	3,00	0,11	4,65	3,90	0,30	Breznic ravine	L
12	8	46,86	13,59	4,08	3,30	0,11	9,33	10,92	1,04	2,47	0,16	0,11	5,00	2,82	Right tributary Gramensca Valley	L

AP = Aplites; GP = Granodiorite-porphyries; PG = Pegmatites; L = Lamprophyres
* = Analyses according to Lidia Bîrlea (1975)

Analyst: Elena Colios

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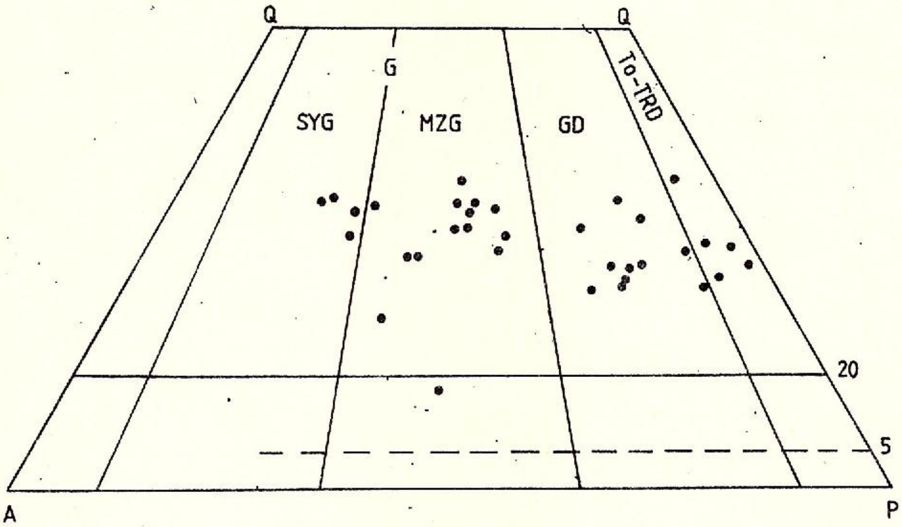

Fig. 1 – QAP diagram: normative composition (Rittmann method)
G, granites (SYG, syenogranites; MZG, monzogranites); GD, granodiorites;
TO-TRD, tonalites-trondhjemites.

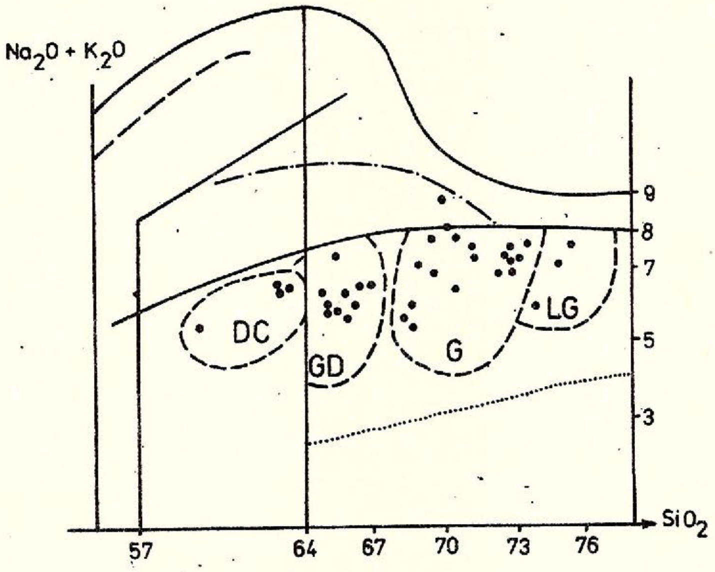

Fig. 2 – $Na_2O+K_2O-SiO_2$ diagram (Andreeva et al., 1981)
DC, quartz diorites; GD, granodiorites; G, granites;
LG, leucogranites.

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TABLE 3

Minor Elements (ppm) of Sichevita Granitoid Rocks

No.	Sample	Pb	Cu	Ga	Sn	Ni	Co	Cr	V	Sc	Y	Yb	La	Nb	Zr	Be	Ba	Sr
1	7	27	6,5	13	4	2,5	2	3,5	6	<2	15	1,2	<30	<10	80	2,3	950	170
2	13	28	5	12	4	2	<2	2	5	<2	11	1,3	<30	<10	75	3,4	1500	200
3	135	25	3,5	18	6	3	2	6	9,5	4	18	1,8	<30	13	75	4,6	800	160
4	13T	38	9	19	5,5	<2	<2	1,5	6,5	<2	10,5	<,0,5	<30	<10	60	3,6	820	190
5	0165	55	15	17	5	3,5	<2	4	10	3	17	1,4	<30	12	60	2,8	950	150
6	11	38	15	19	6,5	2,5	2	2	9,5	<2	9	<0,5	32	<10	78	3,6	900	210
7	0103	19	4	18	5,5	2	2	2	8	<2	10	0,5	<30	10	88	3,2	1100	220
8	122	30	20	18	6,5	5	2	6,5	14	3	21	1,4	<30	13	125	3,0	750	140
9	14T	21	38	17	8	8	3,5	12	29	8	20	1,1	48	16	160	5,0	1000	210
10	12T	32	9,5	15	4	2,5	2,5	2,5	13	<2	9	<0,5	32	<10	110	2,8	1200	330
11	0173	30	15,5	18	4	7,5	3,5	11	20	6	37	2,5	30	12	110	3,0	700	170
12	57	17	11	16	4	13	4,5	18	25	5,5	16	1,1	46	16	135	2,0	1600	400
13	54	18	18	18	4,5	9	5	23	29	7	18	1,2	48	15	160	2,3	900	180
14	87	24	20	20	5	17	8	22	43	7	10,5	<0,5	44	<10	230	1,8	800	320
15	126	18	14	28	3,5	12,5	6	14	32	5,5	21	1,4	<30	<10	190	2,4	700	350
16	90	15	10	17	4,5	16	8	19	40	6	10	0,5	<30	<10	190	3,0	650	220
17	0180	18	12,5	26	5,5	15	6,5	15	40	7	19	0,9	<30	13	140	5,5	460	230
18	79	17	19	19	3,5	13	7	15	38	6,5	13	0,6	48	13	160	2,5	1000	400

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Table 3 (continued)

19	113	17	36	17	3,5	17	8	30	48	5.5	12,5	0.5	<30	10	200	1,6	900	420
20	40	13	22	21	4	11	7	24	46	8	15	0,9	38	<10	210	1,9	850	430
21	34	21	30	18	7	15	8	17	43	5,5	10	0,6	<30	<10	200	2,5	800	260
22	41	14	24	19	4	15	7,5	15	50	7	16	1,0	<30	<10	220	1,7	750	500
23	0168	21	25	21	5	21	9	13	55	6	12	0,6	30	<10	210	4,3	1000	330
24	39	14	34	19	4	18	8,5	14	62	7	18	0,7	<30	<10	220	2,1	650	360
25	74	16	25	20	4	11	8	11	52	8	19	1,1	42	10	270	1,7	650	450
26	36	18	36	18	3,5	23	9,5	19	60	9	15	0,7	<30	<10	200	2,6	600	280
27	27	12	27	20	4	9,5	9	6,5	53	6	8	<0,5	<30	<10	180	1,8	800	450

Minor Elements (ppm)
in Vein-like Rocks Associated withe Sichevița Granitoids

28	12	40	6	14	7	<2	<2	<1	2	2	13,5	1,3	<30	15	36	4,0	190	25
29	15	50	5,5	13	4	<2	<2	<1	4	<2	17	2,0	<30	<10	42	3,4	900	120
30	118	36	5,5	8,5	<2	<2	<2	1	2	<2	9	0,6	30	11,5	65	1,4	1200	120
31	53	37	9	17	7,5	3,5	3	3	13	3	15	1,5	<30	21	95		950	1100
32	124	40	7,5	13	4	2,5	<2	3,5	4,5	<2	9,5	0,6	<30	<10	55	2,8	1100	220
33	94	21	31	21	3	15	3,5	10	15	2	5,5	<0,5	<30	<10	130	1,8	480	220
34	111	22	8,5	14,5	3,5	18	3,5	16	15	2,5	9	<0,5	<30	11	170	2,1	800	240
35	8	5	50	10	3,5	210	32	730	200	36	36	1,1	<30	<10	80		600	1100

Analyst: Constanța Udrescu

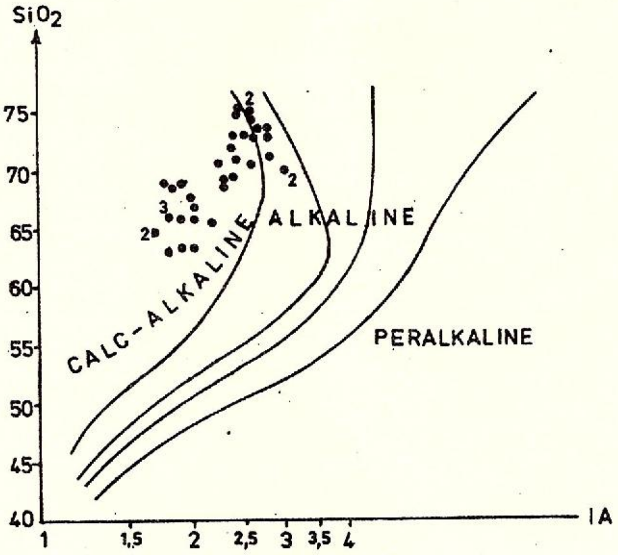

Fig. 3 – SiO_2–IA diagram
IA, alkalinity index.

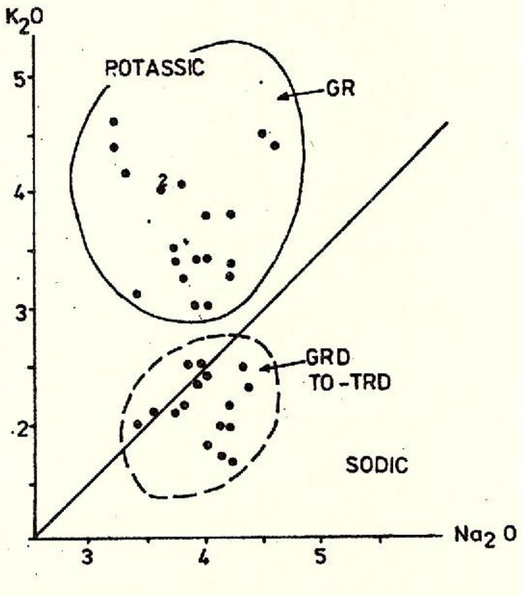

Fig. 4 – K_2O–Na_2O diagram
GR, granites (monzogranites and syenogranites);
GRD, granodiorites; TO-TRD, tonalites-trondhjemites.

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Minor Elements

The minor elements contents of granitoids and vein-like rocks determined by emission spectrography are enlisted in Table 3.

The Nockolds-Allen differentiation diagrams (Figs. 5, 6) point out the following features of the Sicheviţa granitoids: Sc, Co, Cr, Mg, Ni, Fe_{tot}, V, Ca values decrease during the differentiation process (tonalites→granodiorites→granites); Sr and Na values are almost constant; while K value is increasing. Ba behaves rather unexpectedly.

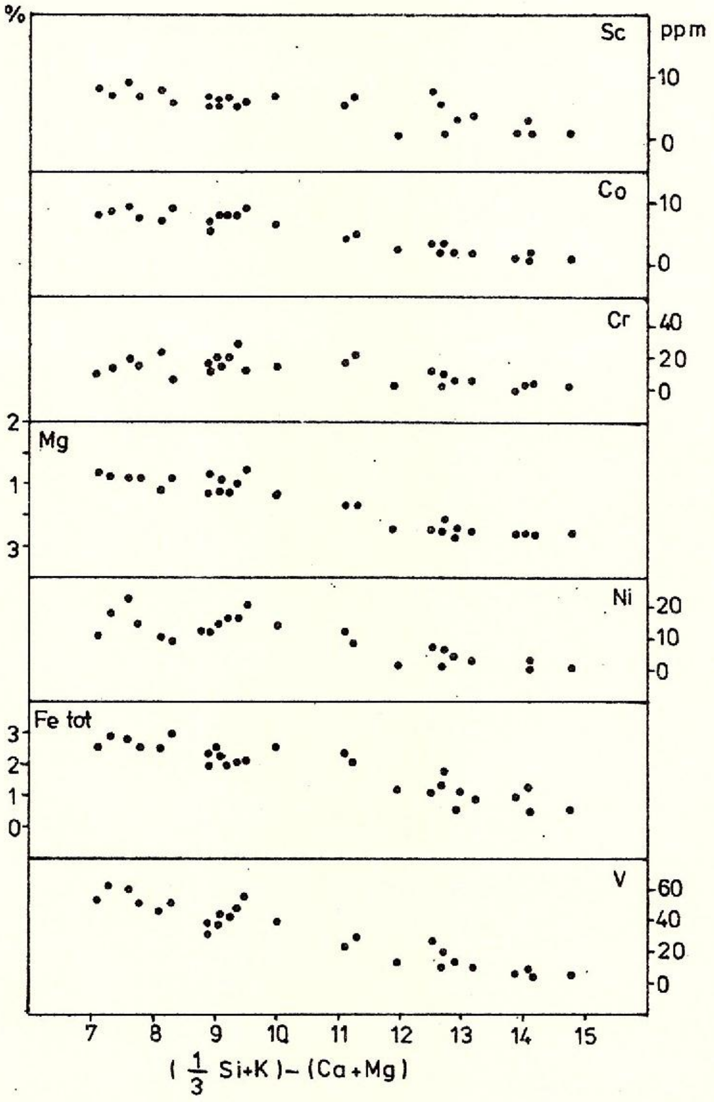

Fig. 5 – Nockolds-Allen diagram.

A comparison between minor elements yielded by the Sicheviţa granitoid and the mean values of minor elements reported for the standard granitoid, calculated by Kraft and Schindler (1961) is illustrated in Figure 7. Be, Sn, Pb, Ga, Zr, Ba mean values are very close to the mean values calculated by Kraft and Schindler for the standard granites. Only Y values are lower, when comparing the Sicheviţa granitoid to the standard one.

The femafile minor elements occur in small amounts in granitoid rocks. As regards the Sicheviţa granitoid they range as follows (ppm): Ni=<2–23; Co=<2–9.5; Cr=<1–38; V=5–62; Sc=<2–12; Cu=3.5–38.

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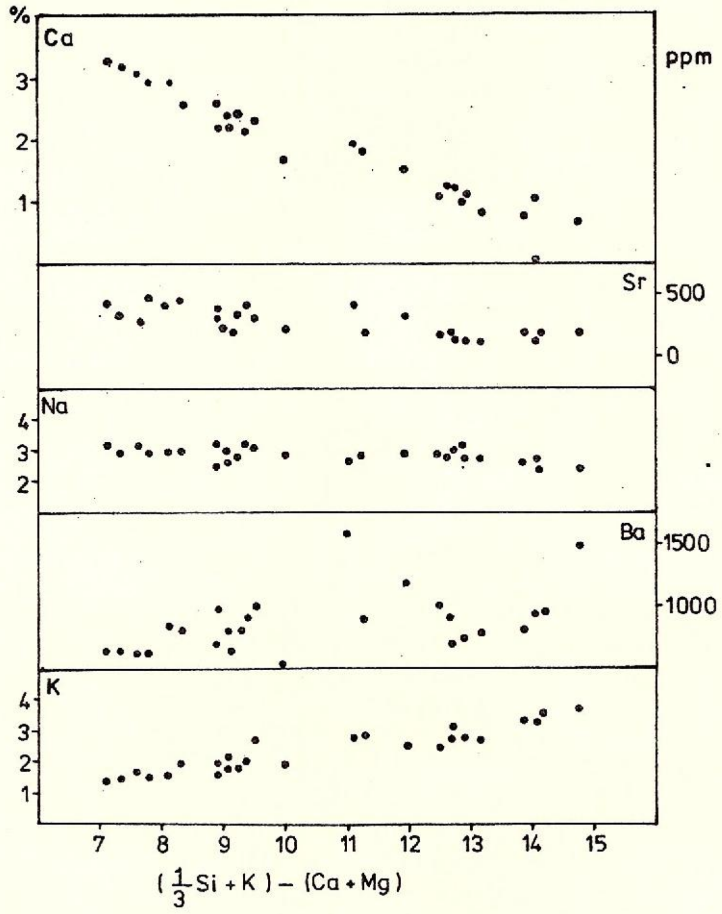

Fig. 6 – Nockolds-Allen diagram.

Radioactive Elements

By means of emission gamma spectrometry U and Th contents and by non-dispersive X-ray fluorescence Rb contents of 84 granitoid rocks and 11 aplites have been analysed. The ppm values are: U=0.1–6.2; Th=0.7–23.4; Rb=68–209 for granitoid rocks and U=0.5–4.5; Th=1.4–13.2; Rb=83–129 for aplites (analysts I. Tiepac and G. Grabari).

Hypotheses on the Petrogenesis and Age of Granitoid Rocks

The north-southward elongated shape of the granitoid accounts for its emplacement on a former tectonic line. In this low resistance zone prolonged northwards the Poneasca granitoid was emplaced and south of the Danube, in Yugoslavia, intrusive acid magmatites exhibiting similar petrochemical features do occur. The entire north-southwards trending alignment of granitoid rocks is exposed intermittently along almost 100 km.

The obvious contacts between the Sicheviţa granitoid and the Sebeş-Lotru Group or the Miniş series as well as the thermal contact metamorphism show that the granitoid is an allochthonous bounded intrusive body. Therefore, it is inferred that the anatectic calc-alkaline magmas ascended from their source to upper areas subsequently to the metamorphism of crystalline schists. According to available data there are important age differences among the granitoid rocks, the Sebeş-Lotru Group and the Miniş Formation. Thus, the Sebeş-Lotru Group is considered of Precambrian age (Savu, 1979; Kräutner, 1980; Iancu, Mărunţiu, 1989) and the Miniş Formation of ante-Upper Carboniferous, probably Lower Paleozoic age (Iancu and Mărunţiu, 1989). The radiometric ages determined by different methods on different minerals range within a relatively large interval in

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the case of the Sicheviţa granitoid: 250–350 Ma, as inffered from Table 4 (Bîrlea, 1975, unpubl. report). Thus, the Sicheviţa granitoid could not be younger than 350 Ma, i.e. Devonian-Carboniferous boundary, according to Odin scale (1982), but not older than the Lower Paleozoic (the age of the Miniş Series, sensu Iancu, Mărunţiu, 1989) which is crossed by the Sicheviţa granitoid. Therefore, this intrusive massif is of Hercynian, possibly Caledonian age. The Hercynian age agrees with the opinions of Maier (1974) and Bîrlea (1975, unpubl. report). South of the Danube, the age of the granitoid is determined by Yugoslavian researchers by Sr/Rb method as 259-272 Ma, i.e. Hercynian too (Cissar, 1969, fide Bîrlea, 1975, unpubl. report).

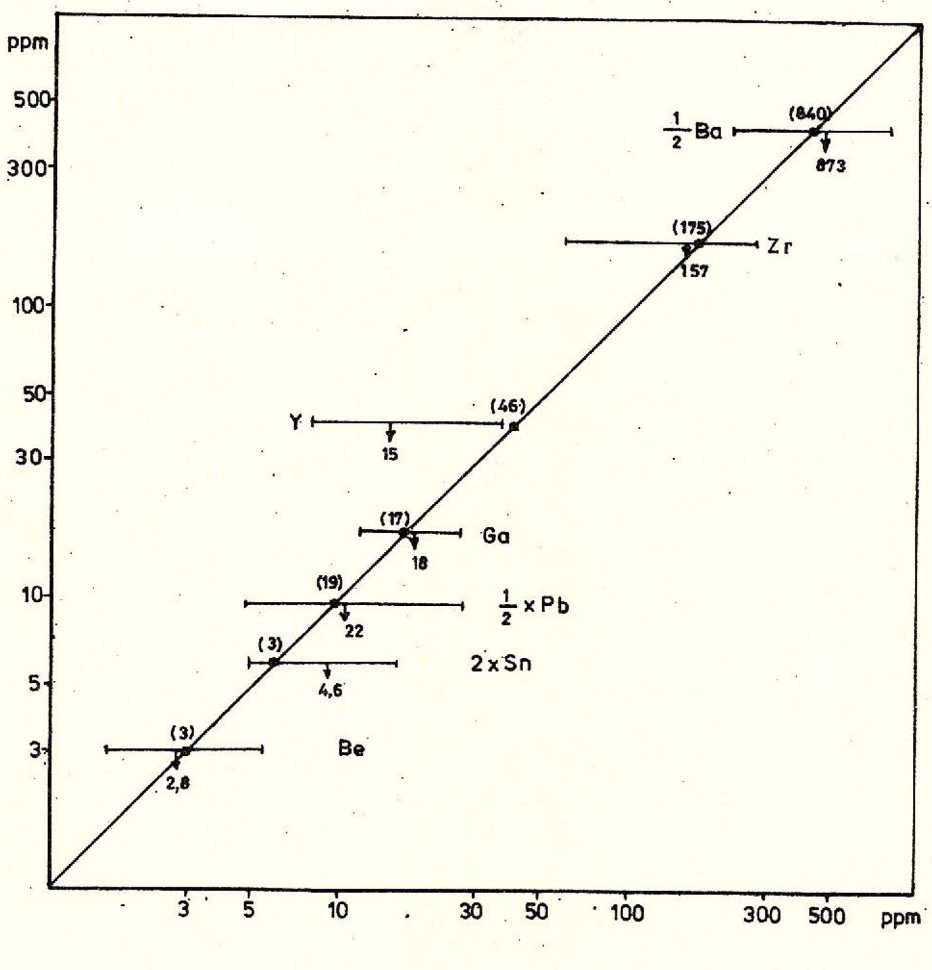

Fig. 7 – Kraft-Schindler diagram (1961)

It is quite probable that the granitoid batholith be generated by anatectic magma. The diagrams envisaged by Pearce et al. (1984) suggests that these melts resulted from the collision of lithospheric plates (Fig. 8).

Considering the petrographic types and the chronologic relations among them – more or less obviously exposed – one might suppose a pulsatile differentiation of anatectic magmas as follows: tonalites (quartz diorites)→trondhjemites→granodiorites →granites→aplites. These differentiations to more and more acid rocks are also suggested by the radiogenic ages illustrated in Table 4. The highest values are yielded by biotite granites and quartz diorites; the decreasing order implies granites and aplites. However, radiogenic ages are insufficient for this purpose.

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TABLE 4

Radiometric Ages

No.	Location	Rock type	Analysis	Method	M.a.
1	*Liuborajdea Valley*	*aplite*	*complex*	*K/Ar*	*250*
2	*Sichevița*	*granite*	*microcline*	*K/Ar*	*276*
3	*Crusovita Valley*	*quartz diorite*	*biotite*	*K/Ar*	*292*
4	*Liuborajdea Valley*	*pegmatoid granite*	*microcline*	*K/Ar*	*310*
5	*Danube Vally*	*biotite granite*	*monazile*	*U Th/Pb*	*328*
6	*Danube Vally*	*biotite granite*	*monazile*	*U/Pb*	*350*

Samples 1-2 analyst I. Tiepac (Nancy 1970);
Sample 6 - analyst Grünenfelder (Zürick 1971)

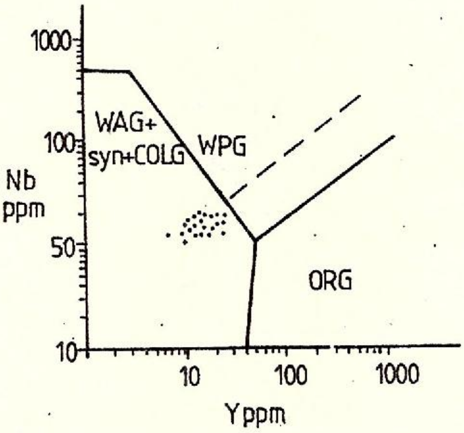

Fig. 8 – Nb-Y diagram (Pearce et al., 1984) WAG+sin-COLG, volcanic arc granites+syncollision granites; WPG, withinplate granites; ORG, ocean rift granites.

References

Andăr P. (1976) Calculul automat al normelor Rittmann. *D. S. Inst. Geol. Geofiz.*, LXII/1, p. 59–62, Bucureşti.

Andreeva E. D., Bogatikov O. A., Borodaevskaia M. B. (1981) Klassificaţia i nomenklatura magmaticeskih gornîh porod. Ed. Nedra, 159 p., Moskva.

Codarcea Al. (1940) Vues nouvelles sur la tectonique du Banat Méridional et du plateau de Mehedinţi. *An. Inst. Rom.*, XX, Bucureşti.

Iancu V., Mărunţiu M. (1989) Toroniţa Zone and problems of the pre-alpine metamorphic basement of the Getic and Danubian Realms. *D. S. Inst. Geol. Geofiz.*, 74/1, p. 223–237, Bucureşti.

Jung J., Brousse R. (1959) Classification modale des roches éruptives. Masson, Paris.

Kraft M., Schindler R. (1961) Periodisches System der Elemente Zusammengefasst von Zentralen Geologischen Institut, Berlin.

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

Maier O. (1974) Studiul geologic şi petrografic al masivului cristalin Locva. *St. tehn. econ.*, I, 5, Inst. Geol., Bucureşti.

Institutul Geologic al României

Năstăseanu S., Mărunțiu M., Stancu I., Mărunțeanu M., Întorsureanu I. (1981) Harta geologică a R.S.R., scara 1:50000, foaia Sichevița, I.G.G., București.

Odin G. S. (1982) The Phanerozoic Time Scale Revisited. *Episodes*, 3.

Pearce J. A., Harris N. B. W., Tindle A. G. (1984) Trace Element Discrimination Diagrams for the Tectonic Interpretation of Granitic Rocks. *Jour. of Petrol.*, 15, 4, p. 956–983, Oxford.

Savu H. (1979) Crystalline schists Precambrian granitoid rocks and associated metallogenesis from the Getic nappe unit (Banat). *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 23, 2, p. 123–136, București.

Streckeisen A. (1934) Sur la tectonique des Carpathes Méridionales. *An. Inst. Geol.*, XVI, București.

Streckeisen A. (1973) Classification and Nomenclature of Plutonic Rocks, Recommendation. *N. Jahrb. f. Mineral. Monatshefte*, Stuttgart.

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PROFESORUL ALBERT STRECKEISEN – 90 DE ANI

La 8 noiembrie 1991 Profesorul Albert Streckeisen a aniversat nouă decenii de viaţă, înconjurat de un cerc intim de prieteni, colegi, foşti elevi şi studenţi din care nu au lipsit reprezentanţi ai geologiei româneşti, căreia i-a dăruit o parte importantă din prodigioasa sa activitate ştiinţifică.

În 1927 Albert Streckeisen a fost chemat la Scoala Politehnică din Bucureşti pentru a preda cursul de mineralogie–petrologie după încetarea din viaţă a profesorului Gh. Munteanu–Murgoci. De la început tînărul doctor în geologie din Elveţia, în vîrsta de 26 ani, a simţit o atracţie pentru România, afecţiune pe care a cultivat-o şi după întoarcerea în ţara natală, în 1934. Intr-un timp deosebit de scurt şi-a însuşit graiul românesc şi a prezentat de la catedră cursul în limba română.

Pe lîngă obligaţiile didactice, în calitate de membru al Institutului Geologic al României, profesorul Albert Streckeisen a adus o contribuţie substanţială la desluşirea unor probleme esenţiale din geologia Carpaţilor. Această activitate de cercetare a fost îndreptată spre două direcţii principale. Prima, cuprinde Carpaţii Meridionali, abordată la solicitarea prof. L. Mrazec, cu scopul de a verifica ipoteza structurii în pînze de şariaj a acestora, emisă de Gh. Munteanu–Murgoci. Rezultatele investigaţiilor, publicate în mai multe lucrări, îşi menţin valabilitatea după mai bine de o jumătate de veac. Trebuie menţionat în acest sens identificarea unităţilor Supragetice, şi separarea zonelor de metamorfite din Banat, pentru care-i revine întîietatea.

A doua direcţie de cercetare s-a bazat pe opţiunea proprie şi priveşte una din problemele petrologice speciale ale Carpaţilor – masivul alcalin de la Ditrău – al cărui interes ştiinţific depăşeşte limitele geologiei naţionale. Bazele concepţiei despre originea magmatică a masivului, elaborate cu 60 de ani în urmă, şi găsesc o confirmare în investigaţiile recente, după ce au înfruntat cu succes modelul metasomatic lansat în jurul anilor '60.

Pe plan internaţional profesorul Albert Streckeisen şi-a cîştigat o notorietate meritată, în special prin coordonarea timp de peste 20 de ani a comisiei pentru sistematica rocilor magmatice. Prin recomandările elaborate în urma consultării a sute de specialişti şi-a legat numele de sistematica rocilor magmatice care la ora actuală se află în uz în lumea întreagă. Această activitate i-a adus şi unele distincţii academice, ca de exemplu Medalia Werner (1984) şi alegerea în Academia dei Lincei (Roma, 1984).

Pe lîngă activitatea ştiinţifică prodigioasă desfăşurată în timpul şederii în ţară (1927–1934), profesorul Streckeisen a fost preocupat de formarea tinerei generaţii de geologi romani. Astfel, în urma unei selecţionări judicioase, s-a înconjurat la catedră de doi tineri petrografi, G. Manolescu şi G. Paliuc, pe care i-a îndrumat în pregătirea tezelor de doctorat, plasate în inima Carpaţilor Meridionali. Din nefericire timpurile de restrişte postbelică au curmat prea devreme viaţa acestor discipoli ai profesorului.

Preocupat continuu de a păstra legătura cu geologia românească, Prof. Streckeisen foloseşte întrunirile internaţionale din 1968, 1977, 1981 pentru a revedea locurile îndrăgite şi pentru a cunoaşte noua generaţie de geologi români. După revoluţia din 1989 vizitează anual masivul Ditrău, înconjurat de numeroşi geologi tineri, dornici să cunoască personal pe cel ştiut vreme îndelungată din literatura de specialitate.

După o întrerupere impusă timp de o jumătate de secol Prof. Streckeisen reia preocuparea pentru promovarea geologiei româneşti, atît prin facilitarea cu sprijin, uneori chiar material, a contactului cu viaţa geologică internaţională, cît şi prin supunerea atenţiei internaţionale a problematicii masivului Ditrău. Totodată prevede trecerea în viitor a părţii cele mai însemnate din biblioteca sa personală, în patrimoniul Institutului de Geologie şi Geofizică din Bucureşti.

Graţie descătuşării de restricţiile trecute, geologii români au avut posibilitatea să omagieze în mod oficial contribuţia profesorului Albert Streckeisen la progresul geologiei româneşti în 1990, cînd a fost numit membru de onoare al Societaţii Române de Geologie şi în 1991 cînd i s-a conferit titlul de membru de onoare al Academiei Române. La aceste distincţii se adaugă pe plan personal reînnodarea contactului cu geologia română în special prin generaţia tînără, care în ultimii ani a avut şanse să-l cunoască pe profesorul Albert Streckeisen atît ca om de ştiinţă cît şi ca om de suflet.

În numele acestor petrologi români pentru care sînteţi un exemplu demn de urmat, D-le Profesor Albert Streckeisen, cu ocazia împlinirii a 90 de ani de viaţă,

La mulţi ani!

H. G. Kräutner

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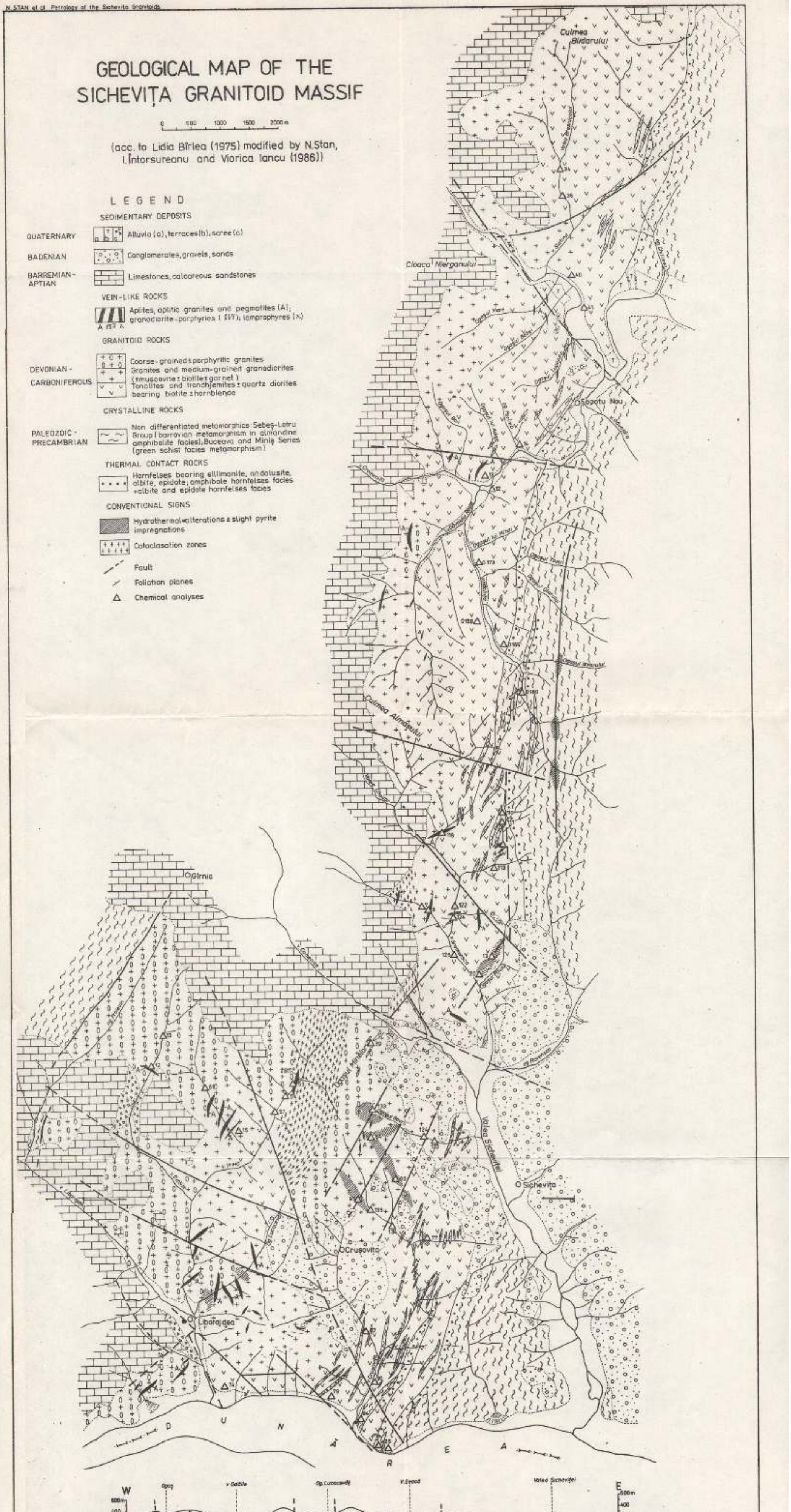
GEOLOGICAL MAP OF THE SICHEVIȚA GRANITOID MASSIF

(acc. to Lidia Bîrlea (1975) modified by N. Stan, I. Întorsureanu and Viorica Iancu (1986))



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Rom. J. Petrology, 1992, 75, p. 17–38

MAJOR AND TRACE ELEMENT GEOCHEMISTRY OF RHYOLITES FROM NORTHERN DOBROGEA. PETROGENETIC IMPLICATIONS

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Key words: Rhyolites. Major elements. Minor elements. Petrology. Dobrogea – Northern Dobrogea – Măcin Area, Tulcea and Consul-Niculiţel Areas.

Résumé: *La géochimie des éléments majeurs et en traces des rhyolites de la Dobrogea septentrionale. Implication pétrogénétiques.* Tenant compte de la distribution de leurs éléments majeurs et en traces, les volcanites et sous-volcanites acides, prépondérant rhyolitiques, des différentes unités tectoniques de la Dobrogea septentrionale sont divisées, selon l'âge, le cadre géotectonique, la source du magma et les processus de différenciation, de la manière suivante: (1) trachytes paléozoïques, associées aux roches basiques, dans l'unité de Tulcea, générées par la fusion partielle d'une source du manteau au milieu d'un rift; (2) rhyolites carbonifères, calco-alcalines, associées aux dépôts terrigènes, dans l'unité de Măcin, générées par la fusion partielle d'une source quartzo-feldspathique dans la croûte, situées dans un environement géotectonique de type 'back arc'; (3) rhyolites triasiques calco-alcalines, associées aux basaltes, dans les unités de Măcin, Consul et Tulcea, générées par la fusion partielle d'une source dans la croûte et, parfois, par la cristallisation fractionée dans des chambres magmatiques intermédiaires, developpée au début d'une riftogenèse d'intraplaque continentale; (4) rhyolites d'âge triasique-jurassique (?) calco-alcalines–alcalines associées aux granites ayant les mêmes traits chimiques, sur 'l'alignement du sud' de l'unité de Măcin, générées par l'action d'un 'dome thermique', suivies, par places, par des processus de cristallisation fractionnée dans une intraplaque continentale; (5) peu de rhyolites alcalines jurassique supérieures associées aux basaltes, au sud-ouest de l'unité de Măcin, générées par la cristallisation fractionnée d'un magma résultée de la fusion partielle du matériel parental dans le manteau, ou à la base de la croûte.

1. Introduction

The relationship between igneous rock bodies and Paleozoic and Mesozoic metamorphic and sedimentary formations is one of the major problems implied in the study of the geologic structure of Northern Dobrogea. Igneous rocks are represented by basaltic and rhyolitic volcanics and granitic plutonics. Taking into account their wide areal extent and their generation within certain geotectonic environments, a geochemical study of the igneous rocks is highly relevant to the reconstitution of the Hercynian and Alpine evolution of Northern Dobrogea.

Until present the petrographical and geochemical (mainly major elements) features of Northern Dobrogea rhyolites ("quartz porphyries") have been repeatedly investigated. Among the published works worth mentioning are Peters (1867), Pascu (1904), Mrazec (1912), Cantuniari (1912), Murgoci (1914), Cădere (1925), Savul (1931, 1935), Streckeisen (1931), Dimitrescu (1959), Ianovici et al. (1961), Mirăuţă, Mirăuţă (1962), Mutihac (1964), Mirăuţă (1966a, 1966b), Stiopol et al. (1978), Grădinaru (1981), Savu et al. (1982, 1986), Întorsureanu (1987), Seghedi et al. (1987). The unpublished studies (including trace element data, too) to be cited belong to Bălan (1966, 1967), Stiopol et al. (1975), Constantinescu et al. (1978, 1981, 1982), Caraveţeanu in Vîlceanu et al. (1979) and in Manea et al. (1983), Baltreş et al. (1984), Ştefan, Roşu in Berbeleac et al.(1985) and in Nedelcu et al. (1986, 1987, 1988), Seghedi, Szakács in Seghedi et al. (1985, 1986) and in Mirăuţă et al. (1986).

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A comprehensive study of both major and trace element geochemistry for rhyolites in Northern Dobrogea has not been carried out so far. The present study is the first attempt in this respect, involving both petrogenetic and geotectonic implications.

2. Geology

2.1 Geologic and tectonic setting

The Northern Dobrogea is an intracratonic orogenic belt completed during the Alpine orogeny. From west to east the following major structural units have been delimited (Mirăuţă, in Patrulius et al., 1973, unpubl. report): (1) Măcin Unit, (2) Consul-Niculiţel Unit and (3) Tulcea Unit. They are overlapping one another and trend north-eastwards. These units were reconsidered by Săndulescu (1985) as (1) Măcin Nappe, (2) Niculiţel Nappe including (2a) Consul and (2b) Sarica digitations and (3) Tulcea Nappe.

The Măcin Unit consists of several entities: pre-Silurian metamorphic formations, Silurian-Carboniferous very low grade metamorphic rocks (Carapelit Formation included) and Mesozoic (Triassic) sedimentary formations with associated acid and basic igneous rocks.

The Consul Unit includes mainly calcareous Triassic sedimentary formations overlying a crystalline basement of Măcin type (Boclugea Series) and associated acid and basic eruptive rocks.

The Niculiţel Unit consists of a basic volcano-sedimentary formation and Upper Triassic flysch deposits.

The Tulcea Unit is built of prealpine metamorphic formations (Precambrian, Silurian and Devonian) and Mesozoic sedimentary formations, both associated with acid and basic igneous rocks.

An Upper Cretaceous post-tectonic sedimentary cover occurs in the southern part of all these structural units.

2.2 Igneous rocks: occurrence and age

2.2.1 Măcin Unit

a) In the NW of the unit Paleozoic igneous rocks are represented by important granitic intrusions, some older than the Carapelit Formation, others subsequent to it (Rotman, 1914). Acid volcanics related to the Carapelit Formation are also known (Mrazec, Pascu, 1896; Pascu, 1904; Murgoci, 1914; Mirăuţă, Mirăuţă, 1962). They occur mainly in the Horia-Balabancea area and they are interlayered with continental terrigenous sedimentary deposits of alluvial fan type (Drăgănescu, in Russo-Săndulescu et al., 1975, unpubl. report; Seghedi, Oaie, 1986; Seghedi et al., 1987). They include mainly subaerial pyroclastics (ignimbrites), epiclastics and subordinately lava flows. Their petrographic features were described by Russo-Săndulescu et al. (1975, unpubl. report), Seghedi et al. (1985, 1987), Ştefan, in Berbeleac et al. (1985, unpubl. report) and in Nedelcu et al. (1986, unpubl. report).

b) The Mesozoic igneous rocks are represented by both acid (granites, alkaline rhyolites, rhyolites) and basic rocks. In the north-eastern half of the Măcin Unit numerous basic dykes and veins piercing pre-Mesozoic formations have been mapped. Their trend is generally NW-SE; conformably to the main structural trends of this unit. The Triassic age is inferred from both geometrical relations to the Paleozoic formations (Seghedi et al., 1980; Seghedi, Oaie, 1986) and their petrographical and petrochemical similarity with the more accurately dated rocks of the Consul Unit (Seghedi et al., 1985). In the south-western area of the Măcin Unit the rhyolite occurrences show an elongated pattern (the "southern alignement") parallel to the Peceneaga-Camena fracture zone; from north to south they are grouped in three zones: Turcoaia, Cîrjelari and Camena.

In the Turcoaia-Iacobdeal area the rhyolites are intimately associated to granites exhibiting similar general chemical features (alkaline rhyolites and granites). This genetic relation between these rhyolites and granites was unanimously accepted by the previous authors (Mrazec, 1899; Cantuniari, 1912; Ianovici et al., 1969; Întorsureanu, 1987; Întorsureanu et al., 1989). Their common petrographic feature is the presence of alkaline mafic minerals, as riebeckite and aegirine. Rhyolites occur as dykes, small shallow intrusive bodies or marginal facies in some granitic intrusions. Their age as well as the age of all the acid igneous rocks in the NW of the Măcin Unit is controversial. They were initially considered as Hercynian (Cantuniari, 1912; Mirăuţă, Mirăuţă, 1962) and this was also supported by some K/Ar ages of rhyolite and alkaline granite samples from Turcoaia-Iacobdeal (Minzatu et al., 1975; Mureşan, 1975). According to more recent results they seem to be younger, i.e. Mesozoic. A Rb-Sr isochrone of 193±15 Ma points to the Triassic-Jurassic boundary as the time of their emplacement (Pop et al., 1985; Întorsureanu et al., 1989).

In the Cîrjelari area the rhyolites are associated with alkaline granitic rocks which occur in more limited areas than in the Turcoaia area. Their petrographic and chemical features are presented by Dimitrescu (1959), Constantinescu et al. (1985, unpubl. report), Stiopol, Drăghici (1978), Ştefan, Roşu in Berbeleac et al. (1985,

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unpubl. report), in Lupu et al. (1986, unpubl. report) and in Mounazih (1988, unpubl. thesis of doctor's degree). The rhyolites occur as both subvolcanic bodies and effusive as well as explosive products. Like in the Turcoaia area, the age of the acid rocks is poorly elucidated. Rb-Sr isochrones show 197±10 Ma (Pop et al., 1985), almost identical to the Rb-Sr age of Turcoaia rocks exhibiting similar chemical and petrogenetic features. According to Ştefan and Roşu (in Nedelcu et al., 1987, 1988, unpubl. reports) the Upper Jurassic age is very likely for the Cîrjelari rhyolites due to their resemblances to certain rhyolites from Camena. On the other hand rhyolitic tuffs are interlayered with Upper Jurassic sedimentary deposits (Grădinaru, 1981).

In the Camena area the rhyolites occur along the Peceneaga-Camena fault zone. According to several authors these rhyolites are grouped in two parallel alignments of similar or different evolution (Cădere, 1925; Constantinescu et al., 1978, unpubl. report; Ştefan in Berbeleac et al., 1982, 1985, unpubl. reports). Unlike the Turcoaia and Iacobdeal areas here no granitic rocks occur in association with rhyolites. However, a basaltic rock occurrence is noticed in the Başpunar quarry (Grădinaru et al., 1981). The mentioned authors consider the Camena rhyolites of both subvolcanic and effusive or pyroclastic type.

Not all the authors agree on the age of the Camena rhyolites either. Former K-Ar ages of 212–232 Ma (Mînzatu et al., 1975) differ from the recent ones of 147–153 Ma (Ştefan in Nedelcu et al., 1988, unpubl. report). According to micropaleontologic investigations the sedimentary deposits with interlayered rhyolites in the Başpunar quarry are Oxfordian-Kimmeridgian in age (Grădinaru, 1981). In the Camena area the rhyolites associate with Ladinian-Carnian sedimentary deposits (Grădinaru, 1981) and other deposits of uncertain age, assigned – on palynologic basis (Antonescu in Constantinescu et al., 1985, unpubl. report) – to the Spathian and the Middle Jurassic (Constantinescu et al., 1985, unpubl. report), the Middle-Upper Jurassic (Grădinaru, 1984), the Middle Jurassic (Mureşan et al., 1987, unpubl. report), the Spathian (Mirăuţă, Visarion in Nedelcu et al., 1988, unpubl. report).

2.2.2 Consul Unit

Widespread rhyolite occurrences are related to the Lower Triassic mainly carbonate deposits of this unit. They prevail between Valea Teilor to the north and Nicolae Bălcescu to the south. From Valea Teilor northwards the rhyolites associate with an increasing amount of basaltic rocks of Triassic age. The southernmost occurrences are found at Mihai Bravu.

There are several opinions on the nature of the rhyolites. They are assigned as intrusive (Murgoci, 1912; Ştefan in Nedelcu et al., 1986, unpubl. report), effusive (possibly partly intrusive) and pyroclastic (Savul, 1935; Stiopol et al., 1975, unpubl. report; Caraveţeanu in Vîlceanu et al., 1980, unpubl. report) or ignimbritic (Constantinescu et al., 1978, 1981, 1982, unpubl. reports). The mainly effusive and subordinately explosive character of the rhyolitic rocks in this unit has been recently demonstrated (Seghedi et al., 1990). Their Lower Triassic (Spathian-Anisian) age is unanimously accepted at present (Seghedi et al., 1986, unpubl. report; Mirăuţă et al., 1986, unpubl. report).

2.2.3 Niculiţel Unit

A typical feature of this unit is the large development of the Upper Triassic basic igneous rocks accounted as withinplate basalts (Savu et al., 1985, 1986) or ophiolitic rift basalts in plate margin environment (Cioflica et al., 1980). The unit is practically devoid of rhyolites. Although on some maps the Isaccea rhyolites are reported to the Niculiţel Unit (Savu et al., 1985), their assignment to the Tulcea Unit is more appropriate considering the geological background (Săndulescu, 1984).

The rhyolites in the Trestenic area are related to Lower Triassic deposits and constitute an outlier assigned to the Consul Unit, while the acid and basic dykes from the Cilic area belong to an outlier of the Măcin Unit (Mirăuţă in Patrulius et al., 1973, unpubl. report).

2.2.4 Tulcea Unit

The Tulcea Unit contains a small number of rhyolitic rock outcrops which occur mainly in the Somova-Mineri area and east of Cataloi (in the Rediu and Uzumbair hills)(Mirăuţă, 1966; Seghedi, Uricaru, 1985). Minor occurrences are found also at Isaccea and at Muchia Păstorului, a place situated between Izvoarele and Nalbant. Paleozoic granites are also reported. While the rhyolites from Somova-Mineri-Isaccea and Muchia Păstorului are considered of Triassic age because they are associated with carbonate rocks of the same age, those east of Cataloi which pierce Paleozoic formations are implicitly assigned to the Paleozoic (Mirăuţă, 1966).

The rhyolites from Somova-Mineri and Isaccea occur in different facies from subvolcanic to effusive and ignimbritic (Savul, 1931; Stiopol et al., 1975, unpubl. report, 1978; Savu et al., 1982, 1985, 1986; Ştefan, Roşu, Gridan in Nedelcu et al., 1986, unpubl. report). They are related to basic of continental within plate type (Savu, 1986). At Muchia Păstorului the rhyolites are effusive and pyroclastic.

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TABLE 1

MAJOR ELEMENTS

GROUP.	NO	SAMPLE	LOCATION	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	K_2O	Na_2O	P_2O_5	H_2O^+	S	CO_2	TOTAL
IA	1.	159	4 Hotare hill	77.59	0.24	11.24	2.23	0.19	0.02	0.43	0.17	6.30	0.71	0.07	0.96	0.09	-	100.24
	2.	161 A	" "	76.47	0.10	12.24	1.26	0.31	0.02	0.23	0.11	8.25	0.60	0.01	0.64	0.10	-	100.40
	3.	161 B	" "	75.24	0.13	11.90	1.30	0.38	0.01	0.19	0.12	8.70	0.63	0.03	0.85	0.06	-	99.54
	4.	173	Crapoea hill	77.70	0.24	10.84	0.96	0.40	0.03	0.16	0.30	6.64	1.54	0.03	0.60	0.08	-	99.52
	5.	324	4 Hotare hill	80.30	0.22	9.64	1.77	0.29	0.04	0.40	0.23	4.82	0.95	0.08	1.14	0.12	-	100.10
	6.	355	" "	81.50	0.23	10.50	1.08	0.14	0.03	0.09	0.12	0.18	5.62	0.14	0.22	0.10	-	100.22
	7.	358	" "	79.10	0.23	9.80	0.59	0.22	0.06	0.18	0.97	5.42	1.38	0.16	0.69	0.31	0.36	99.74
	8.	364	" "	75.40	0.12	12.17	0.58	0.18	0.01	0.16	0.24	9.58	0.17	0.10	0.42	0.16	0.16	99.59
	9.	401	Crapoea hill	70.25	0.28	12.05	1.55	0.22	0.18	0.24	3.00	2.86	4.63	1.87	0.52	-	1.94	99.59
	10.	2405	Cetate hill	80.00	0.04	12.65	0.09	0.15	0.01	0.09	0.17	0.11	6.65	0.04	0.08	0.09	-	100.17
	11.	3123	Martina V.	79.00	0.02	12.85	0.33	0.52	0.03	0.35	0.18	0.19	6.28	-	0.21	0.14	0.02	100.22
	12.	3124	" "	79.50	-	12.70	0.16	0.18	-	0.09	0.17	0.14	6.42	0.10	0.06	0.15	-	99.67
	13.	3196	Cetate hill	79.50	0.04	12.45	0.16	0.18	0.02	0.15	0.15	0.11	6.45	0.04	0.13	0.09	0.04	99.50
	14.	3197	" "	78.70	0.02	12.80	0.19	0.22	0.01	0.09	0.25	0.11	6.71	0.08	0.21	0.13	-	99.52
	15.	3254	Crapoea hill	72.50	0.27	11.98	2.64	1.71	0.09	0.49	0.20	7.83	0.19	0.06	1.41	0.14	-	99.51
	16.	3255	" "	78.00	0.18	11.30	0.93	0.30	0.02	0.33	0.20	7.61	0.60	0.02	0.68	0.12	-	100.19
	17.	3258	Secaru hill	76.93	0.18	12.64	1.43	0.57	0.06	0.53	0.20	0.06	6.45	0.12	0.50	0.07	-	99.76
	18.	3259	" "	75.95	0.14	13.60	1.66	0.39	0.01	0.18	0.12	0.04	7.24	0.05	0.20	0.06	-	99.64
	19.	3260	" "	76.91	0.13	12.34	2.19	0.43	0.01	0.26	0.20	0.06	6.58	0.03	0.50	0.06	-	99.70
	20.	3261	" "	75.89	0.18	13.04	1.43	0.84	0.02	0.40	0.15	0.16	6.87	0.05	0.50	0.05	-	99.58
	21.	3200	Ioaneşul hill	79.50	0.03	12.35	0.06	0.30	0.01	0.18	0.38	0.14	6.15	0.12	0.23	0.09	0.08	99.62
IB1	22.	113	Geaferca Rusă	74.05	0.39	11.84	0.61	1.69	0.06	0.59	0.30	6.73	2.87	0.08	0.50	0.12	-	99.83
	23.	114	" "	74.13	0.34	11.69	-	2.20	0.05	0.50	0.28	6.91	2.69	0.03	0.60	0.10	-	99.52
	24.	120	" "	75.29	0.35	11.64	1.53	0.63	0.05	0.22	0.32	5.69	3.37	0.05	0.40	0.06	-	99.60
	25.	122	Islam Geaferca	75.60	0.31	11.14	1.42	0.49	0.03	0.14	0.18	8.66	0.78	0.04	0.70	0.09	-	99.59
	26.	124	" "	73.97	0.45	11.44	1.76	0.92	0.04	0.51	0.24	8.48	1.00	0.10	0.55	0.05	-	99.51
	27.	128	" "	70.21	0.62	12.24	1.83	2.46	0.05	0.77	0.31	9.20	0.64	0.16	1.00	0.09	-	99.58
	28.	131	" "	74.95	0.37	11.70	0.28	1.75	0.05	0.44	0.25	7.44	1.41	0.07	0.80	0.08	-	99.59
	29.	133	" "	70.32	0.54	12.84	-	4.31	0.07	1.48	0.40	6.08	1.97	0.13	1.30	0.13	-	99.57
	30.	136	Pietroiu Mare	71.04	0.55	12.00	2.19	0.68	0.03	0.74	0.24	10.97	0.23	0.13	1.15	0.04	-	99.99
	31.	137	" "	74.77	0.43	11.00	-	2.01	0.04	0.59	0.18	9.59	0.22	0.09	0.50	0.16	-	99.58
	32.	138	" "	72.55	0.44	12.05	2.14	-	0.02	0.15	0.12	11.10	0.20	0.10	0.55	0.09	-	99.51
	33.	178	Stipanu V.	76.20	0.14	10.90	1.78	0.31	0.02	0.47	0.20	6.38	2.36	-	0.76	0.08	-	99.60
	34.	203	Cuanac hill	74.80	0.19	11.35	1.18	0.79	0.02	0.65	0.20	6.81	2.39	-	0.99	0.06	-	99.43
	35.	204	Pietroiul hill	76.30	0.22	12.20	1.19	0.64	0.01	0.17	0.63	3.79	3.92	0.12	0.22	0.22	-	99.63
	36.	208	" "	76.00	0.20	12.10	1.30	0.26	0.02	0.11	0.32	7.09	1.85	0.11	-	0.13	0.35	99.84
	37.	213	Cuanac	76.90	0.22	11.40	0.80	1.26	0.02	0.59	0.19	6.45	1.90	0.01	0.26	0.06	-	100.06
	38.	2985	Martina V.	76.10	0.10	12.94	0.20	0.50	0.01	0.12	0.30	1.13	2.44	0.12	2.92	0.20	0.89	99.73
	39.	3053	Ditcova hill	75.88	0.14	12.24	1.14	0.34	0.02	0.28	0.14	5.04	3.52	0.03	0.70	0.06	-	99.53
IB2	40.	3289	Cilic	68.10	0.82	13.33	4.31	1.74	0.15	0.85	1.11	4.47	3.47	0.18	1.19	0.16	0.16	100.18
	41.	3290	"	62.70	1.14	13.40	7.11	2.22	0.09	1.45	0.99	5.83	2.49	0.17	0.98	0.12	0.72	99.53

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Table 1 (continued)

	42.	214	Bujorul Bulgărесc hill	77.50	0.22	10.30	1.77	1.39	0.07	0.11	0.13	4.05	3.71	0.08	-	0.11	0.33	99.77
C1	43.	215	Negoiu V.	76.50	0.20	10.80	1.71	1.39	0.05	0.10	0.07	4.27	4.00	0.05	0.06	0.11	0.15	99.92
	44.	318	S.Turcoaia	76.40	0.28	11.27	2.04	0.14	0.05	0.22	0.52	4.15	3.58	0.09	0.10	0.38	0.34	99.89
	45.	319	"	76.44	0.32	11.80	1.45	0.29	0.03	0.24	0.22	5.78	2.54	0.10	0.33	0.30	-	100.10
	46.	325	Turcoaia	75.50	0.30	11.25	3.24	0.26	0.09	0.33	0.31	3.76	3.73	0.02	-	0.13	0.67	99.59
	47.	158	Secaru V.	76.33	0.13	12.09	1.77	0.62	0.02	0.11	0.10	3.88	4.13	0.14	0.10	0.12	-	99.53
	48.	351	Cîrjelari	73.20	0.33	11.95	2.51	0.40	0.01	0.25	0.05	9.40	0.28	0.08	0.75	0.35	-	99.86
C2	49.	352	"	75.70	0.24	11.34	3.35	0.18	0.10	0.11	0.31	4.62	3.50	0.07	0.35	0.19	0.04	100.27
	50.	353	"	75.10	0.39	12.45	1.85	0.66	0.07	0.25	0.18	5.16	3.38	0.16	0.26	0.25	-	100.38
	51.	354	"	76.10	0.30	12.00	1.64	0.29	0.03	0.16	0.12	4.56	3.55	0.10	0.55	0.28	-	100.22
	52.	150 A	Taş Bair	75.22	0.19	11.09	2.70	0.26	0.04	0.25	0.12	9.33	0.26	0.04	0.67	0.08	-	100.25
	53.	150 B	" "	78.30	0.18	10.54	2.16	0.16	0.04	0.33	0.14	7.49	0.32	0.07	0.66	0.08	-	100.47
	54.	150 E	" "	72.00	0.32	13.34	3.82	0.07	0.04	0.30	0.02	9.01	0.32	0.03	0.80	0.09	-	100.16
	55.	151	Taş Bair NE	75.09	0.14	11.79	2.69	0.07	0.02	0.35	0.06	8.56	0.31	0.10	0.57	0.06	-	100.41
	56.	152	Camena	77.56	0.12	11.69	1.94	-	0.03	0.10	0.08	5.27	2.91	0.06	0.65	0.05	-	100.46
	57.	155 A	Başpunar	68.58	0.12	13.34	3.51	0.19	0.08	0.17	4.02	2.10	4.76	0.75	0.85	0.05	1.80	100.32
	58.	156	Slava Rusă	80.02	-	10.54	0.95	-	0.09	0.09	0.19	6.46	1.23	0.13	0.70	0.09	-	100.49
	59.	3516	Taş Bair	74.60	0.30	11.65	1.67	0.15	0.03	0.16	0.70	9.69	0.25	0.54	0.32	0.08	0	100.14
	60.	3517	Taş Bair E	77.00	0.20	11.15	1.11	0.18	0.01	0.18	0.36	8.58	0.19	0.28	1.07	0.06	0	100.37
C3	61.	3518	Taş Bair E	75.00	0.26	11.05	1.91	0.18	0.01	0.18	0.21	9.27	0.19	0.78	1.21	0.08	0	100.15
	62.	3521	Taş Bair W	76.70	0.30	10.35	2.09	0.15	0.03	0.11	0.43	8.67	0.18	0.57	0.14	0	0.05	99.77
	63.	3522	Taş Bair W	76.00	0.16	12.35	0.26	0.48	0.02	0.18	0.98	6.79	2.34	0.46	0.24	0.08	0	100.34
	64.	3527	Taş Bair W	76.00	0.21	11.85	0.65	0.36	0.03	0.11	0.70	7.12	1.81	0.36	0	0.05	0.09	99.34
	65.	3531	Holdurmi Bair	75.00	0.22	11.25	1.80	0.26	0.01	0.19	0.34	8.59	0.18	0.18	0.71	0.06	1.09	99.88
	66.	3535	"	75.00	0.30	11.05	1.61	0.26	0.03	0.32	0.59	8.31	0.14	0.50	1.57	0.04	0	99.72
	67.	3538	"	76.50	0.28	10.10	2.77	0.11	0.03	0.26	1.38	7.51	0.16	0.44	0.42	0.07	0	100.03
	68.	3542	Camena V.	75.00	0.34	10.55	2.56	0.15	0.02	0.14	0.64	8.42	0.25	0.90	0.82	0.08	0	99.87
	69.	3542 A	Camena V.	78.50	0.29	10.45	0.61	0.44	0.01	0.15	0.51	8.18	0.18	0.36	0.42	0.08	0	100.18
	70.	3543	Podarniţa V.	76.00	0.24	11.65	1.77	0.33	0.03	0.21	0.43	5.43	2.23	0.73	0.78	0.08	0	99.94
	71.	3544	Uspenia V.	78.00	0.26	8.80	3.10	0.18	0.01	0.06	0.22	7.60	0.14	0.61	0	0.09	0.73	99.80
	72.	10	Consul hill	79.15	0.14	10.89	0.32	0.13	0.01	0.05	0.18	5.25	2.74	0.07	0.60	0.06	-	99.59
	73.	12	"	77.09	0.18	11.69	1.00	0.37	0.01	0.11	0.16	5.16	3.21	0.05	0.50	0.06	-	99.59
	74.	14	"	79.90	0.20	10.69	0.89	0.65	0.02	0.26	0.25	0.93	4.90	0.04	0.80	0.05	-	99.58
	75.	16	"	74.96	0.15	11.04	-	4.05	0.08	0.60	0.12	4.66	2.44	0.05	1.33	0.06	-	99.54
	76.	20	"	79.40	0.13	10.34	0.72	0.25	0.02	0.13	0.17	4.66	3.00	0.04	0.60	0.07	-	99.53
	77.	36	Lozova hill	77,90	0.27	11.34	0.91	0.90	0.02	0.10	0.33	2.67	4.60	0.07	0.40	0.07	-	99.58
	78.	37	"	75.43	0.22	12.19	1.33	0.48	0.01	0.16	0.15	6.32	2.64	0.06	0.55	0.05	-	99.59
	79.	42	"	77.61	0.22	10.89	0.47	0.67	0.01	0.24	0.16	7.41	0.67	0.06	1.09	0.06	-	99.56
	80.	44	"	73.97	0.43	12.74	0.59	1.06	0.02	0.16	0.22	5.59	4.23	0.07	0.40	0.05	-	99.52
II	81.	46	"	76.02	0.20	11.84	0.27	1.25	0.02	0.18	0.15	6.75	2.10	0.05	0.70	0.07	-	99.61
	82.	56	Consul hill	79.40	0.14	10.44	1.32	0.35	0.02	0.13	0.23	4.57	2.35	0.05	0.70	0.08	-	99.78
	83.	62	Eschibalik hill	76.20	0.16	11.54	0.26	1.21	0.05	0.12	0.21	7.94	2.14	0.05	0.45	0.15	-	100.48
	84.	78	Malciu hill	74.75	0.25	11.44	1.83	0.83	0.05	0.11	0.24	6.07	3.11	0.05	0.70	0.10	-	99.53
	85.	82	" "	74.78	0.38	11.64	-	2.46	0.07	0.51	0.35	6.53	1.73	0.04	1.10	0.07	-	99.63
	86.	98	Consul hill	78.92	0.32	10.59	0.36	1.66	0.06	0.26	0.30	3.03	3.10	0.05	0.80	0.09	-	99.54
	87.	102	" "	75.08	0.34	11.59	-	2.25	0.06	0.36	0.52	6.59	2.55	0.08	0.40	0.09	0.50	100.41
	88.	141	Eschibalik hill	72.36	0.47	12.60	0.39	2.15	0.08	0.53	0.31	8.45	4.52	0.09	0.55	0.08	-	99.58

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Table 1 (continued)

	89.	147	N.Bălcesou	82.15	0.13	9.84	2,26	-	0.05	0.15	0.09	0.39	4.73	0.08	0.53	0.06	-	100.46
	90.	148	M.Bravu	77.21	0.26	11.74	1.59	-	0.02	0.49	0.18	6.48	1.06	0.14	0.93	0.07	-	100.17
	91.	149	M.Bravu	76.36	0.25	12.19	1.57	0.04	0.02	0.51	0.18	4.93	2.39	0.02	1.17	0.05	-	99.68
	92.	266	Eschibalîk hill	72.40	0.44	12.60	2.80	1.08	0.04	0.59	0.34	5.71	3.12	0.06	0.57	0.14	0.08	100.09
	93.	293	Lozova hill	74.35	0.20	11.98	1.30	0.90	0.04	0.72	0.34	8.96	0.22	0.08	0.52	0.13	0.31	100.16
	94.	312	Malciu hill	76.45	0.20	11.50	1.69	0.22	0.02	0.09	0.24	5.90	2.82	0.06	0.10	0.14	0.10	99.65
	95.	347	Delictaş hill	78.30	0.18	10.30	1.20	0.93	0.02	0.31	0.22	5.07	2.55	0.10	0.57	0.12	-	99.97
III A	96.	146	Isaccea	72.48	0.54	13.24	1.90	0.88	0.02	0.72	0.80	2.81	5.22	0.12	0.71	0.07	-	99.51
	97.	367	Muchia Păstorului hill	75.20	0.21	11.77	0.13	0.14	0.02	0.28	0.31	10.18	0.23	0.10	0.26	0.12	0.29	100.14
	98.	143A	Cîşla hill	74.48	0.52	12.25	0.74	1.57	0.03	0.48	0.50	5.50	2.23	0.12	1.00	1.11	-	99.53
	99.	143B	Cîşla hill	71.22	0.54	13.30	-	2.72	0.03	0.57	0.24	7.14	2.21	0.13	1.30	0.15	-	99.55
	100.	145A	Somova	77.20	0.15	10.04	1.10	0.18	0.04	1.08	0.10	7.04	0.54	0.12	1.26	0.08	-	99.53
	101.	145B	"	79.30	0.13	8.44	1.57	0.01	0.05	0.06	0.01	8.91	0.23	0.14	0.78	0.06	-	99.79
III B	102.	103	Tulcea Monument	61.95	0.61	19.31	1.89	0.36	0.04	0.34	0.15	12.20	0.46	0.09	1.90	0.20	0.36	99.46
	103.	402	Mahmudia	70.10	0.44	15.45	3.56	0.36	0.01	0.64	0.28	3.64	0.56	1.30	3.45	0	0	99.79

The Tulcea Unit also contains some small acid rock occurrences as dykes and veins cutting the Paleozoic formations at Tulcea (Monument) and on the Mahmudia hills; they are assigned to the "porphyries" (Murgoci, 1914; Ianovici et al., 1961; Mirăuţă, 1966). Both mineralogical and chemical features point to an intermediate alkaline composition of these rocks. They are also associated with basic dykes.

3. Petrochemistry

For a geochemical characterisation of Northern Dobrogea rhyolites the following analyses have been carried out: (1) major elements for 103 samples (chemical analyses)(Tab. 1); (2) spectral analyses (emission spectrography) for Pb, Cu, Zn, Ga, Sn, Ni, Co, Cr, V, Sc, Y, Yb, La, Nb, Zr, Be, Ba, Sr for 99 samples; (Tab. 2) (3) non dispersion F X-ray spectrometric analyses for Y, Nb, Zr, Rb, Sr for 91 samples; (Tab. 2)(4) U, Th, K analyses by gamma spectrometry for 93 samples; (Tab. 2)(5) rare earth elements (La, Ce, Sm, Eu, Tb, Lu) analyses by neutron activation method for 23 samples (Tab. 3). A relevant number of analyses was envisaged for all the structural units and all the occurrence types both Paleozoic and Mesozoic. Excepting the Cîrjelari area, all the other rhyolite occurrence areas in Northern Dobrogea are represented by a number of analyses roughly proportional to their outcrop areas.

In order to facilitate the presentation of analytical data the following notation is proposed:

I. Măcin Unit
I.A. Intracarapelitic rhyolites
I.B.1. Triassic rhyolites dykes from the east of the unit
I.B.2. Triassic dacite dykes from the Cilic outlier
I.C. Triassic-Jurassic rhyolites from the west of the unit ('southern alignment')
I.C.1. Turcoaia area
I.C.2. Cîrjelari area
I.C.3. Camena area
II. Consul Unit - Triassic rhyolites of the Consul Nappe
III. Tulcea Unit
III.A. Triassic rhyolites

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III.B. Paleozoic trachytes

3.1 Major elements

As far as the major chemistry of rhyolites from the main occurrence areas is generally known from previous studies, the chemical data will be discussed by comparison. Thus, most of the analyses are grouped on the TAS diagram (Fig. 1 a, b, c) in the rhyolite field. The plotting fields of rhyolites assigned to different units and to different areas within them are practically superposed with no relevant differences between Paleozoic and Mesozoic rocks on the one hand and between different structural units and occurence areas on the other hand.

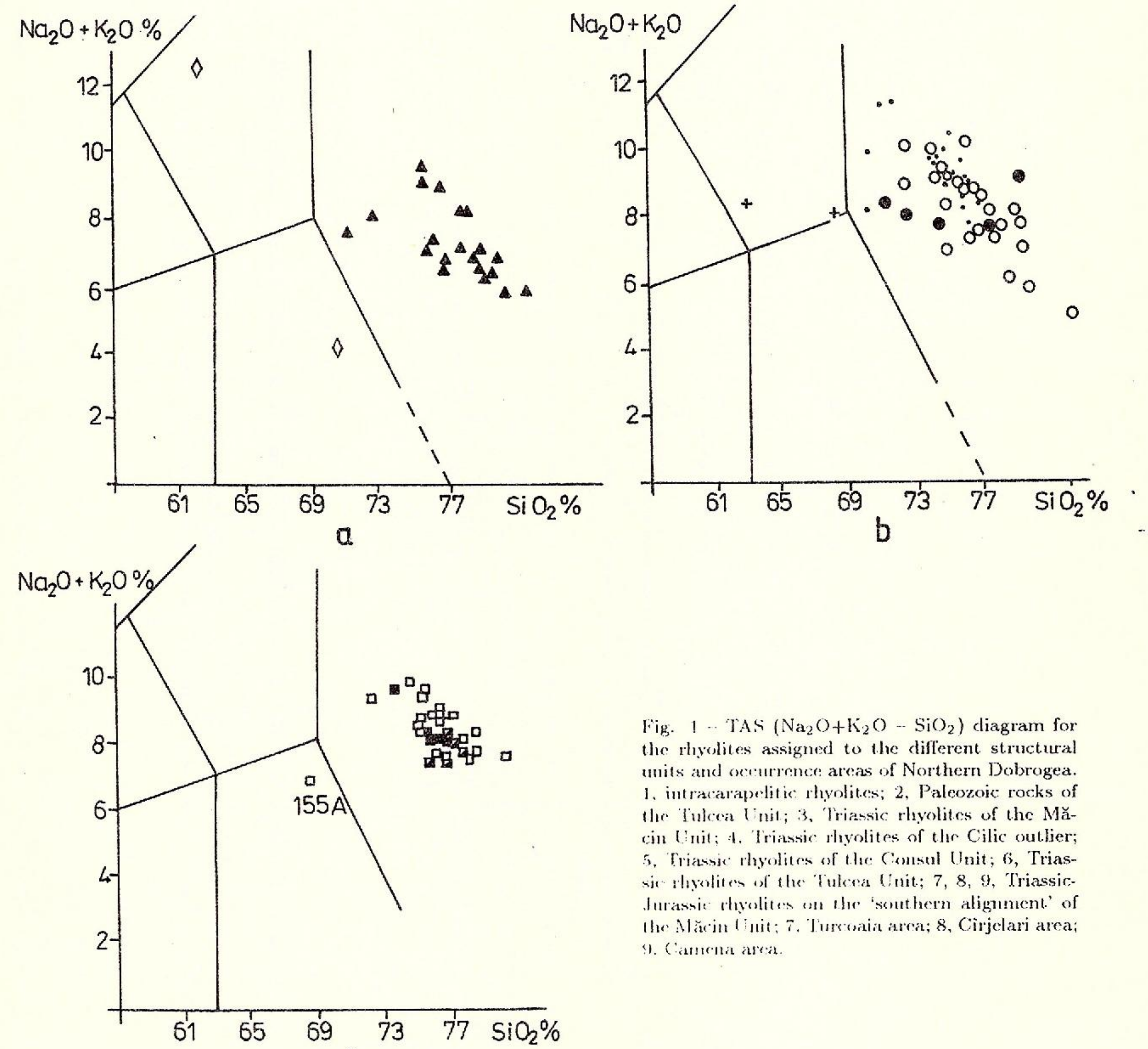

Fig. 1 – TAS (Na_2O+K_2O – SiO_2) diagram for the rhyolites assigned to the different structural units and occurrence areas of Northern Dobrogea. 1, intracarapelitic rhyolites; 2, Paleozoic rocks of the Tulcea Unit; 3, Triassic rhyolites of the Măcin Unit; 4, Triassic rhyolites of the Cilic outlier; 5, Triassic rhyolites of the Consul Unit; 6, Triassic rhyolites of the Tulcea Unit; 7, 8, 9, Triassic-Jurassic rhyolites on the 'southern alignment' of the Măcin Unit; 7, Turcoaia area; 8, Cîrjelari area; 9, Camena area.

Most of the analysed rocks are marked by high silica contents and by rather high alkalinity. There is a negative correlation between SiO_2 content and the amount of alkalis. If we extend the calc-alkaline/alkaline field delimiting line to the field of rhyolites on this diagram, we find that most of the rocks under discussion enter the calc-alkaline field, but close to the alkaline one. The Paleozoic rocks of the Tulcea unit are plotted on the field of trachytes (Monument) or dacites (Mahmudia Hill). The Triassic rocks of the Cilic outlier are marked by their low silica and high alkaline composition entering the field of trachydacites. They obviously

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TABLE
TRACE

Group	No.	Sample	Pb	Cu	Zn	Ga	Sn	Ni	Co	Cr	V	Sc	Y	Y*	Yb	La	Nb
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
I A	1	159	19	10	140	22	6	<2	<2	1	<2	<2	110	59	7	67	23
	2	161 A	15	9	<30	15	5.5	<2	<2	1	<2	2	24	52	2.2	32	13
	3	161 B	70	13	40	17	6.5	<2	<2	1.5	2	3.5	44	52	3.4	34	14
	4	173	60	44	30	14	6	<2	2.5	1.5	5	3	50	64	3.7	44	12
	5	324	<2	3	–	10	<2	4	2	1	8.5	4	68	67	6.8	45	28
	6	355	<2	4	–	3	2	6.5	<2	5	10	4	60	57	6	42	23
	7	358	<2	6.5	–	8.5	<2	<2	<2	2	9	4	55	52	6.5	45	16
	8	364	10	6	–	8.5	<2	2	<2	2	5	4	33	37	5.2	55	32
	9	401	5.5	2.5	–	6.5	<2	3	<2	2	7.5	4	27	29	3.6	50	14
	10	2405	<2	9.5	–	17	3.5	3	<2	1.5	2.5	4.5	55	65	7.2	36	44
	11	3123	3.5	13	–	18.5	11	5	<2	1.5	6	4	55	67	6	72	22
	12	3124	<2	23	–	22	3	2	<2	1.5	<2	4	42	73	6	<30	34
	13	3196	3	26	–	27	7	<2	<2	2	5	12	135	106	8	33	24
	14	3197	3	9	–	19	4	2.5	<2	2.5	5	3.5	70	64	5.5	40	36
	15	3200	3	10	–	26	<2	5.5	<2	3	12	3	46	41	4	30	18
	16	3254	75	37	–	16	5.5	4.5	4	2	8.5	7	70	54	5	95	10
	17	3255	4	12	–	11	5	2	<2	1.5	3.5	3.5	21	28	2.6	40	15
	18	3258	4.5	7	<30	13	3	4.5	3.5	4	10	4	34	52	3	40	13
	19	3259	100	50	30	16	3	<2	2	1	6	3.5	19	25.4	2.2	<30	12
	20	3260	5.5	11	<30	13	3	2.5	2.5	2.5	7	2	24	28	2.4	<30	11
	21	3261	36	10	<30	165	5	4.5	6	7.5	12	4	52	64	4	50	15
I B 1	22	113	15	11	70	30	8	3.5	<2	2.5	4	2.5	53	70	6	60	29
	23	114	17	16	75	31	6.5	4	<2	4	4	2.5	56	84	7	62	30
	24	120	6.5	8	<30	27	6	4	<2	2.5	5.5	3	72	81	9	95	32
	25	122	9	44	230	24	6	2	<2	2.5	4	2.5	47	63	6	30	24
	26	124	10	25	230	18	4.5	2.5	2	5	15	5	40	41	3.6	38	14
	27	128	10	70	<30	24	3.5	4.5	5	5	28	6.5	40	39	4	30	17
	28	131	28	27	40	27	5	2.5	2	4.5	8	4	55	67	7	55	20
	29	133	26	23	38	32	9	6	5	10	28	8	50	65	5.2	52	18
	30	136	13	13	30	17	4.5	5	3	5	23	6	45	50	4	46	15
	31	137	170	24	60	15	4	3.5	2.5	6.5	12	3.5	28	39	2.6	30	13
	32	138	15	12	35	15	4	3	3.5	4.5	14	3.5	20	25	2.3	30	10
	33	178	18	10	–	18	5.5	6.5	<2	7	8	2.5	65	–	8.6	30	36
	34	203	25	10	–	16.5	2	5	3	5.5	13.5	5	72	–	7.2	68	19
	35	204	3.5	24	–	30	6.5	4.5	<2	2	5	<2	70	75	65	6	24
	36	208	14	18	–	18	5.5	3.5	<2	3	5	<2	75	64	7	60	30
	37	213	13	4	–	17	2.5	3	8	2.5	65	–	8.6	30	36	–	300
	38	2985	3.5	16	–	17	7	2.5	<2	2	8	5	60	61	5.2	30	30
	39	3053	19	9.5	38	19	4	<2	<2	1.5	2.5	3	32	46	2.6	56	15
	40	3199	20	25	–	20	3	2	<2	2.5	5.5	5	70	51	5	52	28
I B 2	41	3289	12	20	–	18	5.5	5	6	5.5	36	11	70	51	4	55	14
	42	3290	55	28	–	17	7.5	4.5	6	8.5	38	11	70	50	3.7	45	14
I C 1	43	214	20	14	–	28	7.5	4	<2	3	2.5	<2	115	84	14	70	30
	44	215	25	18	–	26	7	2.5	<2	2	4	<2	135	92	8.5	52	30
	45	318	3	4	–	16	2.5	3.5	<2	<2	5	6.5	110	85	10.5	88	36
	46	319	4	3.5	–	18	3.5	3	<2	2	3.5	5.5	90	87	8.5	75	34
	47	3256	3	6.5	–	18	4.5	6.5	<2	4	5.5	2.5	135	110	14.5	90	40
I C 2	48	158	19	10	146	22	6	<2	<2	1	<2	<2	110	97	7	67	23
	49	351	<2	9	–	20	2	3	<2	<2	3	9	85	73	7.7	62	26
	50	352	5	3.5	–	19	3	5.5	<2	<2	3.5	4	110	94	16	65	34

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2
ELEMENTS

Nb *	Zr	Zr *	Be	Ba	Sr	Sr *	Rb *	TiO$_2$(%) *	U	Th	K(%)
18	19	20	21	22	23	24	25	26	27	28	29
23	520	225	2	70	26	24	237	0.260	3.3	12.6	6.4
28	60	142	<1	1250	40	60	311	0.314	4.7	28.3	8.8
26	80	116	<1	1300	24	43	292	0.323	4.3	23.9	7.7
29	150	250	1.6	1300	28	54	217	0.242	5.9	14.1	6.4
22	245	200	-	-	-	13	148	0.356	2.2	14	4.8
20	200	179	-	-	-	61	4	0.317	-	-	-
21	215	182	-	-	-	42	156	0.318	3.1	12.3	4.9
25	135	122	-	-	-	54	275	0.211	3.1	32.3	9.5
20	210	196	-	-	-	121	63	0.207	1.1	11.9	2.5
31	150	129	2.1	30	65	77	-	0.349	4.1	35.2	-
32	330	121	3.8	105	460	108	-	0.201	-	-	-
50	36	69	4.6	16	39	60	-	0.195	-	-	-
21	91	101	-	11	32	48	-	0.091	2.6	21	-
32	115	125	-	12	40	71	-	0.376	2.2	38.5	-
19	150	148	-	16	100	154	3	0.114	2.2	14.6	4
15	230	278	2.4	1000	< 10	26	258	0.275	3.1	13	7.8
19	90	126	2.5	820	< 10	27	242	0.220	0.71	27.4	7.5
24	130	166	<1	28	42	79	10	0.222	1.7	9.7	-
18	150	181	1.1	55	48	101	6	0.150	1	8.7	-
14	140	160	1.3	26	38	80	7	0.345	3.2	12.2	-
16	150	179	1	42	40	99	13	0.152	1.8	13.2	-
28	380	359	3.8	300	36	47	235	0.227	4.9	25.5	5.3
33	480	493	3.8	260	24	41	274	0.184	5	24.2	5.3
33	450	437	4	400	28	45	174	0.266	4	23.3	4.4
28	250	284	1.6	950	36	43	260	0.463	5.3	27.3	8.6
17	280	308	1.3	1700	50	59	211	0.449	4	22.5	8.2
21	320	342	1	1500	72	38	210	0.653	5.2	18.7	8.7
25	300	335	2.5	720	30	41	253	0.391	6.5	23.6	7.7
23	380	394	1	800	46	50	259	0.602	4	22.4	6.9
20	280	303	<1	1800	40	47	255	0.363	1.8	23.4	9.1
18	220	227	<1	2900	55	54	229	0.340	2.9	18.7	8.5
17	200	233	<1	1400	20	29	264	0.351	4	18	10.1
-	300	-	2.6	345	17	-	-	-	3.1	11.8	3.4
-	290	-	3.2	480	23	-	-	-	-	-	-
33	270	333	-	310	44	82	144	0.423	6	27	4
29	210	291	-	380	36	57	252	0.190	6	27	6.8
-	2.6	345	17	-	-	-	-	-	-	-	-
36	80	116	4.7	1300	46	79	129	0.213	6	36.9	3.8
24	100	180	1.3	700	42	90	112	0.218	4.2	27.4	4.4
23	220	185	-	135	50	91	148	0.250	2.1	14.1	3.7
24	320	356	3	700	80	103	159	0.749	3.9	18.2	4.2
22	280	347	2.7	950	65	893	163	0.895	3.7	18.7	5.8
31	1100	1191	-	18	<10	11.7	177	0.221	5.2	24.1	3.9
30	850	971	-	27	<10	10	209	0.258	4.1	26.3	4
30	750	582	-	-	-	32	133	0.226	4.7	20	4
28	610	605	-	-	-	24	163	0.261	3.1	20.5	5.6
28	1000	934	-	-	-	22	132	0.284	1.8	25.9	3.7
32	520	865	2	70	26	42	125	0.256	3.3	13.4	3.7
24	435	492	-	-	-	24	276	0.401	3.2	10.8	8.3
31	740	776	-	-	-	17	156	0.063	3	18.7	4.3

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	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
	51	353	19	4.5	-	15	3.5	3.5	<2	2.5	4.5	7	100	90	13.5	45	27
	52	354	2.5	3	-	17	2.5	3	<2	4.5	3	3.5	78	82	7.7	60	25
	53	150 A	16	25	48	18	4	<2	<2	2	3	3	85	71	6.8	60	22
	54	150 B	17	28	55	19	9.5	<2	<2	1.5	3.5	3	72	78	6.5	74	24
	55	150 E	95	34	63	22	5	<2	<2	<1	2	6	90	75	6	70	20
	56	151	55	8	33	19	6	<2	<2	1	<2	<2	115	96	8	<30	26
	57	152	12	12	68	23	7.5	<2	<2	1	2	<2	90	98	9	<30	26
	58	155 A	30	9	55	15	5	5	<2	1.5	2	<2	42	46	3.7	60	40
	59	156	24	12	30	18	6.5	<2	<2	1	2.5	<2	32	48	3.2	<30	47
	60	3516	15	8	-	12	5.5	5	2	3.5	8	6	85	-	9.5	46	19
	61	3517	7.5	2.5	-	21	6	2.5	<2	1	2.5	<2	90	-	9	<30	30
	62	3518	12	2.5	-	21	6	2	<2	<1	<2	<2	125	-	12	<30	33
IC 3	63	3521	20	3	-	17	6	2	<2	1	<2	3	82	-	9	<30	27
	64	3522	23	3	-	18	2.5	2	<2	<1	<2	3.5	32	-	4.5	34	44
	65	3523	29	2.5	-	20	3.5	2.5	<2	<1	<2	3	23	-	3	<30	42
	66	3531	10	2	-	18	5.5	2.5	<2	1	<2	<2	98	-	10	33	30
	67	3535	12	2.5	-	16	6	2	<2	1	2	4	82	-	9	<30	27
	68	3538	24	3	-	22	6	2	<2	1	<2	3.5	115	-	13	<30	30
	69	3542	17	5	-	18	4	<2	<2	1	3.5	4	90	-	9.5	52	24
	70	3542 A	5	11	-	22	3.5	2	2	1	2.5	3	72	-	8	36	19
	71	3543	5	3	-	15	4	<2	<2	<1	<2	5	85	-	10	<30	30
	72	3544	13	2.5	-	11	4.5	<2	<2	<1	<2	3	120	-	8.5	<30	24
	73	10	14	20	42	20	6	2.5	<2	3	<2	<2	53	64	5.5	55	20
	74	12	9.5	28	60	20	7.5	2	<2	3	2	<2	50	71	4.5	60	17
	75	14	18	30	70	20	8.5	2	<2	2	<2	2	46	67	5.5	<30	17
	76	16	24	21	85	24	6.5	2.5	<2	1.5	2	<2	85	83	7	65	18
	77	20	14	18	48	20	7.5	3	<2	3	2.5	<2	52	74	5	58	20
	78	36	42	22	<30	16	9	3	2	5	9	3.5	50	63	3.8	80	<10
	79	37	18	24	32	17	7	3	<2	4.5	8	3	29	31	2.6	40	<10
	80	42	12.5	19	<30	15	6.5	3	<2	4	14	2.5	21	39	2	42	<10
	81	44	14	25	38	15	6	4	2	4	13	3	32	44	2.8	48	<10
	82	46	34	24	92	20	7.5	2	<2	1.5	3	3	60	67	5.5	50	17
	83	56	60	27	125	18	7.5	3	<2	2	2	<2	52	59	6	50	17
	84	62	100	33	220	17	7	2.5	2	2.5	2	3	46	63	3.6	35	14
II	85	78	18	22	55	23	7	2.5	<2	2.5	<2	2.5	100	105	6.5	55	14
	86	82	28	21	70	25	7.5	<2	<2	2	2	2.5	65	79	5.2	55	17
	87	98	26	22	80	18	7.5	2.5	<2	4	<2	2.5	60	73	5	58	16
	88	102	13	16	50	16	7	<2	<2	3	4.5	3	55	77	5.3	32	13
	89	141	17	17	75	23	5	4.5	3.5	5.5	12	4	33	37	3.2	40	14
	90	147	50	13	38	13	6.5	<2	<2	2	4.5	2.5	95	91	7	90	30
	91	148	36	9	<30	16	4	<2	<2	3	8	4	55	61	5	52	17
	92	149	26	13	32	18	5	2	<2	3	9	5	76	63	5.2	62	16
	93	266	-	-	-	-	-	-	-	-	-	-	-	74	-	-	-
	94	293	-	-	-	-	-	-	-	-	-	-	-	36	5.5	-	12
	95	312	-	-	-	-	-	-	-	-	-	-	-	57	-	-	-
	96	347	-	-	-	-	-	-	-	-	-	-	-	56	-	-	-
	97	143 A	38	21	<30	24	7	5	3	7	26	6	50	51	5	60	15
	98	143 B	30	19	38	24	6.5	6	4.5	13	26	7	52	40	4.7	62	14
IIIA	99	145 A	20	18	30	14	8	<2	<2	1.5	5	2	27	38	2.5	56	18
	100	145 B	180	28	38	7	3.5	<2	<2	1.5	3	<2	23	36	2.2	47	16
	101	146	30	8	44	16	5	<2	3	3	31	5.5	38	46	2.5	55	13
	102	367	45	3	-	9	<2	7	<2	5.5	6	4.5	36	39	4.5	34	17
IIIB	103	103	22	8.5	<30	14	3.5	2.5	<2	3	75	3	13	19	<1	50	<10
	104	402	<2	5.5	-	16	<2	16	3	4	20	4.5	18	27	2.5	50	22

* analysed by non dispersion FX-ray spectrometry

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Table 2 (continued)

18	19	20	21	22	23	24	25	26	27	28	29
30	725	724	-	-	-	21	149	0.443	3.5	20.9	4.9
30	535	616	-	-	-	22	118	0.350	2.6	16.1	4.6
32	570	702	1.7	200	115	29	321	0.232	3.7	19	9.2
32	400	447	2.2	150	110	11	381	0.323	2.1	20.1	7.9
28	520	681	2.2	160	74	48	364	0.399	4	14.6	9.7
37	500	693	2.1	42	<10	13	324	0.222	3.8	18.8	8.4
37	560	774	3.2	36	<10	13	216	0.206	4.2	21.1	4.3
58	140	191	2	170	16	197	80	0.058	4.5	20.8	1.7
61	120	180	2,6	130	13	22	289	0.239	7,5	20.9	5.97
-	440	-	1.4	420	20	-	-	-	3.2	22.5	9.3
-	870	-	2	46	<10	-	-	-	2.7	25.2	8.8
-	900	-	2.3	36	<10	-	-	-	2.8	22.5	9.1
-	700	-	1.9	1300	<10	-	-	-	-	-	-
-	120	-	2.1	100	<10	-	-	-	2.6	29	7.9
-	130	-	2.8	125	<10	-	-	-	2.1	28.5	7.8
-	800	-	2.3	42	<10	-	-	-	4.8	28.4	8.5
-	730	-	1.9	90	<10	-	-	-	30.7	7	6.98
-	880	-	1.5	170	<10	-	-	-	-	-	-
-	620	-	1.5	320	<10	-	-	-	-	-	-
-	480	-	1.4	200	<10	-	-	-	1.6	22	8.1
-	370	-	1.5	280	<10	-	-	-	-	-	-
-	650	-	1	380	<10	-	-	-	-	-	-
32	400	368	2.8	165	38	32	204	0.377	4.5	25.5	5
30	380	395	4	160	46	38	247	0.319	2.9	26.4	4.7
32	310	329	4.4	100	160	191	58	0.260	5.3	25.5	0.5
31	400	358	3.5	160	55	52	166	0.129	1.8	25.3	4.3
31	300	307	4	115	27	25	196	0.273	2.5	22.6	4.2
16	300	234	2.3	170	125	135	92	0.319	4.5	23.5	2.4
16	220	218	1.4	450	62	119	272	0.141	4.9	22.5	5
19	320	239	2.3	950	115	55	223	0.349	2.9	19.5	6.7
20	240	248	1.9	625	48	54	237	0.320	4.5	24.6	4.9
25	400	355	4	600	72	75	257	0.249	6.5	26	6.2
31	460	398	2.7	400	105	110	128	0.414	2.9	21.1	3.7
24	360	313	1.7	1100	43	29	229	0.107	63	21.6	6
31	460	367	3.2	560	66	53	140	0.275	3.3	23.7	5.2
32	540	423	4.2	280	240	199	300	0.321	5.2	24	5.6
26	360	305	4.6	300	60	62	162	0.347	5.4	24.5	2.5
290	350	331	3.8	280	35	26	253	0.376	5.7	25.7	5.3
18	200	228	1.4	1150	26	39	244	0.140	4.2	21.2	7.2
36	750	757	1.7	44	20	37	18	0.229	-	5.8	-
26	220	242	2.1	320	20	31	239	0.588	3	15	5.2
28	250	244	2.2	460	19	44	195	0.346	-	3.9	1
30	-	520	-	-	-	40	203	0.374	4.4	21.3	5.5
16	200	175	-	-	-	75	260	0.332	5.2	23.4	8.3
29	-	334	-	-	-	36	154	0.544	2	24	5.4
29	-	275	-	-	-	39	170	0.238	6.4	24.5	4.9
18	320	327	3.3	700	65	89	251	0.488	7.2	31.4	5.3
18	370	331	4.6	500	55	57	256	0.385	5	33.9	0.7
14	140	170	3.8	480	32	41	326	0.194	10.7	38.3	6.2
15	130	173	1.1	2200	28	41	256	0.443	5.2	33	7.3
30	340	300	2.5	360	75	159	162	0.612	4.7	18.3	2.3
21	165	171	-	-	-	32	152	0.285	-	-	-
21	220	281	1	2000	80	60	514	0.412	4.5	80.4	12.9
23	260	300	-	-	-	72	165	0.505	4.2	31	3

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TABLE 3

RARE EARTH ELEMENTS

GROUP	NO	SAMPLE	La	Ce	Sm	Eu	Tb	Yb	Lu
I A	1	159	24	66	6.2	0.75	0.38	7.0	0.64
	2	173	45	81	8.6	1.40	0.59	3.7	0.24
	3	324	37	59	7.7	1.00	0.81	6.8	0.57
	4	364	45	108	12.7	0.67	0.92	5.2	0.37
	5	355	11	63	2.0	1.29	0.62	6.0	0.36
I B	6	120	35	89	10.8	0.47	0.75	9.0	0.88
	7	128	36	79	6.3	0.80	0.32	4.0	0.41
I C1	8	318	48	122	6.1	1.20	1.04	10.5	0.85
I C2	9	158	44	116	11.6	0.60	0.76	7.0	0.95
	10	351	58	92	12.8	1.11	0.89	7.7	0.65
	11	354	20	109	5.5	0.90	1.00	7.7	0.74
I C3	12	150B	42	58	6.9	1.0	0.63	6.5	0.77
	13	155A	37	85	6.5	0.30	0.61	3.7	0.52
II	14	20	45	107	10.9	0.39	0.77	5.0	0.65
	15	42	47	57	5.3	0.51	0.35	2.0	0.39
	16	78	36	94	9.4	0.29	0.79	6.5	0.81
	17	141	46	73	7.3	0.61	0.50	3.2	0.22
	18	293	45	79	9.4	0.48	0.90	5.5	0.41
	19	147	39	43	11.6	0.92	0.86	7.0	0.62
III A	20	143A	52	81	6.4	0.87	0.75	5.0	0.34
	21	146	25	80	6.0	1.10	0.40	2.5	0.23
	22	367	40	66	10.0	1.07	0.62	4.5	0.31
III B	23	103	46	57	3.0	1.13	0.52	1.0	0.22

differ from the other Triassic rhyolite bodies of the Măcin Unit to which they have been assigned on tectonic grounds.

Quite puzzling is the plotting in the dacite field of one sample collected from the Camena area, different from the other rocks occurring there. This sample is from Başpunar quarry, the only place in which the Jurassic age of some acid volcanics on the 'southern alignment' has been proved (Grădinaru, 1981).

The K_2O/Na_2O ratio reveals the usually potassic character of most of the studied rhyolites. The only areas of relatively high Na_2O content are at Turcoaia and Cîrjelari where the rhyolites are associated with comagmatic alkaline granites. This chemical feature was pointed out by Ştefan (in Nedelcu et al., 1986, unpubl. report) who calls them "alkali-sodic rhyolites" in contrast to all the other rhyolites from Northern Dobrogea ("alkali-potassic rhyolites"). The subunitary K_2O/Na_2O ratio is also characteristic for some intracarapelitic rhyolites and for some rhyolitic rocks in the Consul Unit. The highest K_2O/Na_2O ratio characterizes most of the Camena rhyolites. A typical feature for both Paleozoic and Mesozoic rhyolites from Northern Dobrogea is the usually low Fe_2O_3, FeO, MnO, MgO and CaO contents.

The study of major chemical composition of the Northern Dobrogea rhyolites shows that, except for the high

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potassic character of the Camena rhyolites and the higher Na_2O content of Turcoaia and Cirjelari ones, there are no relevant chemical differences between the rhyolites assigned to various structural units or between Paleozoic and Mesozoic rhyolites. It is worth mentioning that, with respect to the alkali sum and other chemical features the Turcoaia and Cirjelari rhyolites do not differ from the other rhyolite occurrences in Northern Dobrogea (Fig. 1 a, b, c).

3.2 Trace elements

Because some trace elements such as Y, Yb, together with TiO_2, are less mobile during alteration, diagenesis and metamorphism, Winchester and Floyd (1977) proposed a systematic petrographic classification of old igneous rocks based on these elements. The diagrams thus drawn (Figs. 2, 3) lead to the slight modification of the rock assignment by comparison to their assignment according to major chemistry, but point to certain differentiation tendencies among various rock groups. Therefore on the SiO_2-Zr/TiO_2 diagram (Fig. 2) most of the samples belong to rhyolites and rhyodacites with a few exceptions inferred from their major chemical composition. Only some samples collected from the 'southern alignment' of the Măcin Unit belong to alkaline rhyolites, most of them being assigned to rhyolites. However, on the Zr/TiO_2-Nb/Y diagram (Fig. 3) almost all the samples including those from the 'southern alignment' are plotted in the field of rhyolites and rhyodacites. On both diagrams the exceptions are represented by the Paleozoic volcanics of the Tulcea Unit (trachyandesites), the Jurassic rhyolite in the Başpunar quarry (the single sample with alkaline character) and the rocks of the Cilic outlier (dacites). This points out the calc-alkaline chemistry of most of the Northern Dobrogea rhyolites, with special reference to the obvious alkaline tendency of rocks from the 'southern alignment' of the Măcin Unit.

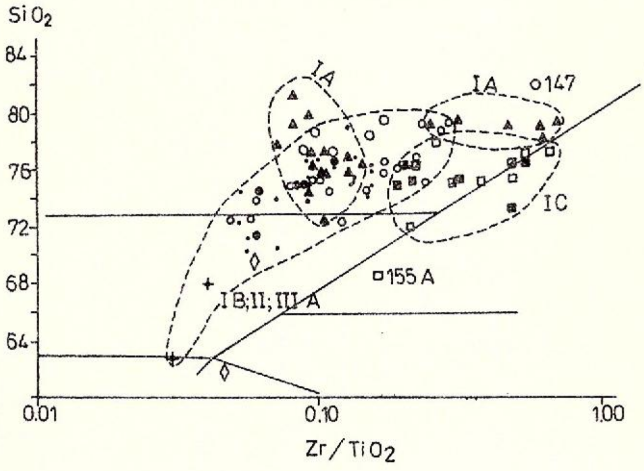

Fig. 2 – SiO_2-Zr/TiO_2 diagram; for symbols see Fig. 1.

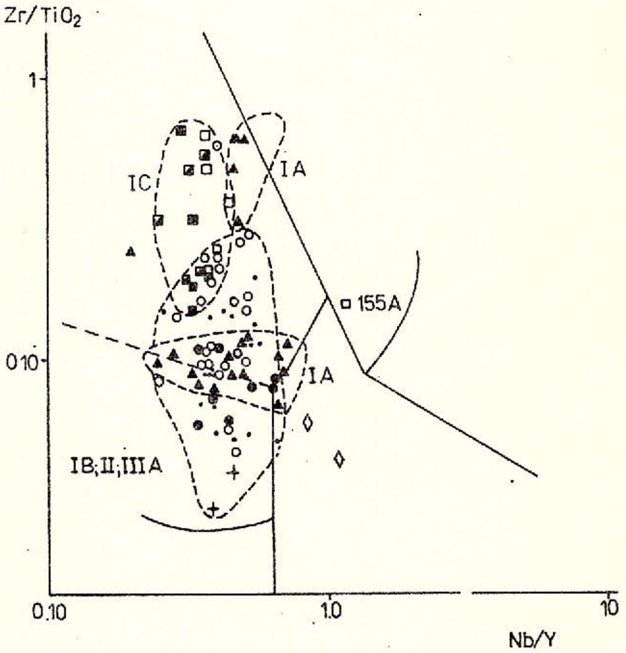

Fig. 3 – Zr/TiO_2-Nb/Y diagram; for symbols see Fig. 1.

The same diagrams (Figs. 2, 3) show the slightly different position of the Paleozoic and Mesozoic rhyolites inferred from the different correlation tendency of the element ratios: the former are marked by a relatively constant Zr/TiO_2 ratio by comparison to the Mesozoic rhyolites with varying Zr/TiO_2 ratio and low Nb/Y ratio. The Mesozoic rhyolites stand out by the superposition of plotting areas of the rhyolites assigned to the dykes and veins of the Măcin Unit and those of both Consul and Tulcea Unit. The rhyolites of the 'southern alignment' of the Măcin Unit show the same correlation trend and have highest Zr/TiO_2 ratio. This seems to be a common feature of all three occurrence areas (Turcoaia, Cirjelari and Camena).

The rhyolitic rocks from the different structural units of Northern Dobrogea also differ with respect to their Rb and Sr contents. Thus, the Sr-Rb diagram (Fig. 4) points to usually lower Sr contents for the rhyolites of

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the 'southern alignment' by comparison to the others. Quite illustrative are the Camena rhyolites exhibiting Rb contents higher than those from Turcoaia and Cîrjelari.

The Triassic rhyolites of groups I B 1, I B 2, II and III A have the same reverse correlation trend for Rb and Sr within their superposed plotting areas. Unlike this, the intracarapelitic rhyolites show positive Rb-Sr correlation with lower Sr values, partly overlapping the plotting area of the Triassic rhyolites. Some of these rhyolites and the Triassic rhyolitic tuff at Nicolae Bălcescu are remarkable by their anomalously low Rb values or even the lack of this element; the former have a higher Sr content. It is also to note the particular plotting of the Jurassic rhyolite from Başpunar quarry (high Sr) and of the Paleozoic trachyte from Tulcea (maximum Rb).

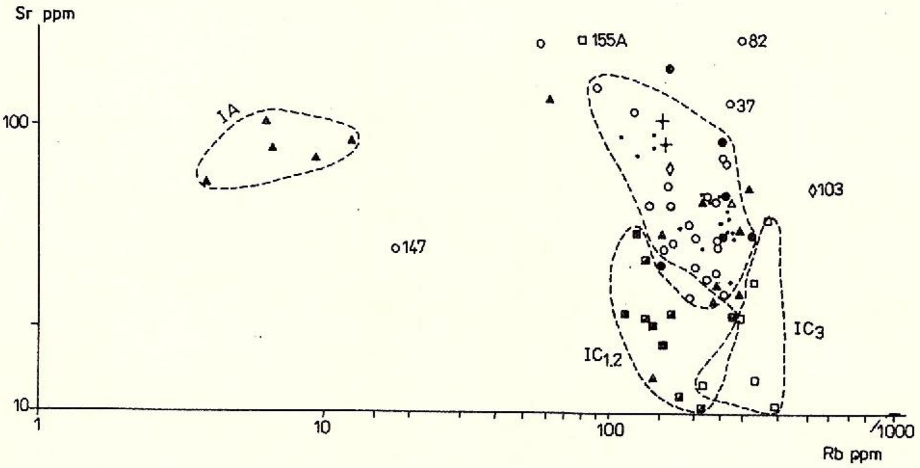

Fig. 4 – Sr – Rb diagram;
for symbols see Fig. 1.

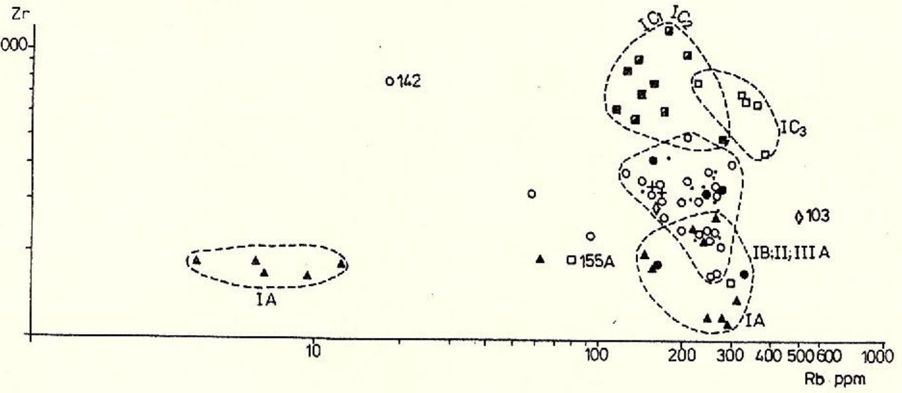

Fig. 5 – Zr – Rb diagram;
for symbols see Fig. 1.

A more accentuated discrimination among the studied rock groups is obtained on the Zr-Rb diagram (Fig. 5) which shows: similar Rb values on the whole, except for certain intracarapelitic rhyolites and the Camena rhyolites (low Zr content in the intracarapelitic rhyolites, practically identical Zr content of Triassic rhyolites for groups IB, II, III), maximum Zr contents in the 'southern alignment' and the peculiar character of the Jurassic rhyolite and the Tulcea trachyte.

Significant differences of trace element distribution within the investigated rock groups are also proved by the Rb–Y+Yb (Fig. 6) and Nb–Y (Fig. 7) diagrams used by Pearce et al. (1984) in order to discriminate the geotectonic environments of the granitic rocks. On both diagrams the Paleozoic trachytes of the Tulcea Unit

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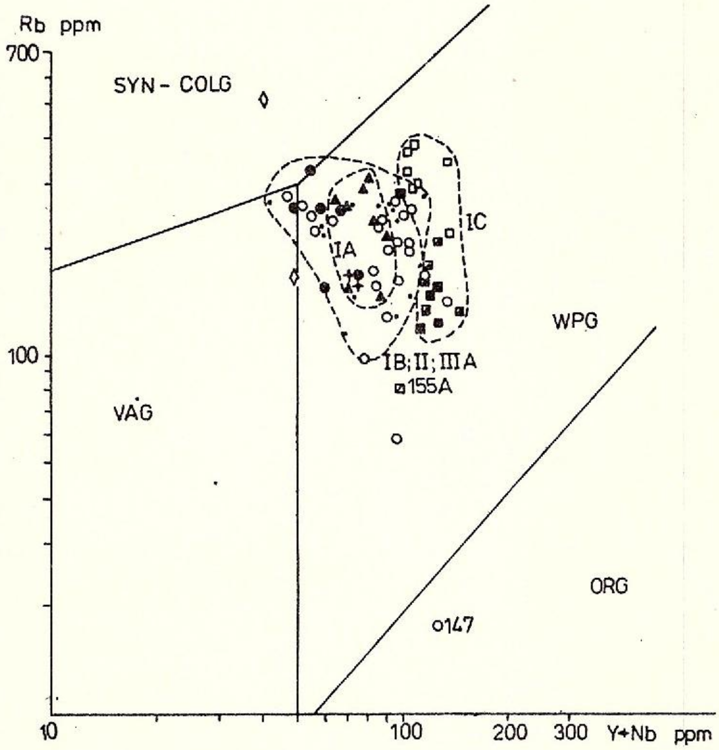

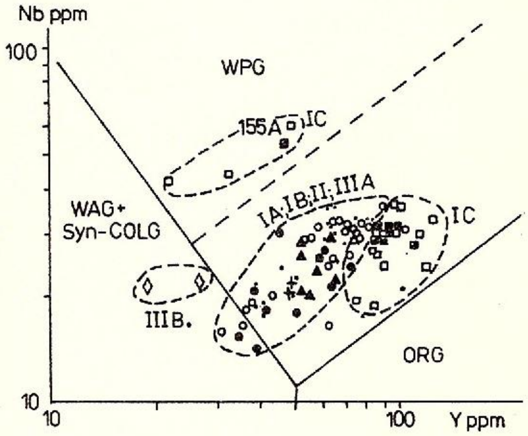

Fig. 6 – Rb – Y+Nb diagram (Pearce et al., 1984); for symbols see Fig. 1. Abbreviations: syn-COLG, syncollisional granites; VAG, volcanic arc granites; WPG, withinplate granites; ORG, oceanic rift granites.

Fig. 7 – Nb – Y diagram (Pearce et al., 1984); for symbols see Fig. 1, for abbreviations see Fig. 6.

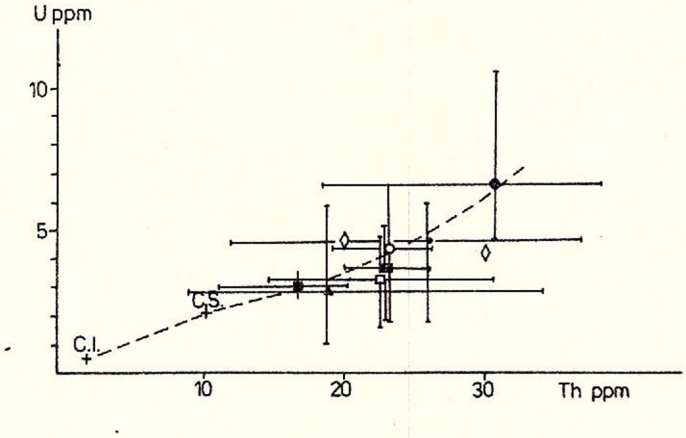

Fig. 8 – U – Th diagram; for symbols see Fig. 1. Abbreviations: CI, lower crust; CS, upper crust.

and the 'southern alignment' Mesozoic rhyolites are standing apart, while the Triassic rhyolite groups (IB1, IB2, II, IIIA) are almost perfectly superposed similarly to the other diagrams used. The intracarapelitic rhyolites are plotted on the same area.

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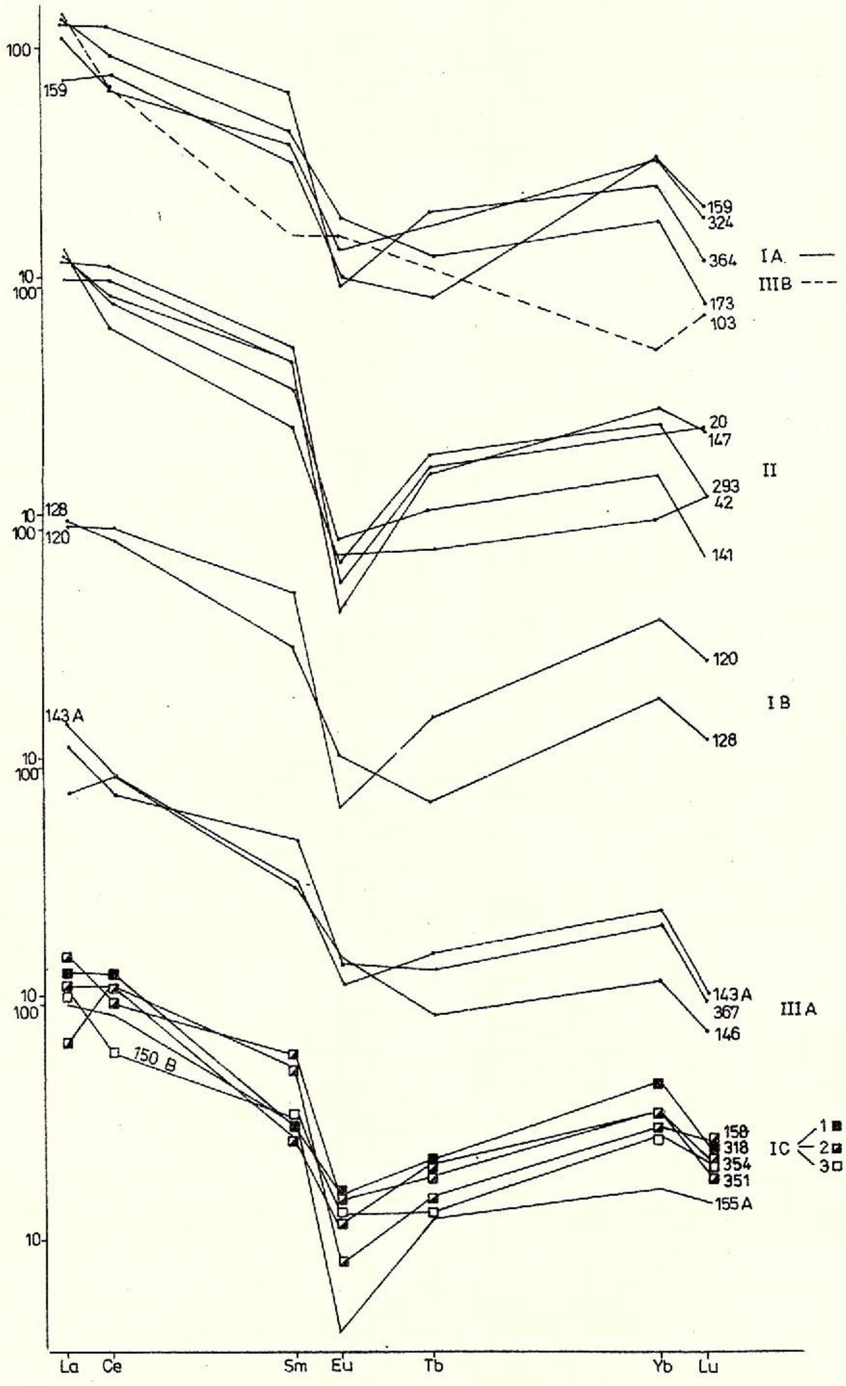

Fig. 9 – Chondrite normalized REE distribution diagrams. IA, intracarapelitic rhyolites; I B, Triassic rhyolites of the Măcin Unit; II, rhyolites of the Consul Unit; III A, Triassic rhyolites of the Tulcea Unit; III B, Paleozoic rocks of the Tulcea Unit.

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Ba is distributed within a wide variation range (18–2 900 ppm) which in the case of these old rocks is accounted for by high Ba mobility during the geological processes. The Triassic rhyolites have higher Ba contents than the other rocks.

U and Th distribution shows that the maximum content of these elements occurs in rhyolites assigned to the Tulcea Unit and the minimum ones within the Cirjelari rhyolites, excepting the intracarapelitic rhyolites with anomalous behaviour on most of the diagrams. Rhyolites of the other units show intermediate values similar for both U and Th; only the intracarapelitic rhyolites have a somewhat lower Th content (Fig. 8). The average U/Th ratios of the studied rhyolite groups, compared to the lower and upper crust values (Taylor and Mc Lennan, cited by Tischendorf et al., 1987) are assigned approximately to the same correlation line which suggests their crustal origin.

The chondrite-normalized REE distribution (Tab. 3, Fig. 9) points to major similarities between the distinct rock groups and to less important differences. The general REE distribution pattern is practically identical for all rocks except for the trachyte from Monument (Tulcea), showing the same light REE fractionation and the lack of fractionation or even the slight increase of heavy REE. The high Yb content is their common feature. The trachyte sample from Monument exhibits fractionation of both light and heavy REE. There are some differences concerning Eu and Tb behaviour: some samples show negative Eu anomaly, others lacking this feature, depending on higher or lower Tb values. These two instances occur differently in all the mentioned groups. Another difference regards the wider variation range of heavy REE, the maximum values being typical of the 'southern alignment' rhyolites. The Paleozoic intracarapelitic rhyolites (IA) with a similar general REE distribution pattern differ however from the Mesozoic rhyolites due to the absence of/or attenuated anomalies. Due to its low Sm content one of the samples (355) which was not plotted on the diagrams shows a false positive Eu anomaly, difficult to interpret.

The minor element distribution within Northern Dobrogea rhyolites allows significant remarks and discriminations which seem to solve some controversial problems of the chronological and genetic systematisation of acid rocks from different occurrences. Therefore, considering that on most of the diagrams the rhyolites of groups IB$_1$, IB$_2$, II and IIIA assigned to different structural units plot on the same fields, their strong resemblance is indicated and similar ages and petrogenesis are suggested (the same source and the same processes of magma generation and evolution).

The rhyolite dykes and veins of the Măcin Unit intruding only Paleozoic formations have been long time considered Paleozoic in age (Mirăuţă, Mirăuţă, 1962; Mureşan, 1975). These rocks show similar chemical features as the rhyolites of the Consul Unit and their association with basic – usually contemporaneously – rocks suggest the same Triassic age for both.

As can be seen on the presented diagrams, most of the intracarapelitic rhyolites show similar trace element distribution to the Triassic rhyolites suggesting a commun crustal source. The divergent trends of the plotting fields on some diagrams (Figs. 2, 3, 4) as well as some differences in the Sr, Zr, Th and Eu contents account however for different evolution trends of these common source magmas. Some of the samples of these groups show strong differences of the trace element distribution with respect to all the acid rocks of Northern Dobrogea, as observed on several diagrams (Figs. 2, 4, 5, 8). This can be explained either as the result of some processes subsequent to rhyolite emplacement or of special processes connected with the magma evolution.

The rhyolites of the 'southern alignment' (IC$_1$, IC$_2$, IC$_3$) show a trace element distribution slightly different from the Triassic rhyolites (the highest Zr and Y, the lowest Sr contents). On most of the diagrams these rhyolites plot on a distinct field. This could be a proof for their different age and/or different source and genetic processes.

In the 'southern alignment' the rhyolites from Turcoaia and Cîrjelari have similar trace element distributions while the Camena rhyolites are significantly different by their high Rb contents, the maximum values reported for the rhyolites of Northern Dobrogea.

The Başpunar rhyolite differs from both the 'southern alignment' rocks and the other rhyolites of Northern Dobrogea through its alkaline character, REE distribution, the lowest Rb, high Nb, the highest Sr values. These features point to slightly different petrogenetic processes than for the other rhyolites and possibly a different source. The Başpunar rhyolite-associated with basalts – is the only occurrence with a demonstrated Jurassic age (Kimmeridgian-Oxfordian). The generalisation of this age for the other rhyolites with clearly different chemical features is highly doubtful.

Compared to the Paleozoic and Mesozoic rhyolites of Northern Dobrogea, the Paleozoic trachytes of the Tulcea Unit show a clearly different trace element distribution (see all the diagrams) suggesting a different origin.

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Table

Synthesis of petrogenetic data on rhyolites

Occurrences(groups)				facies	Acid magmatites		Associated magmatites		Age
Măcin Unit	Consul Unit	Niculiţel Unit	Tulcea Unit	Deposit	Type	Series	Type	Series	
			IIIB	S,V	T	A	B	Th	Pz
IA				P,E	R	CA			Pz(C)
IB				S	R,D	CA	B,D	Th	T_{1-2}
	II			E,P	R	CA	B,D	Th	T_{1-2}
		+		E,P,S			B,D	Th	T_{1-3}
			IIIA	E,P,S	R,RD	CA	B,D	Th	T_{1-2}
IC1				H,S	R	CA-A	G	CA-A	T/J(?)
IC2				H,S,E,P	R	CA-A	G	CA-A	T/J(?)
IC3				E,P	R	CA-A		CA-A	T/J(?)
IC3				P,E	Ra	A	B	Th	J_3(Ox-Km)

Deposit: V, volcanic; S, subvolcanic; E, effusive; P, pyroclastic; H, hypabyssal. Rock type: R, rhyolite; D, dacite; RD, rhyodacite; Ra, alkaline rhyolite; B, basalt; D, dolerite; G, granite; T, trachyte.

4. Petrogenesis

Enough geological and geochemical data are now available to allow a petrogenetic approach based on the presented petrochemical information. Particularly the trace element distribution is relevant to interpretations concerning the source of the rhyolite generating magmas and the different processes of magma evolution. Chemical data also enable us to reconstruct the geotectonic setting in which these magmas were generated and evolved.

Both major and mainly trace element geochemistry of the studied rocks suggest the crustal origin of magmas which generated most of the acid igneous rocks of Northern Dobrogea. Arguments for such an origin are the low Sr and high Rb, Y, Yb, Nb, U, Th contents, the high REE sum, the low Fe_2O_3, FeO , MgO, MnO, Ni, Co, Cr, Sc, V contents of rhyolites. Their derivation from basaltic melts through fractional crystallisation and/or crustal contamination seems to contradict most of the analytical and observation data. Such a model could not explain the absence of rocks with intermediate composition. A deeper magma source is easily acceptable only for the Paleozoic rhyolites of the Tulcea Unit and the Jurassic Başpunar rhyolite showing higher Sr contents and a specific REE distribution. For both the Triassic and intracarapelitic rhyolites (with 50-60 ppm Sr, Fig. 4) the correlation between Sr contents and magma generation depth (Hart et al., 1970) points to a shallow source situated in the lower crust. For similar reasons the 'southern alignment' rhyolites may be related to a more shallow source located in the upper crust. Certain high Sr contents in the Triassic rhyolites of the Măcin, Consul and Tulcea units, closely associated to basic dykes could be explained by some magma mixing processes with deep seated basaltic melts. The presence of partly assimilated basic xenoliths in some rhyolites (Seghedi et al., 1990) supports this interpretation. The distinct plotting fields of the analysed rock groups in the Zr-Rb diagram (Fig. 5) can be interpreted as the result of the crustal source heterogeneity or, alternatively, due to different level of the source within the crust.

As known the three main processes which control the trace element distribution within magma derived rocks are partial melting, fractional crystallisation and magma mixing (Allègre, Minster, 1978).

The trace element (Sr, Rb, Zr, Nb, Y, Yb and REE) distribution sugests the importance of both partial melting and fractional crystallisation as petrogenetic processes for the studied rocks. There are significant

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4
and their associated rocks in Northern Dobrogea

Geotectonic environment	Source Acid volcanics	Petrogenesis Associated rocks	Processes
Rift-oceanic crust (?)	mantle	mantle	TP
Intracontinental back-arc	lower crust		TP+CF
Intracontinental incipient rift	lower crust	mantle	TP+CF+(AM)
Intracontinental incipient rift	lower crust	mantle	TP+CF+(AM)
Intracontinental rift		mantle	
Intracontinental incipient rift	lower crust	mantle	TP+CF+(AM)
Continental intraplate	upper crust	upper crust	TP+CF
Continental intraplate	upper crust	upper crust	TP+CF
Continental intraplate	upper crust		TP+CF
Intracontinental incipient rift	lower crust (upper mantle)	mantle	(TP)+CF

Magmatic series: A, alkaline; CA, calc-alkaline; Th, tholeiitic.
Petrogenetic processes: TP, partial melt; CF, fractional crystallisation; AM, magma mixture.

differences with respect to the relative participation of these two processes for different rock groups. For the intracarapelitic rhyolites the fractional crystallisation seems to have a less important contribution to their petrogenesis because these rocks lack significant negative Eu anomalies which obviously result from plagioclase and K-feldspar extraction from the melt (Hanson, 1978). This is valid for the group IIIA rocks, too. Within the other groups (IA, IC, II) there are either cases of strong Eu anomalies or cases of no Eu anomaly. This suggests that fractional crystallisation occurred only within magma volumes with long residence times in intermediate magma chambers.

The trace element distribution is controlled by both partial melting and fractional crystallisation in 'southern alignment' rhyolites, too. A different degree of partial melting for the Turcoaia and Cîrjelari rhyolites and the Camena ones is accounted for by their observed trace element distributions. All these rocks derived from the same Sr depleted and Zr enriched source but through a lower degree of partial melting for the Camena rhyolites. This also accounts for the K_2O/Na_2O ratio: the "alkalisodic" composition of the Turcoaia and Cîrjelari rhyolites is not necessarily explained by their different source, as suggested by Ştefan and Roşu (in Nedelcu et al., 1988, unpubl. report). Some information concerning the occurrence of feldspar and quartz xenocrysts as restites in these rocks (Constantinescu et al., 1981, unpubl. report, Ştefan, Roşu, in Berbeleac et al., 1985, unpubl. report, Nedelcu et al., 1986, 1988, unpubl. reports) are in good agreement with the origin of the generating magmas within a felsic crustal source.

The petrogenetically significant trace element distribution in the Başpunar quarry rhyolite suggests its formation by fractional crystallisation of a magma generated by a low degree of partial melting of an upper mantle or lower crustal source.

In the case of the Paleozoic trachyte from the Tulcea Unit the fractionation of both light and heavy REE can be explained only by the implication of garnet and, possibly, orthopyroxene, in the partial melting process. This involves great depths and high pressure condition for magma genesis (Hanson, 1978).

5. Discussions

Using the diagrams for discriminating between different geotectonic setting of granitic rocks (Pearce et al., 1984; Figs. 6, 7) one confirms the withinplate setting of the investigated rocks which is also suggested by their

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regional geological environment. The petrochemical study of the basaltic rocks of the Niculiţel and Tulcea Units associated with rhyolites in the other units (Savu et al., 1986) too revealed the same geotectonic setting, i. e. withinplate, for a large part of them. The Carboniferous evolution of the Măcin Unit characterized by the association of rapid continental sedimentation with calc-alkaline acid extrusive activity favours the interpretation in terms of an active continental margin with back-arc features as the result of subduction processes related to Hercynian orogeny.

The Triassic magmatic activity of polarised – acid and basic – character ('bimodal volcanic activity' according to Vlad, 1984, Savu et al., 1986) was generated during the Triassic riftogenesis. According to the geodynamic model proposed by Liu (1980) riftogenesis occurs due to the action of a 'thermal dome' situated in the mantle generating magmatic processes and 'horst and graben' structures. A similar model may account for the riftogenesis and related magmatic activity in Northern Dobrogea as it has already been mentioned (Vlad, 1978, 1984, Savu et al., 1985, 1986). The ascent of a 'thermal dome' accounts for the almost simultaneous generation of basaltic magmas in the upper mantle and of acid magmas by crustal anatexis, resulting in spatially and temporally associated basic and acid eruptive areas in agreement with the model proposed by Eichelberger (1978). A direct correlation between the amount of rhyolites generated by 'bimodal volcanism' and the continental character of the crust which supplies the magmas is suggested by this model. This seems to apply to Northern Dobrogea too, when considering the numerous rhyolite occurrences assigned to Măcin and Consul Units by comparison to Tulcea Unit and their absence from Niculiţel Unit.

The geotectonic significance of the Triassic/Jurassic rhyolites occurring in the area of the "southern alignment" results from their calc-alkaline chemistry, the crustal source of magmas (with $^{87}Sr/^{86}Sr=0.707\text{-}0.708$, Pop et al., 1985), their continental (withinplate) location and their association with cogenetic granites. All these features account for crustal anatexis initiated in continental (withinplate) environment. Its causes are difficult to state according to the available data. However, taking into account the small amounts of acid and basic igneous rocks generated in an incipient riftogenesis during the Upper Jurassic, as previously suggested (Grădinaru, 1981), a 'thermal dome' might be supposed in the upper mantle.

The occurrence of abundant intrusions might be due to the action of the 'thermal dome' during a compressional tectonic stage. The variation of the relative amounts of rhyolites and granites along the alignment seems to reflect the variation of the compressional tectonic regime.

Table 4 is a synthetic representation of the sequence, evolution and geotectonic background of the rhyolitic rocks from Northern Dobrogea.

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References

Allègre C.J., Minster J.F. (1978) Quantitative models on the trace element behaviour in magmatic processes. In "Trace element in igneous petrology", ed. C.J. Allègre, S.R. Hart; Developments in petrology, 5, Elsevier Publ. Comp., Amsterdam.

Cantuniari S. N. (1912) Masivul eruptiv Muntele Carol-Piatra Roşie (jud. Tulcea). *An. Inst. Geol. Rom.*, VI, 1, p. 1–157, Bucureşti.

Cădere D. (1925) Rocile eruptive de la Camena (Dobrogea, jud. Tulcea). Studiu geologic, petrografic şi chimic. *An. Inst. Geol. Rom.*, X, Bucureşti.

Cioflica G., Lupu M., Nicolae I., Vlad Ş. (1980) Alpine ophiolites of Romania: tectonic setting, magmatism and metallogenesis. *An. Inst. Geol. Geofiz.*, LVI, Bucureşti.

Dimitrescu R. (1959) Observaţii asupra geologiei regiunii Cîrjelari. *D. S. Com. Geol.*, XLII, Bucureşti.

Eichelberger J. C. (1978) Andesitic volcanism and crustal evolution. *Nature*, 275, p. 21–27.

Grădinaru E. (1981) Rocile sedimentare şi vulcanitele acide şi bazice ale Jurasicului superior (Oxfordian) din zona Camena (Dobrogea de nord). *An. Univ. Buc., XXX, Geol.*, p. 89–110, Bucureşti.

– (1984) Jurasic rocks of North Dobrogea. A depositional tectonic approach. *Rev. Roum. Géol., Géophys. Géogr., Gèol.*, 28, p. 61–72, Bucureşti.

Hanson G. N. (1978) The application of trace elements to the petrogenesis of igneous rocks of granitic composition. *Earth. Planet. Sci. Lett.*, 53, p. 255–266, Amsterdam.

Institutul Geologic al României

Hart S. R., Brooks C., Krogh T. E., Davis G. L., Nava D. (1970) Ancient and modern volcanic rocks: a trace element model. *Earth. Planet. Sci. Lett.*, 10, p. 17–28, Amsterdam.

Ianovici V., Giuşcă D., Mutihac V., Mirăuţă O., Chiriac M. (1961) Privire generală asupra Dobrogei. Asoc. Geol. Carp.-Balc., Congr. V, Ghid Dobrogea. Ed. Inst. Geol., Bucureşti.

– , Ionescu J., Bălan M. (1969) Étude minèralogique des amphiboles alcalines contenues dans les roches du massif éruptif de Iacobdeal-Dobrogea. *Rev. Roum. Géol. Géophys. Géogr., Géol.*, 13/2, p. 123–135, Bucureşti.

Întorsureanu I. (1987) Consideraţii privind geneza granitelor alcaline din masivele Iacobdeal şi Piatra Roşie (Dobrogea de nord). *D. S. Inst. Geol. Geofiz.*, 72-73/1, p. 81–96, Bucureşti.

– , Colios E., Grabari G., Popescu G., Şerbănescu A. (1989) Petrologia asociaţiei granitelor alcaline din masivele Iacobdeal şi Piatra Roşie (Dobrogea de nord). *D. S. Inst. Geol. Geofiz.*, 74/1, p. 67–86, Bucureşti.

Liu Han-Shan (1980) Convection generated stress field and intraplate volcanism. *Tectonophysics*, 65, p. 225–244, Amsterdam.

Mirăuţă O., Mirăuţă E. (1962) Observaţii asupra structurii geologice Başpunar-Camena-Ceamurlia de sus (Dobrogea). *D. S. Com. Geol., Inst. Geol.*, XLIV, Bucureşti.

– , Mirăuţă E. (1962) Paleozoicul din partea de sud a munţilor Măcin (regiunea Cerna-Hamcearca). *D. S. Com. Geol.*, XLVI, Bucureşti.

– (1966) Contribuţii la cunoaşterea formaţiunilor paleozoice din partea nordică a munţilor Măcin. Acad. R.S.R. *St. cerc. geol. geofiz. geogr., Geol.*, 11/2, p. 497–512, Bucureşti.

– (1966) Paleozoicul de la Cataloi şi cuvertura lui triasică. *D. S. Inst. Geol.*, LII/1 (1964 - 1965), p. 275–289, Bucureşti.

Mirăuţă E. (1982) Biostratigraphy of the Triassic deposits in the Somova-Sarica Hill zone (North Dobrogea) with special regard on the eruption age. *D. S. Inst. Geol. Geofiz.*, LXVII/4, p. 63–78, Bucureşti.

Mînzatu S., Lemne M., Vâjdea E., Tănăsescu A., Ionica M., Tiepac I. (1975) Date geocronologice obţinute pentru formaţiunile cristalofiliene şi masive eruptive din România. *D. S. Inst. Geol. Geofiz.*, LXI/5, Bucureşti.

Mrazec L., Pascu R. (1896) Note sur la structure gèologique des environs du village d'Ortachioi D. Tulcea. *Bul. Soc. St. fiz.*, 12 , Bucureşti.

– (1899) Note preliminaire sur un granite à riebeckite et égirine des environs de Turcoaia. *Bul. Soc. Ing. Ind. Mine Rom.*, III, Bucureşti.

– (1912) Discussion sur la géologie de la Dobrogea septentrionale. *C. R. Inst. Gèol. Roum.*, III, Bucureşti.

Mureşan M. (1975) Privire de ansamblu asupra succesiunii de formare a rocilor magmatogene paleozoice sinorogene şi subsecvente din Dobrogea de nord. *D. S. Inst. Geol. Geofiz.*, LXI/5, Bucureşti.

Murgoci M. G. (1914) Cercetări geologice în Dobrogea nordică, cu privire specială la rocile paleozoice şi eruptive. *An. Inst. Geol. Rom.*, VI, Bucureşti.

Mutihac V. (1964) Zona Tulcea şi poziţia acesteia în cadrul structural al Dobrogei. *An. Com. Geol.*, XXXIX/1, Bucureşti.

Pascu R. (1904) Studii geologice şi miniere în judeţul Tulcea (Dobrogea). *Bul. Min. Agr. Ind. Comerţ şi Dom. Serv. Minelor*, Bucureşti.

Pearce A. J., Harris B. W. N., Tindle G. A. (1984) Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *Jour. of Petrol.*, 25, 4, p. 956–983.

Peters K. F. (1867) Grundlinien zur Geographie und Geologie der Dobrodscha. *Denk. der K. Ak. d. wiss. math. nat. wiss. K. l.*, XXVII, Wien.

Pop C., Buzilă A., Cioloboc D., Catilina R., Popescu G. (1985) Isotopic Rb/Sr ages for establishing the emplacement of some granitoids of North Dobrogea. Proceeding reports of the XIII-th Congres of KBGA, Additionaly received report, p. 108–111, Kracow.

Rotman D. (1914) Comunicare preliminară asupra întinderii, clasificării, repartizării şi originii rocilor care alcătuiesc stratele de Carapelit în Dobrogea de nord-vest. *D. S. Inst. Geol.*, V, p. 1–8, Bucureşti.

Savu H., Udrescu C., Neacşu V. (1982) Structure and genesis of the diabase complex from the Luncaviţa-Isaccea-Mănăstirea Cocoş zone (North Dobrogea). *D. S. Inst. Geol. Geofiz.*, LXVII, p. 135–153, Bucureşti.

– , Udrescu C., Neacşu V., Stoian M. (1985) Vulcanismul bimodal de intraplacă continentală triasic de la Somova (Dobrogea de nord). *St. cerc. geol. geofiz. geogr., Geol.*, 30, p. 62–69, Bucureşti.

– (1986) Triassic continental intra-plate volcanism in North Dobrogea. *Rev. Roum. Géol. Géophys. Géogr., Géol.*, 30, p. 21–29, Bucureşti.

Savul M. (1931) Porfirul de la Isaccea. *D. S. Inst. Geol.*, XVII, Bucureşti.

– (1935) Porphyres quartzifères de la région Meidanchioi-Consul (Dobrogea du nord). *C. R. Inst. Geol. Rom.*, XX(1931-1932), Bucureşti.

Săndulescu M. (1984) Geotectonica României. Ed. Tehnică, Bucureşti.

Seghedi A., Ghenea C., Ghenea A., Mirăuţă E. (1980) Harta geologică a RSR, sc. 1:50000, foaia Măcin, Arh. IGG, Bucureşti.

– (1980) Consideraţii privind succesiunea de formare a masivelor granitoide din unitatea de Măcin a Dobrogei de Nord. *D. S. Inst. Geol. Geofiz.*, LXI/1, p. 65–78, Bucureşti.

Institutul Geologic al României

– , Uricariu V. (1985) Metamorphic history of the Uzum Bair Formation. *D. S. Inst. Geol. Geofiz.*, LXIX/1, p. 279–282, Bucureşti.

– , Oaie G. (1986) Formaţiunea de Carapelit (Dobrogea de Nord): faciesuri şi structuri sedimentare. *D. S. Inst. Geol. Geofiz.*, 70-71/4, p. 19–37, Bucureşti.

– , Seghedi I., Szakács A., Oaie G. (1987) Relationschip between sedimentation and volcanism during deposition of the Carapelit formation (North Dobrogea). *D. S. Inst. Geol. Geofiz.*, 72-73/1, p. 191–202, Bucureşti.

Seghedi I., Szakács A., Baltreş A. (1990) Relationships between sedimentary deposits and eruptive rocks in the Consul Unit (North Dobrogea) – implications in tectonic interpretations. *D. S. Inst. Geol. Geofiz.*, 74/5, p. 125–136, Bucureşti.

Stiopol V., Jude L., Drăghici I. (1978) Magmatitele acide din Dobrogea de Nord şi relaţiile acestora cu mineralizaţiile. *Peuce*, V, Muzeul "Deltei Dunării", p. 23–32, Tulcea.

Streckeisen A. (1931) Asupra petrografiei Dobrogei. *D. S. Inst. Geol.*, XVIII, p. 67–86, Bucureşti.

Tischendorf G., Geisler M., Gerstenberger A., Budzinski H., Volger P. (1987) Geochemistry of variscan granites of the Westerzgebirge-Vogtland region – an example of tin deposits-generating granites. *Chem. Erde*, 46, p. 213–235.

Vlad Ş. (1978) Metalogeneza triasică din zona Tulcea (Dobrogea de Nord). *St. cerc. geol. geofiz. geogr., Geol.*, 23/2, Bucureşti.

– (1984) Triassic mineralization in North Dobrogea (Romania). *Radovi Geoinstituta*, 17, p. 137–143, Beograd.

Winchester J. A., Floyd D. A. (1977) Geochemical discrimination of different magma series and their differentiation products using immobile elements. *Chem. Geol.*, 20, p. 325–343.

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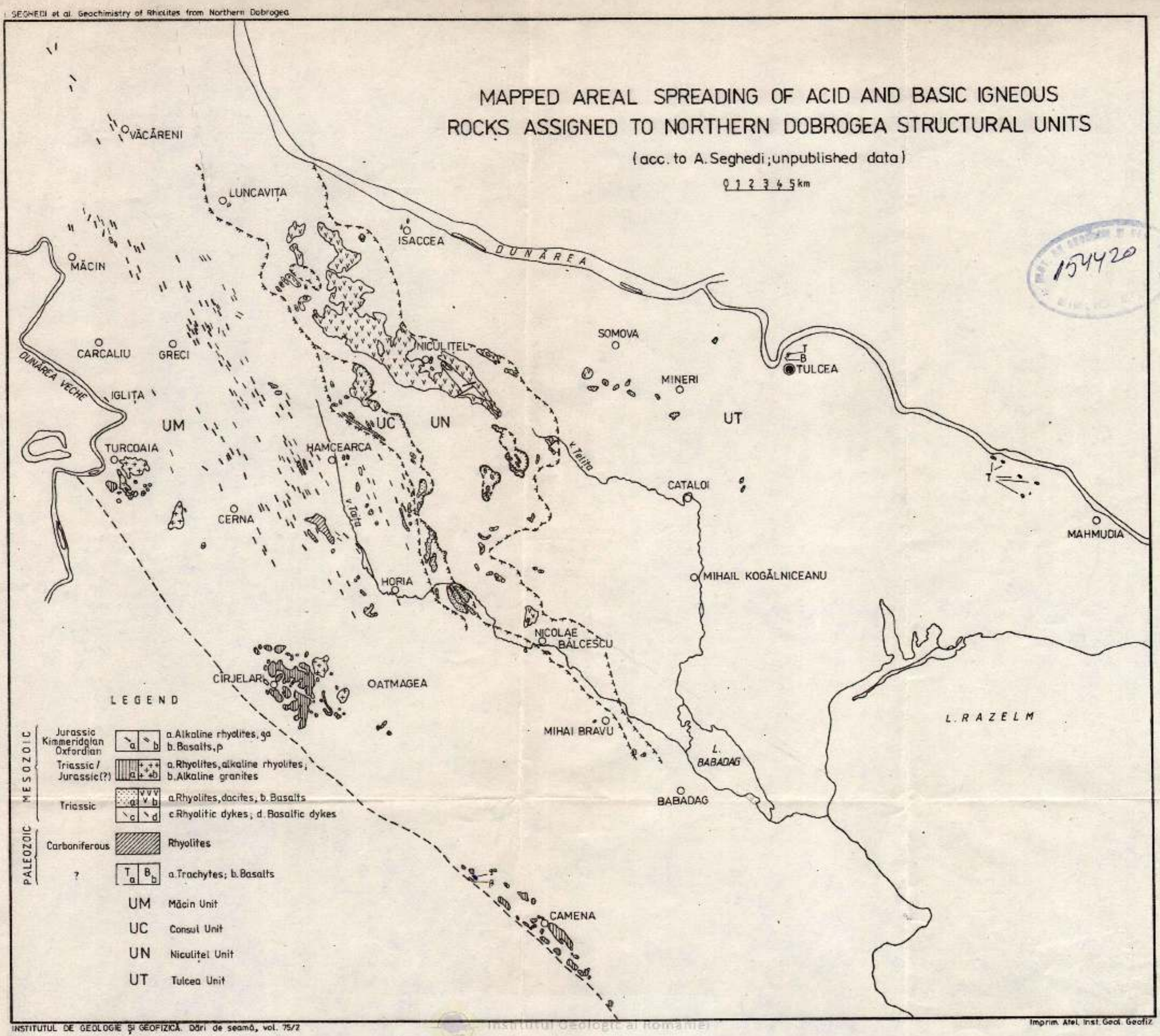

MAPPED AREAL SPREADING OF ACID AND BASIC IGNEOUS ROCKS ASSIGNED TO NORTHERN DOBROGEA STRUCTURAL UNITS

(acc. to A. Seghedi; unpublished data)

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CAMENA RHYOLITES (NORTH DOBROGEA)

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Résumé: *Rhyolites de Camena (Dobrogea de Nord).* Les roches rhyolitique de Camena, jalonées en direction NW-SE le long de deux alignements parallèles sont representées par plusieurs variétés: des rhyolites granophyriques, des granophyres, des rhyolites fluidales et/ou eutaxitiques, des rhyolites vitroclastiques (tufs soudés), des rhyolites perlitiques, des rhyolites – rhyolites microgranitiques à biotite. Dans l'alignement oriental prédominent des roches à caractère intrusif (des rhyolites granophyriques, granophyres, etc.) tandis que dans celui occidental se trouvent specialement des volcanites (des rhyolites perlitiques). Les particularités texturales, structurales et minéralogiques, ainsi que la forme de gisement des rhyolites infirment l'existence d'une formation ignimbritique dans l'alignement ouest de Camena; dans celui oriental de telles produits volcaniques ayant seulement des apparitions locales. Les caractères minéralogiques, pétrographiques et géochimiques semblables des rhyolites des duex alignements de Camena, des rapports avec les roches environnantes, ainsi que les données d'âge isotopique K-Ar (qui indique 151,7 Ma) attestent leurs origine commune et le même âge, oxfordien. L'origine crustale du magma par l'anatexie d'un granitoïde calco-alcalin riche en K_2O est soutenue tant par les données minéralogiques (restites de roches ou de cristeaux) aussi bien que par celles géochimiques. L'anomalie négative en Eu ne s'explique pas par le fractionnement des plagioclases pendant le processus de cristallisation, qui n'a pas fonctionné dans ce cas, mais par la conservation des plagioclases dans de résidu granitoïde après l'extorsion du magma rhyolitique très riche en SiO_2 et K_2O et pauvre en FeO, MgO, CaO et Na_2O. Dans les conditions des mêmes valeurs grandes de SiO_2 et respectivement constamment réduites et comparable de CaO, la relation Na_2O/K_2O nettement différente dans les rhyolites de Turcoaia-Cîrjelari, par rapport à celles de Camena, exclut leurs origine commune.

Introduction

Rhyolites in Camena zone appear from under the loess cover as a series of NNW-SSE trending bodies, from north of Slava Rusă Valley towards south-east up to nearby the locality of Baia. They are magmatic rocks which belong to the North Dobrogea Orogen. Along the Peceneaga-Camena Fault these rocks come into tectonic contact with the crystalline formations of Altın-Tepe and Green schists Series of Central Dobrogea; they occur within a narrow strip which only locally exceeds 1 km in width (Pl. I).

There are numerous references on the Camena rhyolites, but detailed information on their mineralogical and/or chemical composition, the structural and textural variations, their body morphology and age is given only in a few papers (Pascu, 1904; Murgoci,1914; Cădere, 1925; Bălan, 1966, 1967, unpubl. reports; Stiopol et al., 1975, unpubl. report; Constantinescu et al., 1978, 1981, 1985, unpubl. reports; Grădinaru, 1981, 1984, 1988; Ştefan, 1982; Ştefan, in Berbeleac et al., 1982, unpubl. report; Ştefan and Roşu, in Nedelcu et al., 1986, 1987, 1988, unpubl. reports; Ştefan, in Mureşan et al., 1987, unpubl. report; Seghedi et al., 1992).

Our new data make possible a unitary and more complete presentation of the acid magmatites in Camena zone; some conclusions on the origin of the magma from which the North Dobrogea rhyolites with a high-K_2O content crystallized are also inferred.

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Rhyolites Relationships with Surrounding Rocks and Age of Magmatites

In the first thorough paper dealing with the Camena rhyolites Cădere (1925) distinguished two lineaments:

– an eastern lineament, in which the porphyritic massif is more deeply exposed by erosion, within which "microgranitic, micropegmatitic and microgranulitic quartz porphyries, granophyres and felsophyre-microgranites" can be distinguished;

– a western lineament, in which the porphyritic massif is less eroded, within which "devitrified felsophyres, vitrophyres, breccias and porphyritic tuffs" are developed.

Cădere (1925) and Bălan (1966, 1967, unpubl. reports) considered that the rhyolites from the two mentioned lineaments originate in the same magma and are essentially of the same age, penetrating the "Carapelit Formation" and being reworked into Cretaceous conglomerates. The same age of the rhyolites from the two lineaments was also asserted by Constantinescu (1978, 1981, 1985, unpubl. reports) and Ştefan (1982), as well as by Ştefan and Roşu (in Berbeleac et al., 1985, unpubl. report and in Nedelcu et al., 1986, 1987, 1988, unpubl. reports).

When dealing with the rhyolites relationships with the surrounding rocks, Cădere (1925) pointed out that rhyolites can be Upper Carboniferous, Permian or Mesozoic in age.

The Paleozoic age of the quartziferous porphyries in Camena zone was asserted by Mureşan (1975) on the basis of the K/Ar isotopic data of 212 Ma for Camena rhyolites and 236 Ma, respectively, for Turcoaia porphyries (acc. to Mînzatu et al., 1967, unpubl. report).

Mînzatu et al. (1975) obtained on a whole sample the K/Ar age of 220 Ma for Camena rhyolites and on orthoclase 190 Ma, therefore Triassic age. Also of Triassic age, as all the rhyolites in North Dobrogea, are considered the acid magmatites occurring in this zone (Constantinescu, 1978, 1981, 1985, unpubl. reports).

Grădinaru (1981, 1984, 1988) is the first and also the only author who considers the rhyolites from the two lineaments of different ages, i.e. Permian age for rhyolites in the eastern lineament, and Oxfordian age for rhyolites in the western one. He also compared incorrectly the rhyolites in the eastern lineament at Camena with the Cîrjelari rhyolites – isotopic Rb/Sr age = 196 Ma (Întorsureanu et al., 1989) – and named them all "Cîrjelari rhyolites", although thirty years ago Dimitrescu (1959) and, more recently, Ştefan and Ştefan and Roşu, in all the above-mentioned papers, pointed out the petrographic and chemical differences between the two rhyolite varieties.

As the rhyolites in the eastern lineament do not come into direct contact with the limestones of the "Uspenia Formation" and the relationships between them are unclear (Cădere, 1925) there are no arguments in favour of an age older than the Ladinian-Carnian, as mentioned by Grădinaru (1981, 1984, 1988).

The rhyolites in the eastern Camena lineament include numerous xenoliths from the sedimentary deposits referred in the past to the Carapelit Formation (Bălan, 1966, 1967, unpubl. reports; Ştefan, in Berbeleac et al., 1982, unpubl. report); at present these deposits are considered of Middle Jurassic (Mureşan et al., 1987, unpubl. report) or Lower Triassic age (Mirăuţă, in Nedelcu et al., 1987, unpubl. report).

The presence of rhyolite fragments in the microconglomerates in Camena Valley, referred either to the Cretaceous (Cădere, 1925) or to the Lower Jurassic (Mirăuţă, Mirăuţă, 1962, 1964 and Mirăuţă, 1967, fide Grădinaru, 1981) or recently to the Oxfordian (Grădinaru, 1981, 1984, 1988) shows that the granophyric rhyolitic magmatites are older than these sedimentary rocks. As the age of these deposits is not well enough specified and taking into account the quasi-identical mineralogical and chemical composition of the rhyolites from the two mentioned lineaments we considered (Ştefan, in Berbeleac et al., 1982, unpubl. report; Ştefan and Roşu, in Nedelcu et al., 1987, 1988, unpubl. reports) that the rhyolites in the eastern lineament are of the same age as those in the western one; this opinion is maintained by other authors, too. The age could be probably Upper Jurassic, as stated by Grădinaru (1981, 1984, 1988) although the relations in the Başpunar quarry, lying in a brittle shear zone, are quite confused. The aspect of "primary" alternations of "tufites" in which rhyolitic fragments surrounded by a carbonatic matrix can be recognized might be false. From these "alternations" with high-calcite contents a rhyolite block with 4–5 cm was collected which, under the microscope, proved to be granophyric, similar to those in the eastern lineament. On the other hand, the "tuffs" from the thicker alternation (ca. 0.80 m) in the northwesternmost part of the Başpunar quarry displays a mixed composition, visible in thin sections and also indicated by the results of the chemical analysis (Grădinaru, 1988).

In order to clear up the problem of the age of the Camena rhyolites, especially those in the eastern lineament, K/Ar radiometric ages were determined by means of the isotopic dilution method (Tab. 1), with ^{38}Ar as

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standard, for three rhyolite samples (the isotopic measurements were performed with an AEI MS-20 mass spectrometer in static regime):

– a sample of grey, massive rhyolite, practically aphyric, vitroclastic with reworked shards, collected from Movila Goală, indicated 153.1±6.2 Ma on a whole sample;

– a sample of rhyolite, mostly porphyritic, fluidal, with highly corroded perthitic potash feldspar and quartz crystals, collected from Văcăria Hill (Holdurmi), pointed to 146.8±6.2 Ma, on a whole sample;

– a sample of porphyritic, granophyric rhyolite, with corroded crystals of alkali feldspar and quartz, displaying undulatory extinctions, collected from the northeasternmost part of Văcăria Hill, when leaving the locality of Camena, indicated 155.7±6.9 Ma, on a whole sample.

TABLE 1

K/Ar apparent radiogene ages (M A)

No.	Sample no.	K%	^{40}Ar rad $x10^{-10}$mol/g	^{40}Ar rad %	Apparent age MA ± 1
1	2434 - B	6.35	17.5972	69.12	153.1 ± 6.2
2	2450	6.46	17.0488	62.38	146.8 ± 6.2
3	2488	6.19	17.4568	57.41	155.7 ± 6.9

The analysed samples represented as a system of coordinates $^{40}Ar/^{36}Ar$ $^{40}K/^{36}Ar$ plot on an isochrone of 151.7 Ma with a ratio of initial argon $^{40}Ar/^{36}Ar=294.4$, close to the ratio of the atmospheric argon (295.5). In these circumstances the isochrone age is valid in geologic respect even without verifying with isochrone ^{40}Ar rad – ^{40}K, which is not suitable due to the very close K contents (6.19; 6.35; 6.46 per cent).

Structural-Textural and Mineralogical Particularities of Rhyolites, Their Body Morphology and Emplacement Way

In his paper on Camena rhyolites, Cădere (1925) pointed out among others: the trending of the "phenocrysts" from microgranites and the presence of granophyric lenticular-elongated and bent schlieren, with the same composition as the host rock, due to the magma movement during the consolidation. The same author also emphasized the presence of perlites and of pisolites as well as the existence of breccias along the fractures of the Peceneaga-Camena Fault System.

More recently, Bălan (1966, 1967, unpubl. reports) underlined the presence of perlites and insisted on the tectonic origin and spreading of the green breccias formed especially at the expense of the rhyolites from the western and, to a less extent, the eastern lineaments. The same author also pointed out that rhyolites form dykes; he rejected the existence of ignimbrites at Camena.

Constantinescu (1978, 1981, unpubl. reports), when discussing the relationships between different types of quartziferous porphyries, referred the rhyolites in the whole North Dobrogea to ignimbrites.

Grădinaru (1981) considered ignimbrites a great part of the "Camena rhyolites" from the western lineament, the other rocks from the same lineament being described as rhyolitic hyaloclastites and rhyolites.

As already mentioned, detailed information on the mineralogical composition of the Camena rhyolite was presented by Cădere (1925) who pointed out the presence as "phenocrysts" of quartz, orthoclase, Na-K feldspar and anorthoclase, more rarely of oligoclase and albite.

Bălan (1966, 1967, unpubl. reports) described the biotite-bearing rhyolites as grey-yellowish porphyries; he assigned the alkali feldspar to the sanidine-anorthoclase-high-albite series.

Stiopol et al. (1975, unpubl. report) specified that in the Camena zone, especially in eastern rhyolite lineament, potash feldspar is represented by orthoclase ($2V_{Np}=60\text{-}78^0$ and reduced optic triclinicity $\Delta_c=0.38$) as well as by sanidine ($-2V=20^0$) in the Movila Goală rhyolites with a vitroclastic texture. Sanidine is mentioned especially in the vitrophyric rhyolites from the western lineament, where the mentioned authors also found orthoclase (Cuorlic).

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Considering the presence of granitoid xenoliths in rhyolites, the anatectic character of the magma which yielded rhyolites was pointed out (Constantinescu, 1978, 1981, unpubl. reports; Grădinaru, 1981; Ştefan, 1982), stress being laid on the restitic nature of the crystals of quartz, orthoclase and biotite which give the porphyritic character of the rock (Ştefan, 1982; Ştefan in Berbeleac et al., 1982, unpubl. report).

Our field and laboratory observations emphasized varied structural and textural aspects of the rhyolitic rocks from the two lineaments at Camena; sometimes differences of mineralogical composition are also noticed but they are not contradictory to the consanguinity and the contemporaneous age of these rocks. Considering all this, we ploted on the geological map (sc. 1:25 000): granophyric rhyolites and granophyres, fluidal and/or eutaxitic rhyolites, vitroclastic rhyolites, perlitic rhyolites and biotite-bearing rhyolites-microgranitic rhyolites.

Granophyric rhyolites and granophyres clearly predominate in the eastern lineament and are also found in the western one (Cuorlîc). Generally, the rocks are pink-coloured, massive or banded; locally, especially in case of the latter, they are aphyric. They consist of a granophyric, spherulitic or microgranitic quartz-feldspar intergrowth. Both quartz and K-feldspar crystals, when present, and granophyric mass do not present any flow trending. Generally, large restitic quartz and feldspar crystals are intensely corroded magmatically; the rock groundmass and the restites of turbid orthoclase, more or less perthitic, are finely pigmented with hematite, as already mentioned by Cădere (1925) and Bălan (1966, 1967, unpubl. reports). Locally, in the rocks where the groundmass is more largely crystallized, K-feldspar in hipidiomorphic narrow prismatic crystals becomes more limpid, iron oxides being expelled outside the aggregate. The abundance of iron oxides in gritty xenoliths included into granophyric rhyolites points out that oxides are taken over by the rhyolite melt mainly from the rocks they penetrated.

Fluidal rhyolites are not always eutaxitic, often the vitreous or poorly devitrified groundmass being homogeneous. Sometimes the groundmass includes fragments of fluidal rhyolitic matrix with different degrees of crystallization, or "shards" (convex-concave glass fragments) resulting from the films of volatile bubbles from the melting which locally had vesiculation conditions.

The crystals, this time much more frequent, which give the porphyritic aspect to the rock, are also restitic, being represented by quartz and K-feldspar ($-2V=52-79^0$); alkali feldspar can include sericitized plagioclase. Contours of opacitized amphiboles were rarely observed in these rocks. Fluidal and/or eutaxitic rhyolites are developed both in the eastern and the western lineament, in the latter case the mentioned aspects can be more or less obvious due to the influence of subsequent tectonic movements.

Vitroclastic rhyolites (welded tuffs) are found only in the eastern lineament, in Movila Goală (as first mentioned by Mureşan, 1971) and especially in Holdurmi Summit (Văcăria Hill). They include xenoliths of sandstones and clays; such rocks were also found northwest of Camena. In such places glass fragments with concave-convex contours (shards) show no trending (north of Văcăria), in others their reworking took place which, beside the devitrification phenomena, made the vitroclastic character less obvious (Movila Goală). The quartz and potash feldspar crystals, represented by sanidine ($-2V=20^0$) and by anorthoclase and orthoclase ($-2V=44-78^0$) are corroded by the groundmass, which is generally quite intensely devitrified.

Perlitic rhyolites, of white-yellowish colour, with lower contents of iron oxides, often devitrified, occur only in the western lineaments. Generally, they are poorer in crystals than the rhyolites from the eastern lineament and although sanidine ($-2V=28-30^0$) is present, orthoclasic potash feldspar ($2V_{Np}=50-78^0$) occurs too; quartz predominates. Oligoclasic plagioclase is more rarely found. Crystals are intensely fissured, often crushed, like the groundmass which, in places, by recrystallization displays a "mortar" aspect; however, it preserved relict islands or perlitic or granophyric rhyolites. Subsequent breccifiation phenomena make difficult the recognition of primary hyaloclastic characters of the rhyolitic rock. Usually perlitic rhyolites seem to include fragments of granophyric rhyolites; but some granophyric intergrowths seem to succeed to the perlitic separations.

Rhyolites-microgranitic rhyolites are massive rocks occurring in the western lineament as slightly curved dykes. These rocks display a massive, microcrystalline up to microgranitic groundmass, formed of quartz, feldspar and biotite, within which larger crystals of the same type are included. Three groups of values of the -2V angle – $22-36^0$, $41-52^0$, $56-78^0$ – were determined with U-stage for the potash feldspar crystals. Although the restitic character of some crystals is presumed on the basis of fissures which do not exceed their limit, a subsequent cicatrization phenomenon more likely accompanied by a change of the optical symmetry is also probable, as suggested by the wide range of values of the -2V angle too.

Biotite rhyolites which penetrate the perlitic and fluidal varieties are included into the tectonic breccia (more or less green coloured): their apparent isotopic age 63.2 Ma proves that the fractures from the Peceneaga-Camena Fault System had been active during a very long time-span.

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The above-mentioned rhyolite varieties, with accessory minerals represented by iron oxides and zircon, are generally fresh; locally silicifiations are, however, quite intense. Chloritizations occur especially at the expense of brecciated perlitic rhyolites from the western lineament, whereas zeolites were found only in two cases and fluorine only at Sacar Bair.

The relationships between rhyolites in different facies are not always very clear: sometimes vitroclastic or perlitic rhyolites include angular or lenticular fragments of granophyric rhyolites, other times granophyric intergrowths seem to have crystallized concomitantly with the emplacement of fluidal rhyolites, the latter usually including vitrophyric and, more rarely, perlitic varieties. Mesoscopically and microscopically, the relationships between the rhyolite sequences in different facies frequently indicate a laminary flow, in places, turbulent, of a quite viscous rhyolitic magma.

The spatial distribution and the structural-textural and mineralogical particularities of the rhyolitic rocks varieties presented on the map made possible the estimation of the emplacement way and of their body morphology. Thus, one inferred that the rhyolitic rocks, especially those from the eastern lineament consolidated under subvolcanic conditions – either subvolcanoes proper or enrootment zones of old linear volcanic edifices. In this respect, the vitroclastic rhyolites separated at Movila Goală and Văcăria Hill represent either local peripheral facies, especially apical, of this intrusive body or remains of the volcanic suprastructure.

On the other hand mostly perlitic rhyolites from the western alignment represent volcanic products; at present, it is difficult to specify whether their emplacement was subaerial or subaquatic. Better crystallized rhyolite facies, comparable with those from the eastern lineament from the textural and mineralogical point of view, are found at Cuorlic.

As regards the ignimbritic character of the Camena rhyolites, it is worth mentioning that, generally, the melt which yielded the rhyolites was only locally vesiculated and consequently such pyroclastic products have a limited and unsignificant spreading.

In our opinion there are no ignimbrites on the western lineament because most of the rhyolites from this lineament are perlitic more or less intensely brecciated tectonically, as well as because on the one hand such products are not typical of ignimbrites and on the other hand the structural and textural particularities characteristic of such pyroclastic flows are also absent.

Geochemical Characteristics of Rhyolites

The geochemical characterization of the rhyolites from the two Camena lineaments was based on 16 chemical analyses for major elements (7 from the western lineament and 9 from the eastern one) effectuated in the laboratories of I.P.G.G. Bucureşti (Tab. 2) and on 16 spectral analyses effectuated in the laboratories of I.G.G. Bucureşti (Tab. 3), using the emission spectrography method. A smaller number of samples were analysed by nondispersive XRF spectrometry method (Tab. 4) as well as by thermic neutron activation, the nondestructive variant (Tab. 5); these two types of analyses were also performed in the laboratories of I.G.G.

Table 2 showing the major elements contents of the rhyolites points out: high and very high values of SiO_2 and K_2O, as well as of H_2O for perlitic rhyolites; low or very low contents of FeO, MgO, CaO, Na_2O; very high values of the K_2O/Na_2O ratio and, consequently the high potash character of the Camena rhyolites from either lineament. The higher contents of Fe_2O_3 (1.19-2.56 per cent) are specific to the rhyolites from the eastern lineament.

As the contents of CO_2 and H_2O, except for perlitic rhyolites, were low when plotting the rocks on diagrams the oxides values were not recalculated in percentages.

The analysed magmatites plot on the TAS diagram (Fig. 1) mostly in the rhyolite field; the perlitic rhyolites samples plot in the dacite field, nearby the rhyolite field. Although rhyolites display a high K_2O content only two samples have the peralkalinity index > 1; most of the samples have the parameter close to the unit.

On the QAP diagram (Fig. 2) the analysed rocks, except for one sample (1 152 A, Tab. 2), plot in a relatively unitary field; the samples from the western lineament fall in the rhyolite field and those from the eastern lineament in the alkaline rhyolite field.

The alkaline and peralkaline character of the analysed rocks is also suggested by the al/alk ratio (Fig. 3). On the Na_2O+K_2O/SiO_2 diagram (Fig. 4), Camena rhyolites mostly plot in the alkaline rocks field or nearby it, in the calc-alkaline field.

Due to the higher contents of K_2O and SiO_2, Camena rhyolites fall in the extension of the shoshonitic series, outside the diagram proposed by Peccerillo and Taylor (1976).

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TABLE 2

Chemical analyses of the Camena rhyolites

No.	Sample no	OXIDES %														Total
		SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	CO_2	S	H_2O^+	
1.	152A	84,88	0,13	6,50	1,21	0,63	0,01	0,04	0,24	0,04	5,56	0,06	0,16	0,05	0,18	99,68
2.	114	80,62	0,20	9,50	0,16	0,30	0,01	0,07	0,15	0,03	7,94	0,05	0,11	0,06	0,26	99,64
3.	2430	79,90	0,00	9,54	0,16	0,44	0,00	0,12	0,12	0,92	7,71	0,05	0,00	0,14	0,24	99,52
4.	627A	79,11	0,07	11,00	0,07	0,28	0,01	0,01	0,13	1,74	6,86	0,05	0,00	0,05	0,31	99,68
5.	147D	78,69	0,20	9,69	2,56	0,40	0,00	0,24	0,22	0,13	8,09	0,05	0,00	0,20	0,35	100,82
6.	2434A	77,80	0,18	10,19	1,63	0,34	0,03	0,15	0,15	0,18	8,18	0,09	0,00	0,17	0,34	99,54
7.	2450	77,66	0,16	8,73	2,89	0,14	0,00	1,45	0,63	0,22	7,37	0,03	0,00	0,02	0,51	99,84
8.	2434B	77,51	0,16	8,80	1,88	0,14	0,03	1,60	0,84	0,20	7,32	0,02	0,06	0,00	0,72	99,87
9.	557	76,58	0,19	10,20	2,03	0,86	0,01	0,06	0,21	0,03	8,86	0,03	0,14	0,05	0,28	99,53
10.	157C	76,40	0,20	11,66	0,84	0,39	0,02	0,13	0,25	1,25	8,82	0,00	0,00	0,09	0,31	100,38
11.	2454	76,00	0,13	11,06	1,00	0,28	0,02	1,25	0,77	1,80	6,81	0,06	0,00	0,02	0,57	99,91
12.	2449C	75,80	0,15	7,65	4,34	0,18	0,01	1,65	0,70	0,27	8,19	0,02	0,00	0,00	0,85	99,89
13.	2457	75,48	0,16	11,14	0,95	0,28	0,02	1,10	0,84	1,47	7,84	0,05	0,04	0,01	0,34	99,84
14.	2402B	74,95	0,22	11,54	2,66	0,73	0,01	0,58	0,30	0,24	8,03	0,05	0,00	0,15	0,40	99,86
15.	2417	71,98	0,00	11,98	0,52	0,59	0,01	0,18	0,18	0,96	7,87	0,03	0,00	0,19	3,98	99,48
16.	2458	71,65	0,11	11,42	1,21	0,28	0,01	0,90	1,12	1,27	5,80	0,07	0,24	0,00	4,64	99,93

Analysts: M.David: samples 1,2,4,9,10; V.Neacşu: samples 3,5,6,14,15; A.Dănescu: samples 7,8,11,12,13,16.
1, rhyolite, Movila Goală; 2,granophyric rhyolite, South Camena; 3, rhyolite, NW Movila Goală; 4, rhyolite, Sacar Bair I; 5,rhyolite, Camena; 6, vitroclastic rhyolite,Movila Goală; 7,fluidal rhyolite, NW Văcăria Hill; 8,vitroclastic rhyolite, Movila Goală; 9, rhyolite, height 213.3 NW Camena; 10, biotite rhyolite, W Movila Goală; 11, biotite rhyolite NW height 152.3; 12,rhyolite, NW Camena; 13, rhyolite, W Movila Goală; 14, vitroclastic rhyolite, Văcăria Hill; 15, perlitic rhyolite Sacar Bair I; 16, brecciated rhyolite, Sacar Bair I.

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TABLE 3

Trace elements(ppm)

No	Sample no	Pb	Cu	Zn	Sn	Ga	Ni	Co	Cr	V	Sc	Nb	Zr	Be	B	Yb	Y	La	Sr	Ba
1.	152A	16,5	21	9o	3,5	9	3	<2	7	3,5	3	11	28o	<1	<3o	3,2	17,5	<3o	15	135
2.	114	5	34,5	57	3,5	17	6	3	9	3	4,5	2o	6oo	2,2	<3o	1o	65	46	14	13o
3.	243o	26	24	3o	2	13	3	<2	3	7,5	<2	48	13o	3,2	<3o	3,7	4o	<3o	<1o	65
4.	627A	27	9	55	4	15	4	<2	5	3	<3	65	14o	3,3	<3o	3,8	17,5	<3o	12	4o
5.	147D	16	15	4o	3,5	17	3,5	<2	2,5	4,5	2,5	32	8oo	2	<3o	8,8	85	<3o	<1o	16o
6.	2434A	13	15	4o	3	14	2	<2	3	4	3	3o	52o	3	55	8	8o	58	<1o	2oo
7.	245o	19	4	<3o	5	16	2,5	<2	<2	<2	3	35	1ooo	1,8	<3o	9	1o5	<3o	<1o	15o
8.	2434B	28	4	32	3,5	16	2	<2	<2	2	4	19	33o	2,1	<3o	6,2	65	35	<1o	22o
9.	557	26	13	44	6	17	4	<2	7	<3	5,5	28	95o	2,2	<3o	14	12o	<3o	2o	16o
1o.	157C	18	8	54	3,5	14	3	<2	7	<3	4	28	17o	1,9	<3o	5,4	33	39	17	23o
11.	2454	18	<2	<3o	6	14	2	<2	<2	<2	4	5o	15o	2,4	<3o	5	34	<3o	<1o	12o
12.	2449C	7	<2	<3o	6	2o	2	<2	<2	<2	<2	27	72o	2	<3o	1o	95	<3o	<1o	42
13.	2457	16	<2	38	2,5	14	2,5	<2	<2	3,5	5	4o	16o	1,9	<3o	4,8	52	6o	<1o	38o
14.	24o2B	84	22	34	5	19	5	<2	4	5	3	4o	93o	3,3	<3o	12	12o	<3o	2o	15o
15.	2417	22	8	45	4	21	2	<2	<2	2	<2	57	155	7,5	48	5,5	4o	38	7o	<1o
16.	2458	11	<2	44	2,5	13	<2	<2	<2	<2	2	37	12o	3,1	<3o	3,8	35	4o	24o	16

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TABLE 4

X-ray fluorescence analyses

No	Sample no	Rb ppm	Sr ppm	Y ppm	Zr ppm	Nb ppm	Rb/Sr
1	2430	210	24	58	160	58	8.58
2	147 D	310	14	92	735	30	21.80
3	2434 A	366	17	82	415	30	21.00
4	2450	304	12.80	80.9	646	28.9	23.60
5	2434 B	295	12.90	77.1	376	26.1	22.80
6	2454	213	11.10	42.8	154	56.6	19.10
7	2449 C	298	7.39	89.2	673	33.2	40.30
8	2457	344	17.10	59.1	179	45.6	14.20
9	2402 B	345	31	120	900	35	11.00
10	2458	213	175	61.6	206	77.2	1.21

TABLE 5

REE and Hf distribution (ppm)

No	Sample no	La	Ce	Sm	Eu	Tb	Hf
1	2430	23	62	6.1	0.70	0.82	4.00
2	147 D	6	8	4.2	0.50	0.87	10.00
3	2434 A	49	53	9.0	0.99	0.85	8.40
4	2402 B	9	15	6.2	0.48	0.90	11.00

The table with the trace elements distribution in the studied rhyolites points out small contents of Sr and siderophile elements, on the one hand, and generally high values of Zr, particularly in samples coming from the eastern lineament, on the other hand. However, quite high values of Zr (200–900 ppm) were recorded for Sacar Bair rhyolites from the western lineament (Beşuţiu et al., 1983, unpubl. report). Higher values of Sr for the brecciated perlitic rhyolites from the western lineament are due to the plagioclase from the rock composition and the contamination with carbonatic material, not to the origin in a more deep-seated magma (Seghedi et al., 1992). High contents of Y, Yb (Tab. 3) and of Rb (Tab. 4), whose maximum quantities were recorded in the same high-K_2O rhyolites from the eastern lineament, were determined in correlation with high-Zr values.

Sr and Rb distribution versus SiO_2, although not plotted on diagrams, shows a wide dispersion, whereas Rb variation versus K_2O displays a positive correlation; a tendency of the rhyolite grouping in each lineament was also observed.

The plottings of the contents and/or ratio of trace elements or oxides (for samples of rhyolites, generally fresh) on the diagrams proposed by Winchester and Floyd (1977) point out disagreements from one type of diagram to another as regards the petrotypes discrimination. Thus, on the Zr/TiO_2–Ga diagram (Fig. 5), except for one sample, which plots in the dacite field, all the other samples fall in the rhyolite field; on the

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SiO_2–Nb/Y diagram (Fig. 6) although most of the rhyolites fall in their characteristic field, four samples plot in the comendite-pantellerite field. On the SiO_2–Zr/TiO_2 diagram (Fig. 7) nine samples fall in the rhyolite field, four in the pantellerite field and one sample in the dacite-rhyodacite field. On the Zr/TiO_2–Nb/Y diagram (Fig. 8), ten samples plot in the rhyolite field, two samples in the trachyte field and two in the trachyandesite field. Considering that the same samples plot in different fields on the above-mentioned diagrams, their use in petrographic nomenclature of the North Dobrogea rhyolites (as suggested by Seghedi et al., 1992) is not indicated until their improvement.

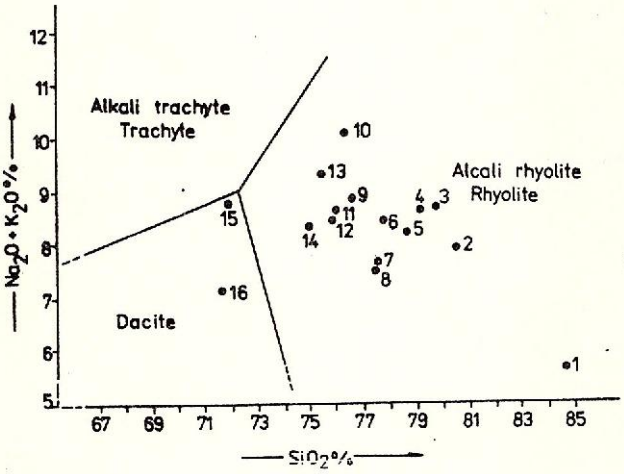

Fig. 1 – Rhyolite plotting on TAS diagram.

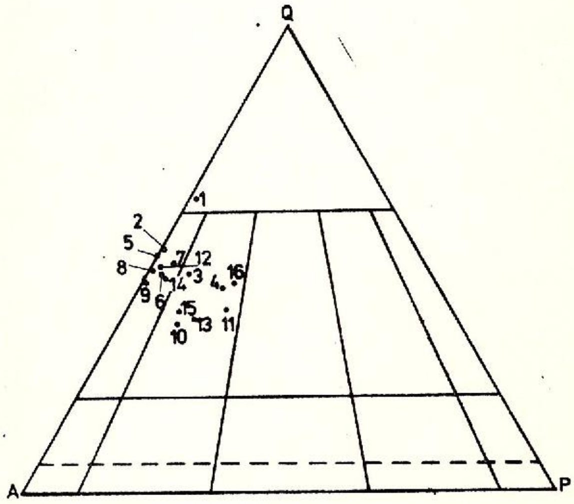

Fig. 2 – QAP diagram.

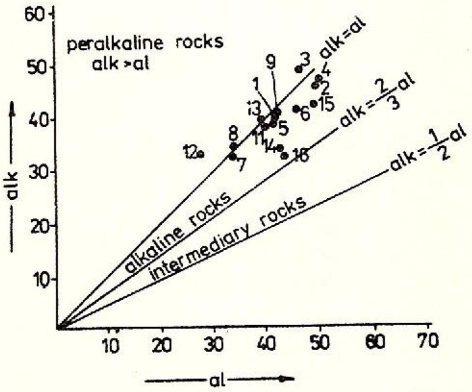

Fig. 3 – alk–al diagram.

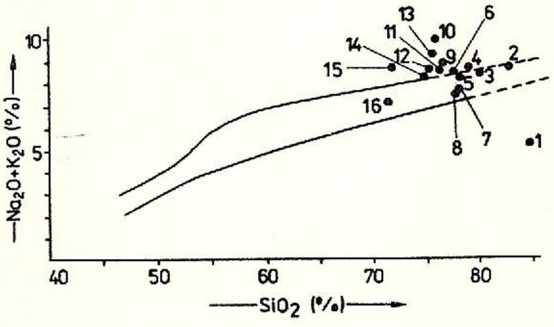

Fig. 4 – Na_2O+K_2O – SiO_2 diagram.

REE and Hf contents were determined for four samples (Tab. 5) out of 16 samples analysed chemically (Tab. 2), three from the eastern lineament and one from the western one.

The distribution of REE (Fig. 9) chondrite-normalized (Henderson, 1984) points out an enrichment in LREE the higher it is the most evolved the melt is (sample 2430) or the higher the participation of the glassy fraction in the rock constitution is (sample 2434 A).

Yb contents (Tab. 3) chondrite-normalized point to higher values (correlatable with the Zr ones) for rocks with a high amount of monocrystalline restites (2402 B) and to lower values in rocks originating in a more evolved melt (2430).

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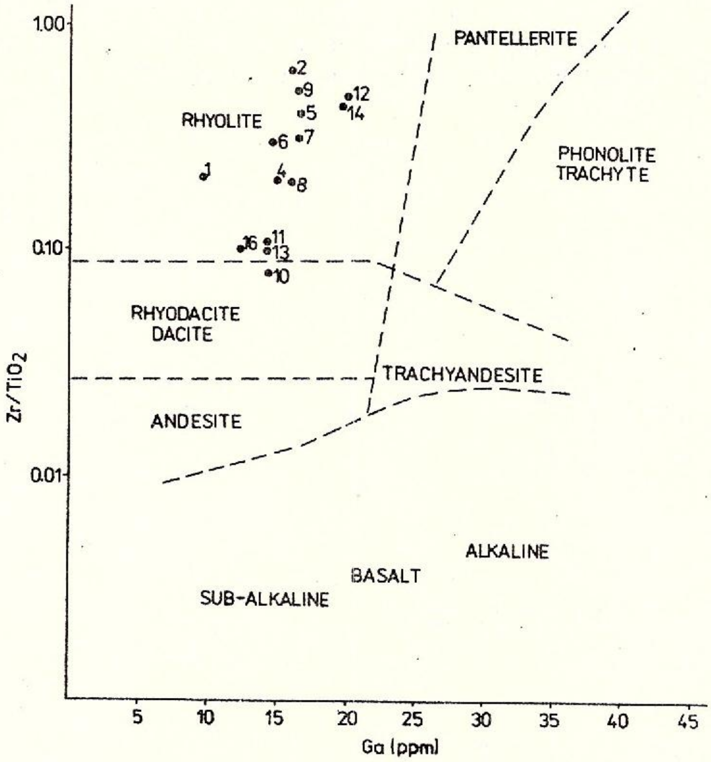

Fig. 5 - Zr/TiO$_2$ - Ga diagram.

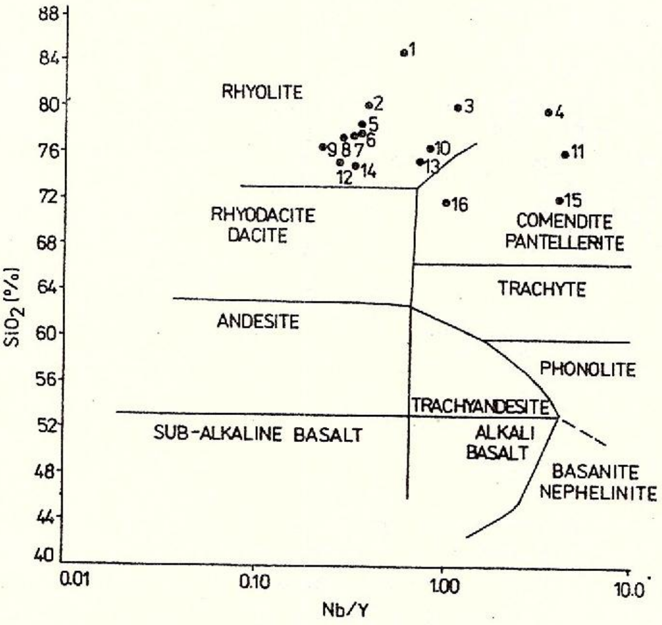

Fig. 6 - SiO$_2$ - Nb/Y diagram.

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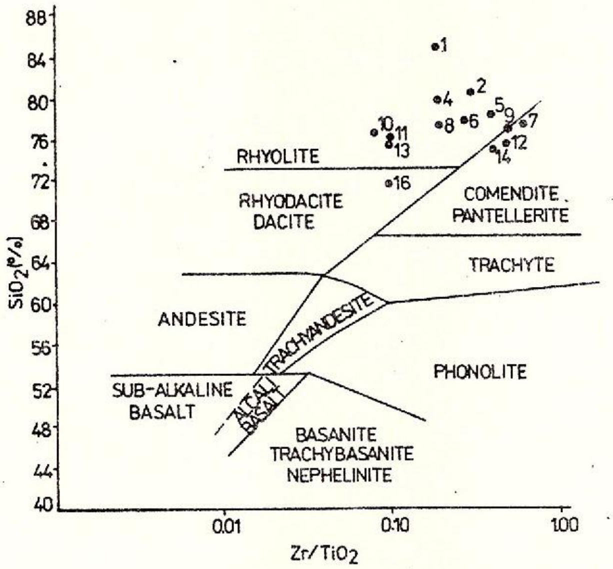

Fig. 7 – SiO_2 - Zr/TiO_2 diagram.

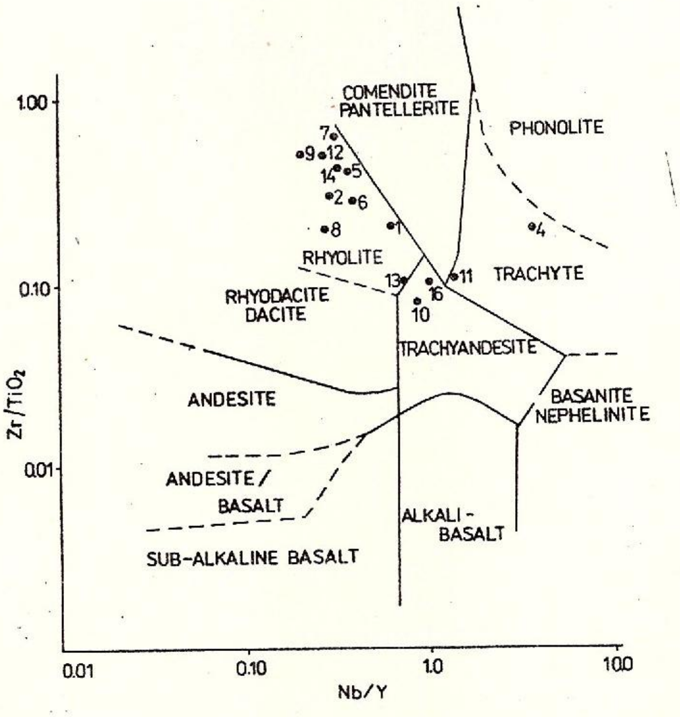

Fig. 8 – Zr/TiO_2 – Nb/Y diagram.

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The diagram with the REE distribution in Camena rhyolites, on which the fields characteristic of Greci (GG) and M. Kogălniceanu (GMK) granitoids are outlined for comparison, emphasizes comparable values, with a slightly marked Eu negative anomaly and higher values and/or tendencies of HREE enrichment in volcanics, as compared with granitoid calc-alkaline rocks.

Petrogenetic Remarks and Conclusions

The crustal origin of the magma from which North Dobrogea rhyolites crystallized is supported by all the workers who carried out minute studies on these rocks; their formation under withinplate conditions, starting from the diagrams presented by Pearce et al. (1984) for fresh and nonporphyritic granitoids, was asserted by Seghedi et al. (1992).

The presence of granitoid fragments, consisting of quartz and K-feldspar± plagioclase±biotite±amphiboles, in Camena rhyolites (Bălan, 1966, 1967, unpubl. reports; Constantinescu et al., 1978, 1981, unpubl. reports; Grădinaru, 1981, 1984, 1988; Ştefan, 1982; Ştefan in Berbeleac et al., 1982, unpubl. report; Ştefan and Roşu in Berbeleac et al., 1985, unpubl. report and in Nedelcu et al., 1986, 1987, 1988, unpubl. reports; Şeclăman, Lupulescu, 1986; Seghedi et al., 1992) and the identical composition of the monocrystals from rhyolites emphasize the anatectic origin of these volcanics originating rather by the anatexis of a calc-alkaline granitoid with a high-K_2O content than by the anatexis of an alkaline granitoid (Şeclăman, Lupulescu, 1986). The yielded magma did not undergo a noticeable differentiation.

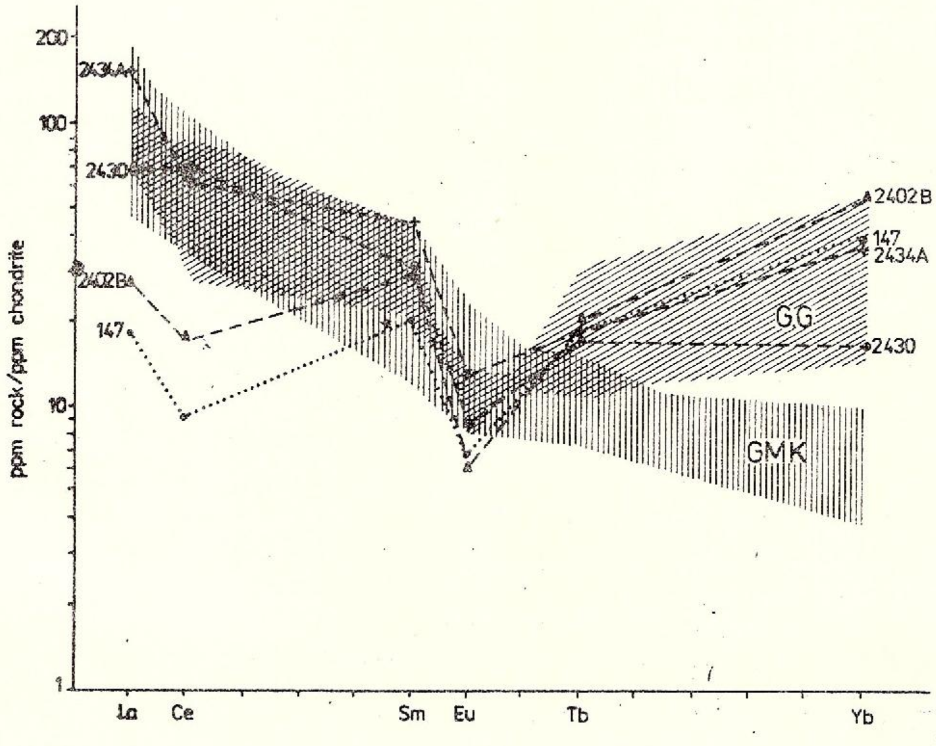

Fig. 9 – REE distribution.

The Eu negative anomaly cannot be explained in this case by the plagioclase removal by fractional crystallization in intermediary magmatic chambers (Seghedi et al., 1992), but by their preservation in the granodioritic residuum after the extraction of the high-K_2O and high-SiO_2 melt, resulting from a fusion at a temperature close to the minimum point of granites.

The great amount of Zr, Y, Yb and Rb in the rhyolites from the eastern lineament is explained by the melting of the granites at the lowest temperature by which, to a reverse order of the crystallization succession, the components crystallized at the end of the granitoids consolidation process were extracted initially. In this respect the high-Zr contents in the Turcoaia–Cîrjelari rhyolites and granitoids, connected with minerals

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crystallized in the pegmatito-pneumatolitic phase cannot be compared with those from the Camena rhyolites, where zirconium is found as rare restites or where zirconium forms probably complex compounds, hard to be determined.

The fact that the rhyolites from the western lineament also contain rare crystals of oligoclase and biotite, but especially the higher Na_2O and CaO contents suggest a more advanced melting process of the same calc-alkaline granitoids.

The high K_2O contents and the K_2O/Na_2O ratios, obviously different in Camena rhyolites versus those in Turcoaia–Cîrjelari rhyolites, correlated with comparable CaO contents, equally low for the rhyolites in the two zones, are contradictory to the origin of these rocks in a common source, as a result of different fusion degrees (Seghedi et al., 1992). However, the quite low values (0.707–0.708) of the $^{87}Sr/^{86}Sr$ isotopic ratios (Pop et al., 1985; Întorsureanu et al., 1989) as well as the mesoperthitic character of the alkaline feldspar in the Turcoaia–Cîrjelari rocks seem to suggest only a contamination with sialic material of a melt which is affiliated rather with basic products than with quartz-feldspathic ones in the upper crust.

The identical origin of the granitoid fragments and of the restitic monocrystals and the constantly high K_2O contents for most of the North Dobrogea rhyolites, except for the Turcoaia–Cîrjelari ones, are in favour of at least a common source, if not of the same age.

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References

Cădere D. (1925) Rocele eruptive de la Camena (Dobrogea, jud. Tulcea). Studiul geologic, petrografic şi chimic. *An. Inst. Geol. Rom.*, X, p. 121–299, Bucureşti.

Dimitrescu R. (1959) Observaţii asupra regiunii Cîrjelari (Dobrogea de Nord). *D. S. Com. Geol.*, XLII, p. 115–133, Bucureşti.

Grădinaru E. (1981) Rocile sedimentare şi vulcanitele acide şi bazice ale Jurasicului superior (Oxfordian) din zona Camena (Dobrogea de nord). *An. Univ. Buc.*, XXX, p. 89–110. Bucureşti.

– (1984) Jurassic rocks of the North Dobrogea. A depositional-tectonic approach. *Rev. Roum. Géol., Géophys. Géogr., Géologie*, 28, p. 61–72, Bucureşti.

– (1988) Jurassic sedimentary rocks and bimodal volcanics of the Cîrjelari–Camena outcrop belt: evidence for a transtensile regime of the Peceneaga–Camena Fault. *Stud. cerc. geol. geofiz., geogr., Geologie*, 33, p. 97–121, Bucureşti.

Henderson P. (1984) Rare Earth Element Geochemistry. Ed. P. Henderson, Elsevier, Amsterdam.

Întorsureanu I., Colios E., Grabari G., Popescu Gh., Şerbănescu A. (1989) Petrologia asociaţiei granitelor alcaline din masivele Iacobdeal şi Piatra Roşie (Dobrogea de nord). *D. S. Inst. Geol. Geofiz.*, 74/1, p. 67–86. Bucureşti.

Mînzatu S., Lemne M., Vâjdea E., Tănăsescu A., Ioncică M., Tiepac I. (1975) Date geocronologice obţinute pentru formaţiuni cristalofiliene şi masive eruptive din România. *D. S. Inst. Geol. Geofiz.*, LXI/5 (1973–1974), p. 85–111, Bucureşti.

Mureşan M. (1971) Asupra prezenţei unei ferestre tectonice în zona şisturilor verzi din Dobrogea centrală (regiunea Altîn-Tepe). *D. S. Inst. Geol.*, LVII/5 (1969–1970), Bucureşti.

– (1975) Privire de ansamblu asupra succesiunii de formare a rocilor magmatogene paleozoice sinorogene şi subsecvente din Dobrogea de Nord. *D. S. Inst. Geol. Geofiz.*, LXI/5, Bucureşti.

Murgoci M. G. (1914) Cercetări geologice în Dobrogea nordică, cu privire specială la rocile paleozoice şi eruptive. *An. Inst. Geol. Rom.*, VI, Bucureşti.

Pascu R. (1904) Studii geologice şi miniere în judeţul Tulcea (Dobrogea). *Bul. Min. Agr. Ind. Comerţ. şi Dom.*, Serviciul Minelor, Bucureşti.

Pearce A. J., Harris B. W. K., Tindle G. A. (1984) Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *Jour. of Petrol.*, 25, 4, p. 956–983.

Pecerillo A., Taylor S. R. (1976) Geochemistry of Eocene calc-alkaline volcanic rocks from Kastamonu area, Northern Turkey. *Contrib. Mineral. Petrol.*, 58, 1, p. 63–81.

Pop G., Buzilă A., Cioloboc D., Catilina R., Popescu Gh. (1985) Isotopic Rb/Sr ages for establishing the emplacement of some granitoids of North Dobrogea. Proceeding reports of the 13th Congress of KBGA Additionally received report, p. 108–111, Cracow.

Seghedi I., Szakács A., Vlad C., Udrescu C., Grabari G., Stoian M., Tănăsescu A. (1992) Major and trace element geochemistry of rhyolites from Northern Dobrogea. Petrogenetic implications. *Rom. J. Petrology*, this volume, Bucureşti.

Institutul Geologic al României

Şeclăman M., Lupulescu M. (1986) Anatectic restites in magmatic rocks and their genetic significance (with reference to the Romanian territory). *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 30, p. 31–40.

Ştefan A. (1982) Geneza riolitelor de la Camena (Dobrogea). *Bul. Inf. Şt. Teh. C.I.T.*, II/1, p. 21–22, I.G.G. Bucureşti.

Winchester J. A., Floyd P. A. (1977) Geochemical discrimination of different magma series and their differentiation products using immobiles elements. *Chem. Geol.*, 20, 4, Amsterdam.

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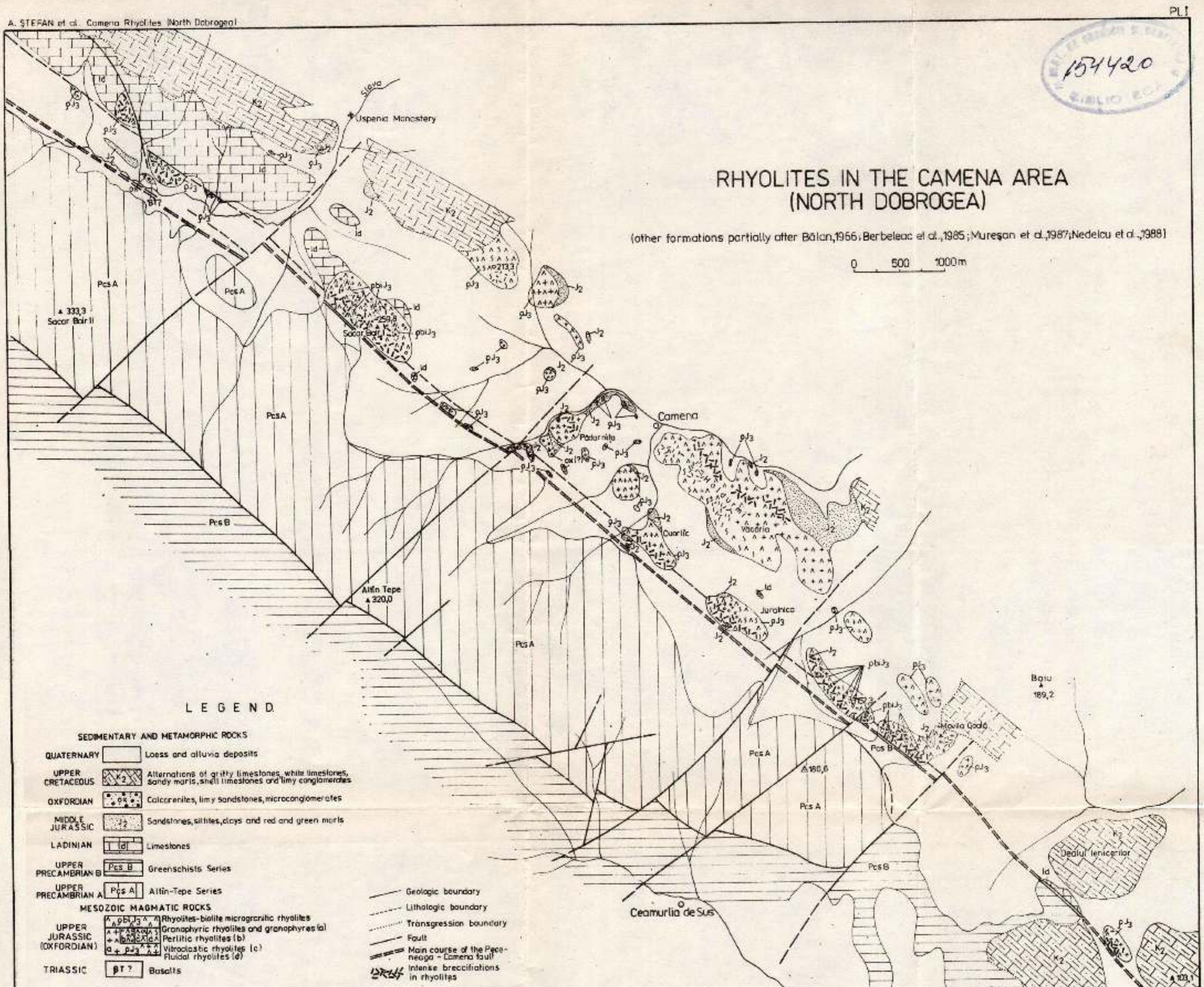

RHYOLITES IN THE CAMENA AREA
(NORTH DOBROGEA)

(other formations partially after Bălan, 1966; Berbeleac et al., 1985; Mureşan et al., 1987; Nedelcu et al., 1988)

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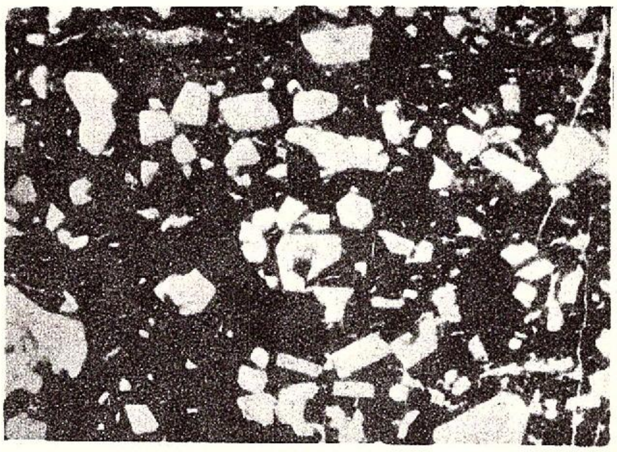

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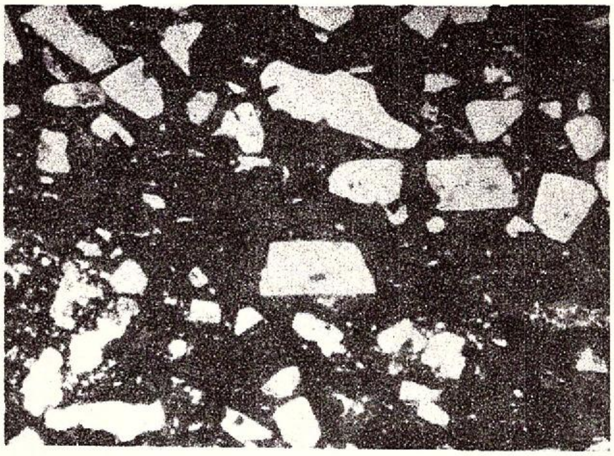

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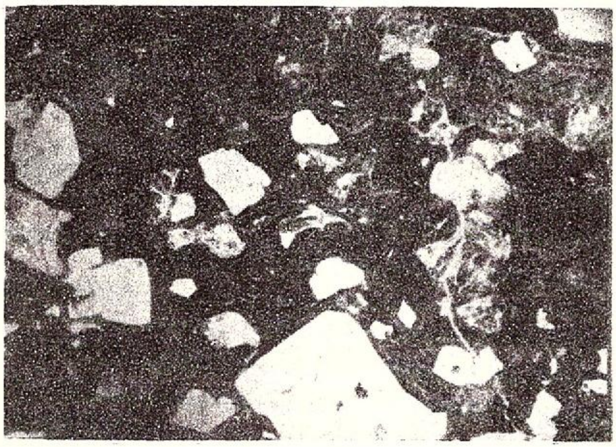

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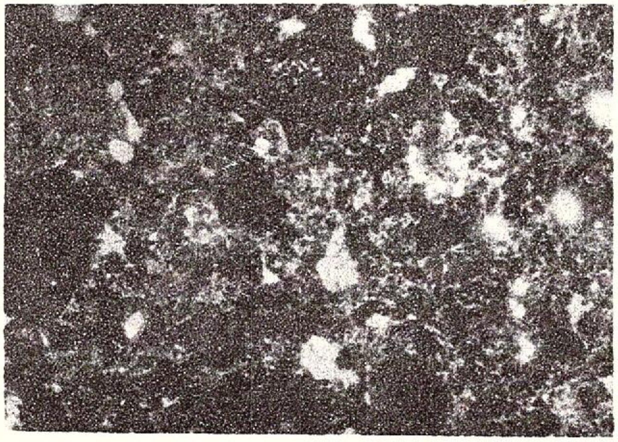

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Plate II

Fig. 1 – Fluidal structure in rhyolite Văcăria Hill, N II, 8 x.
Fig. 2 – Details from Figure 1; N II, 11 x.
Fig. 3 – Eutaxitic structure in rhyolites, Văcăria Hill, N II, 15 x.
Fig. 4 – Perlitic rhyolite in rhyolitic breccia, Sacar Bair I, N II, 20 x.
Fig. 5 – Rhyolitic breccia, Sacar Bair I; N +, 10 x.
Fig. 6 – Xenoliths of opacitized sandstones in vitroclastic rhyolites; North Holdurmi; N +, 9 x.

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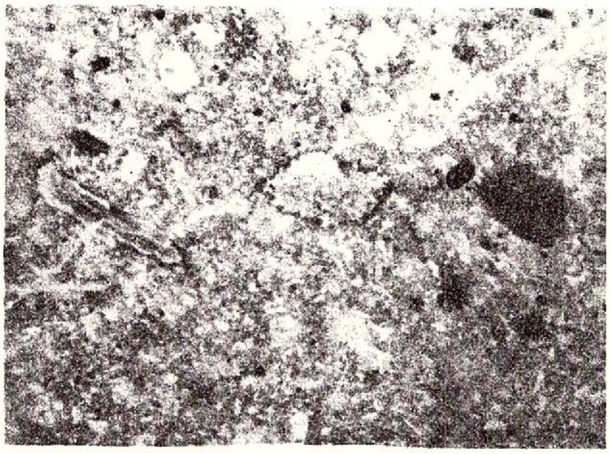

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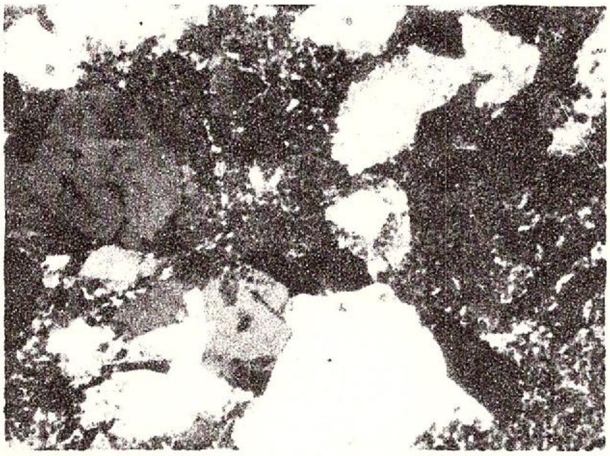

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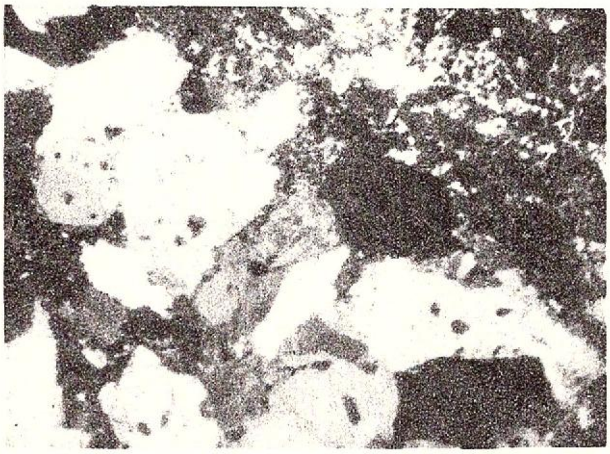

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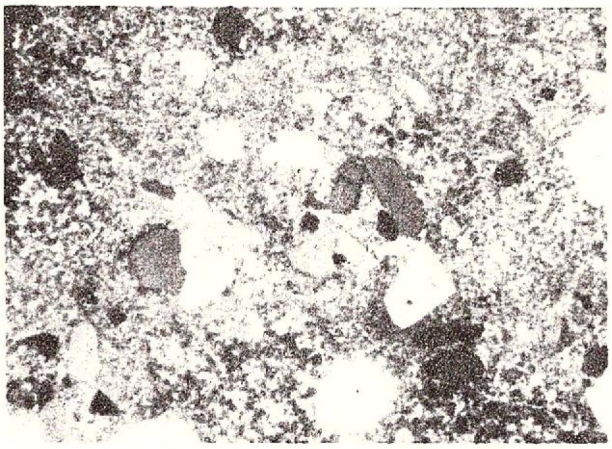

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Plate III

Fig. 1 – Fragments of fine sandstones and fluidal rhyolitic glass in vitroclastic rhyolites, north-east Holdurmi, N II, 12 x.

Fig. 2 – Graphic intergrowths of quartz and K-feldspar, and monocrystalline restites of the same minerals in eutaxitic rhyolites, north height 213; N II, 19 x.

Fig. 3 – Granitoid restites consisting of quartz and orthoclase in rhyolites, south Camena; N +, 14 x.

Fig. 4 – Restitic granitoids in Holdurmi rhyolites; N +, 14 x.

Fig. 5 – Monocrystalline restites with angular contours in fluidal rhyolites, Văcăria Hill; N II, 12 x.

Fig. 6 – Nonhomogeneous structures of the groundmass and granitoid fragments in rhyolites, Văcăria Hill; N +, 12 x.

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Rom. J. Petrology, 1992, 75, p. 53–62

ON THE PRESENCE OF OCEAN FLOOR ROCKS (LIASSIC OPHIOLITES) IN THE TRASCĂU MOUNTAINS (MUREŞ ZONE). THEIR PETROLOGY AND GEOCHEMISTRY

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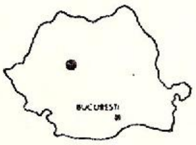

Key words: Ophiolite. Lower Jurassic. Oceanic crust. Obduction. Major elements. Minor elements. Apuseni Mountains – South Apuseni – Trascău Mountains.

Résumé: *La présence des roches de fond océanique (ophiolites liasiques) dans les monts Trascău (zone de Mureş); leur pétrologie et géochimie.* Les roches ophiolitiques de fond océanique liasiques des monts Trascău se présentent en trois situations tectoniques distinctes: (1) dans l'unité de Podeni, (2) dans la cordillière d'Ampoi et (3) comme olistolites dans le mélange à matrice volcanique des unités du système de la nappe de Bedeleu. Elles ont résulté de la désagrégation du mégaslabe de croûte océanique chévauché de l'Océan de Mureş qui a formé la nappe initiale de Mureş. Ces ophiolites sont représentées par des roches liasiques du complexe des basaltes de fond océanique (O_1) et le complexe des sheeted dykes (O_2) connus dans la partie occidentale de la zone de Mureş. Elles sont des roches tholèiitique, fait indiqué par les teneurs en éléments majeurs et mineurs, ainsi que par la valeur des rapports entre les éléments chimiques. Les valeurs de ces rapports indiquent que les ophiolites sont des roches de marge de plaque et les situent sur les diagrammes dans le champ des roches de fond océanique.

Introduction

The eruptive rocks from the Trascău Mts were studied in 1987 on the occasion of the resuming of the researches on the geologic and metallogenetic evolution in the Metaliferi Mountains with a view to the correlation of their eastern part with the western one. Thus, the presence of the Liassic ophiolites – widespread in the Drocea Mts in the west of the Mureş Zone – was revealed in this area, too.

Although the Mesozoic magmatites were described in the Trascău Mts since the last century (Hauer and Stache, 1863; Szentpétery, 1916; Papp, 1914; Ilie, 1935, 1940; Ghiţulescu and Socolescu, 1941) no ocean floor rocks were rendered evident up till now. Rădulescu and Săndulescu (1973), Bleahu (1974) and Herz and Savu (1974) mentioned in their papers on the geotectonic evolution of the Romanian territory that ophiolites are ocean floor rocks. According to the above-mentioned authors this term included all the Mesozoic eruptive rocks from the Southern Apuseni Mountains. In recent years the Mesozoic eruptive rocks from the southern Trascău Mts were considered by Cioflica et al. (1980), Cioflica and Nicolae (1981), Nicolae (1983, 1985) and Lupu (1983) volcanics which erupted in an active marginal basin concomitantly with the deposition of the Feneş Beds (Lupu, 1983; Nicolae, 1985) of Barremian-Lower Aptian age. This age was pointed out by other authors (Ilie, 1935; Ghiţulescu and Socolescu, 1941; Giuşcă et al., 1963)[1]. Gandrabura (1981) mentioned that the volcanics from the north of the Trascău Mts belong to the second stage of the Mesozoic volcanism in the Southern Apuseni Mountains according to the classification of Giuşcă et al. (1963). This author also admitted the possibility that eruptions from the first stage might also exist in this area, as well. More recently, Mârza (1987) represented on his map all the Mesozoic volcanics from the Trascău Mts as a Jurassic-Cretaceous (J-Cr_1) ophiolitic complex.

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[1] The conclusion that the Mesozoic eruptive rocks in the south of the Trascău Mts are Barremian-Aptian in age was inferred from the fact that the post-Neocomian flysch in the study area includes, beside other olistoliths types, numerous exotic blocks of Late Kimmerian island arc volcanics (J_3-Cr_1), considered as eruptions synchronous with its sedimentation.

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Savu (in Savu et al., 1988) pointed out the presence and position of the ocean floor rocks in the Trascău Mts and indicated that the basement of this region includes an ocean crust fragment formed of Liassic ophiolitic rocks obducted from the Mureş Ocean crust, from which olistoliths were torn off and insedimented in a volcano-sedimentary mélange.

Ophiolitic Rock Occurrence and Petrographic Data

The Liassic ophiolitic rocks (180 Ma, Herz et al., 1974) – which in the southwest of the Mureş Zone constitute the allochthonous basement unconformably overlain by Late Kimmerian island arc volcanics and flysch deposits – occur in the Trascău Mts in three different tectonic positions: (1) in the Podeni Unit, (2) in the northeastern segment of the Ampoi Cordillera and (3) as olistoliths in Late Kimmerian island arc volcanics of the tectonic units of the 'Bedeleu Nappe System' (Pl. I).

1. Podeni Unit (Savu, 1987 – unpublished field evidence) thrusts from the southeast over the Late Kimmerian island arc volcanics from the north of the Trascău Mts, included in the Rimetea Unit separated by Lupu (in Bleahu et al., 1981), which is situated between Poiana Ampoiului and Turda. This unit also comprises Stramberg limestones (J_3-Cr_1) which overlie the above-mentioned volcanics, pointing to an Upper Jurassic age.

Podeni Unit seems to be an equivalent of the Căpîlnaş-Techereu Unit described by Lupu (1965), possibly its prolongation in the Trascău Mts; both units represent fragments from the big Mureş ophiolitic sheet initially obducted from the crust of the ocean with the same name (Savu, 1983; Savu et al., 1987) whose closing was achieved at the end of the Neocomian. This unit develops between Pietroasa and Podeni. It consists – as one can see from the formations cropping out from under the Tertiary sedimentary deposits of the Transylvania Basin – of Liassic ophiolites, identical in all respects with those from the ocean floor basalt complex (O_1) described in the west of the Mureş Zone (Savu, 1983). In places they underlie small outliers of Late Kimmerian island arc volcanics. Liassic ophiolites are represented by submarine flows of basaltic rocks in pillow lava facies, occurring on Porcul (Csumor) and Imbre valleys at Pietroasa, as well as by aphanitic basalt flows which are found on Hidiş (Rachiş) Brook, south of Podeni. In petrographic respect the rocks were mentioned by Gandrabura (1981) and referred to the first stage of the Mesozoic volcanism in the southern Apuseni Mts established by Giuşcă et al. (1963). He also mentioned microgabbros at Pietroasa.

The basic flows consist of basalts with an intersertal texture, rarely of spinifex type up to variolitic, and of spilites. The 1–2 cm thick crust of the pillow separations is formed of hyalobasalt. Between these separations occurs a glassy matrix, devitrified and chloritized under conditions of ocean floor metamorphism. Basalt volcanic breccias appear as thin intercalations between the basic rock flows. At the confluence of Hidiş (Rachiş) Valley with Vadu Pleş Brook basalts are intruded by a dolerite dyke.

2. Ampoi Cordillera represents (acc. to Savu, 1987 – unpublished field evidence) the tectonic rise of the basement of the southern part of the Trascău Mts which took place in the Barremian-Aptian sea during the sedimentation of the upper part of the "Feneş Beds". It extends from the Ampoi upper basin (Zlatna), north of Feneş towards the east and then from Ampoiţa it trends NNE, describing an arc of a circle, thus reaching the north of the Tibru Valley (Pl. I). This cordillera consists, in the west and southwest, mostly of Tithonian-Neocomian flysch in association with Late Kimmerian island arc volcanic tuffs and Stramberg limestones. In places these formations are slightly metamorphosed (deformed) regionally (Bleahu, Dimian, 1967). From Ampoiţa Fault towards NNE the cordillera consists of Mesozoic eruptive rocks, figured by Nicolae (1985) on his map for the Barremian with a question mark. This cordillera yielded several olistoliths of magmatic and sedimentary rocks from Barremian-Aptian flysch formations in the Trascău Mts and especially from the formations of the Feneş Unit described by Lupu (1975) which include the cordillera nowadays.

In the northeastern segment of the Ampoi Cordillera, lying between Tibru and Ampoiţa valleys, the ocean floor Liassic ophiolites are represented by an ocean crust fragment torn off the sheeted dyke complex (O_2) described by Savu (1983) in the west of the Mureş Zone. This ophiolitic fragment (ca. 2 km long and 1 km wide) is in tectonic relationships with all surrounding formations. The ophiolitic fragment is comprised between two longitudinal faults, with a NNE-SSW trending and a steep dip that is the western fault, which separates it from the Late Kimmerian island volcanics associated with Upper Jurassic limestones, tuffs and jaspers belonging to the cordillera, and the eastern one, which separates it from the flysch formations – and two transversal faults with a WNW-ESE trending, among which Ampoiţa Fault. All this shows that the Mesozoic eruptive rocks in the northeastern part of the Ampoi Cordillera do not intrude the sedimentary deposits as presumed by certain workers, they penetrated the flysch formations being pushed upwards by tectonic forces (Fig. 1) as indicated by the almost vertical position of the tuff and jasper beds on Valea Mare (Ţelna) or

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by Tithonian limestones intercalated in the Late Kimmerian island arc volcanics (Pl. II, Fig. 1) from the cordillera.

The sheeted dyke fragment (100 % dykes),which might belong to the Podeni Unit, occurs immediately north of the Ampoiţa Fault, the transversal southern fault lying along the valley with the same name. Sheeted dykes trend N20^{0}E/74^{0}S - N40^{0}E/70^{0}S (Pl. II, Fig. 2). They consist of basalts with an intergranular texture, dolerites and albite-bearing dolerites (orthospilites) with subophitic or ophitic texture and fine-grained gabbro-dolerites, rocks often albitized, with chloritized clinopyroxene. Dykes of fine-grained rocks resembling the albitic quartz-diorites described by Savu (1985) in the Drocea Mts and albitic felsites occur as well. Basic dykes display 'chill margins' either on one side or on both sides (Pl. III, Fig. 1).

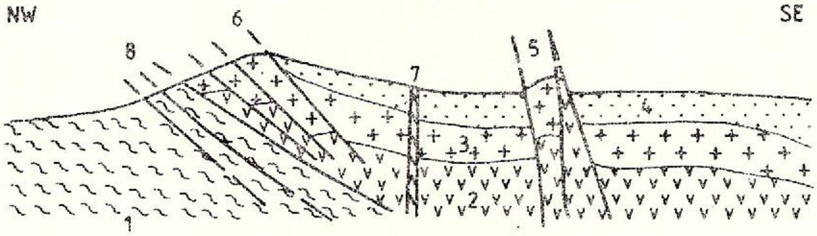

Fig. 1 - Model concerning the formation of the Ampoi Cordillera during the sedimentation of the Barremian-Aptian flysch and the tectonization of the Late Kimmerian island arc at the contact with the sialic part of the Apuseni Mountains Plate.
1, crystalline schists from the sialic part of the Apuseni Mountains Plate; 2, Liassic ophiolitic basement obducted from the crust of the Mureş Ocean; 3, Late Kimmerian island arc volcanics (J_3-Cr_1); 4, flysch; 5, Ampoi Cordillera (liassic ophiolites + Late Kimmerian volcanics) in process of rising in the Cretaceous sea; 6, tectonized Late Kimmerian island arc; 7, Laramian rocks intrusions; 3, old Benioff plane.

At Ampoiţa, where exposures are quite good, sheeted dykes were tectonized during the rise of the cordillera through the sedimentary deposits (Pl. III, Fig. 2). They are penetrated by some small intrusions of Late Kimmerian granites (as also mentioned by Ştefan, 1986), similar with certain facies of the Săvîrşin granite. Their radiometric age (K/Ar) determined by Vâjdea and Tănăsescu (I.G.G.) is 92.5 ±12.7 Ma. These rocks were regarded as banatites. They were tectonized, a phenomenon which according to the K/Ar datings occurred during the Austrian orogenesis[2].

3. As olistoliths (exotic blocks) Liassic ophiolites are found in the pyroclasts of the Late Kimmerian island arc volcanics which enter into the constitution of the imbricated tectonic units of the 'Bedeleu Nappe System', described by Bleahu et al. (1981), situated in the northwestern part of the Trascău Mts (Pl. I). Here, the Late Kimmerian volcanics with ophiolitic rock olistoliths constitute a mélange with pyroclastic matrix (coloured mélange), similar to that described by Savu (1984) in the southern part of the Metaliferi Mountains. Such olistoliths were found in pyroclasts occurring between Inzel Valley and Munte Valley (Colţeşti). They are represented by basalts in pillow lava facies and dolerites at Izvoarele and by dolerites and gabbro-dolerites on the Inzel, Bedeleu and Munte valleys. The presence of dolerites as intrusions on Munte Valley was mentioned by Russo-Săndulescu et al. (1976), who referred them to the Kimmeridgian-Tithonian volcanic (calc-alkaline) series (acc. to Lupu, 1972). The same age and origin were assigned to these rocks by Nicolae (1985). He described in the area north of Valea Mînăstirii - where the above-mentioned mélange develops - dykes of microgabbros, dolerites, anamesites, and basalts, which would penetrate the keratophyres from the Bedeleu Nappe and the tectonic units from its system.

On Bucerdea Valley, in the north-eastern segment of the Ampoi Cordillera, at the eastern contact of the volcanics with the Barremian-Aptian flysch, dolerites are found in a block of 10 m, which is in trenchant contact with the Late Kimmerian pyroclasts. Due to unfavourable observation conditions one could not establish whether dolerites represent only an olistolith included into pyroclasts - a possibility closer to reality - or a part from the western margin of the fragment torn off the sheeted dykes occurring at Ampoiţa.

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[2] The radiometric age was established on a sample of tectonized granite collected by Savu from the intrusion which penetrates the sheeted dykes in the southern part of the slab, nearby the primary school in Ampoiţa.

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Table 1. Chemical composition of the ophiolitic rocks.

No^{+}	1	2	3	4	5	6	7	8	9	10
SiO_2	44.72	45.20	45.68	48.52	48.95	49.93	50.35	50.80	51.05	51.33
TiO_2	1.12	1.04	0.88	2.14	1.62	1.22	2.14	1.10	2.60	2.34
Al_2O_3	15.95	16.80	14.24	14.29	16.56	15.85	15.50	16.44	14.44	14.29
Fe_2O_3	4.39	3.47	3.46	6.33	7.62	4.37	4.36	3.01	8.08	8.81
FeO	4.86	5.42	3.75	6.05	6.36	4.56	6.64	3.97	6.41	4.98
MnO	0.12	0.15	0.15	0.19	0.15	0.12	0.18	0.19	0.20	0.23
MgO	6.47	8.58	5.12	5.23	6.42	6.66	6.28	6.76	4.28	4.73
CaO	13.58	9.73	10.63	9.71	2.59	8.20	7.06	8.49	5.51	4.30
Na_2O	2.66	3.41	5.42	3.65	4.56	3.92	4.69	3.66	5.48	5.36
K_2O	0.35	0.47	0.11	0.62	0.09	0.96	0.28	0.50	0.25	0.11
P_2O_5	0.08	0.09	0.13	0.21	0.15	0.15	0.20	0.10	0.25	0.40
S	0.14	0.20	0.20	0.22	0.22	0.22	0.10	0.14	0.09	0.10
CO_2	3.12	1.45	7.54	0	0	0	0	2.26	0.70	1.92
H_2O^+	2.33	3.47	3.29	2.90	4.32	3.28	1.96	2.76	0.63	1.70
Total	99.89	99.48	100.60	100.06	99.61	99.44	99.74	100.18	99.97	100.60
Ni (ppm)	140	70	80	46	37	40	30	70	14	12
Co	38	31	23	31	7.5	22	34	30	34	17
Cr	340	240	300	115	65	230	44	160	1.5	2.5
V	310	240	320	420	270	190	270	280	500	270
Sc	40	36	30	40	32	31	30	40	32	27
Zr	68	65	65	125	70	110	155	75	220	220
Y	20	20	18	40	30	23	42	20	55	55
Yb	2.5	2.8	3	5.2	4.4	3.7	4.0	3.4	5.5	4.8
Ba	10	15	10	60	13	170	32	30	28	14
Sr	120	380	125	260	145	360	275	300	141	79.9
Rb	-	-	-	-	-	-	3.29	-	13.0	9.00
Pb	2	2	2	2	5.5	3	6.5	2	9	8
Cu	85	130	60	65	4	18	3	30	3	2
Zn	32	30	30	44	32	30	42	30	38	90
Ga	14	12	13	17	19	18	12	12	17	17
Rb/Sr	-	-	-	-	-	-	0.011	-	0.091	0.123
Sr^{87}/Sr^{86}	-	-	-	-	-	-	0.703	-	0.703	0.705

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Table 1 (continued)

No	11	12	13	14	15
SiO_2	51.40	54.09	55.83	60.90	62.16
TiO_2	1.98	2.64	2.76	1.38	1.08
Al_2O_3	14.80	13.79	14.54	14.04	13.84
Fe_2O_3	5.65	5.38	6.53	6.44	5.88
FeO	7.26	6.40	5.16	3.75	2.40
MnO	0.23	0.18	0.16	0.07	0.13
MgO	5.13	3.05	3.90	2.27	1.69
CaO	6.00	5.21	4.29	3.56	4.35
Na_2O	4.37	5.06	4.52	5.94	6.50
K_2O	0.23	0.17	0.23	0.08	0.15
P_2O_5	0.21	0.42	0.25	0.48	0.36
S	0.22	0.09	0.14	0.20	0.28
CO_2	0	1.27	0	0	0.28
H_2O^+	1.85	2.22	1.36	1.10	0.58
Total	99.55	99.97	99.67	100.21	99.68
Ni (ppm)	29	9.5	16	7.5	3.5
Co	19	23	22	12	9
Cr	50	2.5	9.5	8	1.5
V	310	310	420	33	55
Sc	32	29	36	18	15
Zr	85	240	240	250	320
Y	34	60	80	88	72
Yb	5.2	4.6	7.0	7.5	7.0
Ba	38	24	10	10	15
Sr	290	111	75	105	64.3
Rb	-	2.95	-	-	4.75
Pb	4.5	6.0	9.0	5.5	2
Cu	2	4	2	2	3
Zn	30	65	30	30	38
Ga	18	16	19	28	16
Rb/Sr	-	0.026	-	-	0.073
Sr^{87}/Sr^{86}	-	0.706	-	-	0.705

$^+$The analyses from this table represent: basalts - 1, Rachiş Valley; 2, 3, Porcu (Csumor) Valley; dolerites - 4, Bucerdea Valley; 6, Muntele Valley (Colţeşti); 8, Rachiş Valley; albite basalts and dolerites from the sheeted dyke complex (orthospilites) - 5, 7, 9, 10, 11, Ampoiţa; albite microdiorite - 12, Ampoiţa; albite felsites - 13, 14, 15, Ampoiţa.

We mention that in all samples Nb is lower than 10 ppm; Sn is lower than 2 ppm, excepting sample 9 which contains 2.5 ppm Sn; La is lower than 30 ppm.

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Geochemistry and Tectonic Setting

With a view to the geochemical study of the Liassic ophiolitic rocks 15 samples of rocks collected from different places in the study area were analysed (see the table). The table shows that the Liassic ophiolites are basic rocks with a content of SiO_2 lower than 52 per cent. Four samples – one albitic microdiorite and three albitic felsites – represent acid differentiates of the normal tholeiitic magma (Savu et al., 1984). Except for sample 1 the content of Na_2O from the basic rocks is much higher than 2.50 per cent, a value established by Miyashiro (1975) for tholeiitic rocks, which points out characteristics of albite-bearing rocks (spilites) of the Liassic ophiolites in the study area. In the four rocks from the acid differentiate group the content of Na_2O is comparable with that of similar rocks from the sheeted dyke complex of the western part of the Mureş Zone (Savu et al., 1985). Correspondingly CaO is generally lower than in normal basic rocks. These features can be observed in rocks from the sheeted dykes, which might point to primary spilites (Piirainen, Rauhunkovski, 1974; Reinhardt, 1974) or orthospilites, as well as in basalts in pillow lava facies, whose albitization could be also the result of the lava reaction with the ocean water.

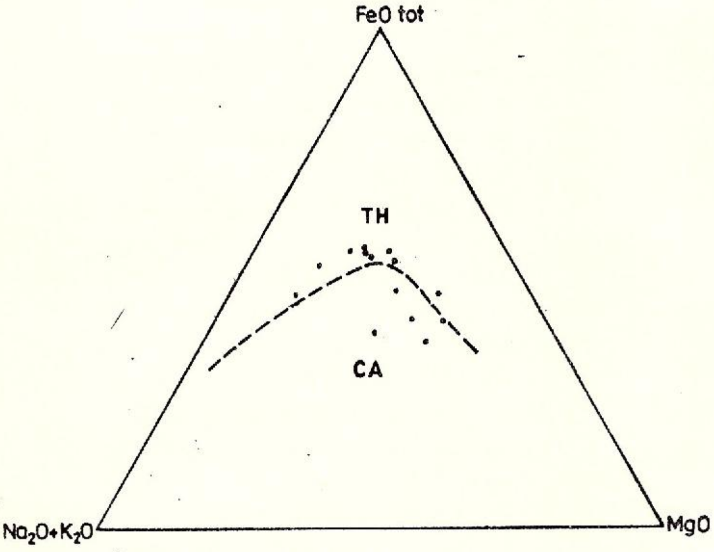

Fig. 2 – $FeO_{(tot)}$–MgO–Na_2O+K_2O diagram (Irvine, Baragar, 1971).
TH, tholeiitic rocks field. CA, calc-alkaline rocks field.

The diagrams in Figures 2 and 3 show clearly that the analysed ophiolitic rocks are normal tholeiites, comparable with the abyssal tholeiites described by Miyashiro (1975). This conclusion is confirmed by the contents of TiO_2 specific to ocean floor tholeiitic series (Pearce, Cann, 1973). In agreement with the trends of the ocean floor parental tholeiitic magma differentiation (acc. to Savu et al., 1984) most of the rocks from the acid differentiate group also display the highest contents of TiO_2 in the whole series of ophiolitic rocks analysed (Tab.). Exceptions are the albitic felsites (nos. 14 and 15) which are very poor in TiO_2.

The content of K_2O, which in some cases is low due probably to albitization, represents 0.50 per cent in the other rocks, a value indicated by Miyashiro for normal tholeiites.

The trace elements (Tab.) also characterize the rocks as representing a tholeiitic series. Siderophile elements (Ni, Co, Cr, V) display high contents specific to ocean floor rocks. Values characteristic of such a series are also displayed by Zr. The Zr/Y and Ti/Y ratios (Pearce, Gale, 1977) show that ophiolites are plate margin magmatic rocks (Fig. 4). As shown on the diagram in Figure 5 (Shervais, 1982) ophiolites clearly fall in the ocean floor tholeiites field, which ascertains the above-mentioned observations. An exception are the two already mentioned acid differentiates (nos. 14 and 15), with a low-V content, which fall outside OFT field, at low-Ti values.

The value of the ratio between Ti and Zr contents determined the plotting of most of the Liassic ophiolitic rocks from the Trascău Mts in the MORB field (Fig. 6). In this case, too, the above-mentioned acid differentiates richer in Zr are an exception beside other four rocks among which some belong to the basic differentiates with higher contents of Ti and Zr. These rocks fall outside the MORB field, in the WPB field, although they do not pertain to this magmatic series.

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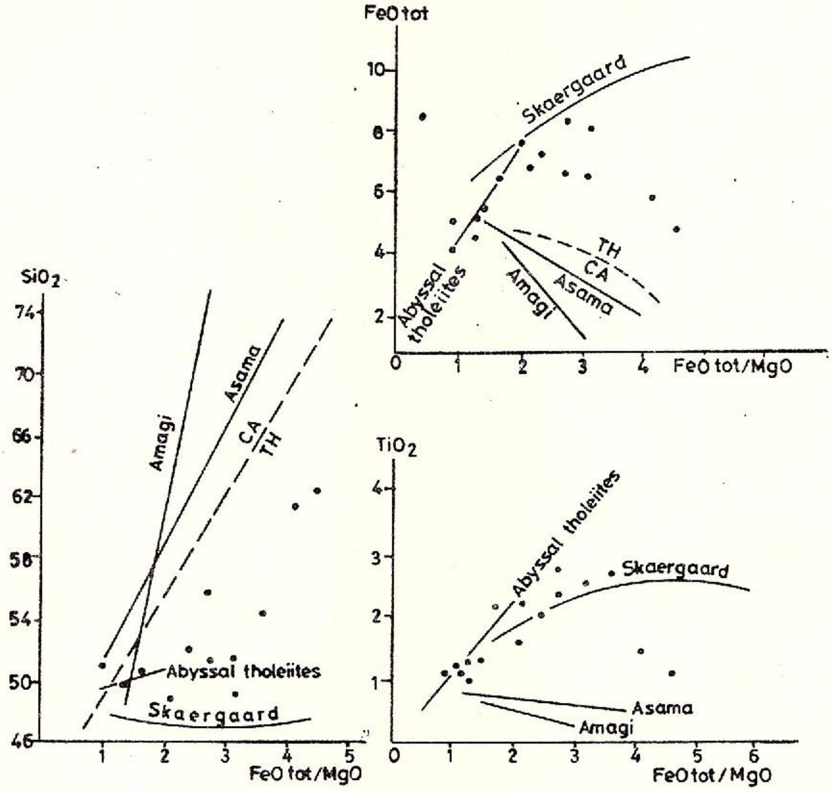

Fig. 3 – SiO_2, $FeO_{(tot)}$, TiO_2–$FeO_{(tot)}/MgO$ diagrams (Miyashiro, 1975). TH, tholeiitic rocks field; CA, calc-alkaline rocks field.

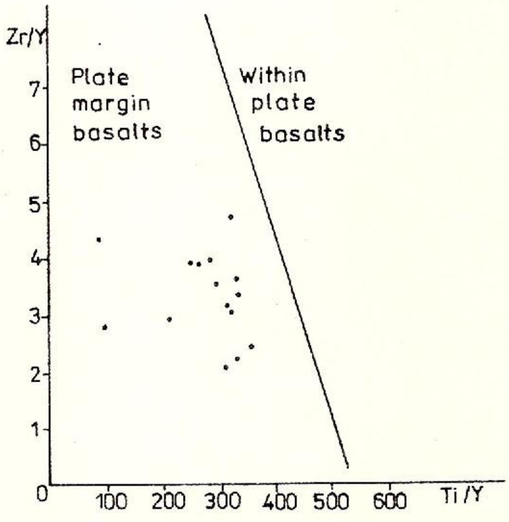

Fig. 4 – Zr/Y–Ti/Y diagram (Pearce, Gale, 1977).

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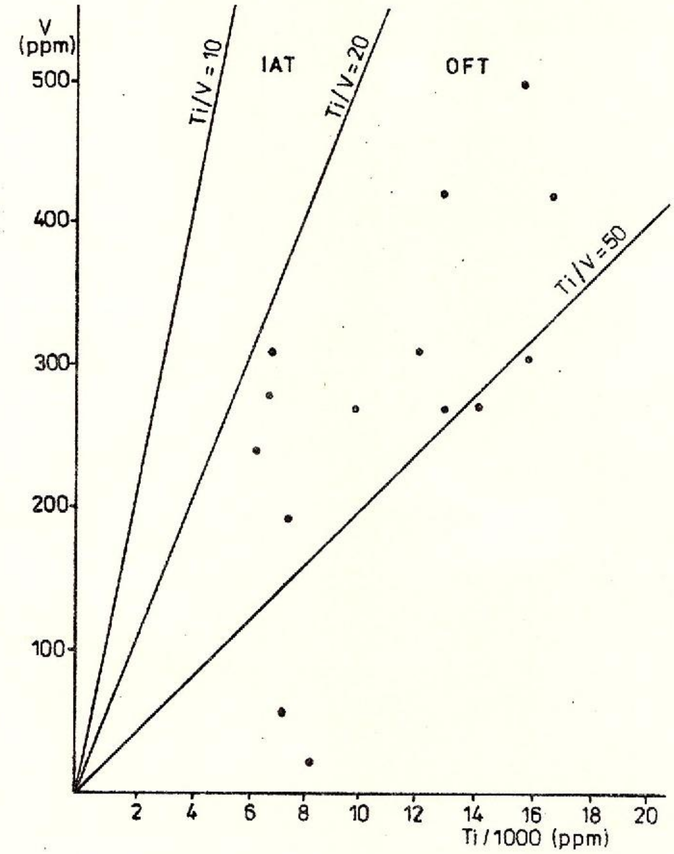

Fig. 5 - V-Ti/1000 diagram.
OFT, ocean floor tholeiites field;
IAT, island arc tholeiites field.

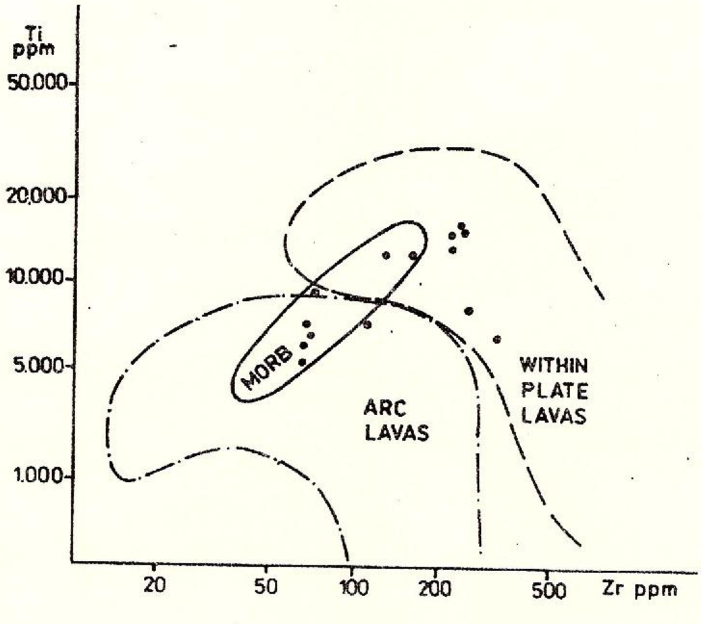

Fig. 6 - Ti-Zr diagram (Pearce, 1980).
MORB, field of basalts from median ocean ridges.

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The contents of Ba, with one exception, are much lower than 100 ppm, a characteristic of the abyssal tholeiites (Miyashiro, 1975), the ocean floor tholeiites, respectively. Values specific to such a rock series are observed in case of Sr, as well (Tab.).

The contents of Rb determined by Grabari (IGG) in five samples (Tab.) are typical of the ocean floor basic rocks; they are comparable with the contents determined by Herz et al. (1974) in such rocks in the west of the Mureş Zone. The $^{87}Sr/^{86}Sr$ ratio determined by Popescu (IFIN-Măgurele) on the same five samples displays values specific to the ocean floor tholeiitic rocks (see also Herz et al., 1974) in case of the first two samples (nos. 7 and 9), which represent normal tholeiites. In the other three samples representing rocks belonging to the acid differentiate group, which have been albitized, the initial $^{87}Sr/^{86}Sr$ ratio seems to have changed due probably to metasomatic processes. For this reason its value increased, as also observed by Herz et al. (1974) who studied rocks from the acid differentiate group in the Mureş Zone, e. g. granophyres. However, the high-Y content (60–88 ppm) and the low-Nb content (< 10 ppm) points to ocean floor rocks (Pearce et al., 1984), like the rocks in the Drocea Mts.

Conclusions

The Liassic ocean floor ophiolitic rocks from the Trascău Mts are allochthonous rocks; they are found in three different tectonic situations: (1) in the Podeni Unit; (2) in the northeasternmost part of the Ampoi Cordillera; (3) as olistoliths in the mélange with pyroclastic matrix from the imbricated units of the 'Bedeleu Nappe System', formed as a result of the tectonization of the northwestern arc of the Mureş Zone.

All the Liassic ophiolitic rocks come from the dismembering of the ocean crust megaslab obducted from the Mureş Ocean which constituted the initial nappe of the Mureş (Savu, 1983; Savu et al., 1987); at present its remains are found in the basement of the Trascău Mts and under the Transylvania Basin.

These ophiolites are represented by rocks from the ocean floor basalts complex (O_1) and sheeted dykes complex (O_2) occurring in the west of the Mureş Zone.

Petrochemically, the studied ophiolites are tholeiitic rocks as also indicated by their contents of trace elements and by the value of the ratios between some of them and the major chemical elements.

The values of these ratios point out that ophiolites are plate margin rocks; according to them these ophiolites plot in the ocean floor rocks field.

All this shows that the geological constitution of the Trascău Mts is similar to that in the rest of the Mureş Zone, except for the flysch deposits with olistoliths of eruptive and sedimentary rocks which are well represented.

References

Bleahu M. (1974) Zone de subducţie în Carpaţii româneşti. *D. S. Inst. Geol.*, LX/5, p. 6–25, Bucureşti.

– , Dimian M. (1967) Studii stratigrafice şi tectonice în regiunea Feneş-Ighiu-Intregalde (Munţii Metaliferi). *D. S. Inst. Geol.*, LIII/1, p. 281–304, Bucureşti.

– , Lupu M., Patrulius D., Bordea S., Ştefan A., Panin S. (1981) The structure of the Apuseni Mountains. *Carp.-Balk. Geol. Assoc., XII Congr., Bucharest 1981, Guide B3*, 106 p.

Cioflica G., Lupu M., Nicolae I., Vlad Ş. (1980) Alpine ophiolites of Romania: Tectonic setting, magmatism and metallogenesis. *An. Inst. Geol. Geofiz.*, LVI, p. 79–85, Bucureşti.

– , Nicolae I. (1981) The origin, evolution and tectonic setting of the Alpine ophiolites from the South Apuseni Mountains (Romania). *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 25, p. 19–29, Bucureşti.

Gandrabura E. I. (1981) Studiul mineralogic, petrografic şi geochimic al eruptivului mezozoic din Munţii Trascău. *An. Inst. Geol. Geofiz.*, LVIII, p. 5–121, Bucureşti.

Ghiţulescu T. P., Socolescu M. (1941) Étude géologique et minière des Monts Métallifères. *An. Inst. Geol. Rom.*, XXI, p. 181–463, Bucureşti.

Giuşcă D., Cioflica G., Savu H. (1963) Vulcanismul mezozoic din Masivul Drocea (Munţii Apuseni). *Asoc. Geol. Carp.-Balk., Congr. V, 1961*, II, p. 31–44, Bucureşti.

Hauer Fr., Stache G. (1863) Geologie Siebenbürgens. Wien.

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

– , Jones L. M., Savu H., Walker R. L. (1974) Strontium isotope composition of ophiolitic and related rocks, Drocea Mountains, Romania. *Bull. Volc.*, XXXVIII, p.1110–1124, Napoli.

Ilie M. (1935) Recherches géologiques dans les Monts du Trascău et dans le Basin de l'Arieş. *Ann. Inst. Geol. Roum.*, XVII, p. 329–466, Bucureşti.

– (1940) Structure géologique de la région aurifère de Zlatna (Roumanie). *An. Inst. Geol. Rom.*, XX, p. 75–146, Bucureşti.

Institutul Geologic al României

Irvine T. N., Baragar W. R. (1971) A guide to the chemical classification of the common volcanic rocks. *Can. J. Earth Sci.*, 8, p. 523–548, Ottawa.

Lupu M. (1965) Quelques considerations sur les phases du diastrophisme dans le sillon des Monts Métallifères. *Carp.-Balk. Geol. Assoc. VII Congr.*, I, Sofia.

– (1972) Stratigrafia şi structura formaţiunilor mezozoice din masivul Trascău. Teză de doctorat.

– (1975) Einige Bemerkungen zur Tektonik des Südlichen Apuseni Gebirges. *Rev. Roum. Géol., Géogr., Géologie*, 19, Bucureşti.

– (1983) The Mesozoic history of the South Apuseni Mountains. *Asoc. Geol. Carp.-Balk. XII Congr. An. Inst. Geol. Geofiz.*, LX, p. 115–124, Bucureşti.

Mârza I. (1987) Districtul petrometalogenetic (banatitic) Gilău-Trascău. *Stud. cerc. geol., geofiz., geogr. (Geol.)*, 32, p. 37–44, Bucureşti.

Miyashiro A. (1975) Volcanic rock series and tectonic setting. *Ann. Rev. Earth Planet. Sci.*, 3, p. 251–260, Palo Alto.

Nicolae I. (1983) Consideraţii pe marginea interpretării cadrului tectonic al ofiolitelor din Munţii Apuseni de Sud. *Stud. cerc. geol., geofiz., geogr. (Geol.)*, 28, p. 33–45, Bucureşti.

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

Papp K. (1914) Das taube Sedimente von Zlatna. *Jahresb. d. k. ung. geol. R.- A. f. 1914*, Budapest.

Pearce J. A. (1980) Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites. In A. Panayiotou Ophiolites - Proc. Intern. Ophiol. Symp. Cyprus, 1979, p. 261–272, Nicosia.

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

– , Gale G. H. (1977) Identification of ore deposition environment from trace-element geochemistry of associated igneous host rocks. In M. J. Jones Volcanic processes in orogenesis. Inst. Mining and Metallurgy and Geol. Soc. Special Publ. 7, p. 14–24, London.

– , Harris N. B. W., Tindle A. G. (1984) Trace element discrimination diagram for the tectonic interpretation of granitic rocks. *Jour. Petrology*, 25/4, p. 956–983, Oxford.

Piirainen T., Roumhunkoski P. (1974) General features of the spilitic rocks in Finland. In G. C. Amstutz, Spilites and Spilitic Rocks. Springer–Verlag, p. 191–206, Berlin.

Rădulescu D., Săndulescu M. (1973) The plate tectonics concept and the geological structure of the Carpathians. *Tectonophysics*, 16, p. 155–161, Amsterdam.

Reinhardt B. (1974) The relations between spilites and other members of the Oman Mountains ophiolite suite. In G. C. Amstutz, Spilites and Spilitic Rocks. Springer-Verlag, p. 207–228, Berlin.

Russo-Săndulescu D., Berza T., Bratosin I., Ianc R. (1976) Contribuţii la studiul petrologic al unor magmatite alpine din nordul Munţilor Trascău. *D. S. Inst. Geol.*, LXII/1, p. 165–194, Bucureşti.

Savu H. (1983) Geotectonic and magmatic evolution of the Mureş Zone (Apuseni Mountains). *An. Inst. Geol. Geofiz.*, LXI, p. 253–262, Bucureşti.

– (1984) Mélange-ul cu matrice piroclastică asociat arcului insular sudic al Zonei Mureş. *Stud. cerc. geol., geofiz., geogr., Geologie*, 29, p. 36–43, Bucureşti.

– , Udrescu C., Neacşu V. (1981) Geochemistry and geotectonic setting of ophiolites and island arc volcanics of the Mureş Zone (Romania). *Ofioliti*, 6 (2), p. 269–286, Bologna.

– , Udrescu C., Neacşu V., Stoian M. (1984) Trends of the tholeiitic magma differentiation in the sheeted dyke complex from the Mureş Zone (Romania). *An. Inst. Geol. Geofiz.*, LXIV, p. 121–131, Bucureşti.

– , Udrescu C., Neacşu V. (1985) Petrology and geochemistry of the sheeted dykes complex in the Mureş Zone, Dumbrăviţa-Baia-Bătuţa-Juliţa region. *An. Inst. Geol. Geofiz.*, LXIX/1, p. 129–148, Bucureşti.

– , Udrescu C., Neacşu V., Nacu D. (1987) Mid-ocean characteristics of ophiolites in Baia-Lupeşti-Vărădia area (Drocea Mountains), their tectonics and petrology. *D. S. Inst. Geol. Geofiz.*, 72-73/1 (1985; 1986), p. 161–180, Bucureşti.

– , Udrescu C., Lemne M., Neacşu V. (1988) Petrology, geochemistry and tectonics of the ophiolites from the oceanic paleoridge of the Mureş Zone between Pîrneşti, Troaş and Valea Zoldiş (Drocea Mountains). *D. S. Inst. Geol. Geofiz.*, 72-73/1 (1985; 1986), p. 237–257, Bucureşti.

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

Szentpétery S. (1916) Der Melaphyr und seine Rolle im Siebenbürgischen Erzgebirge. Földt. közl. XLVII, 148 p., Budapest.

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

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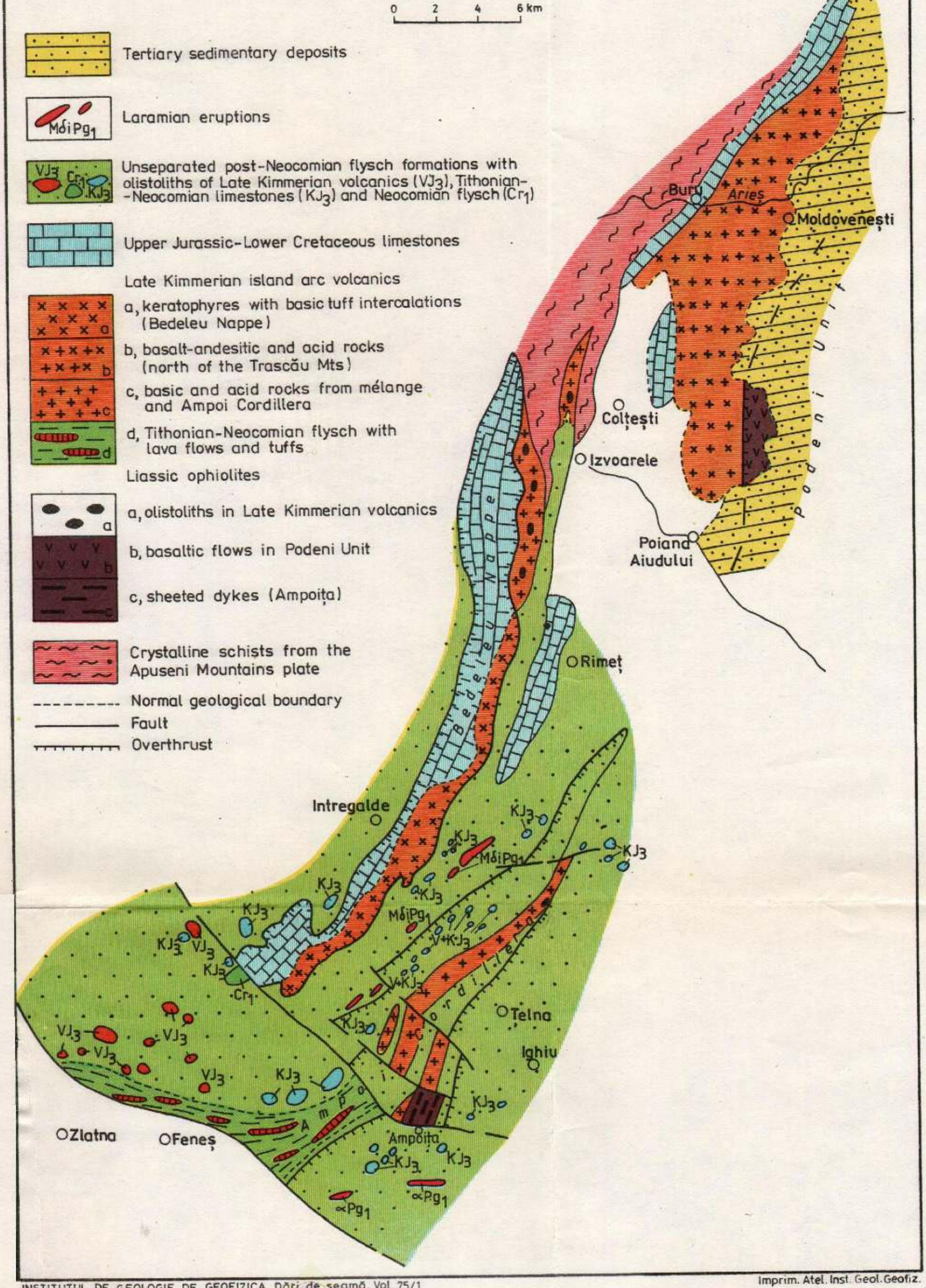
H.SAVU

GEOLOGICAL SKETCH-MAP WITH THE MESOZOIC ERUPTIVE ROCKS FROM THE TRASCĂU MTS

(after the I.G.G. map sc. 1:200000; E.Gandrabura, 1981; I.Nicolae, 1985 and author's data)



Imprim. Atel. Inst. Geol. Geofiz.

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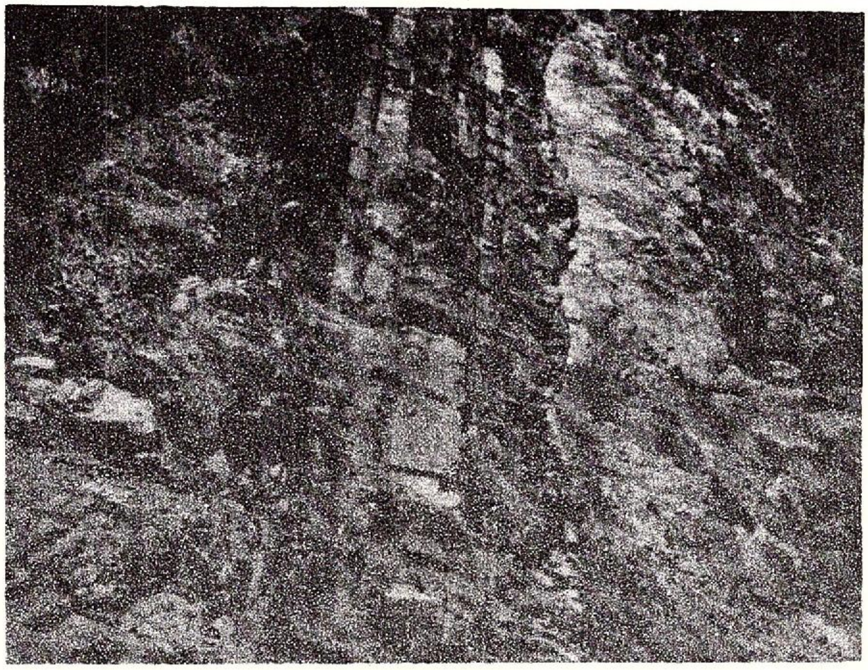
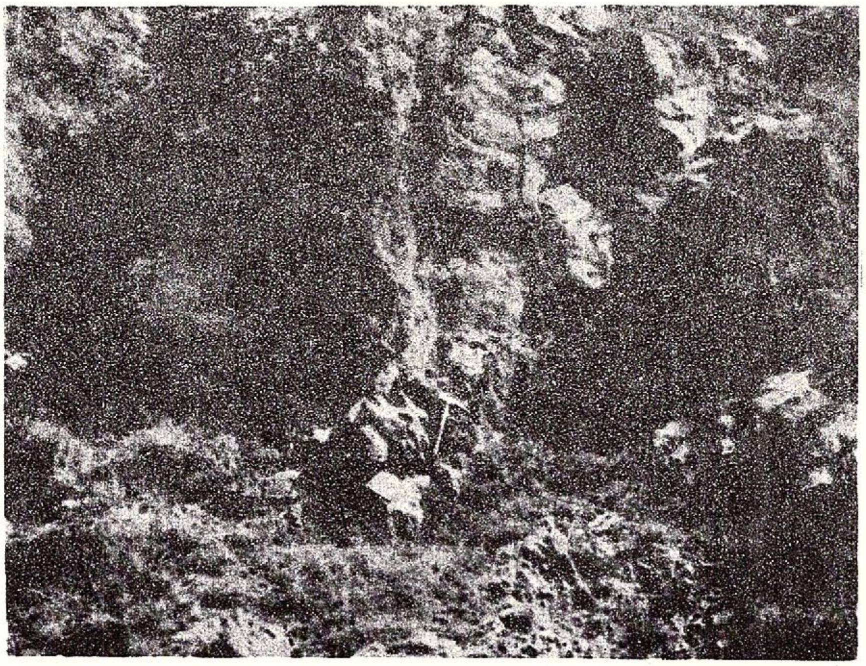


1



2

Plate II

Fig. 1 – Andesitic tuffs levels in alternation with jaspers, intercalated in pyroclastics of the Late Kimmerian island arc volcanics, in an almost vertical position (N 60°E/88°S). Valea Mare–Ţelna.

Fig. 2 – Sheeted dykes on the Ampoiţa Valley, west of the village.

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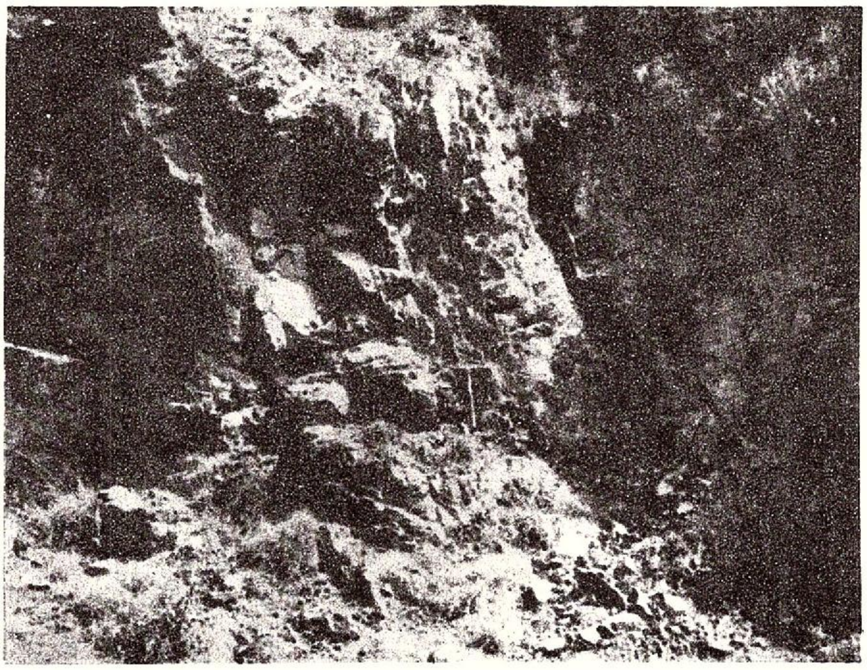

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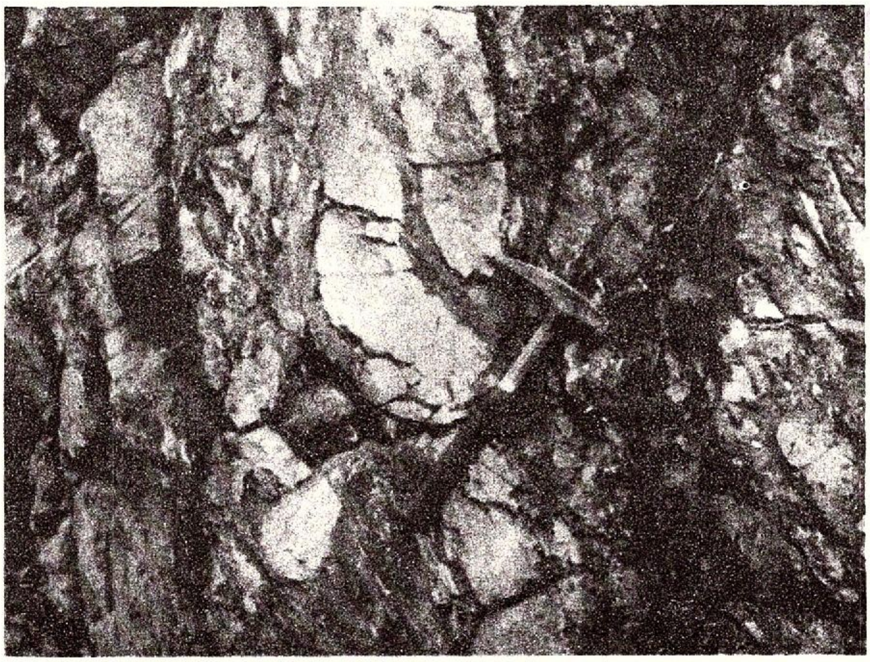

2

Plate III

Fig. 1 – Sheeted dykes displaying 'chill margins'. Ampoiţa Valley, west of the village.

Fig. 2 – Sheeted dykes tectonized (deformed) during the uplift of the Ampoi Cordillera through the Barremian-Aptian wildflysch. Ampoiţa Valley, west of the village.

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Rom. J. Petrology, 1992, 75, p. 63–74

PETROLOGY AND GEOCHEMISTRY OF THE LATE KIMMERIAN ISLAND ARC VOLCANICS FROM THE TRASCĂU MOUNTAINS (MUREŞ ZONE)

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Key words: Volcanic roks. Island arcs. Neocomian. Subduction. Major elements. Minor elements. Apuseni Mountains – South Apuseni Mountains – Trascău Mountains.

Résumé: *Pétrologie et géochimie des volcanites d'arc insulaire néokimmériennes des monts Trascău (zone Mureş).* Dans les monts Trascău sont présents les deux complexes volcaniques d'arc insulaire néokimmériens (IAV_1 et IAV_2) connus dans la zone de Mureş, les deux complexes comportant des roches calco-alcalines. Les roches se sont formées le long de l'arc volcanique du NW de la zone de Mureş, étant ainsi équivalentes aux vulcanites néokimmériennes (J_3–Cr_1) des monts Drocea. Elles ont fait leur apparition dans la zone de l'eugéosynclinal de Mureş parallélement avec la sédimentation des dépôts tithonique-néocomiens et reposent d'une manière discordante au-dessus du soubassement ophiolitique allochtone liasique, qui a été emplacé par obduction dans l'aire de la suture ophiolitique de Mureş. Pendant la sédimentation des dépôts de flysch post-néocomiens, le soubassement néokimmérien s'est élevé dans le miogéosynclinal en formant des soulèvements tectoniques ou cordillières, telle la cordillière d'Ampoi. L'érosion sous-marine rapide a y rompu de nombreux olistolites de roches volcaniques et calcaires de Stramberg qui ont été englobés dans les dépôts détritogènes post-néocomiens de l'unité de Feneş (faciès détritique) et des unités tectoniques adjacentes. Se forment aussi des niveaux de conglomérats grossiers où sont remaniées toutes les roches volcaniques et sédimentaires néokimmériennes (J_3-Cr_1). Le magma originel résulte de la fusion du manteau surjacent du plan de Benioff, pendant que celui-ci était métasomatisé (hornblenditisé) par des substances émanées par la croûte océanique liasique enfoncée. Le magma a subi un large processus de différentiation fractionnée de basaltes à des rhyolites, série calco-alcaline qui s'associe à une autre série alcaline collatérale qui renferme toute une gamme de roches, de basaltes à des rhyolites alcalines sodiques. Les vulcanites basiques forment, dans des différents diagrammes, un champ des roches mélanocrates et celles acides un champ des roches leucocrates. Dans ces champs se projetent tant les roches calco-alcalines, aussi bien que celles alcalines, correspondantes par leur teneur en SiO_2. Ces particularités des vulcanites sont imprimées par le caractère du magma originel, déterminé par l'évolution du processus de subduction, la composition de la croûte océanique liasique, le degré de métasomatisation du substratum éclogitique et le degré de fusion de ce dernier pour engendrer le magma originel des volcanites néokimmériennes.

Introduction

The first geologists who investigated the Trascău Mts were concerned especially with their stratigraphy and tectonics, the eruptive rocks being tackled generally. In the last decades studies on the petrology and geochemistry of these rocks and chemical and spectral analyses (Ianovici et al., 1969; Russo-Săndulescu et al., 1976) were effectuated. Subsequently, as numerous types of volcanics from the Trascău Mts (Gandrabura, 1981; Nicolae, Bratosin, 1980; Nicolae, 1985) were investigated, their calc-alkaline and alkaline character and the presence of spilites and keratophyres in this area were rendered evident. The Liassic ophiolites were presented by Savu et al. in another paper (this volume). Consequently, a synthesis study of the Late Kimmerian island

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arc volcanics in the light of the latest observations and their comparison with similar rocks from the Drocea Mts seems of a great interest.

Distribution of Late Kimmerian Island Arc Volcanics

Late Kimmerian volcanics occur in most of the structural units of the Bedeleu Nappe System, separated by Lupu (1983) and Bleahu et al. (1981) in the study area. They are also found in the Ampoi Cordillera (Savu et al., this volume) from the Feneş Beds and as olistoliths in the flysch formations[1].

Volcanics are widely spread in the north of the Trascău Mts (Rimetea Unit), between Turda, Arieş Valley and Poiana Aiudului, where Gandrabura (1981) described a wide petrographic variety. Here occur volcanics from both complexes distinguished in the Drocea and Metaliferi Mts (Savu, 1983). IAV_1 complex – the most important one – consists of basic intermediary rocks, e. g. clinopyroxene porphyritic basalts, locally amygdaloidal, basaltic andesites, augite hypershene andesites and hornblende pyroxene andesites as well as their alkaline equivalents (Tab. 1, Fig. 3). They appear as stratified pyroclasts (agglomerates, volcanic breccias, tuffs), more rarely as lava flows; their position in the Arieş Valley is N65^0E/70^0S. Here, the tuffs comprise huge blocks of pyroxene andesites (Pl. I, Fig. 1). 'Polygenous agglomerates' occur nearby Stramberg limestones. They consist of limestone blocks torn off during the explosion from the reefs formed on the volcanic cones (Savu, 1962 a). It shows that the limestone blocks are only exceptionally thrust, as in case of the Bedeleu Nappe. IAV_2 complex – the upper one – consists of pyroclasts of quartz andesites, dacites, trachyandesites (latiandesites), orthophyres (trachytes) and rhyolites, rocks often overlain by Stramberg limestones in normal position. All this points to an Upper Jurassic age of the volcanic rocks. According to Gandrabura (1981), between Copăceni and Rimetea, the Stramberg limestones rest upon rhyolitic tuffs. It points out that the volcanic rocks under discussion formed in the axial zone of the northwestern island arc of the Mureş Zone. The Na_2O content higher than 3 per cent in some rocks indicated the presence of keratophyres also in the north of the Trascău Mts.

Late Kimmerian volcanics occur in the units of the 'Bedeleu Nappe System', between Rimetea and Mînăstirea Valley. They are included in a system of scales and nappes trending N 20^0-40^0E/50^0-70^0S, in which the following succession of formations is repeated, from northwest to southeast: Paleozoic phyllites and limestones – (± Verrucano) – Late Kimmerian basic pyroclasts with Liassic ophiolite olistoliths (mélange – Savu et al., this volume) – Late Kimmerian acid tuffs + Stramberg limestones (J_3-Cr_1) – Neocomian sedimentary deposits. The volcanics found in the study area were described by Russo-Săndulescu et al. (1976).

Nicolae and Bratosin (1980) and Nicolae (1985) described Upper Jurassic keratophyres (160 Ma – Nicolae et al., 1987) in the Bedeleu Nappe: The lower part of the nappe consists of rhyolitic tuffs and keratophyres similar to those found north of Rimetea, also overlain by Stramberg limestones. These relationships indicate that the Bedeleu Nappe originated in the northwestern island arc of the Mureş Zone – which collided with the Apuseni Mountains plate – being thus strongly tectonized. Nicolae (1985) mentioned that keratophyres and Stramberg limestones recrystallized due to a burial metamorphism. Similar rocks, as regards the age and composition, from other tectonic units did not recrystallize and, consequently, we consider that the recrystallization and mylonitization of these rocks were rather determined by a dynamo-thermic metamorphism which affected the nappe during the Austrian movements, as indicated by the K/Ar dating (106.3±10.6 Ma) carried out by Văjdea and Tănăsescu (I.G.G.) on such a keratophyre from Galda Valley, east of Modoleşti. Other datings (Lupu et al., 1982, unpubl. report; Bleahu et al., 1984) pointed to rejuvenations at the level of the Lower and Upper Cretaceous.

The acid pyroclasts from the Bedeleu Nappe are intercalated with levels of basic tuffs resembling the petrographic assemblage yielded by the bimodal volcanism in the Drocea Mts (Savu et al., 1986b), which developed at the same time and in the same northwestern trough of the Mureş Zone – Drocea trough. The normal position of volcanics under the Stramberg limestones and their petrographic composition reveal their appurtenance to the upper complex (IAV_2) of the Late Kimmerian volcanics, like those from the Rimetea Unit. Uyeda (1981) mentioned that the bimodal arc volcanism developed in case of a Mariana-type subduction; it is more tensional and its Benioff plane shows a steep dip. Therefore, it may be supposed that the episode generating the bimodal volcanics from the Drocea trough and the Trascău Mts was connected with such a subduction. During the Callovian and Oxfordian the subduction plane was normal (Savu, 1983), yielding a Chilean-type calc-alkaline volcanism with andesites (IAV_1). Towards the end of the Upper Jurassic, before the collision, its dipping was very steep, as it was during the Mariana-type subduction, thus yielding bimodal

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[1] See the map appended to the previous paper on the Trascău Mountains (Savu et al., this volume).

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volcanics (IAV_2). The acid intrusions which occurred in the Neocomian, e.g. at Săvîrşin, Ampoiţa etc., are syncollisional intrusions.

The Late Kimmerian volcanics in the northeastern segment of the Ampoi Cordillera are represented by the complexes IAV_1 and IAV_2. Nicolae (1985) considered them as marginal basin volcanics, formed concomitantly with the deposition of the Barremian-Aptian sediments. It is to be noted that the volcanics are separated by faults both from the Ampoiţa sheeted-dike slab and from the Barremian-Aptian flysch deposits (detrital facies) at the upper part of the Feneş Beds (Pl. I, Fig. 2). These tectonic relationships show that the volcanics eruption did not take place concomitantly with the deposition of the sediments. The volcanics penetrated them along the faults, being pushed upwards by tectonic forces during the rise of the Ampoi Cordillera in the Barremian-Aptian sea.

The lower complex consists of agglomerates, volcanic breccias and tuffs of amygdaloidal porphyritic basalts, basaltic andesites and andesites, as well as alkaline rocks (Tab. 1, Fig. 3) including phenocrysts of plagioclase, augite, locally hypersthene and a resorbed amphibole. The crystallovitroclastic tuffs at Valea Mare (Ighiel) often display small globular calcite depositions. They are radially altered yielding spheroidal separations. Flows of porphyritic basalts in pillow lava facies, as at Valea Mare (Ţelna) and Valea Pietroasă (Moldoveneşti), are rarely found. Pillow lava separations differ from MORB ones as they are incompletely formed (Pl. II, Fig. 1), pass gradually to lavas, do not display a hyalobasalt crust and interpillow matrix. The groundmass of this type of basalt is locally oriented, with flow currents. Banded tuffs in alternation with jasper bands, reminding us of the Upper Jurassic Vorţa Formation in the Metaliferi Mts, are also found at Ţelna. These some metres-thick tuff levels have an almost vertical position and in places they form weathering witnesses. Their position is N 60^0-65^0E/86^0S, pointing out that the structure of the Late Kimmerian pyroclasts pile is unconformable with the N40^0E striking of the Ampoi Cordillera and that volcanics are not conformable with the strongly folded Barremian-Aptian sedimentary deposits. In the Valea Mică (Ighiel) tuffs, the Upper Jurassic limestones are intercalated with limestones bearing eruptive elements similar to those along the Tisa Valley, in the Drocea Mts, occurring in the base of the Tithonian-Neocomian deposits (Savu, 1962 a). These limestones also point to the Upper Jurassic age of the volcanics from the Ampoi Cordillera. The Valea Mare (Ţelna) augite andesites contain hornblendite nodules, which probably represent fragments from the metasomatized mantle on account of which the calc-alkaline parental magma of the Late Kimmerian volcanics formed.

The upper complex (IAV_2) is represented in the study area by pyroclasts of dacites, amphibole quartz andesites and keratophyres.

The southwestern segment of the Cordillera also consists of Late Kimmerian volcanics represented by basic tuffs conformably intercalated in the folded and slightly metamorphosed Neocomian sedimentary formations (Bleahu, Dimian, 1967). This association shows that the Late Kimmerian (Neocomian) volcanism extended in the Trascău Mts and in the Drocea Mts.

Numerous olistoliths of volcanic rocks from the post-Neocomian flysch formations of different tectonic units in the Trascău Mts consist of basalts and basaltic andesites (IAV_1), rocks chiefly found also in the primary deposit of the Late Kimmerian volcanics. They occur in association with Stramberg limestones, as in the Dîmbău Hill olistolith, nearby Zlatna. There are also olistoliths constituted only of limestones. The presence of Late Kimmerian volcanics both as conformable intercalations in the Tithonian-Neocomian pelitic-carbonatic formations from the basement of the detrital Barremian-Aptian deposits of the Feneş Unit and as olistoliths in the latter, made Ghiţulescu and Socolescu (1941), Giuşcă et al. (1963), etc., reach the conclusion that the Mesozoic volcanism lasted till the Aptian and Albian, respectively. All this indicates that all the island arc volcanics in the Trascău Mts – excepting banatites – are Upper Jurassic-Neocomian in age. They are reworked in the Barremian-Aptian deposits (Pl. II, Fig. 2).

The Ampoiţa sheeted-dyke slab is intruded by small biotite granite bodies resembling the Săvîrşin Late Kimmerian granite facies of 128 Ma (Savu et al., 1986a). They are formed of quartz, acid plagioclase, orthoclase and biotite, tectonically deformed minerals. These rocks formed in the intrusive stage of the Late Kimmerian island arc magmatism (Mitchel, Bell, 1973), which manifested during the Neocomian. This age was also stated by Ştefan (1986). The rocks are tectonized; the K/Ar datings effectuated by Vâjdea and Tănăsescu (I.G.G.) pointed to 92.5±12.7 Ma. This age shows that both the deformation of the Late Kimmerian granites and of the sheeted dykes and keratophyres took place during the Austrian movements. Their deformation probably started during the rise of the Ampoi Cordillera. Lupu et al. (1982, unpubl. report) also mentioned such a deformation, followed by a Laramian one.

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Table

Chemical composition of

No	1	2	3	4	5	6	7	8	9	10	11
SiO_2	45.30	48.05	48.90	48.98	49.16	49.90	49.98	50.49	50.65	50.94	51.08
TiO_2	0.34	0.60	0.94	0.86	0.60	0.88	0.80	0.74	0.74	0.84	0.78
Al_2O_3	12.30	15.24	18.61	17.26	17.51	17.96	15.94	14.84	14.84	18.46	17.16
Fe_2O_3	5.26	3.46	4.67	4.51	4.91	5.14	5.65	0.66	0.66	3.54	2.51
FeO	2.47	2.57	4.75	3.37	4.46	3.50	2.75	4.74	4.74	5.38	5.08
MnO	0.13	0.10	0.16	0.14	0.17	0.17	0.14	0.16	0.16	0.16	0.11
MgO	7.24	3.83	5.72	5.57	4.39	4.15	5.85	6.20	6.20	5.40	6.45
CaO	10.92	12.59	3.98	9.36	6.77	4.78	10.42	8.07	8.07	5.26	4.94
Na_2O	3.50	5.81	5.22	4.69	4.86	4.89	2.14	3.66	3.66	3.35	4.58
K_2O	0.96	0.14	1.50	0.31	0.28	1.73	0.69	2.46	2.46	1.60	0.19
P_2O_5	0.05	0.25	0.15	0.15	0.09	0.19	0.24	0.14	0.14	0.12	0.14
CO_2	5.92	5.66	0.50	1.22	0.60	3.14	2.83	5.90	5.90		2.03
S	0.19	0.13	0.20	0.23	0.22	0.18	0.13	0.16	0.16	0.19	0.09
H_2O^+	4.92	2.33	4.16	4.16	5.83	2.86	3.06	1.49	1.49	4.38	4.44
Total	99.54	100.76	99.46	100.81	99.85	99.47	100.62	99.87	99.87	99.62	99.52
Ni	70	46	19	90	30	18	34	60	60	38	60
Co	25	17	22	26	30	28	30	34	34	23	30
Cr	400	230	5.5	200	62	30	30	95	95	100	110
V	230	370	420	320	280	320	330	240	240	310	250
Sc	27	22	27	32	26	23	30	27	27	27	27
Y	<10	17	21	24	14	20	13	15	15	24	18
Yb	1.0	2.1	2.0	1.7	1.4	2.4	1.7	1.6	1.6	2.4	3
Zr	22	75	46	62	32	90	65	67	67	52	93
Ba	50	50	420	50	62	160	60	87	87	290	85
Sr	170	250	650	270	310	210	260	130	130	420	180
Ga	6.5	15	10.5	13	6.5	17	12	10	10	17	15
Cu	44	19	68	33	72	85	75	48	48	78	47
Zn	40	46	30	48	32	50	54	52	52	36	54

The analysed rocks represent: porphyritic basalts: 1, Valea Pietroasă (Moldoveneşti); 4, 7, Valea Faţa Furilor (Pătrinjeni); 10, Valea Mică (Ighiel); 11, Valea Vîltori (Zlatna); trachybasalts; 2, Valea Vîltori; 3, Valea Naibii (Zlatna); 5, Valea Faţa Furilor; 6, 8, Valea Feneş; 12, Valea Mare (Ighiel); basaltic andesites: 14, Valea Mică (Ţelna); basaltic trachyandesites: 9, Valea Mare (Ighiel);

Geochemistry and Tectonic Setting

For the geochemical study of the Late Kimmerian volcanics from the Trascău Mts, 23 new samples were analysed, which represent the main petrotypes (Tab. 1). As their eruption occured in submarine conditions, they reacted with the water so that although basaltic pyroxenes are fresh the plagioclase is turbid, as already mentioned, if the rocks contain amygdals filled with calcite, chlorite and zeolites. The presence of these minerals in rocks is reflected by the higher contents of CO_2 and H_2O. Table 1 shows a wide petrographic variety, from basalts with SiO_2 less than 52 per cent to rhyolites with SiO_2 more than 73 per cent, comprising calc-alkaline and alkaline rocks (Fig. 1).

In Figure 2 all rocks are assigned to basalts, andesites, dacites and rhyolites and plot in all the alkalinity fields, from 'low-K' to 'high-K'. The explanation of their behaviour is presented in Figure 3, in which the Late Kimmerian volcanics fall both in the calc-alkaline field and in the alkaline one. Two andesite analyses plotted in this diagram (according to Ianovici et al. 1969; Gandrabura, 1981). The calc-alkaline series starts from a basalt with 2.83 per cent $Na_2O + K_2O$ and reaches a rhyolite with 5.98 per cent alkalis. The basic term of the alkaline series is a basalt with 4.46 per cent alkalis; the final term is an alkaline rhyolite (quartzkeratophyre) with 8.69 per cent $Na_2O + K_2O$. The rocks of this series plot in the field of trachybasalts, basaltic trachyandesites, trachyandesites and trachydacites, alkaline rhyolite being the final term. In the Drocea Mts the basic rocks of the alkaline series were described (Savu, 1962 b; Savu et al., 1986) as limburgites, oligophyres and trachyandesites, and in the north of the Trascău Mts as latiandesites (Gandrabura, 1981), although they also include more basic terms. These rocks are rich in feldspar (oligoclase), as microlites and phenocrysts, partly decalcified, which accounts for the high Na_2O content, higher than in Liassic spilites.

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1

the volcanic rocks

tb tbp ap tbp tbt ta ta td d td r r

12	13	14	15	16	17	18	19	20	21	22	23
51.38	52.04	52.52	52.64	54.14	60.72	60.95	64.96	65.76	68.18	72.71	73.36
0.76	1.46	0.92	0.66	0.76	0.94	0.94	0.34	0.74	0.86	0.52	0.16
17.46	16.21	17.06	15.89	17.35	14.89	16.81	16.20	14.86	15.16	14.04	12.24
4.46	4.59	3.80	3.97	4.84	2.26	3.43	2.60	1.65	2.83	1.99	5.03
4.52	4.40	4.42	4.14	3.13	6.19	2.69	1.38	1.59	1.24	0.77	0.30
0.13	0.12	0.09	0.15	0.14	0.16	0.04	0.14	0.09	0.08	0.02	
5.29	5.02	4.66	3.28	4.26	2.68	2.14	1.50	1.77	1.25	0.68	0.07
5.43	7.50	7.99	5.39	4.95	1.88	2.72	1.70	3.75	1.39	1.66	0.45
4.84	4.74	2.99	7.28	5.34	6.24	5.53	6.68	5.11	7.46	5.70	4.38
1.21	1.36	0.84	0.17	1.72	0.26	1.40	1.93	1.52	0.47	0.28	4.31
0.16	0.15	0.17	0.25	0.20	0.30	0.29	0.11	0.07	0.15	0.14	0.03
	0.56		2.21	0.30	0.20	0.88		2.05			
0.24	0.16	0.10	0.10	0.20	0.10	0.09	0.16	0.23	0.24	0.09	0.15
4.06	2.06	4.74	3.35	2.16	2.80	2.08	1.96	1.53	0.82	1.22	0.31
99.94	100.37	100.30	99.48	99.49	99.62	99.99	99.66	100.72	100.13	99.82	100.79
85	18	26	17	30	13	12	9.5	9	15	6.5	28
23	19	27	26	19	25	9	5	4.5	5	2	<2
110	52	43	26	96	8	9.5	25	31	20	1.5	36
320	190	340	320	300	55	280	53	30	70	25	10
30	21	29	25	22	19	20	10	5.5	5.5	7	3
20	36	18	22	26	65	18.6	29	72	25	26	21
1.6	2.6	2.0	2.3	2.4	6.5	2.5	2.4	5.0	2.0	1.8	1.8
65	115	100	135	52	370	115	75	250	85	185	145
100	350	65	48	155	35	160	245	245	39	65	570
570	240	300	220	160	60	350	170	245	100	260	62
12	16	13	13	12	19	12	15	19	10	11	10.5
62	34	32	20	29	5	25	42	17	18	6	16
43	52	53	85	32	80	92	75	<30	<30	40	< 30

13, Valea Mînăstirea; 15, Valea Mică (Ţelna); 16, Valea Mare (Ţelna); trachyandesites: 17, Valea Mică (Ighiel); 18, Valea Mare (Ighiel); dacites (keratophyres): 20, Valea Mînăstirea; trachydacites (keratophyres): 19, Arieş Valley; 21, Valea Mare (Ţelna); rhyolites (quartz-keratophyres): 22, Valea Mică (Ţelna); 23, Valea Galda (Modoleşti).

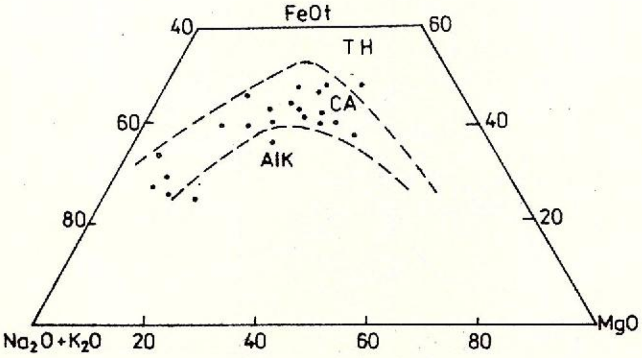

Fig. 1 – FeO(*tot*) – MgO – Na_2O+K_2O diagram (acc. to Irvine and Baragar, 1971, Hutchinson, 1982).

The two volcanic series are equivalents of the two series of Late Kimmerian volcanics found in the Drocea Mts (Savu, 1962 b). It is to be noted that in both areas, both in the calc-alkaline rocks and in the alkaline ones the content of Na_2O is always higher than the K_2O one; it can represent 7.28 per cent in a basaltic trachyandesite at Ţelna, pointing out the sodic alkaline character of the volcanics. In the final term of the alkaline series the two chemical components are in equal percentages and higher than 4 per cent, as in Buceava rhyolites (Drocea Mts). In leucocrate rocks, and particularly in rhyolites, there are cases in which $K_2O>Na_2O$ (Gandrabura, 1981; Nicolae, 1985). In the basic terms of the two series P_2O_5 and TiO_2 are always higher in the alkaline series than in the calc-alkaline ones. The high Na_2O character of the alkaline volcanics is also given by

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the value of the Niggli parameters, calculated by Gandrabura (1981) for the volcanics from the northern part of the Trascău Mts, among which some rocks correspond to the sodic syenitic magma type. On the diagram in Figure 3 these rocks are assigned to basaltic trachyandesites and trachyandesites. The sodic character is specific to acid rocks, among which Nicolae and Bratosin (1983) described keratophyres. The presence of high-Na_2O rocks in all alkaline compartments corresponding to the calc-alkaline ones points out their consanguinity. Alkaline rocks separated as a magmatic series collateral to the calc-alkaline one, which gives the character of the Late Kimmerian volcanism in the Mureş Zone (Fig. 1).

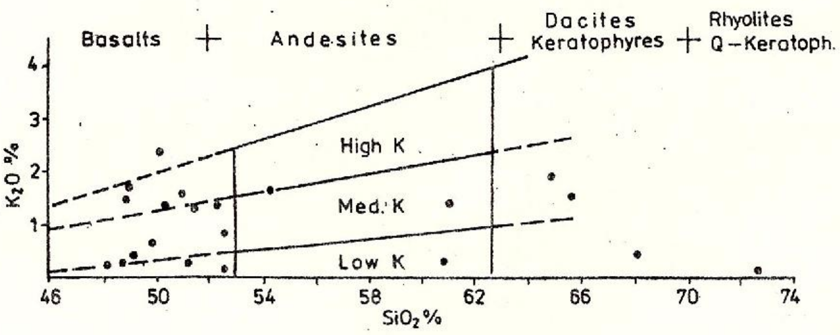

Fig. 2 – K_2O – SiO_2 diagram (acc. to Marriner and Millward, 1984).

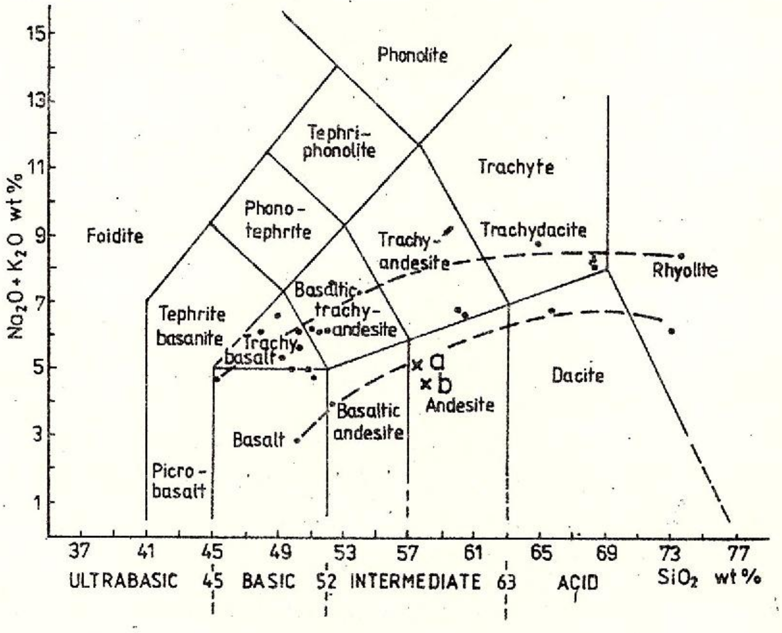

Fig. 3 – Na_2O+K_2O – SiO_2 (TAS) diagram (acc. to Le Bas et al., 1986; a and b – acc. to Ianovici et al., 1969 and Gandrabura, 1981).

The minor elements in the Late Kimmerian volcanics (Tab. 1) are typical of the calc-alkaline and alkaline series. The elements which are akin to iron and magnesium display decreasing contents from the basic to the acid rocks. In Figure 4, which shows Ni and Co variation versus MgO, several relationships can be observed. In case of Ni variation, the basic and calc-alkaline rocks are distributed along the differentiation line from basalts to rhyolites. Due to the process of collateral differentiation which took place at the level of each rock compart-

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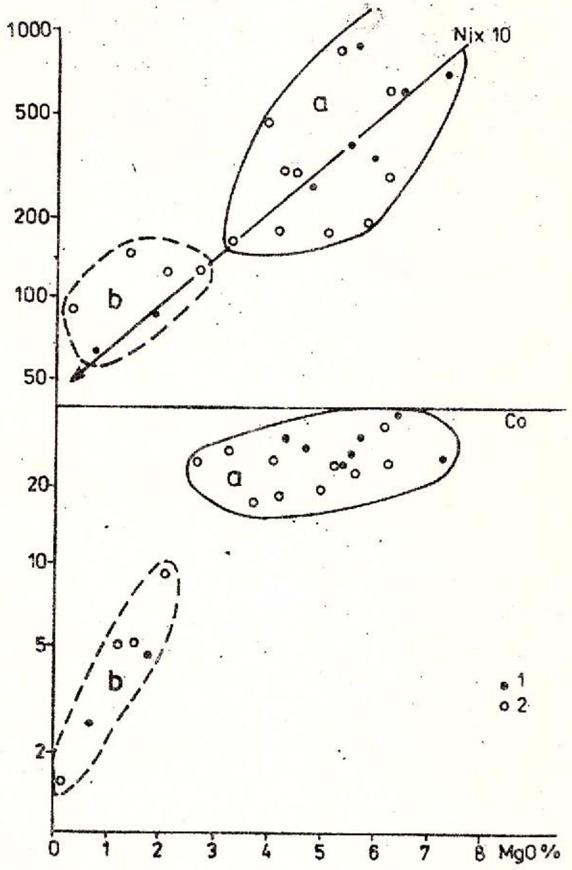

Fig. 4 – Ni x 10, Co – MgO diagram. 1, calc-alkaline rocks, 2, alkaline rocks: a, melanocrate rocks; b, leucocrate rocks.

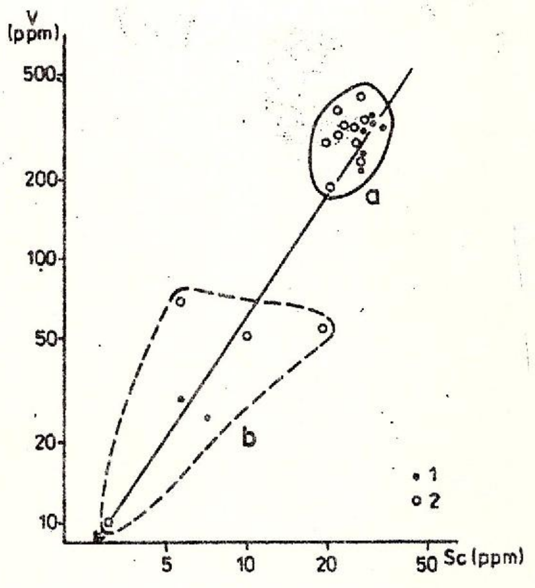

Fig. 5 – V – Sc diagram. 1, calc-alkaline rocks; 2, alkaline rocks: a, melanocrate rocks; b, leucocrate rocks.

ment (Fig. 3), alkaline volcanics corresponding to calc-alkaline ones are dispersed beyond the differentiation line. Thus, two characteristic fields result: the melanocrate rocks field and the leucocrate rocks field. This behaviour reminds us of the geochemical particularities of the Late Kimmerian island arc bimodal volcanics in the Drocea Mts (Savu et al., 1986b). The two fields in which calc-alkaline and alkaline rocks fall, in case of Co

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variation (Fig. 4), are more clearly emphasized. This situation is determined by the scarcity of the andesitic intermediary terms. Therefore, on whole, the Late Kimmerian volcanic activity in the Trascău Mts displays a bipolar character.

Figure 5 presents the V-Sc correlation. Here, calc-alkaline and alkaline melanocrate rocks are grouped in a small field, at high V and Sc values. The leucocrate rocks form a larger field, situated at lower values of the two elements. On this diagram the basic and acid calc-alkaline rocks are distributed along a marked differentiation line.

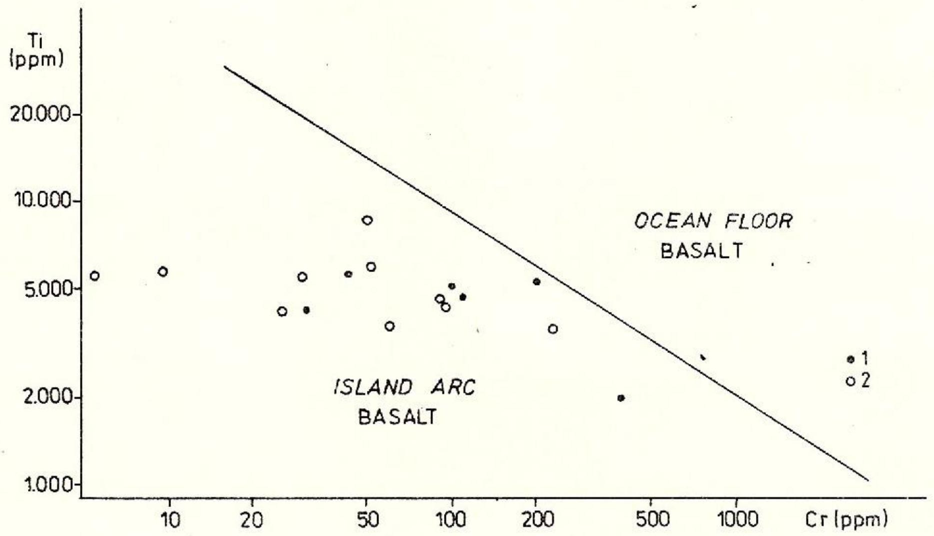

Fig. 6 – Ti – Cr diagram for basic rocks (acc. to Pearce, 1975).
1, calc-alkaline rocks; 2, alkaline rocks.

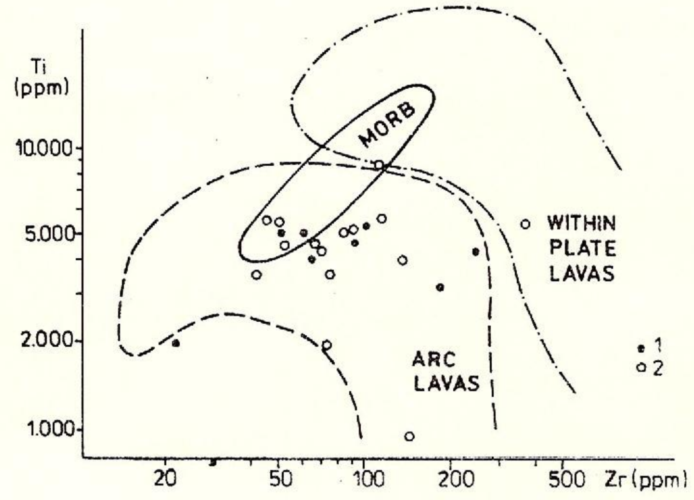

Fig. 7 – Ti – Zr diagram for all Late Kimmerian volcanics from the area (acc. to Pearce, 1980).
1, calc-alkaline rocks; 2, alkaline rocks.

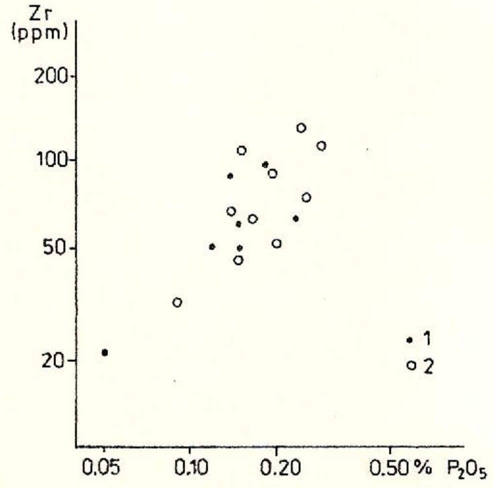

Fig. 8 – Zr – P_2O_5 diagram for basic rocks.
1, calc-alkaline rocks; 2, alkaline rocks.

Chrome has a quite large variation domain; even in melanocrate rocks its content varies much both in calc-alkaline rocks and in alkaline rocks. On the Ti-Cr diagram (Fig. 6) the basic rocks in the whole area of the Trascău Mts plot in the island arc volcanics field. The same tectonic situation is emphasized on the Ti-Zr

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diagram (Fig. 7) on which plot all the volcanics, both from the north and the south of the Trascău Mts. The same conclusion was reached by Cioflica and Nicolae (1981) for the Late Kimmerian volcanic rocks in the north of the area. In melanocrate rocks the Zr variation correlates with the P_2O_5 variation (Fig. 8), pointing also to the differentiation process underwent by the parental magma of the volcanics.

Origin of Volcanics

Late Kimmerian volcanics occurred in the Trascău Mts along the northwestern island arc of the Mureş Zone, like in the Drocea Mts (Savu, 1983). Volcanism developed as an intraeugeosyncline one, its products unconformably overlying the Liassic oceanic crust obducted from the basement of the Trascău Mts and partially intercalated in Tithonian-Neocomian sedimentary deposits. For this reason they present certain similarities with volcanics from 'intraoceanic' island arcs, e.g. those in the Mariana arc, directly overlying the oceanic crust (Meijer, Reagan, 1981). As in those island arcs (Ewart, 1970), in the Mureş Zone (Trascău Mts inclusive) the basic rocks predominate, especially basalts and basaltic andesites, the acid ones being found only in small amounts. Unlike the volcanics from the 'intraoceanic' arcs, in the Mureş Zone the island arc tholeiitic character is less evident; on the diagram in Figure 1 only a single rock plots in the TH field, calc-alkaline rocks being in association with alkaline rocks.

As the Late Kimmerian volcanism developed under eugeosyncline (intraoceanic) conditions, we cannot say that the magma formed in a sialic crust or it contaminated during the penetration of such a crust. Therefore, it is ascertained that these petrological peculiarities develop on the initial character of the parental magma which, in its turn, was influenced by several factors: e.g. evolution of the subduction process, composition of the subducted Liassic oceanic crust, metasomatisation degree of the eclogitic substratum and the degree of its melting. The modification of the character and probably of the subduction rate could influence the PT parameters in the formation area of the calc-alkaline basaltic magma in the mantle suprajacent to the Benioff plane. The appearance of the bimodal volcanics represents a first effect.

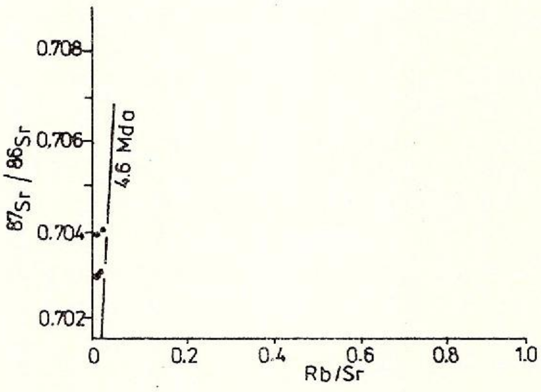

Fig. 9 - $^{87}Sr/^{86}Sr$ - Rb/Sr diagram. (acc. to Hedge and Noble, 1971).

The composition of the subducted Liassic oceanic crust determined the composition of the fluids bearing H_2O, CO_2, Cl and F, emanated during its eclogitization on the Benioff plane (Lloyd, Bailey, 1975; Pearce, Norry, 1979). These substances influenced the metasomatosis of the mantle which generated the parental magma and, later on, determined a more marked fractionation of this magma (Hole et al., 1984). In this respect it is to notice that in the basaltic complex of the oceanic crust slab from the Mureş Zone are intercalated levels of chemical precipitation limestones and levels of abyssal red argillites (Savu et al., 1988 a and b). Under subduction conditions these rocks released higher amounts of CO_2 and H_2O, respectively, than a simple basalt. The metasomatosis degree of the mantle depends on the amount of substances released (Mysen, 1979; Roden, Murthy, 1985). Hornblendite nodules from the Valea Mare volcanics (Ţelna) indicate the probable composition of the mantle in the melting zone. It is obvious that a magma resulting from the melting of a hornblendite will be more alkaline than a magma formed at the expense of a eclogite less altered by metasomatosis.

According to Pearce and Norry (1979) island arc basalts can be related to the same source as MORB, which had been modified by the above-mentioned processes. Nincovich and Hayes' diagram (1971) made us establish that the parental magma formed at a depth of ca. 125 km, a depth at which MORB often formed. However,

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the parental magma of the Late Kimmerian volcanics differs from it, as it is calc-alkaline and consequently its differentiation was different, following the differentiation trend of this magma type (Fig. 4).

Table 2
The variation of Rb/Sr and $^{87}Sr/^{86}Sr$ ratios

No	No in Tab.1	Rb ppm	Sr ppm	Rb/Sr	Age Ma	$^{87}Sr/^{86}Sr$ m	$^{87}Sr/^{86}Sr$ 0
1	8	20	970	0.02	140	0.703	0.703
2	16	20	165	0.12	140	0.703	0.703
3	17	20	220	0.09	140	0.704	0.704
4	18	40	540	0.07	140	0.704	0.704

The diagram in Figure 9 shows that the rocks analysed for the Rb/Sr ratio by Grabari (I.G.G.) and the $^{87}Sr/^{86}Sr$ ratio by Popescu (I.F.I.N.) are situated to the left of the 4.6 Ma isochrone, which represents the Earth's age. This position indicates that the source from the metasomatized mantle which generated the parental magma had low contents of Rb and other chemical elements and high contents of other elements due either to previous partial meltings (Gast, 1968; Carter et al., 1979) or to the metasomatosis processes, or to both phenomena. The values 0.703-0.704 of the last ratio indicates intraeugeosynclinal island arc rocks.

Conclusions

Late Kimmerian island arc volcanics (J_3-Cr_1) from both complexes (IAV_1 and IAV_2) of the Mureş Zone are present in the Trascău Mts. They formed in the area of the northwestern volcanic arc (Savu, 1983) and are equivalents with the Late Kimmerian volcanics from Drocea Trough in the west of the Mureş Zone, whose north-eastern prolongation is represented by the Trascău Mts[2]. Volcanics erupted in the area of the Mureş eugeosyncline, partially concomitantly with the deposition of Tithonian-Neocomian sedimentary formations; they unconformably overlie the Liassic allochthonous ophiolitic basement, obducted in the area of the Mureş ophiolitic suture.

During the sedimentation of the post-Neocomian flysch deposits, the Late Kimmerian basement was uplifted in miogeosyncline, yielding tectonic rises or cordilleras, e.g. Ampoi Cordillera. The rapid submarine erosion torn off numerous olistoliths of volcanic rocks and Stramberg limestones, which were included in the post-Neocomian detrital deposits of the Feneş Unit (detrital facies) and the adjacent units. In the detrital deposits of the Feneş Unit occur coarse-grained conglomerate levels in which all the Late Kimmerian volcanic and sedimentary rocks of the Ampoi Cordillera are reworked.

Except for the Liassic ocean crust fragments previously mentioned (Savu et al., this volume) and of banatites, all the Mesozoic eruptive rocks in the north and south of the Trascău Mts are Late Kimmerian island arc volcanics.

Parental magma formed by the melting of the mantle suprajacent to the Benioff plane, whereas it was metasomatized (hornblenditized) with substances released from the subducted Liassic oceanic crust which was being eclogitized. This magma underwent an intense process of fractional differentiation from basalts to rhyolites, a calc-alkaline series occurring in association with a collateral alkaline series (Fig. 3). Due to the differentiation process basic volcanics form on diagrams a field of melanocrate rocks and the acid ones a field of leucocrate rocks (Figs. 4, 5) in which fall the calc-alkaline rocks and their alkaline correspondents (Figs. 2, 3), pointing out their consanguinity.

The differentiation processes and the resulting rocks are similar with those from the Drocea Mts, in the west of the Mureş Zone.

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[2] At present the Trascău Mts area is separated from the rest of the Mureş Zone by the Valea Verde-Inuri major fault (Savu, 1983), along which the area shifted south-eastwards concomitantly with a counter-clockwise rotation. This shifting is reflected also by the paleomagnetic analyses effectuated by Pătraşcu (1975) on ophiolites from the western block, which had been affected by an opposite movement. Later on the eastern part of the area sunk under the Transylvania Basin, so that Trascău Mts represent the only unit preserved in the eastern part of the Mureş Zone.

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References

Bleahu M., Dimian M. (1967) Studii stratigrafice şi tectonice în regiunea Feneş-Ighiel-Intregalde (Munţii Metaliferi). *D. S. Inst. Geol.*, LIII/1, p. 281–304, Bucureşti.

– , Lupu M., Patrulius D., Bordea S., Ştefan A., Panin S. (1981) The structure of the Apuseni Mountains. *Carp.-Balk. Geol. Assoc. 12th Congr. Bucharest, 1981, Guide B 3*, 106 p.

– , Soroiu M., Catilina R. (1984) On the Cretaceous tectonomagmatic evolution of the Apuseni Mountains as revealed by K-Ar dating. *Rev. Roum. Phys.*, 29, p. 129–130, Bucureşti.

Carter S. R., Evensen N. M., Hamilton P. J., O'Nions R. K. (1979) Basalt magma sources during the opening of the North Atlantic. *Nature*, 281, p. 28–30, London.

Cioflica G., Nicolae I. (1981) The origin, evolution and tectonic setting of the Alpine ophiolites from the South Apuseni Mountains (Romania). *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 25, p. 19–29, Bucureşti.

Ewart A. (1979) A review of the mineralogy and chemistry of Tertiary – recent dacitic, latitic, rhyolitic, and related salic volcanic rocks. In: F. Barker (Edit.) Trondhjemites, dacites, and related rocks. p. 13–122, Elsevier S.P.C., Amsterdam.

Gandrabura E. I. (1981) Studiul mineralogic, petrografic şi geochimic al eruptivului mezozoic din Munţii Trascău. *An. Inst. Geol. Geofiz.*, LVIII, p. 5–121, Bucureşti.

Gast P. W. (1968) Trace element fractionation and origin of tholeiitic and alkaline magma types. *Geochim. Cosmochim. Acta*, 32, p. 1057–1086, Oxford.

Ghiţulescu T. P., Socolescu M. (1941) Étude géologique et minière des Monts Métallifères. *An. Inst. Geol. Rom*, XXI, p. 181–463, Bucureşti.

Giuşcă D., Cioflica G., Savu H. (1963) Vulcanismul mezozoic din masivul Drocea (Munţii Apuseni). *Asoc. Geol. Carp.-Balc., Congr. V*, 1981, II, p. 31–44, Bucureşti.

Hedge C. E., Noble D. C. (1971) Upper Cenozoic basalts with high $^{87}Sr/^{86}Sr$ and Rb/Sr ratios, southern Great Basin, western United States. *Bull. Geol. Soc. Am.*, 83, p. 3503–3510, Boulder.

Hole M. J., Saunders A. D., Marriner G. F., Tarney J. (1984) Subduction of pelagic sediments: implications for the origin of Ce–anomalous basalts from the Mariana Island. *J. Geol. Soc. London*, 141, p. 453–472.

Ianovici V., Giuşcă D., Ghiţulescu T. P., Borcoş M., Lupu M., Bleahu M., Savu H. (1969) Evoluţia geologică a Munţilor Metaliferi. 743 p., Bucureşti.

Le Bas M. J., Maitre R. W., Streckeisen A., Zanettin B. (1986) A chemical classification of volcanic rocks based on the total alkali-silica diagram. *Jour. Petrology*, 27, p. 745–750, Oxford.

Lloyd F. E., Bailey D. K. (1975) Light element metasomatism of the continental mantle. The evidence and the consequence. *Phys. Chem. Earth*, 9, p. 389–416.

Lupu M. (1983) The Mesozoic history of the South Apuseni Mountains. *Congr. 12th Asoc. Geol. Carp.-Balk. An. Inst. Geol. Geofiz.*, LX, p. 115–124, Bucureşti.

Marriner G. T., Millward D. (1984) The petrology and geochemistry of Cretaceous to Recent volcanism in Colombia: the magmatic history of an accretionary plate margin. *J. Geol. Soc. London*, 141, p. 473–486, Oxford.

Meijer A., Reagan M. (1981) Petrology and Geochemistry of the island of Sarigan in the Mariana arc; calc-alkaline volcanism in an oceanic setting. *Contrib. Mineral Petrol.*, 77, p. 337–354, Berlin.

Mitchell A. H., Bell J. D. (1973) Island arc evolution and related mineral deposits. *J. Geol.*, 81, 4, p. 381–405, Chicago.

Mysen B. O. (1979) Trace element partitition between garnet peridotite minerals and water-rich vapor: experimental data from 5 to 30 kbar. *Am. Mineral.*, 64, p. 274–287, Washington.

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

– , Bratosin I. (1980) Petrochemical investigations of the Mesozoic spilite and keratophyre rocks from the Trascău and North-Eastern Metaliferi Mountains (Apuseni Mountains). *Rev. Roum. Géol. Géophys., Géogr., Géologie*, 24, p. 99–114, Bucureşti.

– , Cuna S., Soroiu M. (1987) Preliminary K-Ar investigations of ophiolites from the South Apuseni Mountains (Romania). *Stud. cerc. geol. geofiz., geogr., Geofizică*, 25, p. 43–49, Bucureşti.

Nincovich D., Hayes J. D. (1971) Tectonic setting of Mediterranean volcanoes. Acta Int. Sci. Congr. Volcano thera (Greece), 1, p. 111.

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

– , Norry M. J. (1979) Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks. *Contrib. Mineral. Petrol.*, 69, p. 33–47, Berlin.

– (1980) Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites. In A. Panayiotou (edit.): Ophiolites, Proc. Intern. Ophiol. Symp. Nicosia, Cyprus, 1979, p. 261–272.

Roden M. F., Murthy V. R. (1985) Mantle metasomatism. *Ann. Rev. Earth Planet. Sci.*, 13, p. 269–296, Palo Alto.

Institutul Geologic al României

Russo-Săndulescu D., Berza T., Bratosin I., Ianc R. (1976) Contribuţii la studiul petrologic al unor magmatite alpine din nordul Munţilor Trascău. *D. S. Inst. Geol.*, LXII/1, p. 165–194, Bucureşti.

Savu H. (1962a) Cercetări geologice şi petrografice în regiunea Troaş-Pîrneşti din Masivul Drocea. *D. S. Com. Geol.*, XLVI, p. 137–158, Bucureşti.

– (1962b) Chimismul vulcanitelor jurasice superioare-cretacic-inferioare din Munţii Drocea. *D. S. Inst. Geol.*, XLVII, p. 199–220, Bucureşti.

– (1983) Geotectonic and magmatic evolution of the Mureş Zone (Apuseni Mountains). *Carp.-Balk. Geol. Assoc. 12th Congr. Bucharest, 1981, An. Inst. Geol. Geofiz.*, LXI, p. 253–262, Bucureşti.

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

– , Udrescu C., Neacşu V. (1986b) Bimodal volcanism in the north-western island arc of the Mureş Zone. *D. S. Inst. Geol. Geofiz.*, 70–71/1 (1983; 1984), p. 153–170, Bucureşti.

– , Udrescu C., Lemne M., Neacşu V. (1988a) Petrology, geochemistry and tectonics of the ophiolites from the ocean paleoridge of the Mureş Zone between Pîrneşti, Troaş and Valea Zeldiş (Drocea Mountains). *D. S. Inst. Geol. Geofiz.*, 72–73/5 (1985; 1986), p. 237–257, Bucureşti.

– , Udrescu C., Lemne M., Romanescu O., Neacşu V. (1988b) Petrology, geochemistry and tectonics of ophiolites and Late Kimmerian island arc volcanics from the Glodghileşti-Săliştioara tectonic rise (Mureş Zone). *D. S. Inst. Geol. Geofiz.*, 72–73/5 (1985; 1986), p. 259–281, Bucureşti.

– , Udrescu C., Neacşu V. (in this volume) On the presence of ocean floor rocks (Liassic ophiolites) in the Trascău Mountains (Mureş Zone); their petrology and geochemistry.

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

Uyeda S. (1981) Some thoughts on geodynamics of Asia. *Geol. Surv. Japan. Rep.*, 261, p. 1–6, Higashi.

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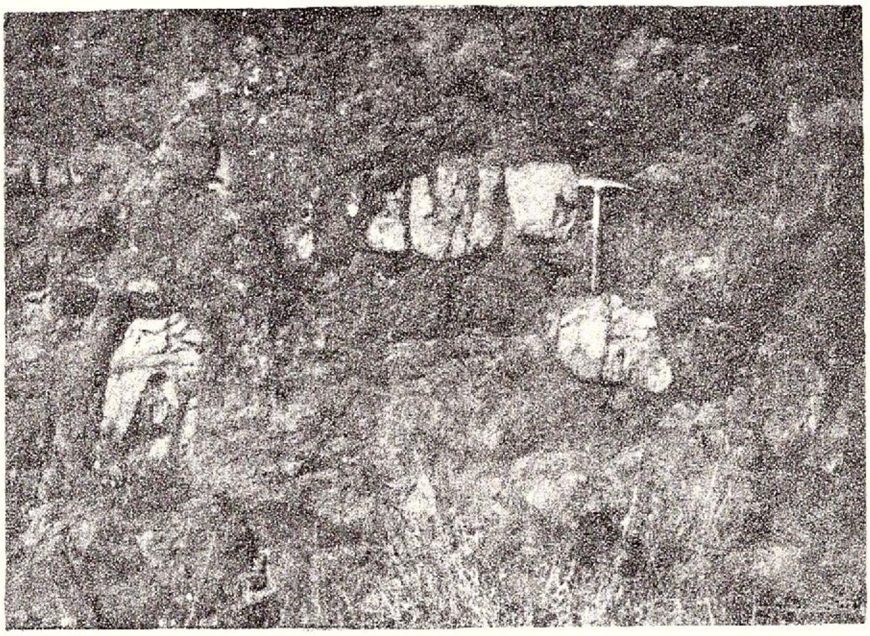

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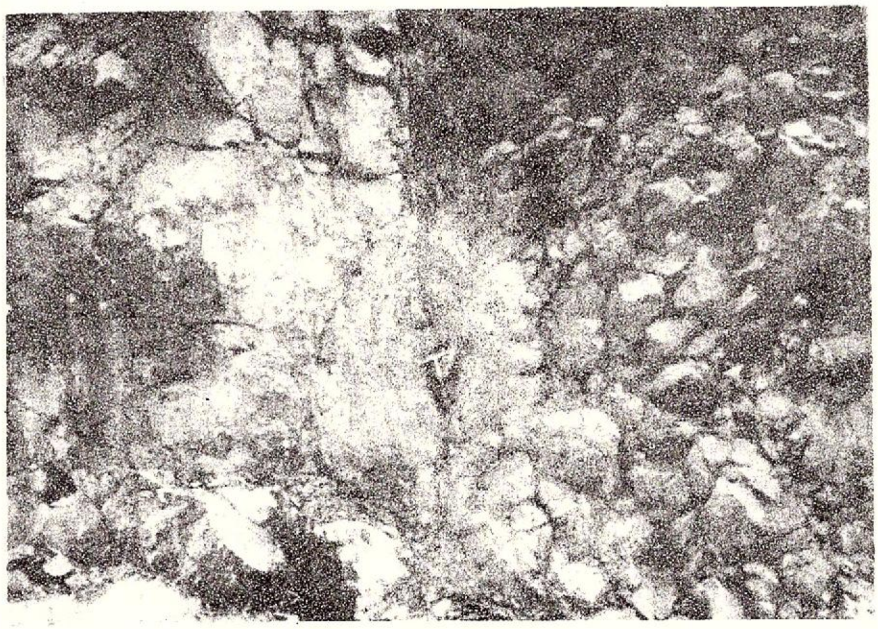

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Plate I

Fig. 1 – Andesite tuffs including pyroxene andesite blocks. Arieş Valley.

Fig. 2 – Tectonic contact along an almost vertical fault between flows of amygdaloidal porhyritic basalts, in pillow lava-like separations (to the right) and Barremian-Aptian flysch (to the left). Valea Mare. Ţelna.

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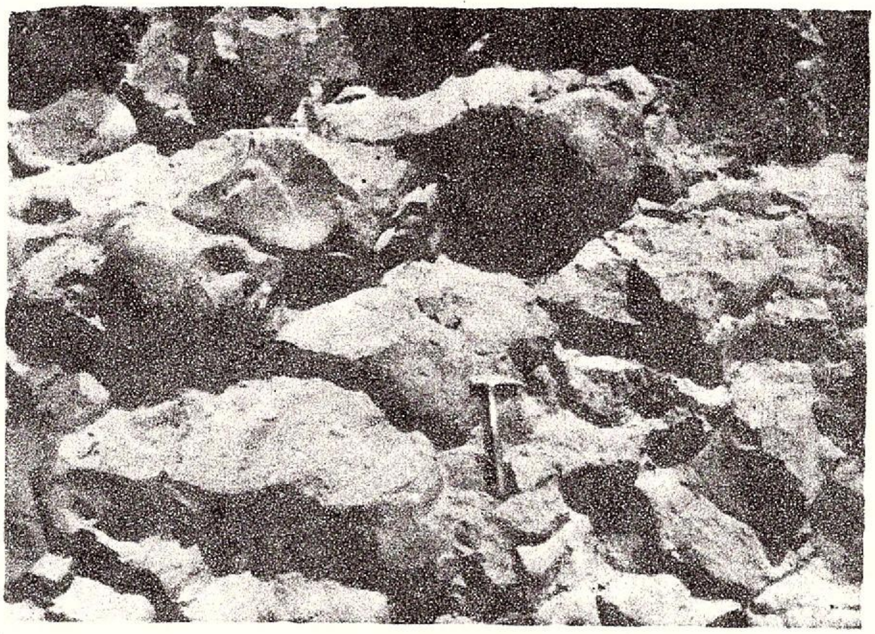

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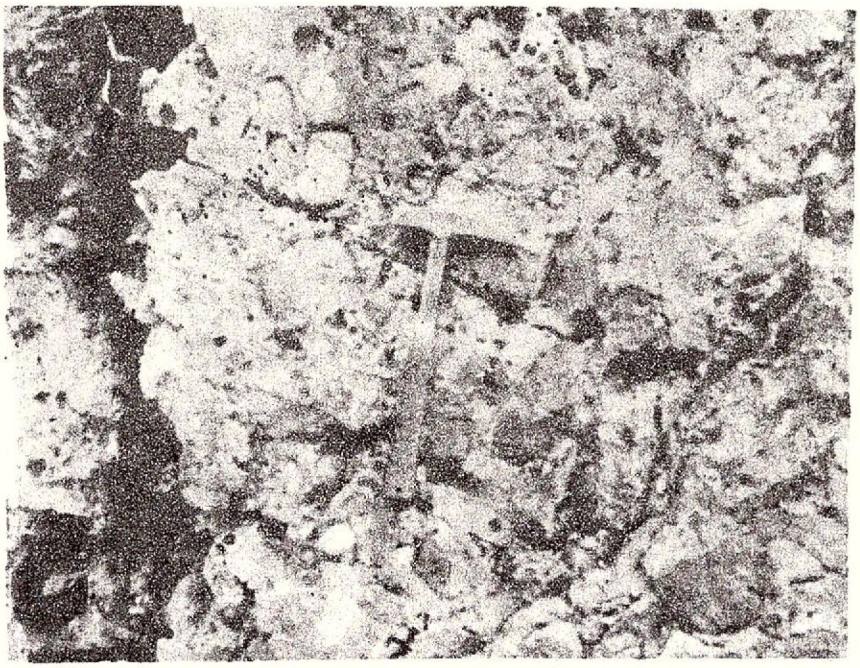

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Plate II

Fig. 1 – Flows of amygdaloidal porphyritic basalts in pillow lava-like separation. Valea Mare. Ţelna.

Fig. 2 – Coarse-grained polygenous conglomerate in the detrital facies of the Feneş Beds, whose pebbles are cemented with gritty material and consist of Late Kimmerian island arc eruptive rocks and sedimentary deposits associated with them in the northeasternmost part of the Ampoi Cordillera. Cricău Valley, right side.

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Rom. J. Petrology, 1992, 75, p. 75–84

PETROLOGY, GEOCHEMISTRY, AGE AND ORIGIN OF THE LATE KIMMERIAN LEUCOCRATE DYKE SWARM BETWEEN VĂRĂDIA AND TROAŞ (MUREŞ ZONE)

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Key words: Dyke swarms. Neocomian. Magmas. Granitic composition. Magmatic differentiation. Absolute age. K/Ar. Major elements. Minor elements. Apuseni Mountains – Drocea Mountains.

Résumé: *Pétrologie, géochimie, âge et origine de la série des dykes leucocrates néokimmériens situés entre Vărădia et Troaş (zone de Mureş).* Les roches leucocrates néokimmériennes (néocomiennes) situées entre Vărădia et Troaş forment une série de dykes associée à l'arc insulaire de sud de la zone de Mureş. Elles percent le complexe de sheeted dykes liasiques, mais ne résultent pas du magma tholéiitique et elles n'ont aucun liaison avec les plagio-granites albitiques associés aux roches ophiolitiques. L'âge radiogène (K-Ar) des roches leucocrates est situé entre 116 et 120 Ma. Ces corps hipabyssaux comportent microgranites porphyriques, aplites granitiques et micrognanosyénites porphyriques. Les roches correspondent aux magmas leucogranitiques de type "high-K" et syénogranitiques, résultés d'un magma originaire calco-alcalin, formé dans le manteau surjacent du plan de Benioff de la partie sud de la zone de Mureş. La série de dykes s'est formée pendant la collision de la plaque transylvaine avec l'arc insulaire du sud de la zone de Mureş, situé sur le bord méridional de la plaque océanique de Mureş.

Introduction

The Liassic sheeted dyke complex in the ophiolitic megaslab of the Mureş Zone is penetrated in its south-eastern part by a Late Kimmerian dyke swarm and other island arc leucocrate rocks situated between Vărădia and Troaş (Drocea Mts). Szentpétery (1928) and all the other workers who studied this area (Socolescu, 1944; Papiu, 1953) considered that the mentioned rocks belong to the Laramian eruptions. Savu (1962a, 1962b) pointed out the striking similarity between the rocks of the hypabyssal eruptive bodies and the Late Kimmerian volcanics hosted in the J_3-Cr_1 flysch on the north-western border of the Mureş Zone and established their appurtenance to the Late Kimmerian eruptive province in the Drocea Mts. This conclusion was also maintained by him later on (Savu et al., 1987a, 1987b) and the rocks were accordingly represented on the geological maps scale 1:50000 Săvîrşin and Roşia Nouă (Savu et al., 1979a, 1979b) on the basis of which the annexed map was elaborated (Pl. I). Although petrographic and chemical data on the leucocrate rocks were presented by Savu in his papers in 1962, there is no complete study on these rocks and consequently we consider the present paper useful.

Distribution and Petrography of the Rocks

The dyke swarm, including veins and small irregular bodies of hypabyssal porphyritic rocks and granitic aplites, is developed between Vărădia and Troaş (Pl. I), in the western part of the Mureş Zone. It is to be noted that such acid alkaline bodies occur more isolatedly in the western part of the Mureş Zone and at Roşia Nouă, Pietriş and Ilteu (Savu, 1962b), where they are spread around the Late Kimmerian granitoid massif of Săvîrşin. The trending of these Late Kimmerian bodies is NE-SW, almost parallel to that of the Liassic sheeted dykes only with a very small acute angle between them. The thickness of the dykes varies

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from 0.5 to 10 m, but the irregular bodies – e.g. at Pîrneşti (Pl. I) – present thicknesses greater than 750 m.

As shown on the diagram in Figure 1, the leucocrate rocks in the dyke swarm, veins and small intrusive bodies, plot in the 'alkali granites' field. The chemical composition differentiates the rocks from the alkali granites of the continental intraplate type because they do not contain sodium amphiboles and pyroxenes. They must be regarded as biotite-bearing leucogranites resulting from a calc-alkaline magma, as shown on the diagram in Figure 2, on which most of the rocks plot in the granite field; two rocks correspond to quartz-keratophyres and two are close to the quartz-monzonites field. Therefore, according to their composition, the rocks can be classified as follows: (1) porphyritic microgranites, (2) granitic aplites and (3) porphyritic microgranosyenites.

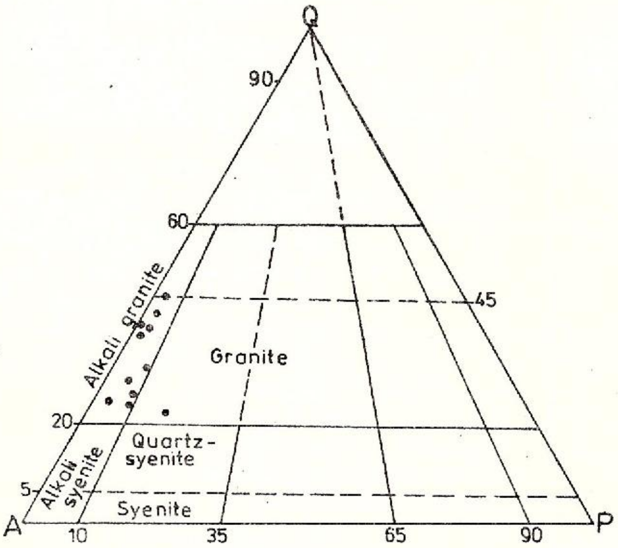

Fig. 1 – QAP diagram (Streckeisen, 1967). The values plotted on the diagram have been obtained computing the data in Table 4.

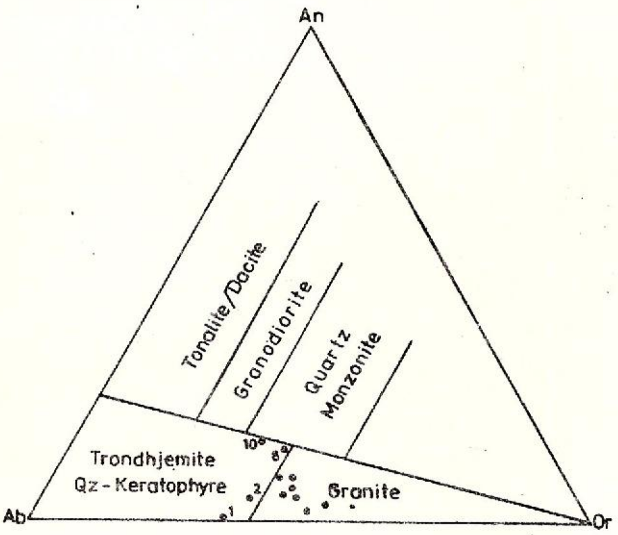

Fig. 2 – AnAbOr diagram (O'Connor, 1965).

1. Porphyritic rocks, among which porphyritic microgranites predominate, which locally pass into quartz-keratophyres, are frequently found north-east of the Pîrneşti Valley (Pl. I). They consist of a fine, holocrystalline groundmass made up of isometric crystals of potash feldspar and plagioclase, dim and cemented with interstitial quartz. Phenocrysts of quartz, potash feldspar and plagioclase occur in the groundmass (Pl. II, Fig. 1). The first are small-sized, locally bipyramidal, surrounded by a reaction rim represented by dim feldspar crystals

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perpendicular to the surface of the quartz phenocrysts (Pl. II, Fig. 2). Potash feldspar phenocrysts are idiomorphic or resorbed and include small plagioclase crystals. They display a perthitic structure, but usually they are dim. Plagioclase phenocrysts (An_{10}) are small-sized and present polysynthetic twins after albite and albite-Karlsbad laws. They are also dim. As a melanocrate mineral, biotite occurs as altered lamellas.

2. Granitic aplites occur in the Troaş Valley, nearby the confluence with the Runcu Brook and in the Bruma Valley, south of Pîrneşti. They consist of a groundmass formed of panallotriomorphic crystals of potash feldspar, albitic plagioclase and quartz which cement them. Microphenocrysts of idiomorphic potash feldspar and dim plagioclase, altered like the feldspars in the groundmass, are rarely found.

3. Porphyritic microgranosyenites correspond to the orthophyres (paleotrachytes) in the bimodal volcanics (Savu et al., 1986b) hosted in the flysch J_3-Cr_1 in the Drocea Mts and are very close to those intruding the Săvîrşin intrusive body, described by Szentpétery (1928) and Savu et al. (1967). These rocks are more often found in the south-western part (Vărădia-Hălăliş) of the occurrence area of the acid and alkaline leucocrate bodies in the region. They represent porphyritic rocks consisting of a microcrystalline groundmass, or more granular in some rocks (porphyritic syenogranites), within which potash feldspar and plagioclase phenocrysts, as well as biotite microphenocrysts, are floating. The potash feldspar phenocrysts (ca. 1–2 cm long) are pink and display twins after Karlsbad law. They contain inclusions of idiomorphic crystals of sphene and twinned albite. These phenocrysts are usually dim being packed with a very fine argillitic mineral. The microperthitic structure is visible in fresher crystals. On margins they are replaced by polysynthetically twinned albite. Plagioclase phenocrysts (An_{18}), also dim and locally impregnated with small crystals of epidote, present polysynthetic twins after albite and albite-Karlsbad law, rarely pericline. In places they are grouped and form a glomeroporphyritic texture. Pseudomorphoses of epidote (pistacite) and chlorite after biotite microphenocrysts (Pl. II, Fig. 3), as well as idiomorphic or broken microphenocrysts of sphene and apatite (Pl. II, Fig. 3), are formed, too. It is to be noted that the epidote and chlorite pseudomorphoses after biotite also contain apatite idiomorphic crystals representing primary inclusions of mica subsequently replaced. In these rocks small magnetite crystals are rarely found.

Radiometric Dating of the Rocks

K/Ar method was used for the rock dating (Tab. 1). The radiogene argon dosage was carried out by isotopic dilution using ^{38}Ar as a tracer. Potassium was determined by the flame photometric method. $^{40}Ar/^{38}Ar$ and $^{38}Ar/^{36}Ar$ ratios were measured with the mass spectrometer AEI-MS 20, operated statically. The apparent

Table 1

Radiometric age (K-Ar)* analyses

No	Sample no	Rock type	K%	$\frac{100\ ^{40}Ar\ rad}{^{40}Ar\ tot}$	^{40}Ar rad x 10^{-10} mol/g	V.a. ± 1 Ma
1	80	Porphyritic microgranite Pîrneşti Valley	4.85	71.71	7.0588	82.0 ± 3.3
2	83	Porphyritic microgranite Pîrneşti Valley	4.77	77.77	8.9269	104.8 ± 6.0
3	81	Porphyritic microgranosyenite Brumii V.	3.55	64.63	6.1048	96.5 ± 4.0
4	90	Porphyritic microgranite Godineşti	2.17	70.38	3.4332	89.0 ± 3.6
5	84	Porphyritic microgranite Pîrneşti Valley	3.29	76.60	6.2879	107.0 ± 3.3

* Analyses were carried out on total rock

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ages were calculated according to the close system hypothesis, using Steiger and Jäger's (1977) constants. The analytical error of determination was calculated after Cox and Dalrymple's (1967) formula. The apparent ages are referred, with the analytical error of 1σ, to the probability level of 68 per cent.

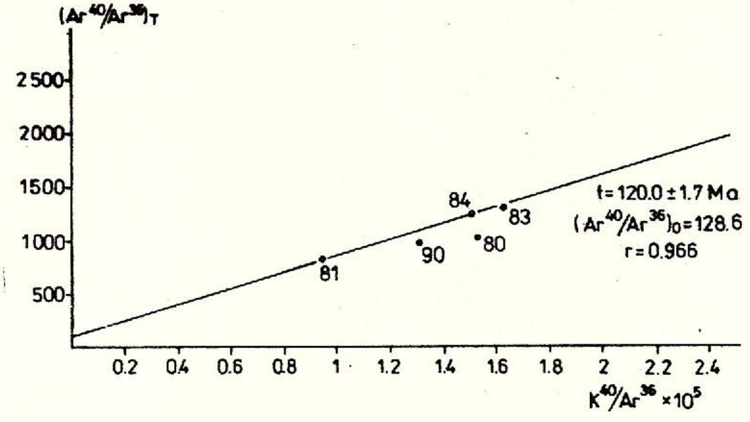

Fig. 3 – $^{40}Ar/^{36}Ar$ versus $^{40}K/^{36}Ar$ for the dyke swarm and a sample from Godineşti.

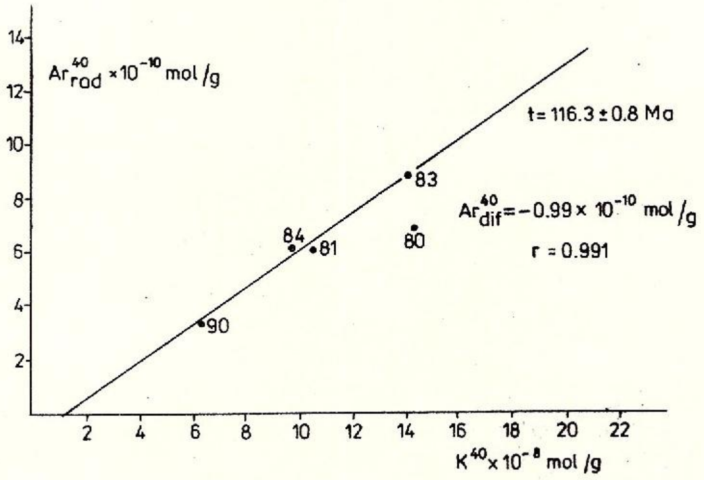

Fig. 4 – ^{40}Ar-rad versus ^{40}K for the dyke swarm and a sample from Godineşti.

Samples nos. 81, 83, 84 and 80 (Tab. 1), with apparent ages ranging between 89.0 and 107.0 Ma, are grouped on an isochrone of 120.0±1.7 Ma within the representation system $^{40}Ar/^{36}Ar$ – $^{40}K/^{36}Ar$. The intersection with axis Y is smaller than the atmospheric argon ratio. The isotopic model (Fig. 3) presumes that the same concentration of ^{40}Ar was missing in the study samples and the nonradiogene component consisting of the atmospheric argon, and the argon lost by diffusion show an isotopic ratio of 128.9 ($^{40}Ar_a/^{36}Ar - {}^{40}Ar_{dif}/^{36}Ar = 128.9$). The radiogene component constitutes the main term of the age calculation equation. Thus, a similar age value (t = 116± 0.8 Ma) was also calculated with the isochrone ^{40}Ar versus ^{40}K (Fig. 4). The intersection of the straight line with axis Y is negative and represents the amount of ^{40}Ar lost by diffusion, established for the samples under consideration and equal to $0.99\text{x}10^{-10}$ mol/g. The isochrone represents the age of the system closing and is higher than the individual apparent ages. The age values indicated by isochrone of both representation systems are concordant with the determination errors limits (120; 116 Ma); thus they can be considered to have geological validity.

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Radiometric dating indicates the emplacement of the leucocrate rocks towards the end of the Neocomian, in a stage close to the closing phase of the island arc magmatism when this magmatism manifested by acid intrusions (Mitchel, Bell, 1973) e.g. at Săvîrşin (128 Ma). Their emplacement was preceded by the intrusion of the more important bodies of Săvîrşin, Cerbia, Pietroasa and Dealul Mare-Vălişoara (Savu et al., 1986a).

Geochemistry and Tectonic Setting

The chemical composition of the porphyritic microgranites and granitic aplites (Tab. 2) indicates a group of leucocrate rocks, mostly hyperacid, within which SiO_2 varies between 70.87 and 78.00 per cent. In this respect they are similar with some facies of the Săvîrşin granite (Savu et al., 1967) and with the acid leucocrate rocks of the bimodal volcanics (Savu et al., 1986b) hosted in the J_3-Cr_1 flysch on the north-western margin of the Mureş Zone. It is to be noted the high-K_2O content in the whole series of acid rocks (3.16–5.68 per cent).

Table 2

Chemical composition of acid and alkaline rocks

No	1	2	3	4	5	6	7	8	9	10	11
Sample no	134	123	135	122	117	130	133	125	81	56	71
SiO_2	78.00	76.33	76.30	75.57	72.88	71.83	71.53	71.15	70.87	66.63	65.90
Al_2O_3	12.00	12.79	12.69	12.64	12.74	14.10	16.10	13.93	13.99	16.10	15.35
Fe_2O_3	0.07	1.83	0.14	1.65	0.06	1.30	1.10	1.13	0.42	1.40	1.91
FeO	0.21	0.21	0.43	0.36	0.86	0.50	0.29	1.07	1.58	1.58	1.58
MnO	0.008	0.006	0.019	0.018	0.02	0.050	0.043	0.042	0.06	0.08	0.07
MgO	0.06	0.26	0.18	0.11	0.21	0.58	0.40	0.68	0.68	1.20	1.08
CaO	0.33	0.49	0.27	0.55	1.72	0.68	0.53	2.10	1.41	2.59	2.56
Na_2O	2.78	3.68	3.22	3.28	3.89	3.85	4.11	3.68	3.99	3.96	3.88
K_20	4.57	3.47	4.58	4.38	3.16	4.97	5.68	4.30	4.68	3.85	4.42
TiO_2	0.14	0.16	0.26	0.20	0.26	0.28	0.32	0.40	0.29	0.42	0.45
P_2O_5	0.03	0.03	0.04	0.02	0.04	0.07	0.04	0.10	0.08	0.22	0.30
CO_2						2.38					0.55
S	0.15	0.13	0.18	0.14	0.16	0.18	0.12	0.16	0.10	0.17	0.12
Fe(S)	0.13	0.11	0.15	0.12	0.14	0.15	0.10	0.14	0.08	0.15	0.10
H_2O^+	0.59	0.40	1.28	0.73	1.02	0.98	0.51	0.69	1.19	1.22	1.26
Total	99.58	99.89	99.74	99.76	99.54	99.52	100.87	99.47	99.47	99.57	99.53
Ni	2.5	2.0	2.5	2.5	2.5	2.5	<2.0	<2	3.5	2.5	2.0
Co	2.0	<2.0	<2.0	<2.0	3.0	3.5	2.0	4.0	3.0	5.0	4.0
Cr	2.0	2.5	1.5	5.0	3.5	2.0	1.5	3.0	4.0	3.5	3.0
V	6.5	7.5	5.5	6.5	12.0	32.0	13.0	28.0	32.0	62.0	60.0
Sc	<2.0	2.0	<2.0	2.0	2.0	3.0	2.0	3.0	6.0	6.0	4.5
Y	12.0	10.0	11.0	13.0	12.5	17.0	10.0	17.0	18.0	21.0	21.0
Yb	2.1	1.5	2.2	2.1	1.6	2.0	1.7	1.8	1.7	1.9	2.0
La	50	50	40	65	56	86	30	80	85	110	100
Zr	120	105	130	140	130	200	105	220	160	210	170
Nb	22	19	21	20	15	20	17	15	<10	<10	<10
Ba	620	600	870	850	850	1500	1650	1900	1150	1000	1400
Sr	60	95	60	130	90	220	140	420	150	360	460
Pb	30	30	10	7	35	7	7	7	10	6	5.5
Cu	44	32	36	42	24	62	34	36	6	4	6.5
Ga	12	12	12.5	12	12	16	13.5	13	10	12	11
Sn	3.5	4	4	4	3	5	3.5	3.5	2	<2	<2

1–9 – porphyritic microgranites; 10, 11 – porphyritic microgranosyenites; 1, 3, 6, 7, 9 – Brumii Valley, Pîrneşti; 2, 4, 8 – Pîrneşti Valley; 5 – Troaş Valley; 10 – Hălăliş Valley; 11 – Ţiganilor Valey (P. lui Bujor) Vărădia.

According to the SiO_2 and K_2O content, the porphyritic microgranites and the granitic aplites fall on Ewart's diagram (1979, Fig. 1) in the 'high-K rhyolites' field. The same conclusion results from Table 3, which presents

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the Niggli parameters (Burri, 1959) of the leucocrate rocks among which the high values of the *si*, *al*, *alk* and *k* parameters are remarkable. These rocks correspond to the leucogranitic magma type.

The normative composition of these rocks (Tab. 4) shows the presence of the normative quartz, which varies between 24 and 43 per cent, and of the orthoclase, which varies between 15 and 33 per cent. Similarly, normative albite varies between 31.27 and 34.96 per cent. The Ab/Or ratio varies between 0.87 and 1.76 (mean value 1.22) for the whole series of leucocrate rocks. The presence of normative enstatite marks the existence of altered biotite in these rocks. Normative sphene is observed in porphyritic microgranites.

Table 3

Niggli parameters of the island arc acid and alkaline rocks

No	Sample	si	al	fm	c	alk	k	mg	ti	o	qz
1	134	577.9	52.4	3.4	2.6	41.6	0.52	0.19	0.78	0.41	+312
2	123	479.8	47.4	13.0	3.3	36.3	0.38	0.19	0.76	0.72	+234
3	135	518.4	50.8	6.2	2.0	41.1	0.48	0.30	1.33	0.29	+254
4	122	476.5	47.0	11.7	3.7	37.7	0.47	0.09	0.95	0.74	+226
5	117	444.9	45.8	7.6	11.2	35.3	0.35	0.25	1.19	0.16	+204
6	130	390.4	45.2	13.4	4.0	37.5	0.46	0.35	1.14	0.46	+140
7	133	367.4	48.7	9.3	2.9	31.1	0.48	0.33	1.24	0.52	+111
8	125	362.1	41.5	15.0	11.4	32.1	0.43	0.34	1.53	0.34	+134
9	81	365.3	42.5	14.4	7.8	35.3	0.44	0.36	1.12	0.14	+124
10	56	290.4	41.4	19.1	12.1	27.4	0.39	0.41	1.38	0.28	+ 81
11	71	287.8	39.5	19.8	12.0	28.7	0.43	0.36	1.48	0.34	+ 73

Magma type: 1–9, leucogranitic; 10, granosyenitic to granitic; 11, granosyenitic.

Table 4

Normative composition of the rocks

No	q	or	ab	an	cor	en	fs	ilm	sfen	he	mt	ap	py
1	43.98	27.18	23.67	1.13	2.08	0.15		0.11	0.25	0.25		0.08	0.28
2	40.54	20.59	31.27	1.94	2.28	0.65		0.16	0.19	2.00		0.08	0.24
3	39.86	27.43	27.61	1.05	2.08	0.45		0.50		0.32	0.05	0.10	0.34
4	38.75	26.08	27.97	2.60	1.57	0.28		0.38		1.74	0.14	0.05	0.26
5[1]	38.38	19.08	33.64		2.98	0.53	0.66	0.50				0.10	0.31
6	28.62	29.67	32.92	2.87	1.36	1.46		0.54		1.31	0.32	0.18	0.34
7	24.49	33.75	34.96	1.98	2.48	1.00		0.42	0.25	1.25		0.10	0.23
8[2]	28.56	25.60	31.37	8.58		1.29		0.77		0.06	1.86	0.26	0.30
9	25.43	28.00	34.18	6.47	0.02	1.71	1.91	0.56			0.78	0.21	0.19
10	22.22	23.04	33.94	11.32	1.34	3.03	0.69	0.81			2.36	0.57	0.32
11[3]	21.93	26.46	33.26	7.04	1.66	2.73	0.36	0.87			3.01	0.77	0.22

1– cc=3.03; 2– diop=0.89; 3– cc=1.27.

For the number of samples see Table 2.

The last two rocks in the mentioned tables display an alkaline character; they represent porphyritic microgranosyenites. These rocks contain smaller amounts of SiO_2 than the porphyritic microgranites (60-67 per cent), K_2O between 3.85 and 4.42 per cent and Na_2O between 3.88 and 3.96 per cent. TiO_2 is also present in these rocks, representing 0.42-0.45 per cent. On Ewart's diagram (1979, Fig. 1) they fall nearby trachyte field. Consequently, according to the magmatic parameters (Burri, 1959) these rocks belong to the granosyenitic microgranites. In this respect the rocks are similar to the granosyenites intruding the Temeşeşti granodiorite and diorite body, as well as to the orthophyres (Savu, 1962a; 1962b), keratophyres respectively, of the J_3-Cr_1 flysch in the north-west of the Mureş Zone (Savu et al., 1986). These rocks also differ from the porphyritic microgranites by the higher values of the parameters *fm* and *c* (Tab. 3).

Table 4 presents the normative composition of the alkaline rocks. It indicates smaller amounts of normative quartz (21.93-22.22 per cent) as compared with the porphyritic microgranites. High amounts of orthoclase

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and normative albite are also presented. The melanocrate minerals participate in greater amounts than in the porphyritic granites, as normative enstatite, pointing out the presence of biotite. There are present also the accessory minerals, e.g. magnetite, sphene and apatite.

The trace element distribution in acid and alkaline rocks (Tab. 2) characterizes a granitoid series in which the siderophile elements occur in very small amounts. Other elements such as Y, Yb, La, Zr, Ba and Sr present high contents. La/Yb ratio increases almost twice along the series – from 33 to 50, and Ba/Sr ratio decreases almost three times – from 10 to 3. However, the highest contents of La, Zr, Ba and Sr are found in the alkaline rocks. The values of these contents are similar to those in the Săvîrşin granite (Savu et al., 1967), whose age has been recently observed (Savu et al., 1986a) to be close to that of the dykes and veins of Late Kimmerian acid and alkaline rocks in the Vărădia-Troaş area.

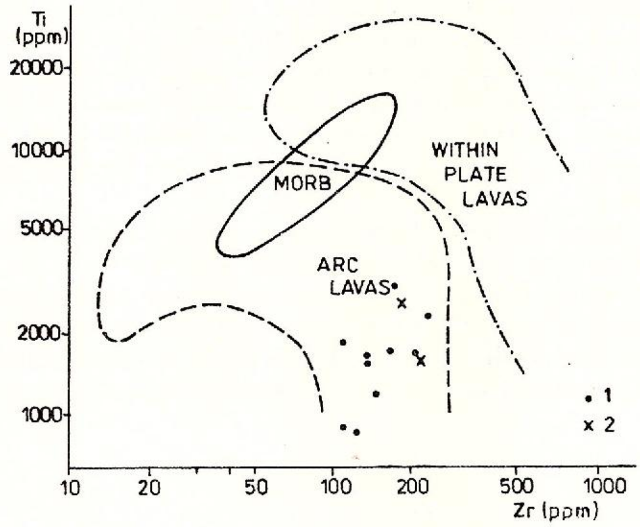

Fig. 5 – Ti–Zr diagram (Pearce, 1980).
1, acid rocks; 2, alkaline rocks.

According to Ti/Zr ratio, all the leucocrate rocks fall on the diagram in Figure 5 in the island arc magmatic field. The same result is obtained if the rocks are plotted on the Nb-Y diagram in Figure 6, where they occur in the field of the volcanic arc granites (VAG) and syncollision granites (syn-COLG), according to Pearce et al. (1984). Unfortunately, there are no other trace elements in order to establish their appurtenance to the last category of granites (syn-COLG) which includes the study leucocrate rocks considering their petrochemical features and the geotectonic environment of their formation (Savu, 1983; Savu et al., 1986a).

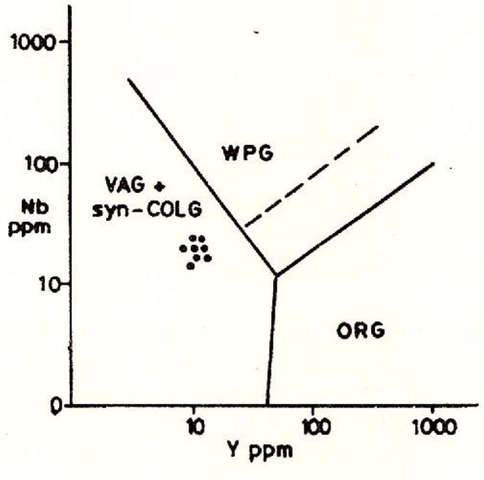

Fig. 6 – Nb–Y diagram (Pearce et al., 1984).

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Origin of the Dyke Swarm

Although the dyke swarm and the small bodies of Late Kimmerian leucocrate rocks in the Vărădia-Troaş region penetrate the Liassic sheeted dyke complex (180 Ma, Herz et al., 1974) of the Mureş Zone, they do not represent the tholeiitic magma differentiates which generated the sheeted dykes and there is no connection between them and the albitic plagiogranites with trondhjemitic character (Savu et al., 1985; Savu, Stoian, 1988) found in the sheeted dykes in the west of the study region. The difference between the two series of acid rocks results clearly from the different values of the radiometric dating, geochemical peculiarities and tectonic setting, the Liassic plagiogranites representing ocean floor rocks and the rocks presented in this paper being island arc intrusions.

The Late Kimmerian leucocrate rocks in the Vărădia-Troaş region form a 'high-K' type eruptive series. They are similar to the leucogranites described by Lameyre et al. (1974). On the QAbOr diagram in Figure 7 they plot in a field situated around the ternary minimum of the ideal granites (Tuttle, Bowen, 1958). It emphasizes their origin in a residual granitic magma, differentiated from a calc-alkaline parental magma. On the mentioned diagram, in the leucocrate rocks field occurs also the average value of the Săvîrşin granitic rocks which represent an intrusion related to the evolution of the southern island arc of the Mureş Zone (Savu et al., 1986a). It clearly indicates that the porphyritic rocks differentiated from the same parental magma which yielded the Săvîrşin granite magma.

Taking into account that the subduction in the Mureş Zone is of ocean plate under ocean plate type, it is out of question that the Late Kimmerian parental magma should have formed by the melting of the lower part of a sialic crust. Moreover, using the Hatherton and Dickinson methods (1969) in order to establish the formation depth of the magmas, which is based on the K_2O content of the rocks with 60 per cent SiO_2, one can observe that the magma was formed at a depth of 240 km. At this depth no Late Kimmerian parental magma could have been formed from a sialic material subducted on the southern Benioff plane of the Mureş Zone.

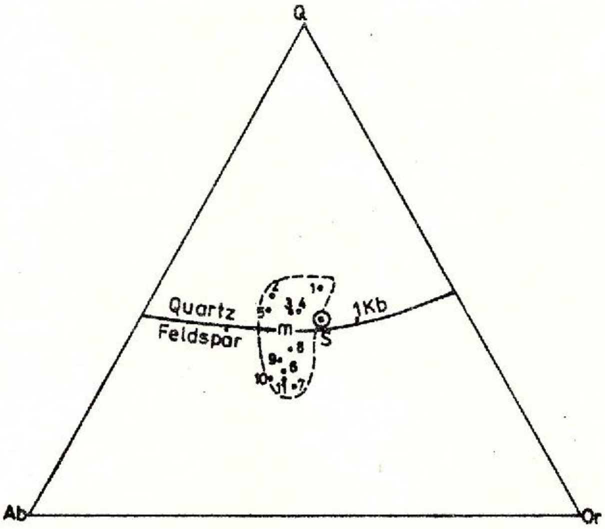

Fig. 7 – QAbOr diagram (Luth et al., 1964).
S represents the plotting of the mean of five granitoid rocks of the Săvîrşin Late Kimmerian massif (data from Savu et al., 1967).

Referring to the origin of the parental magma of the Late Kimmerian island arc volcanics in the Mureş Zone, it is admitted (Savu et al., 1989) that it had been formed by the melting of the mantle above the Benioff plane which was metasomatized with substances released from the subducted Liassic ocean crust which was to be transformed into eclogite. Considering the geotectonic conditions under which the magmatic activity developed in the Mureş Zone during the Neocomian, we infer that the parental magma which generated the Săvîrşin granite, the leucocrate dyke swarm in the Vărădia-Troaş region and the Late Kimmerian island arc rocks in other areas of the Mureş Zone originated in the metasomatized mantle, lying above the Benioff plane.

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Hamilton and Myers (1967) showed that the parental magma which yielded the granitic plutons of California originates in the upper mantle.

The parental magma of the Late Kimmerian magmatic rocks was a calc-alkaline basaltic magma differentiated according to the following scheme: basaltic (gabbroic) magma–dioritic magma–granodioritic magma–'high-K' granitic magma. Rocks derived from the magmas differentiated from the parental magma are found in the Săvîrşin Late Kimmerian granitoid massif, consisting of diorites, quartzdiorites, granodiorites and granites, as well as in the Late Kimmerian volcanics in the Mureş Zone.

Dyke swarm, veins and small leucogranitic bodies resulted from a "high-K" residual magma differentiated from the granitic magma. They were formed, according to the model elaborated by Savu et al. (1986a), in the final stage of the subduction of the Transylvanian Plate under the Mureş Ocean Plate during the collision of the first plate with the southern island arc of the Mureş Zone, situated on the southern margin of the Mureş Ocean Plate. The same conclusion is indicated by the radiometric dating of the dykes which shows that they are a bit younger than the Săvîrşin granite of 128 Ma (Savu et al., 1986a).

Conclusions

The Late Kimmerian (Neocomian) leucocrate rocks in the Vărădia-Troaş region form a dyke swarm associated with the southern island arc of the Mureş Zone.

The radiometric dating of the leucocrate rocks is of 116-120 Ma.

The hypabyssal bodies consist of porphyritic micrograntes, granitic aplites and porphyritic microgranosyenites.

The rocks correspond to 'high-K' leucogranitic and syenogranitic magmas coming from a calc-alkaline parental magma formed in the metasomatized mantle above the Benioff plane.

The dyke swarm was formed during the collision of the Transylvanian Plate with the southern island arc in the Mureş Zone.

References

Burri C. (1959) Petrochemische Berechnungsmethoden auf äquivalenter Grundlage (Methoden von Paul Niggli). Birkhäuser Verl. Basel, 334 p.

Cox A., Dalrymple G. B. (1967) Statistical analysis of geomagnetic reverse data and the precision of potassium-argon dating. *Jour. Geophys. Res.*, 72/10, p. 2603–2614, Washington.

Ewart A. (1979) A review of the mineralogy and chemistry of tertiary – recent dacitic, latitic, rhyolitic, and related salic volcanic rocks. In F. Barker (Edit.) Tronhjemites, dacites, and related salic volcanic rocks. p. 113–122, Elsevier S. P. C. Amsterdam.

Hamilton W., Myers B. (1967) The nature of batholiths, shorter contribution to general geology. *Geol. Surv. Prof. Paper*, 554-C, p. 1–30, Washington.

Hatherton T., Dickinson W. R. (1969) The relationship between andesitic volcanism and seismicity in Indonesia, Lesser Antilles and other island arcs. *Journ. Geophys. Res.*, 74, 29, p. 7776–7783, Washington.

Hertz N., Jones L. M., Savu H., Walker R. L. (1974) Strontium isotope composition of ophiolitic and related rocks, Drocea Mountains, Romania. *Bull. Volc.*, XXXVIII-4, p. 1110–1124, Napoli.

Lameyre J., Rocci G., Didier J. (1974) Granites orogéniques et granites cratoniques. In Centenaire de la Société Géologique de Belgique. Géologie des domaines cristallins, p. 183–221, Liège.

Luth W. C., Jahns R. H., Tuttle O. F. (1964) The granite system at 4 to 10 Kilobars. *Journ. Geophys. Res.*, 64, p. 759–773, Washington.

Mitchell A. H., Bell J. D. (1973) Island arc evolution and related mineral deposits. *Journ. Geol.*, 81, 4, p. 381–405, Chicago.

O'Connor J. T. (1965) A classification of quartz-rich igneous rocks based on feldspar ratios. *U. S. Geol. Surv. Prof. Paper*, 525 B, p. 79–84, Washington.

Papiu V. C. (1953) Cercetări geologice în masivul Drocea (Munţii Apuseni). *Bul. Acad. RPR*, V, I, p. 107–213, Bucureşti.

Pearce J. A. (1980) Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites. In A. Panayiotou, Ophiolites – Proc. Intern. Ophiol. Symp., Cyprus, 1979, p. 261–272, Nicosia.

–, Haris N. B. W., Tindle A. G. (1984) Trace element discrimination diagram for the tectonic interpretation of granitic rocks. *J. Petrol.*, 25, p. 956–983, Oxford.

Savu H. (1962a) Cercetări geologice şi petrografice în regiunea Troaş-Pîrneşti din Masivul Drocea. *D. S. Com. Geol.*, XLVI, p. 137–159, Bucureşti.

Institutul Geologic al României

– (1962) Chimismul vulcanitelor jurasic superioare – cretacic inferioare din Munţii Drocea. *D. S. Inst. Geol.*, XLVII, p. 199–220, Bucureşti.

– (1983) Geotectonic and magmatic evolution of the Mureş Zone (Apuseni Mountains). *Carp.-Balk. Geol. Assoc. XIIth Congr. Bucharest, 1981. An. Inst. Geol. Geofiz.*, LXI, p. 253–262, Bucureşti.

– , Stoian M. (1988) REE contents in the sheeted dyke complex of the Mureş Zone and their petrogenetic significance. *Rev. Roum. Géol. Géophys. Géogr., Géologie*, 32, p. 37–44, Bucureşti.

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

– , Lupu M., Avram E., Marinescu Fl. (1979a) Harta geologică a R.S.R., scara 1:50000, foaia Săvîrşin. I.G.G., Bucureşti.

– , Lupu M., Lupu D., Ştefan A., Istrate Gh. (1979b) Harta geologică a R.S.R., scara 1:50000, foaia Roşia Nouă. I.G.G., Bucureşti.

– , Udrescu C., Neacşu V. (1985) Petrology and geochemistry of the sheeted dyke complex in the Mureş Zone, Dumbrăviţa-Baia-Bătuţa-Juliţa region (Apuseni Mountains). *D. S. Inst. Geol. Geofiz.*, LXIX/1, p. 129–148, Bucureşti.

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

– , Udrescu C., Neacşu V. (1986b) Bimodal volcanism in the north-western island arc of the Mureş Zone. *D. S. Inst. Geol. Geofiz.*, 70-71/1 (1983; 1984), p. 153–170, Bucureşti.

– , Udrescu C., Neacşu V. (1987a) Mid-ocean characteristics of ophiolites in the Baia-Lupeşti-Vărădia area (Drocea Mountains), their tectonics and petrology. *D. S. Inst. Geol. Geofiz.*, 72-73/1 (1985; 1986), p. 161–180, Bucureşti.

– , Udrescu C., Lemne M., Neacşu V. (1987b) Petrology, geochemistry and tectonics of the ophiolites from the oceanic paleoridge of the Mureş Zone between Pîrneşti, Troaş and Zeldiş Valley (Drocea Mts). *D. S. Inst. Geol. Geofiz.*, 72-73/5 (1985; 1986), p. 237–257, Bucureşti.

– , Udrescu C., Neacşu V., Ichim M. (1989) Petrology of the Liassic ophiolites and Late Kimmerian island arc volcanics in the Vaţa-Căzăneşti-Ţebea region (Mureş Zone) and their trace element contents. *D. S. Inst. Geol. Geofiz.*, 74/1 (1987), p. 149–167, Bucureşti.

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

Steiger R. H., Jäger E. (1977) Convention on the use of decay constants in Goo and Cosmochronology. *Earth Planet. Sci. Lett.*, 36, p. 359–362, Amsterdam.

Streckeisen A. L. (1967) Classification and nomenclature of igneous rocks. *N. Jb. Miner. Abh.*, 107 (2, 3), p. 144–240, Stuttgart.

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

Tuttle O. F., Bowen N.L. (1958) Origin of granite in the light of experimental studies in the system $NaAlSiO_3O_8 - SiO_2 - H_2O$. *Geol. Soc. Am., Memoir*, 74, 74 p., Washington.

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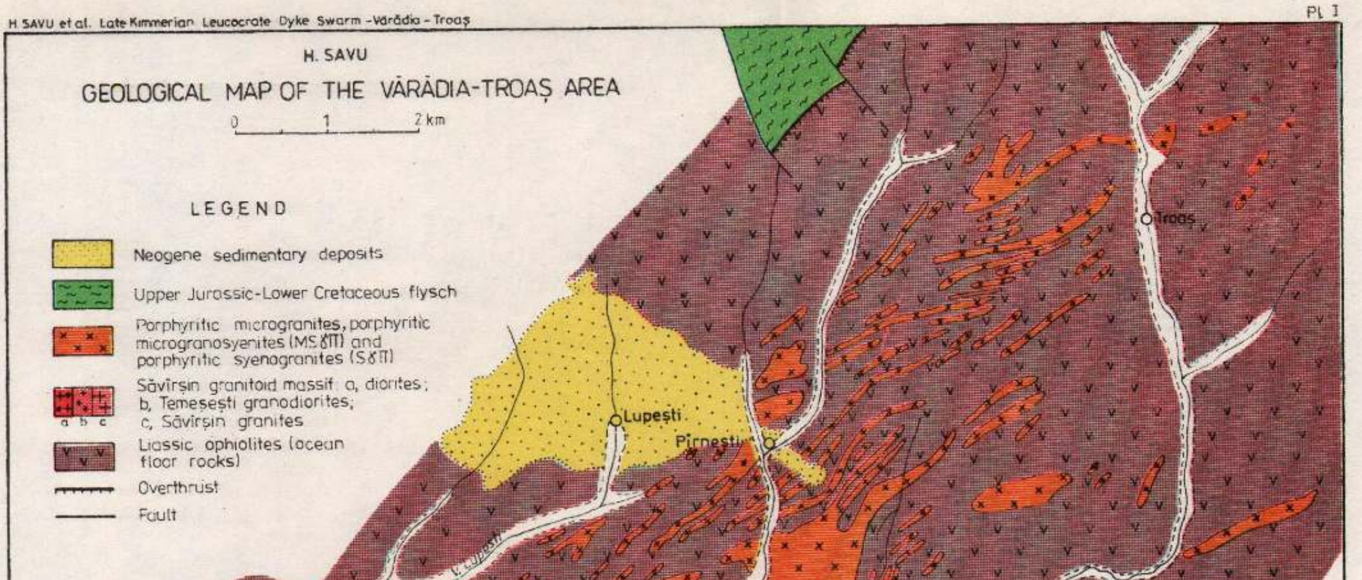

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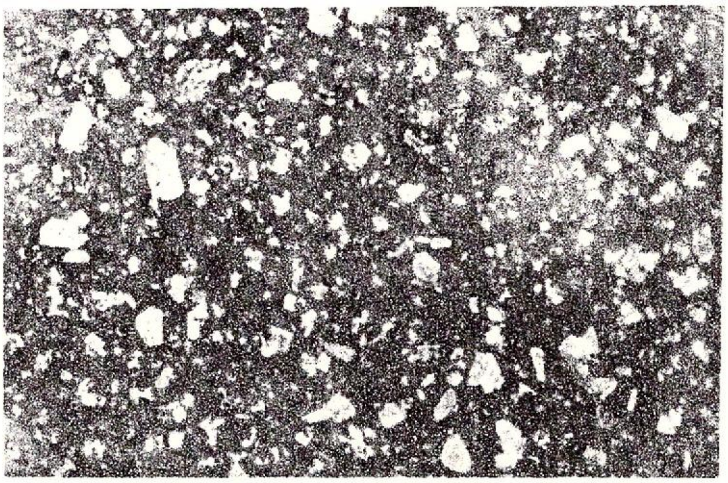

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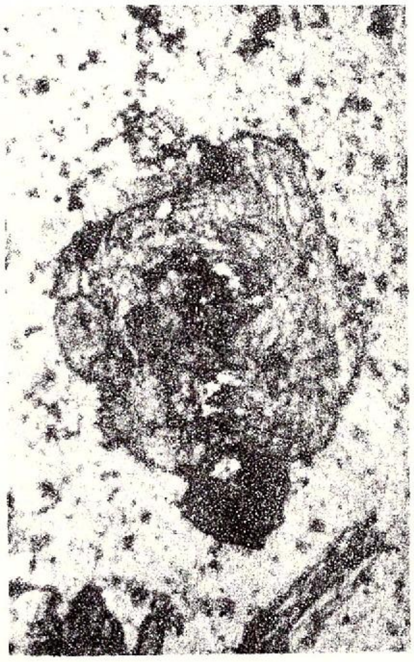

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Plate II

Fig. 1 – Structure of a porphyritic microgranite in the Pîrneşti Valley. Life size.
Fig. 2 – Quartz microphenocrysts with reaction aureoles in a porphyritic granite in the Bruma Valley. N II, x 43.
Fig. 3 – Epidote pseudomorphosis after a biotite phenocryst in a porphyritic microgranosyenite in the Pîrneşti Valley. N II, x 46.

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STRUCTURE, PETROLOGY AND GEOCHEMISTRY OF THE CUIAŞ GABBROIC BODY

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Key words: Gabbros. Magmatic diFferentiation. Tholeiitic composition. Major elements. Minor elements. Apuseni Mts – Drocea Mts.

Résumé: *Structure, pétrologie et géochimie du corps gabbroïque de Cuiaş.* Le corps gabbroïque de Cuiaş est provenu d'un dyke d'alimentation avec magma tholéitique évasé á la partie supérieure. Le magma parental a été affecté par deux processus principaux de différenciation: (1) différenciation dans la chambre magmatique qui a généré des venues successives de: magma gabbroïque normal de fond océanique qui a constitué la masse principale du corps gabbroïque de Cuiaş, magma mélagabbroïque et magma acide riche en Na_2O; (2) différenciation "in situ" du magma gabbroïque du corps, qui a généré une structure à faible stratification rythmique et de nombreux schlieren de gabbros pegmatoïdes. A la base de l'intrusion on a délimité un niveau discontinu de ferrogabbros à magnétite titanifère vanadifère. La composition chimique des roches et leur teneurs en éléments mineurs et terres rares indiquent ces processus. On prouve aussi que le corps gabbroïque s'est formé dans l'Ocean de Mureş près du ridge médian normal (type N), où a été généré le complexe de sheeted dyke (O_2) de la mégaplaque ophiolitique de la zone de Mureş, qui est une suture ophiolitique alpine avec une évolution complexe.

Introduction

The Cuiaş gabbroic body is located in the Mureş Valley, in the southern part of the Drocea Mts (Pl. I). Gabbroic rock occurrences in the Cuiaş area were first reported by Szentpétery (1928) and later by Papiu et al. (1959). In 1967 Savu and Udrescu presented numerous petrographic, geochemical and paleotemperature data on the Cuiaş gabbros. The gabbroic body was subsequently mentioned in different synthesis studies on the geology of the Southern Apuseni Mts.

This gabbroic body is hosted by the oceanic crust fragment of Liassic age (Herz et al., 1974) from the basement of the Căpâlnaş-Techereu Nappe and trends NE-SW-wards. The intrusive body was initially larger, but at present only a part of its north-eastern half is exposed, its extremity being cut by a fault (Pl. I). The south-western half of the gabbroic body was destroyed by the intrusion of Săvîrşin Late Kimmerian granites (128 Ma, Savu et al., 1986) and by the erosion of the Mureş river. Near the south-eastern extremity of the Săvîrşin granite intrusion occur gabbros and basic hornfelses, probably representing the southernmost remnants of the gabbroic body.

Structure and Petrology

Within the hosting Liassic oceanic crust fragment, the gabbroic body lies, just like the Juliţa one (Savu et al., 1982), at the boundary between the ocean floor basalt complex (O_1) and the sheeted dyke complex (O_2). The former complex is well exposed in the south-eastern area, along the national road DN 7 and consists mainly of submarine flows of Liassic basalts, exhibiting intersertal structure, often in pillow lava facies (Pl. II, Fig. 1). They are associated with fine basalt pyroclastics, tachylitic in places, also including large basalt blocks (Pl. II, Fig. 2). The rocks are pierced by calcite and zeolite veinlets, formed in the condition of the zeolitic facies (Savu, 1967) of ocean floor metamorphism (Coleman, 1977). The sheeted dyke complex

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consists of successive parallel intrusions of intergranular basalt, dolerites and gabbros dykes occurring in the north-west of the gabbroic body (Pl. I).

The gabbroic body resembles a trough in shape with asymmetrical slopes formed by the intrusion of a dyke flaring at the top. On the north-western border a Late Kimmerian quartz diorite dyke appeared, while on the south-eastern one the lamination and hydrothermalism of the basalts occurring in the contact area with the gabbroic body point to a fracture system along which the north-western block with the gabbroic body was elevated. On the north-western border, the basalts and the dolerites were thermally metamorphosed in the basic hornfels facies with plagioclase, pyroxene, amphibole and magnetite.

Most of the gabbroic body consists of common gabbros generated by a first intrusion. These rocks contain basic plagioclase (An_{55-65}) and clinopyroxene. They exhibit a heterogeneous, oriented, structure with medium- to fine-grained gabbro bands alternating with pegmatoid gabbro bands or schlieren, rarely nests (Fig. 1), due to an incipient layering (Wager, Brown, 1968) by fractional differentiation of parental tholeiitic magma. The petrographic composition of pegmatoid gabbros is similar or close to that one of common gabbros, the component minerals of the former being more developed. The often uralitized clinopyroxene, poikilitically developed, including numerous fine basic plagioclase crystals, is worth mentioning.

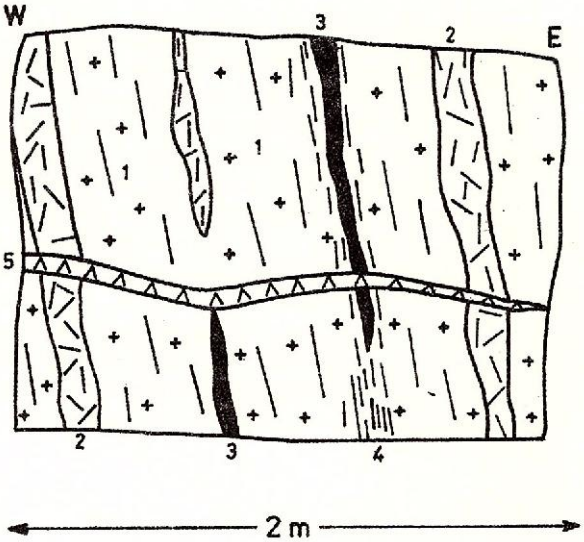

Fig. 1 – Heterogeneous, banded structure of the Cuiaş common gabbros.
1, diopside gabbro; 2, pegmatoid gabbro; 3, hornblende bands; 4, laminated gabbro; 5, dolerite vein. Poiana Brook.

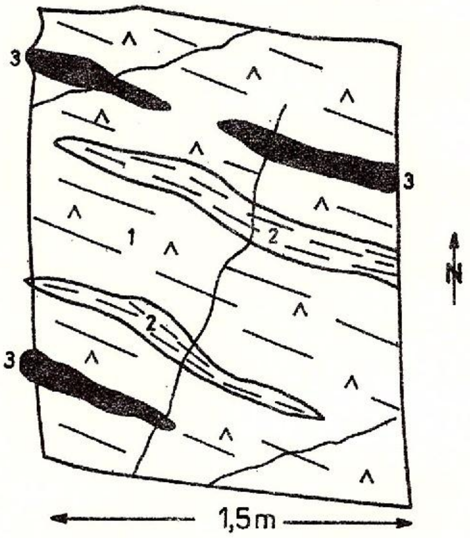

Fig. 2 – Flasergabbro exhibiting hornblende bands, pierced by calcite veins.
1, slightly deformed gabbro; 2, schistous gabbro; 3, hornblenditic bands. Poiana Brook.

The differentiation and layering process at the lower part of the gabbroic intrusion results in the formation of a thin, discontinuous ferrogabbro layer, yielding 5–20 % vanadiferous Ti-magnetite. Remnants of this ferrogabbro layer occur on the north-western border of the body (Pl. I). The consolidation of the ferrogabbroic layer took place at approximately 800° C (Savu, Udrescu, 1967). The component rocks resulted from the differentiation 'in situ' of the first and most important basic magma intrusion. It is probable that prior to the complete cooling of the first gabbroic-tholeiitic magma intrusion from depth, i. e. from the magmatic chamber (Macdonald, 1982), other small intrusions of more basic magma were pulsated, from which the olivine gabbros and hyperites (Tocului Brook), as well as olivine gabbros, troctolites and melagabbros (Pîrîul lui Luca), were generated (Pl. I).

The olivine gabbros in the Tocului Brook are dark coloured rocks, very rich in ferromagnesian silicates. They exhibit a schlieren structure and planar flow bands (Balk, 1937) trending WNW-ESE, sinking 60°-80° southwards, as one may infer from the appended map. The structure of these rocks is represented by alternances of olivine gabbros and troctolites – more melanocratic rocks – with anorthosite bands and light coloured olivine leucogabbros (Irvine, 1982). Some olivine gabbros facies exhibit porphyritic structure resulting from the very large development (1–2 cm length) of some pyroxene crystals (Pl. III, Fig. 1) with a poikilitic structure including olivine crystals.

The basic vein-like rocks which pierce the gabbroic body are represented by dolerite and gabbroporphyrite

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dykes and veins emplaced on longitudinal and cross open fissures (Pl. I). These rocks consist of plagioclase (An_{52}), clinopyroxene, magnetite and pyrite. The plagioclase of gabbro porphyrites constitutes phenocrysts within a hypidiomorphic, granular to fine, doleritic groundmass.

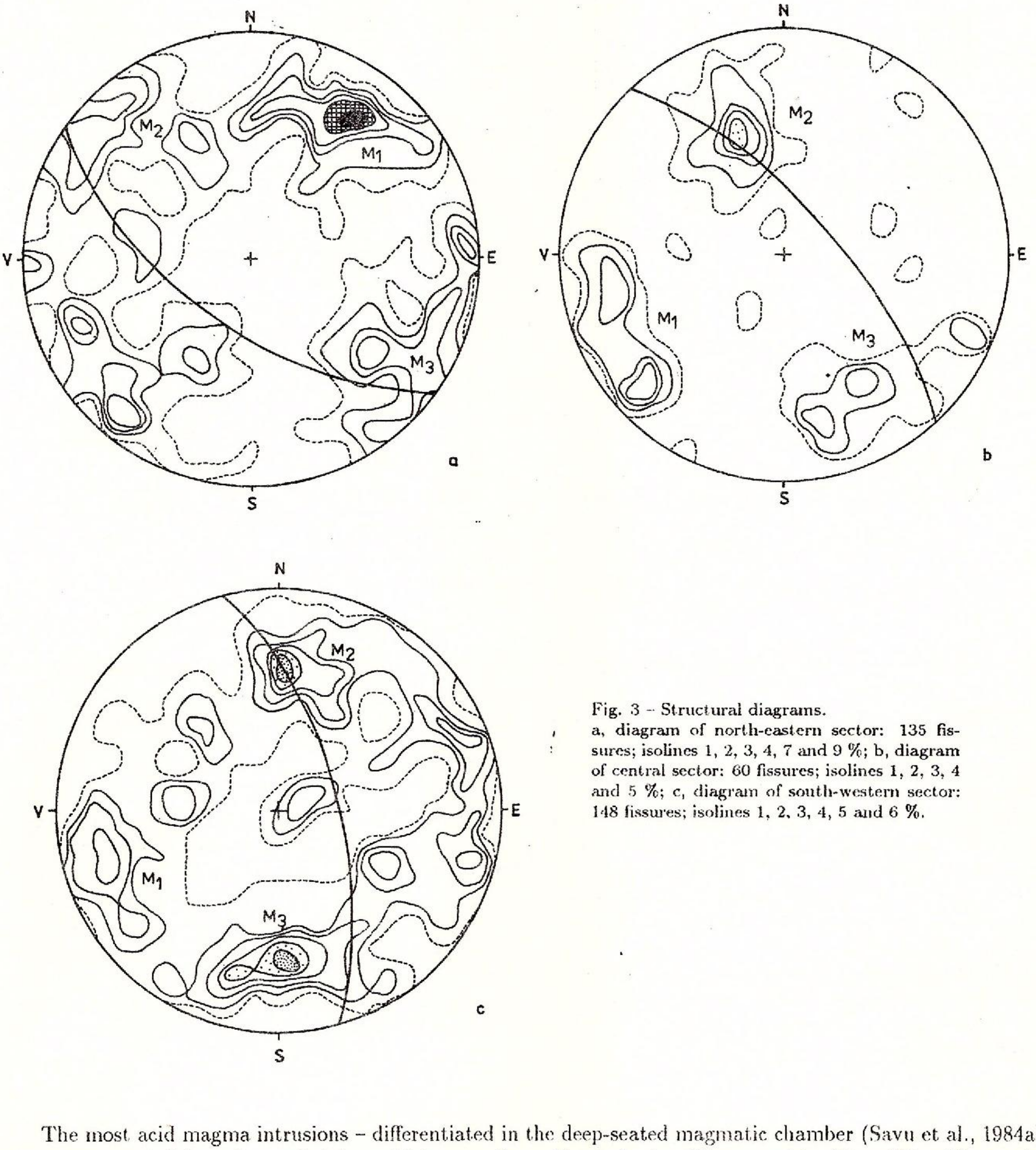

Fig. 3 – Structural diagrams.
a, diagram of north-eastern sector: 135 fissures; isolines 1, 2, 3, 4, 7 and 9 %; b, diagram of central sector: 60 fissures; isolines 1, 2, 3, 4 and 5 %; c, diagram of south-western sector: 148 fissures; isolines 1, 2, 3, 4, 5 and 6 %.

The most acid magma intrusions – differentiated in the deep-seated magmatic chamber (Savu et al., 1984a) – are represented by veins and dykes which cross the gabbroic body. They consist of trondhjemitic quartz diorites, albitic plagiaplites and granophyres, leucocratic rocks, mainly built of albite (An_{8-10}) and quartz, often exhibiting mirmekite or micrographic intergrowth, associated with green amphibole, pistacite, chlorite, apatite and magnetite or ilmenite. These rocks have also been reported from other gabbro bodies in the Mureş

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Table

Chemical composition of

No	1	2	3	4	5	6	7	8	9	10	11
SiO_2	41.97	42.18	42.42	43.32	44.95	47.61	48.01	48.06	48.41	48.70	48.74
TiO_2	3.65	0.74	0.90	0.30	4.49	0.76	0.58	0.22	0.24	1.40	0.24
Al_2O_3	18.85	7.84	5.64	24.88	12.57	16.49	16.10	16.45	14.69	14.44	18.31
Fe_2O_3	11.70	6.29	6.62		5.35	0.90	1.44	0.92	0.72	1.13	0.81
FeO	7.65	6.64	5.84	6.40	10.99	3.97	3.51	3.35	5.05	6.41	2.58
Mn0	0.25	0.19	0.21	0.65	0.34	0.11	3.51	0.09	0.13	0.16	0.08
MgO	6.00	21.95	25.03	9.68	6.89	12.13	12.21	10.88	12.65	7.43	8.78
CaO	8.55	8.77	8.22	13.28	9.99	16.00	16.60	16.20	15.17	13.80	19.13
Na_20	0.10	0.97	0.44	0.74	1.98	0.92	0.80	1.45	1.04	3.15	1.04
K_2O	0.57	0.07	0.02	0.15	0.24	0.14	0.17	0.16	0.16	0.61	0.10
P_2O_5	0.15	0.01	0.17		0.10	0.17	0.14	0.02	0.03	0.12	0.07
S	0.08	0.11	0.09		0.08	0.09.	0.15	0.52	0.23	0.13	0.13
H_2O^+	0.11	3.95	4.85	0.25	1.49	1.00	0.98	1.29	1.07	1.71	0.50
Total	100.03	99.71	100.45	100.30	99.46	100.38	100.78	99.61	99.59	100.64	100.51
Ni		1000	1000		31	270	260	180	350	110	200
Co		170	165		75	65	32	32	60	52	38
Cr		520	460		12	850	820	600	1350	440	580
V		115	110		1100	280	290	125	380	460	320
Sc		30	32		70	65	65	42	80	52	65
Y		10	10		46	12	10	10	10	32	10
Yb		1	1		2.6	1	1	1	1.3	3.6	1.1
La		30	30		30	30	30	30	30	30	30
Zr		19	15		140	125	160	220	120	130	20
Ba		10	10		35	27	26	18	53	22	22
Sr		30	15		140	125	160	220	120	130	150
Nb		10	10		10	10	10	10	10	10	10
Ga		3	2		12	6.5	6	8	7	9	8.5
Cu		90	115		26	180	55	65	140	75	185

* The analysed rocks represent: ferrogabbros: 1, 5 – Sîrbu Brook; melagabbros: 2, 3 – Luca Brook; olivine gabbros: 4, 9, 11 - Toc Valley, 8 – Poiana Brook; pegmatoid gabbros: 6 – Poiana Brook, 20 - Cuiaş Valley; diopside gabbros: 7, 16, 17 - Cuiaş Valley, 12 – Valea Lungă Brook, 13 – Poiana Brook, 15 – Toc Valley, 18 – Valea Rea Brook; gabbroporphyrite: 14 – Poiana Brook; gabbrodolerite: 21 – Valea Rea Brook; basalt: 10 – Cuiaş (DN7); dolerite (dyke): 19 – Toc Valley; albite granophyre: 22 – Crăciuneasa Valley; albite plagiaplite: 23 – Cuiaş Valley.

Zone, such as those at Juliţa, Almăşel, Almaş Sălişte and the sheeted dyke complex in Dumbrăviţa-Juliţa area.

Along some cross (ac) fracture planes the gabbros were deformed (Pl. I and Fig. 2) and flaser gabbros with hornblende bands (Pl. III, Fig. 2) were generated. The deformation and recrystallisation of gabbros took place by the end of the gabbroic body consolidation, at high temperature and under the action of late magmatic solutions. Our conclusion is confirmed by the fact that the deformed rocks are also crossed by basic rock dykes (Fig. 1).

The cooling and the complete crystallisation of magma intrusions have generated a NE-SW elongated gabbroic body with several fissure systems, which, according to the body elongation, are classified into longitudinal, cross and diagonal fissures. The structural diagrams of Figure 3 point to the prevalence and position of these fissures within the three sectors of the gabbroic body – that part which is still preserved – divided in view of measurements (Pl. I). Thus, in the north-eastern extremity of the body (diagram 3a) cross fissures (M_1) trending N65^{0}W/65^{0}S do prevail. On diagram 3b the longitudinal fissures from the central area (M_2 and M_3) and the cross fissures (M_1) trending N42^{0}W/72^{0}N are plotted. In the south-western part of the body, strongly affected by fractures and disturbed, longitudinal fissures with two maxima (M_2 and M_3) and cross fissures (M_1) trending N15^{0}W/67^{0}N are frequent (Fig. 3c). The cross fissures formerly formed (Balk, 1937) often exhibit deuteritic hornblende crystals.

Geochemistry and Tectonic Setting

The elaboration of the present study implied 19 new chemical and spectral analyses (Tab. 1), added to the 4 chemical analyses previously published (Savu, Udrescu, 1967). The table points out the four rock groups delimited petrographically within the gabbroic body; these are also represented on the diagram of Figure 4, drawn up according to Irvine and Baragar (1971).

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1

the gabbroic rocks*

12	13	14	15	16	17	18	19	20	21	22	23
48.92	48.98	49.30	49.62	49.62	49.64	49.84	50.60	51.52	54.48	54.52	63.91
1.10	1.32	1.54	3.20	0.48	1.02	0.62	1.40	0.46	1.42	1.50	0.48
18.81	16.94	15.30	18.31	22.75	15.79	15.64	14.84	15.15	16.44	16.63	19.91
6.60	0.37	2.23	0.93	0.71	0.46	2.74	2.81	0.54	4.59	4.32	0.39
1.20	4.08	6.45	4.90	2.75	6.84	4.57	7.14	4.26	3.26	2.48	0.22
	0.16	0.16	0.12	0.05	0.15	0.15	0.20	0.13	0.15	0.18	0.01
6.03	10.73	8.07	7.01	6.00	9.56	8.25	6.95	8.89	4.95	3.30	0.70
12.17	13.92	12.80	13.05	14.90	12.65	12.17	11.70	13.95	9.18	9.52	10.27
2.14	1.18	2.88	2.26	2.05	2.51	2.28	3.01	3.17	3.18	2.15	3.53
0.27	0.02	0.09	0.12	0.14	0.15	0.16	0.09	0.25	0.13	0.45	0.22
0.54	0.23	0.25	0.20	0.22	0.16	0.05	0.31	0.24	0.22	0.31	0.31
	0.15	0.75	0.19	0.13	0.12	0.36	0.22	0.09	0.70		0.11
0.18	1.52	0.89	0.88	0.82	1.68	2.14	1.35	1.44	1.13	1.75	0.77
97.96	99.60	100.71	100.79	100.66	100.64	99.98	100.62	100.09	99.83	100.06	100.83
	370	230	175	270	150	90	105	110	45		14
	57	42	42	34	60	28	57	38	24		2.5
	2600	460	730	750	550	240	230	185	80		28
	370	280	400	230	420	240	230	185	340		24
	82	40	52	45	70	72	60	82	60		6
	15	27	23	12	19	10	46	18	28		36
	1.4	3.7	2.4	1	1.8	2	3.5	2.2	4		4.4
	30	30	30	30	30	30	30	30	30		55
	29	160	40	24	29	38	210	37	80		220
	65	55	24	20	17	46	36	25	48		30
	105	480	135	200	135	400	210	170	480		280
	10	10	10	10	10	10	10	10	10		10
	7.5	16	12	11	10	12	11	8	14		17
	70	40	115	14	125	40	200	14	10.5		18

The CO_2 content is 0, excepting samples 10 = 1.45 %; 18 = 1.01 %. The Pb content is 2 ppm or lower than 2 ppm, excepting samples: 2 = 7.5 ppm; 4 = 7 ppm; 13 = 3 ppm; 20 = 3 ppm; 21 = 2.5 ppm; 23 = 4 ppm. The Sn content is lower than 2 ppm, excepting samples: 2 = 2.5 ppm,; 19 = 3 ppm.

The first group of rocks plotted on the diagram next to MgO corner consists of melagabbros and some olivine gabbros. The former yield 42 % SiO_2 and 25 % MgO. On this diagram they occur between basic and ultrabasic rocks. These rocks resemble certain ultrabasites of the layered intrusions (Savu, Strusievici, 1987), being plotted on the diagram of Malpas and Stevens (1977) within the field of these rocks. However, they have been supplied by magma intrusions different from the magma which had generated the main part of the Cuiaş gabbroic body, built up of common gabbros.

The SiO_2 content of common gabbros (diopside gabbros) and of some olivine gabbros ranges from 47 to 52 %. The rocks of this type and the vein-like ones are mostly plotted on the tholeiitic field (TH), some of them – Na_2O richer – exceeding its boundaries and entering the field of calc-alkaline rocks (CA). This field includes gabbroic rocks richer in MgO situated at its lower part and some rocks yielding somewhat higher iron content occurring at the upper part. The latter are represented by two basic vein-like rocks (dykes), a basalt and a gabbro (no. 15).

The rocks mostly correspond to tholeiitic basalts, representing, by definition, basic rocks generated by crystallisation of tholeiitic magma supplied by the parental magma, which was less differentiated in the magmatic chamber.

Closer to the FeO_{tot} corner of the diagram of Figure 4 are the ferrogabbros (nos. 1 and 5) the reduced SiO_2 content (42–45 %) of which resembles that one of melagabbros, while their MgO content is lower than the one of the same rocks. The ferrogabbros have increased contents of iron, mainly FeO (7.65–11 %), Fe_2O_3 (11 %) and TiO_2, included in vanadiferous Ti-magnetite. The Ti-magnetite of the ferrogabbro collected from the Sirbu Valley (Toc) exhibits the following elements (ppm) included in its network: 7200 Al, 30000 Ti, 1060 Mn, 4780 V, 51 Ni and 69 Co. The crystallisation of Cuiaş ferrogabbros at approximately 800^0 C (Savu, Udrescu, 1967) corresponds to Norton's (1955) data on the equilibrium $4\ Fe_3O_4=6\ Fe_2O_3$ achieved normally round this temperature. According to this author the mentioned equilibrium is also obtained at less than 800^0 C, as confirmed by the crystallisation temperature (710^0 C) of the Almaş Sălişte ferrogabbros (Savu, Udrescu,

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1967). According to Eugster (1959) hematite is stable below 500⁰ C and magnetite stability is reached as far as 775⁰ C.

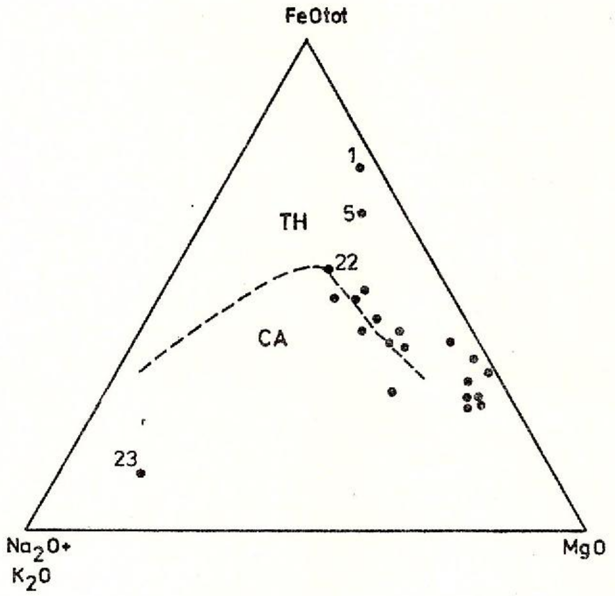

Fig. 4 - FeO_{tot}-MgO-Na_2O+K_2O diagram (acc. to Irvine and Baragar, 1971).
TH, tholeiitic field; CA, calc-alkaline field.

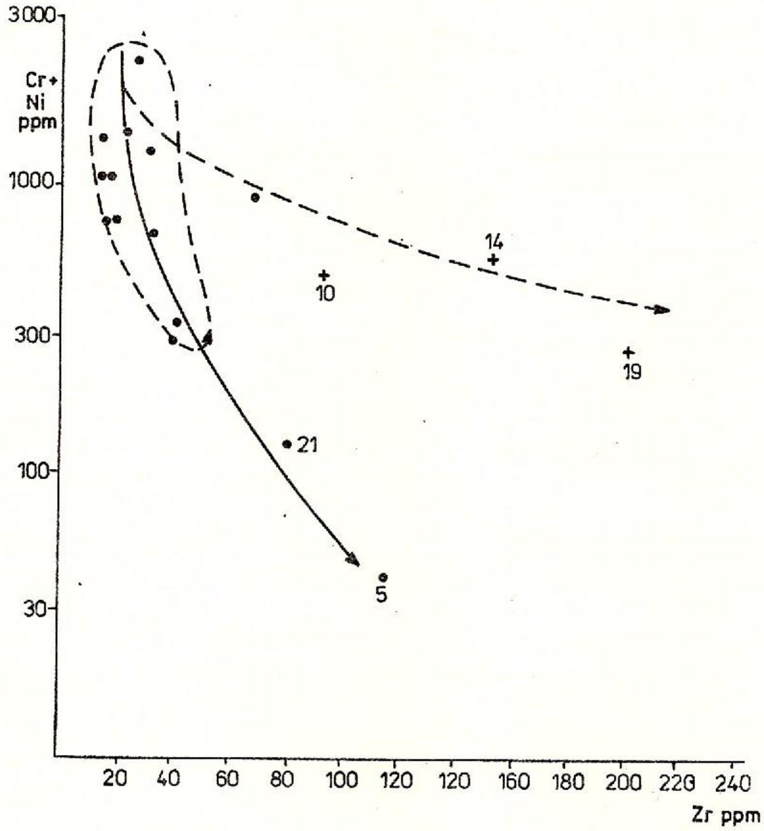

Fig. 5 - Cr+Ni-Zr diagram (acc. to Beccaluva, 1977).

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The tendency of iron to concentrate in ferrogabbroic residual magmas is obviously shown by Cr+Ni–Zr diagram (Beccaluva, 1977) of Figure 5, on which the magma differentiation line crosses the field of gabbroic rocks to ferrogabbros, pointing to Cr and Ni decrease and Zr increase within these residual magmas. On the same diagram, the vein-like rocks (nos. 14 and 19), the basalt (no. 10) and a gabbro occur along a special curve also pointing to Zr enrichment.

The ferrogabbros are the basic differentiates of the main tholeiitic magma intrusion and constitute a characteristic layer at its base. They confirm Fenner's (1929) opinion that the tholeiitic magma concentrates iron, TiO_2 and V (Savu, 1986; Savu et al., 1984a).

The acid differentiates (nos. 22 and 23) show significant SiO_2 (64 %) and Na_2O (3.5 %) increase. On the diagram of Figure 4 these rocks are plotted near the alkali corner and are Na_2O rich.

Most of the rocks presented above are plotted on the field of tholeiitic rocks (Fig. 4), their distribution suggesting their consanguineous character and their generation from the parental magma by fractional differentiation.

Excepting the olivine gabbros, the hyperites, the troctolites and the melagabbros, which are independent basic magma intrusions, one notes that the Cuiaş gabbroic body, as well other gabbroic bodies investigated by us in the Mureş Zone (Savu, 1962; Savu et al., 1982, 1984b), contain the same groups of gabbroic rocks (common gabbros, ferrogabbros and acid-albitic differentiates), also present within the sheeted dyke complex (O_2) in the Mureş Zone (Savu et al., 1981, 1984a). Therefore, the tholeiitic parental magma differentiation trend is preserved both in a major magmatic chamber and in a smaller one which generated the magmas that constituted the Cuiaş gabbroic bodies or other gabbro bodies associated with ophiolites in the Mureş Zone.

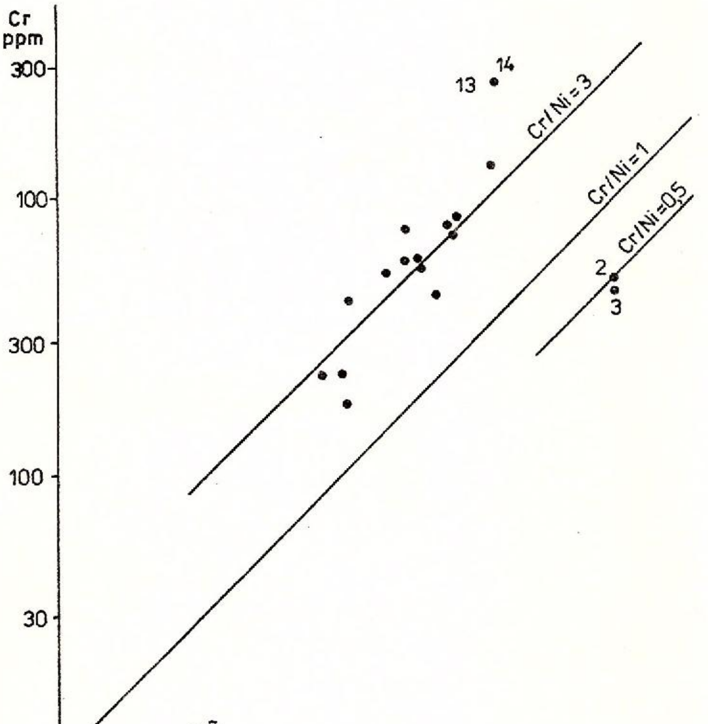

Fig. 6 – Cr–Ni diagram.

The minor elements contents determined by emission spectrography (Tab. 1) have been differently influenced by the two processes of parental tholeiitic magma differentiation, i. e. the one that took place in the magmatic chamber and the rhythmical layering "in situ" within the gabbroic body. Although the gabbroic rocks are assigned to the four rock groups inferred from petrologic data and major chemical elements (Fig. 4) by taking into account their contents and the ratios between certain elements, the minor elements also allow a detailed classification of the rocks plotted on the diagram of Figure 4. Thus, the melagabbros are characterized by the

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highest Ni and also Cr contents, being the olivine-richest rocks (Fig. 6).

The other rocks grouped within the same field as the melagabbros (Fig. 4) are marked by higher Cr values. The latter, as well as the other rocks analysed, occur along the line which corresponds to Cr/Ni=3, as obviously shown by the diagram of Figure 6 on which are plotted the rocks depending on the ratio between the two elements. On this diagram the ferrogabbro (no. 5) is placed within the field of the lowest values. The diagram resembles the one of Figure 7 showing the plotting of Cuiaş gabbroic rocks on the Co-Sc diagram, on which melagabbros (Co/Sc=5) and another important field of most of the gabbroic rocks (Co/Sc=0.9) are delimited. Other gabbros are plotted on this diagram at the lowest values of Co/Sc ratio (0.4).

The ferrogabbros with iron and titanium concentrations prove to be the rocks which yield the highest V contents, reaching 1100 ppm in the analysed rock (no. 5)(Tab. 1).

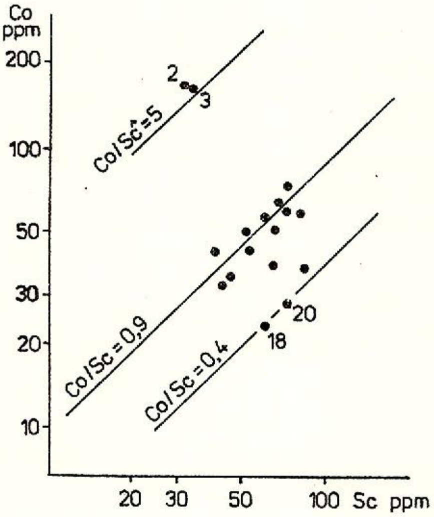

Fig. 7 – Co-Sc diagram.

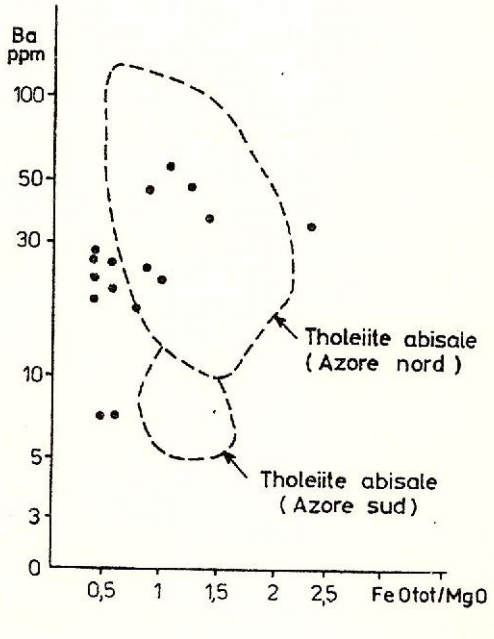

Fig. 8 – Ba–FeO_{tot}/MgO diagram.

It is to be noted that Ba contents of gabbroic rocks are much more stable, with respect to tholeiitic magma differentiation affected by the two mentioned processes – differentiation in the magmatic chamber and rhythmical layering ‘in situ’ – by comparison to the parental tholeiitic magma, of approximately the same

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chemical composition, as ocean floor basalts. The diagram of Figure 8 shows that most of the gabbroic rocks are plotted round or within the field of abyssal basalts from the Azore islands, studied by Miyashiro (1975), which accounts for their character of ocean floor rocks.

The Sr contents are usually lower in melagabbros and ferrogabbros, while in the other rocks they frequently exceed 100 ppm. Herz et al.(1974) determined the $^{87}Sr/^{86}Sr$ ratio of three samples collected from the Cuiaş gabbroic body as 0.7030, while the Rb/Sr ratio average is 0.031. These reduced values are typical of ocean floor rocks and correlate with the geochemical data obtained for all the ocean floor tholeiitic rocks of Liassic age from the Mureş Zone which host the Cuiaş gabbroic body (Savu et al., 1981). According to different authors (Hart, 1971; De Paolo, Wasserburg, 1976; O'Nions et al., 1977; Cohen et al., 1980) the reduced ratios mentioned above account for a mantle source of tholeiitic parental magma, previously affected by melt extraction. The age of these gabbros and of the other ophiolites in the Mureş Zone is of 180 Ma (Herz et al., 1974).

The differentiation process supported by the tholeiitic magma in the magmatic chamber is well rendered by the REE contents of gabbroic rocks (Tab. 2) and mainly by the diagram of Figure 9, on which are plotted the chondrite-normalized values of these elements (Nakamura, 1974). This diagram shows that all the six samples analysed by neutron activation have an important Eu anomaly, which, according to Philpotts and Schnetzler (1968), points to advanced differentiation of initial magma.

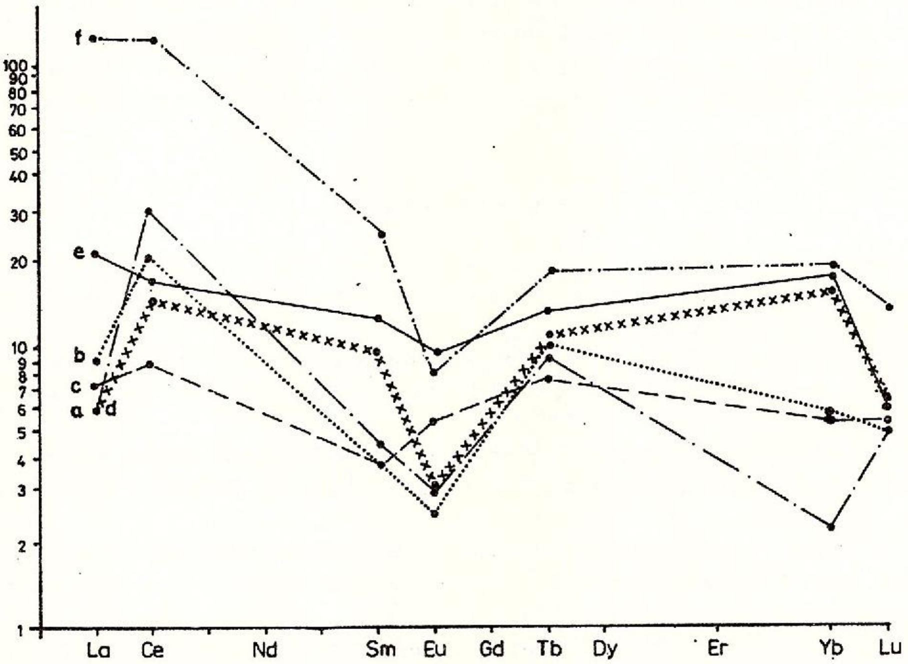

Fig. 9 – Chondrite-normalized REE patterns of the Cuiaş gabbroic rocks.
a, melagabbro (no. 1); b, olivine gabbro (no. 2); c, common gabbros (nos. 3 and 4 – average); d, basalt (no. 5); e, pegmatoid gabbro (no. 6); f, albitic plagiaplite (no. 7).

The average Ce/Yb ratio of the two common gabbros is of 8.05, of gabbrodolerite amounts to 3.75 and of basalt to 3.61 all of them less than 10. In Saunder's (1984) opinion these reduced values account for the origin of magmas in a normal middle oceanic ridge (N-Type MORB), similarly to the ridge in which originates the sheeted dyke complex from the western part of the oceanic crust megaslab of the Mureş Zone (Savu, Stoian, 1988), that hosts the Cuiaş gabbroic body. The highly differentiated rocks, such as melagabbros, ferrogabbros and albitic plagiaplites show a much increased Ce/Yb ratio, ranging from 11 to 54, which accounts for REE concentration in different amounts concomitantly with the differentiation of tholeiitic magma.

Table 1 and 2 show that an increasing Zr content-marking the differentiation of tholeiitic magma (Savu, Udrescu, 1975) – implies the REE increase, as proved by the albitic plagiaplite, the last to be separated from the tholeiitic magma in the magmatic chamber, a rock yielding the highest REE contents (Savu, Stoian, 1988). The high Y content (36 ppm) and the low Nb one (<10 ppm) prove the assignment of this rock to the oceanic ridge granitoids (Pearce et al., 1984).

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Table 2

REE contents in some gabbroic rocks

No*	1 (3)	2 (9)	3 (16)	4 (17)	5 (10)	6 (21)	7 (23)
La (ppm)	2	3	3	2	2	7	44
Ce	27	18	7	8	13	15	110
Sm	0.88	0.90	0.80	0.86	2.0	2.6	5.8
Eu	0.56	0.52	0.67	0.36	0.62	0.73	1.63
Tb	0.45	0.51	0.36	0.40	0.55	0.65	0.95
Lu	0.20	0.19	0.20	0.18	0.20	0.20	0.50
Ce/Yb**	54.00	13.84	11.66	4.44	3.61	3.75	25.00
(Ce/Yb)cn	13.47	3.55	3.00	1.12	0.92	0.96	6.35

* In brackets are indicated the adequate numbers from the Table 1. The REE have been determined by neutron activation analysis.

** The values used for Yb to calculate the two ratios were the ones from the Table 1, determined by emissiom spectrography.

The (Ce/Yb)cn ratio below 1 yielded by basalt and gabbrodolerite, and the mean of the two common gabbros (nos 3 and 8) very close to this value also show (Schilling, 1975) that these rocks, which constitute the groundmass of the gabbroic body, and the adjoining basic rocks, are 'N-Type' tholeiites originating in a normal median ridge (Marriner, Millward, 1984). The REE content and the (Ce/Yb)cn ratio increase in differentiate rocks.

Conclusions

The Cuiaş gabbroic body resembles a trough in shape and originates in a tholeiitic magma supplied dyke, which pierced sheeted dyke complex (O_2) and flared at the top, tending to form a large sill between the two ophiolitic rock complexes from this area.

The parental magma which generated the gabbroic body underwent two differentiation processes (Savu, 1986): (1) differentiation in the deep-seated magmatic chamber and (2) differentiation 'in situ' of the main intrusion.

The magmatic chamber differentiation generated the following successive magma intrusions: (a) gabbroic (tholeiitic) ocean floor magma, (b) melagabbroic magma and (c) SiO_2 and Na_2O-rich acid magma (Savu et al., 1984a).

The gabbroic (tholeiitic) magma generated the main intrusion which constitutes in fact the gabbroic body, poorly rhythmically layered with a Ti-magnetite ferrogabbro level at the base and several gabbroic rocks bands and schlieren of melanocratic or leucocratic nature, for the rest.

Concomitantly with the 'in situ' differentiation, the magma gradually enriched in gas, mainly dissociated water, which generated the banded structure of gabbros by the crystallisation of schliren and nests of frequent pegmatoid gabbros.

The chemical composition of the rocks belonging to the three magma groups and the distribution of minor elements account for these conclusions. The REE distribution patterns represented on the diagram of Figure 9 are illustrative of the tholeiitic magma differentiation both in the magmatic chamber and 'in situ' within the intrusion.

Within the Mureş oceanic zone the gabbroic body formed near the normal median ridge (N-Type MORB) which contained the sheeted dyke complex (O_2) of the ophiolitic megaslab from the Mureş Zone, that is an alpine ophiolitic suture exhibiting complex evolution.

These complex processes generated the Cuiaş gabbroic body, of utterly heterogeneous petrographic constitution, still belonging to the category of differentiated tholeiitic magma products, similarly to other gabbroic from the Mureş Zone.

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References

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

Beccaluva L., Ohnenstetter D., Ohnenstetter M., Venturelli G. (1977) The trace element geochemistry of Corsican ophiolites. *Contrib. Mineral. Petrol.*, 64, p. 11–31, Berlin.

Cohen B. S., Evensen N. M., Hamilton P. J., O'Nions R. K. (1980) U-Pb, Sm-Nd and Rb-Sr systematics of mid-ocean ridge basalt glasses. *Nature*, 283, p. 149–153, London.

Coleman R. S. (1977) Ophiolites – Ancient oceanic lithosphere?. Springer-Verl. p. 229, Berlin.

De Paolo D. J., Wasserburg G. J. (1976) Inferences about magma sources and mantle structure from variation of ^{143}Nd/^{144}Nd. *Geophys. Res. Lett.*, 3, p. 249–253, Washington.

Eugster H. P. (1959) Reduction and oxidation in metamorphysm. In P. H. Abelson. Researches in Geochemistry, p. 397–426, J. Wiley, New York.

Fenner C. N. (1929) The crystallization of basalts. *Am. J. Sci.*, 5, 18, p. 225–253, New Haven.

Hart S. R. (1971) K, Rb, Cs, Sr and Ba contents and Sr isotope ratios of ocean floor basalts. *Philos. Trans. R. Soc.*, A, 268, p. 573–578, London.

Herz N., Jones L. M., Savu H., Walker R. L. (1974) Strontium isotope composition of ophiolitic and related rocks, Drocea Mountains, Romania. *Bull. Volc.* XXXVIII-4, p. 1110–1124. Napoli.

Irvine T. N. (1982) Terminology for layered intrusions. *Jour. Petrology*, 23/2, p. 127–162, Oxford.

– , Baragar W. (1971) A guide to the chemical classification of common volcanic rocks. *Can. J. Earth Sci.*, 8, p. 523–548, Ottawa.

Macdonald K. C. (1982) Mid-ocean ridges: fine scale tectonic, volcanic and hydrothermal processes within the plate boundary zone. *Ann. Rev. Earth Planet. Sci.*, 10, p. 155–191, Palo Alto.

Malpas J., Stevens R. K. (1977) The origin and emplacement of the ophiolitic suite with examples from western Newfoundland. *Geotectonics*, 6, p. 83–102, Moscow.

Marriner G. T., Millward D. (1984) The petrology and geochemistry of Cretaceous to Recent volcanism in Colombia: the magmatic history of an accretionary plate margin. *J. Geol. Soc. London*, 141, p. 473–486, Oxford.

Miyashiro A. (1975) Volcanic rocks series and tectonic setting. *Ann. Rev. Earth Planet. Sci.*, 3, p. 251–270, Palo Alto.

Nakamura N. (1974) Determination of REE, Ba, Fe, Mg, Na and K in carbonaceous and ordinary chondrites. *Geochim. Cosmochim. Acta*, 38, p. 757–775, Oxford.

Norton F. J. (1955) Dissociation pressures of iron and copper oxides. *General Electric Report*, 55 – R. L. – 1248.

O'Nions R. K., Hamilton P. J., Evensen N. M. (1977) Variation in ^{143}Nd/^{144}Nd and ^{87}Sr/^{86}Sr ratios in oceanic basalts. *Earth Planet. Sci. Lett.*, 43, p. 13–22, Amsterdam.

Papiu V. C., Savu H., Romanescu D., Pîrvu O. (1959) Corelaţia dintre alcătuirea geologică şi anomaliile magnetice din zona axială a Masivului Drocea (Munţii Apuseni). *D. S. Inst. Geol. XLII*, p. 49–86, Bucureşti.

Pearce J. A., Harris N. B. W., Tindle A. G. (1984) Trace element discrimination diagram for the tectonic interpretation of granitic rocks. *Jour. Petrology*, 25/4, p. 956–983, Oxford.

Philpotts J. A., Schnezler C.C. (1968) Europium anomalies and the genesis of basalt. *Chem. Geol.*, 3, p. 5–13, Amsterdam.

Saunders A. D. (1984) The rare earth element characteristics of igneous rocks from the ocean basins. In: P. Henderson (edit.) Rare Earth Element Geochemistry. Elsevier, p. 205–236, Amsterdam.

Savu H. (1962) Corpul gabroic de la Almăşel şi contribuţii la cunoaşterea chimismului şi petrogenezei ofiolitelor din masivul Drocea. *An. Com. Geol.*, XXXII, p. 211–248, Bucureşti.

– (1967) Die mesozoischen Ophiolite der rumänischen Karpaten. *Acta Geol. Acad. Sci. Hung.* 11 (1–3), p. 59–70, Budapest.

– (1986) Concentrations of Fe, Ti, and V in the Alpine Ophiolites from the Mureş Zone, Romania. In: Metallogeny of Basic and Ultrabasic Rocks (Regional Presentations), p. 99–120, Athens.

– , Udrescu C., Neacşu V. (1959) Petrology and geochemistry of the sheeted dyke complex in the Mureş Zone, Dumbrăviţa – Baia – Bătuţa – Juliţa region (Apuseni Mountains). *D. S. Inst. Geol. Geofiz.*, LIX, p. 129–148, Bucureşti.

– , Udrescu C. (1967) Paleotemperatura şi geochimia gabrourilor de la Cuiaş (Munţii Drocea). *D. S. Inst. Geol.*, LIII/2, p. 185–217, Bucureşti.

– , Udrescu C. (1975) Distribution of Zr in some basic rocks from Romania and its petrological significance. Proc. 10-th Congr. Carpatho-Balk. Geol. Assoc. (1973), IV, p. 214–221, Bratislava.

– , Udrescu C., Neacşu V. (1981) Geochemistry and geotectonic setting of ophiolites and island arc volcanics of the Mureş Zone (Romania). *Ofioliti*, 6 (2), p. 269–286, Bologna.

– , Vasiliu C., Udrescu C. (1982) Structure, petrology and geochemistry of the Juliţa gabbroic body – Alpine ophiolites of the Drocea Mountains (Apuseni Mountains). *D. S. Inst. Geol. Geofiz.*, LXVI/1, p. 127–152, Bucureşti.

Institutul Geologic al României

– , Udrescu C., Neacşu V., Stoian M. (1984a) Trends of tholeiitic magma differentiation in the sheeted dyke complex from the Mureş Zone, Romania. 27^e Congr. Géol. Intern., Moscow, 1984, *An. Inst. Geol. Geofiz.*, LXIV, p. 121–131, Bucureşti.

– , Udrescu C., Neacşu V. (1984b) Petrology and geochemistry of the layered dyke of Almaş Sălişte (Mureş Zone, Romania). *Mineralia slov.*, 16, p. 43–49, Bratislava.

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

– , Strusievicz R.O. (1987) MgO, Cr, Ni and Co distribution in ultramafic rocks in Romania: classification and origin. *D. S. Inst. Geol. Geofiz.*, 72-73/1, p. 361–379, Bucureşti.

– , Stoian M. (1988) REE contents in the sheeted dyke complex of the Mureş Zone and their petrogenetic significance. *Rév. Roum. Géol. Géophys. Géogr., Géologie*, 32, p. 37–44, Bucureşti.

Schilling J. G. (1975) Rare-earth variation across "normal-segments" of the Reykjanes Ridge, 60–53^{0}N, Mid. Atlantic Ridge, 29^{0}S, and East Pacific Rise, 2-19^{0}S, and evidence of the composition of the underlying low-velocity layer. *J. Geophys. Res.*, 80, p. 1459–1475, Washington.

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

Wager L. R., Brown G. M. (1968) Layered igneous rocks. 588 p., London.

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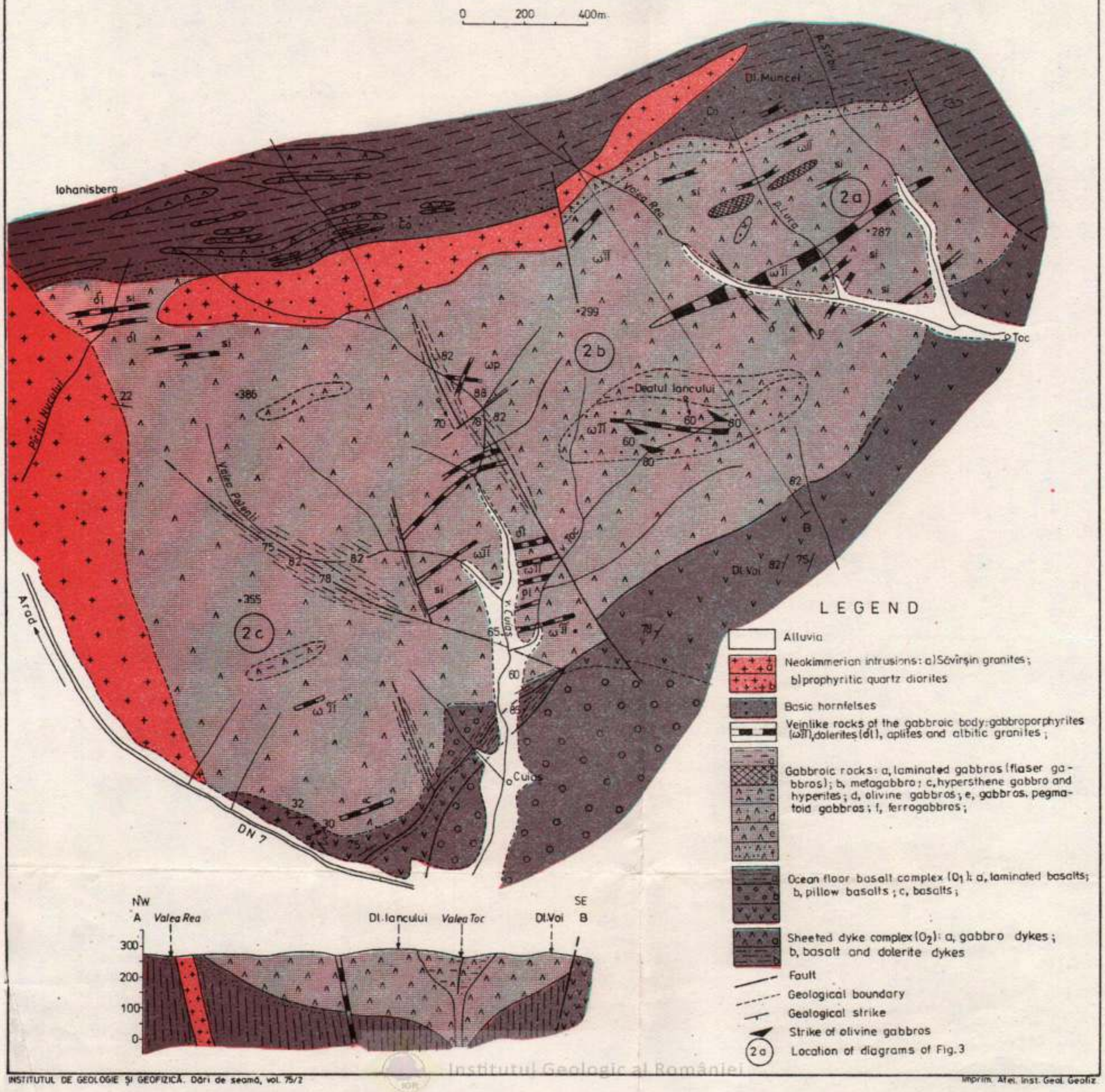
GEOLOGICAL MAP OF THE CUIAŞ GABBROIC BODY



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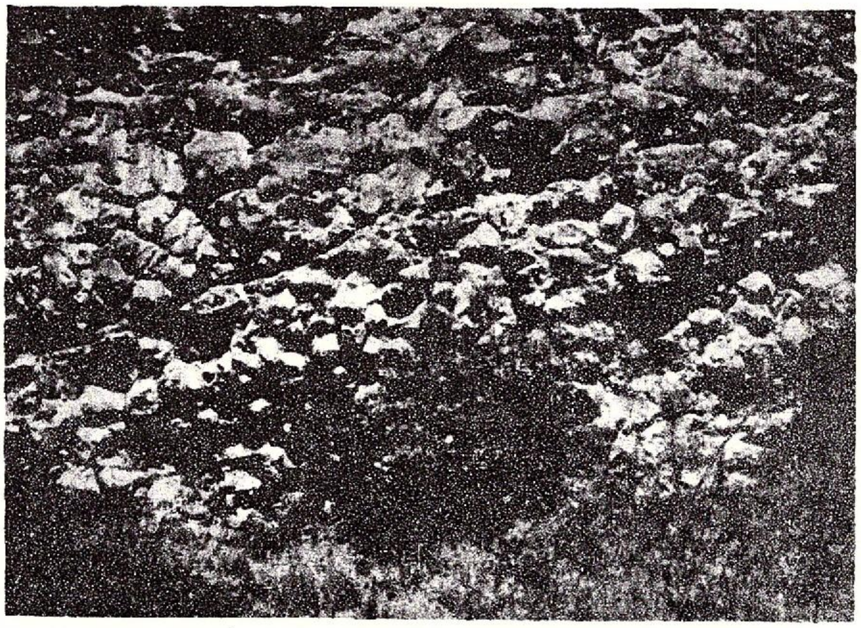

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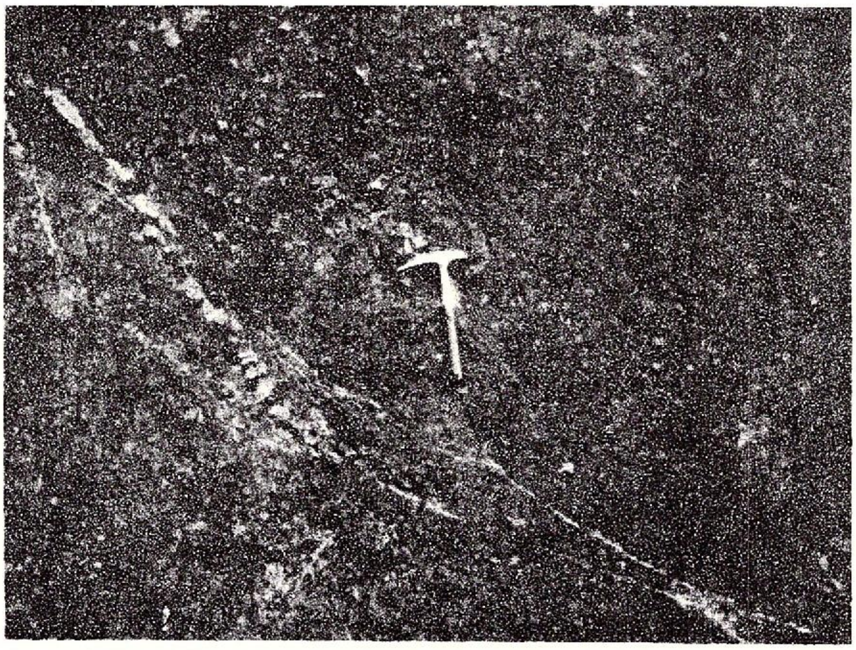

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Plate II

Fig. 1 – Submarine Liassic basalt flows in pillow lava facies, along DN 7 road, between Cuiaş and Toc.

Fig. 2 – Fine (lapillic) pyroclastics also including large basaltic blocks, crossed by calcite and zeolite veinlets. DN 7 road, between Cuiaş and Toc.

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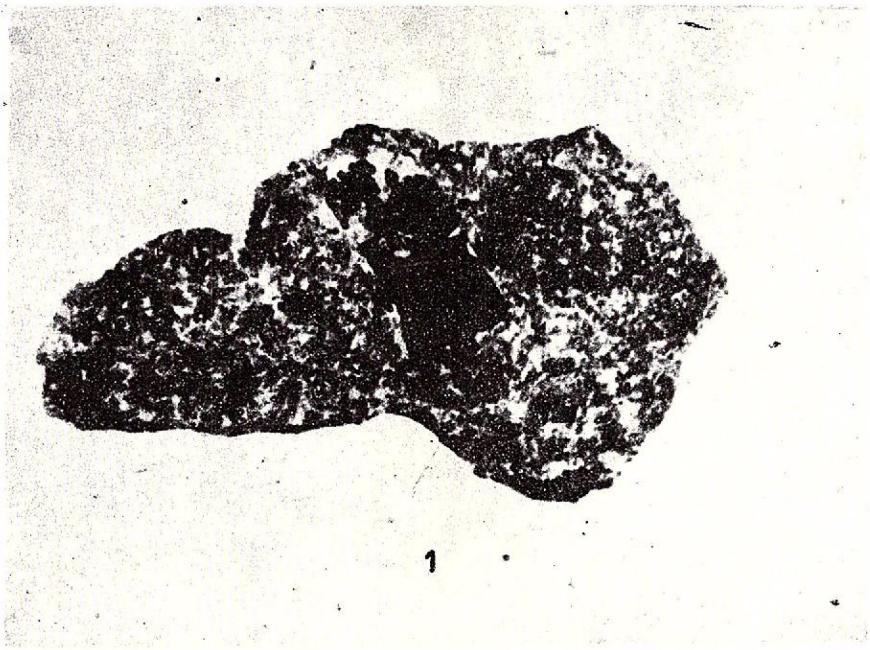

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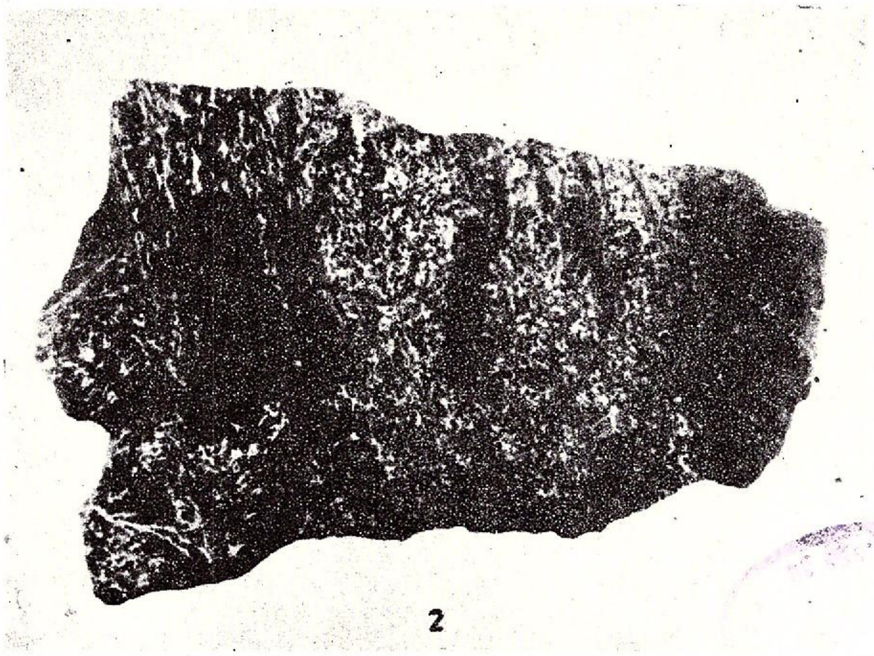

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Plate III

Fig. 1 – Pyroxene (black) megacrystal exhibiting poikilitic structure, within olivine gabbro. Toc Brook.
Fig. 2 – Flasergabbro exhibiting hornblendite bands (black). Poiana Brook.

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PETROLOGICAL AND GEOCHEMICAL FEATURES OF BANATITIC MAGMATITES IN NORTHERN APUSENI MOUNTAINS

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Key words: Banatite. Major elements. Minor elements. Magmas. Calc-alkaline composition. Apuseni Mts – Neocretaceous-Paleogene magmatites.

Abstract: A complex geo-structural unit (Inner Dacides) is crossed or overlain, in Northern Apuseni Mts, by banatitic calc-alkaline magmatites post Lower Maastrichtian-Paleogene in age. The evolution of banatitic magmatism implies two cycles: a first cycle, marked by the emplacement of volcanics (andesites, dacites and rhyolites) and a second cycle, characterized by the emplacement of plutonic and hypabyssal rocks of dioritic, granodioritic and granitic composition, ending with their alkaline differentiates dykes. Basaltic andesites, basalts and lamprophyres dykes, of different origin, partly contemporaneous with the mentioned alkaline differentiates end the magmatic activity in the Northern Apuseni Mts. The geochemical study of banatitic magmatites from Northern Apuseni Mts is based on both new analyses (chemical, emission and gamma-spectrometry, X-ray fluorescence, neutron-activation for REE) and published data; some isotope data ($^{87}Sr/^{86}Sr$ and K-Ar) recently obtained are also presented. The major and minor contents and distribution, REE inclusively, compared to published data and graphically represented on diagrams, show the consanguinity of volcanics and plutonic and hypabyssal magmatites, better arguing the concept concerning the evolution of banatitic magmatism in the Northern Apuseni Mts. The mineralogic-petrographic and geochemical data point out the genesis of banatitic calc-alkaline magma by oceanic crust consumption consequently to a subduction process; the subsequent evolution of magma was complex, through differentiation processes, in intermediate magmatic chambers sometimes, concomitantly with a certain assimilation of rocks from the sialic crust.

Introduction

The present study is based on both lately (these last two decades) published data and new ones (Mureşan, 1971; Lazăr et al., 1972; Istrate, Bratosin, 1976; Istrate, 1978; Ştefan, 1980; Udubaşa et al., 1980; Istrate, Udubaşa, 1981; Ştefan et al., 1985, 1986, 1988; Ştefan et al., 1989, unpubl. report).

The microscopic and chemical analyses, by emission and gamma spectrometry as well as X-ray fluorescence or thermal neutron activation, the non-destructive variant, in order to determine the REE amount, were carried out by the authors on the same samples (30–32 in number) selected from rocks as fresh as possible and representative petrotypes (Pl. I).

The northern part of the Apuseni Mts, where the banatitic calc-alkaline magmatites are present, is a complex geostructural assemblage (Săndulescu, 1984) including several units and groups of units.These structural units consist of crystalline formations of different lithologic features and metamorphic grades, as well as/or Permian sedimentary and igneous formations and Mesozoic pre-Gossau sedimentary deposits. The lithologic features of these Permian-Mesozoic formations are much varied, including basic to acid igneous rocks and detrital or carbonate sedimentary ones.

The nappes from the Northern Apuseni Mts are transgressively overline by Senonian post-tectonic rocks, mostly in Gossau facies; they were usually eroded and preserved only in graben structures (Remeţi, Someşul Cald-Vlădeasa) or as erosion outliers.

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All the mentioned formations are pierced or overlain by the products of the banatitic magmatism which sometimes resulted in strong magmatic contact phenomena or other times in ore deposits, mainly associated with deep-seated igneous bodies.

The younger, Tertiary, epicontinental deposits overlie or contain fragments of banatitic igneous rocks.

Distribution and Emplacement Sequence of Banatitic Magmatites

The banatitic calc-alkaline magmatism in Northern Apuseni Mts had two cycles. The first volcanic cycle starts with andesites and continues with dacites and rhyolites (Pl. I) which constitute lava flows (Vlădeasa, Gilău) associated in places with pyroclastics (Vlădeasa, Mezeş, Mureş Couloir) and shallow subvolcanic bodies (Vlădeasa, Gilău and Bihor).

The pyroxene and hornblende quartz andesites or the amphibole±biotite andesites occur in the Gilău and especially Vlădeasa Mts, where they are located at the periphery of the igneous massif as rooted, linear bodies associated with lava flows and lacking in pyroclastics. These rocks overlie the Upper Coniacian-Lower Maastrichtian sedimentary formations (Dragoş, 1971) and are crossed by or included in younger dacites and rhyolites.

Dacites are largely developed in the north-eastern part of the Vlădeasa massif, from where they are reported as Lunca and Vişag dacites (Giuşcă, 1950), constituting mainly a shallow intrusive nappe (Ştefan, 1969; Giuşcă et al., 1969). The Mărgăuţa dacites belonging to the dyke located south-east of the Vlădeasa massif (Ştefan, 1969) differ from the other varieties by their structural features.

The first cycle volcanic activity continues with the rhyolites developed in the Vlădeasa volcano-plutonic massif (Giuşcă et al., 1969), these rocks occurring less frequently in the Gilău massif and east of the northern boundary of the Mezeş Mts. The rocks assigned to the main Vlădeasa rhyolitic body, often in ignimbritic facies covering an area of ca. 400 km^2, pierce and contain andesites, dacites and two other rhyolite types (Zărna and Ciripa); similar rocks are reported from Măgulicea and Puguiorul hills at Zalău-Moigrad (Ştefan et al., 1986). The vitrophyric rhyolites and the biotite banded rhyolites from the western part of the Vlădeasa massif (Istrate, 1978) make the end of the first cycle of banatitic magmatic activity in Northern Apuseni Mts.

The K-Ar isotopic ages of 61±3 Ma and 61.5±5 Ma obtained on Vlădeasa rhyolite samples collected from Leşului (Bleahu et al., 1984) and Răcadului Valleys do not differ much from the real ages of these rocks. However, the banatitic volcanism ends before the Cuisian as andesite, dacite and rhyolite blocks are contained by the conglomerates occurring at the base of the lower variegated clay complex (Paleocene-Ypresian) at Morlaca.

The second, main, cycle of banatitic magmatism is represented by a wide range of quartz-dioritic and granodiorite-granitic rocks, usually preceded in the crystallisation sequence by their porphyritic varieties which occur at the periphery of plutons and constitute marginal facies or apophyses of the latter.

The diorites, quartz diorites, quartz monzodiorites and their porphyritic varieties, which in the Gilău massif, in Vadului Valley and at Băişoara constitute marginal facies of granodioritic rocks, in Pietroasa, Budureasa and Vlădeasa massifs occur as independent bodies, outside the granodiorite-granitic plutons and as big (sometimes metric) xenoliths within them.

The dacites, granodiorite porphyries and porphyritic microgranodiorites, constituting dykes, sills or apophyses of irregular shapes of some deep-seated granodiorite-granitic bodies are sometimes thermally influenced by the latter and pierce and include quartz andesites-diorite porphyries-porphyritic microdiorites, previously consolidated. The granodiorite-granites, with which the main ore deposits are associated, represent the main mass of the banatitic bodies in Northern Apuseni Mts, even if they are exposed on reduced areas. The rhyodacite, microgranitic rhyolite, granophyre and porphyritic microgranite subvolcanic bodies, which constitute dykes or irregular bodies in the Vlădeasa, Bihor and Mezeş Mts are assigned to the same magmatic event.

The dioritic and granodiorite-granitic rocks pierce both the Lower Maastrichtian sedimentary deposit and the first cycle volcanics (Vlădeasa) and are unconformably overlain by the variegated clays of the lower complex situated on the eastern border of the Gilău massif (Lazăr et al., 1972; Ştefan et al., 1985).

The second cycle ends with alkaline differentiates represented by aplites, microgranitic rhyolites, micropegmatitic rhyolites and porphyritic microgranites which constitute dykes, some times several hundred meters long.

The consolidation of granodiorite-granitic magma was followed by the emplacement, along deep fractures, of basaltic andesites, basalts and lamprophyre dykes generated by a different and deeper magmatic source, partly

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concomitantly with the above mentioned alkaline differentiates and the postmagmatic solutions associated with the granodiorite-granitic main mass.

Mineralogic and Petrogenetic Features of Banatitic Magmatites

The mineralogic composition and the physiographic features of volcanic and plutonic petrotypes are much similar even if they occur in different massifs.

Therefore, the mineralogic composition of andesites is simple: pyroxenes, hornblende, three plagioclase generations and quartz within pilotaxitic or micropoikilitic groundmass. Secondary minerals as albite, epidote, chlorite and silicifiations are present, too.

Table 1

Distribution of opaque minerals in the Northern Apuseni banatitic rocks

No	Sample	%	Mt	Ili	Ilc	Hmt	Mgh	Ru	Lex	Sph	py	Kpy	Gt	Lep	Au
1	9281	-	-	-	-	-	-	-	-	-	(+)	-	-	-	-
2	9329	0.1	-	-	-	-	-	-	-	-	+	-	-	-	-
3	9317	1-2	+	(+)	-	+	+	-	-	-	(+)	-	-	-	?
4	9316	0.6	*	*	(+)	+	-	+	-	-	(+)	(+)	-	-	-
5	9340	0.5	*	-		(+)	(+)	(+)	-	(+)	(+)	-	-	-	-
6	9312	0.2	(+)	(+)	-	*	-	-	-	-	+	(+)	(+)	-	-
7	9314	3.0	(+)	-	-	-	-	-	-	-	*	-	-	-	-
8	9342	0.7	*	-	-	-	(+)	-	-	-	+	(+)	-	-	-
9	9332	0.3	(+)	(+)	-	(+)	-	-	-	-	+	-	(+)	-	-
10	9319	0.5	*	*	(+)	(+)	(+)	*	-	?	+	(+)	-	-	-
11	9333	0.5	+	(+)	-	(+)	(+)	-	-	-	+	-	-	-	-
12	9515	1.0	*	-	-	(+)	-	-	-	-	(+)		-	-	-
13	9318	1.5	*	(+)	-	-	-	-	-	-	(+)	-	-	-	-
14	9330	0.1	+	-	-	-	-	-	-	-	+	(+)	(+)	-	-
15	9356[1]	0.5-1	*	-	(+)	(+)	(+)	(+)	-	-	(+)	?	-	-	-
16	9323	0.3-0.7	(+)	-	-	-	-	-	-	-	*	+	-	-	-
17	9335	0.73-2.0	*	*	+	+	-	-	-	-		-	+	-	-
18	9339[2]	0.44	*	*	+	(+)	-	-	-	-	+	(+)	-	-	-
19	9313[3]	2.5--3	*	*	(+)	(+)	-	(+)	-	(+)	(+)		-	-	-
20	9320	0.15-0.3	*	*	(+)	+	-	(+)	-	-	(+)	(+)	-	-	-
21	9321	1.0	*	+	(+)	-	-	-	-	-	(+)	(+)	+	-	?
22	9324[4]	1.0	*	*	(+)	+	-	+	(+)	-	+	-	-	-	-
23	9326	2-2.5	*	*	(+)	-	-	+	-	-	+	-	-	-	-
24	9345	1.34	*	*	(+)	+	-	+	-	-	+	(+)	-	-	-
25	9327	0.1	+	-	-	(+)	-	(+)	-	-	*	(+)	-	-	-
26	9339A	3.0	*	*	-	*	-	+	-	(+)	+	-	-	-	-
27	9280	1.0	*	(+)	-	-	-	+	-	-	(+)	(+)	(+)	-	?
28	9328[5]	0.5-0.8	*	*	(+)	-	-	+	-	(+)	(+)	+	-	-	-
29	9282	2.5-5	*	+	-	-	(+)	-	-	-	+	(+)	-	-	-
30	9331[6]	3.0	*	+	(+)	*	*	(+)	(+)	?	+	-	(+)	(+)	-
31	9341[7]	4.57	*	*	+	*	-	(+)	(+)	-	+	-	(+)	-	-
32	9334	2-2.5	*	*	*	*	*	-	-	-	(+)	(+)	(+)	-	-

*: very frequent; +: common; (+): sparse; -: lacking; Mt-magnetite; Ili-independent ilmenite; Ilc-composite ilmenite in Mt; Hmt-hematite; Mgh-maghemite; Ru-rutile; Lex-leucoxene; Sph-titanite; Py-pyrite; Kpy-chalcopyrite; GT-goethite; Lep-lepidocrocite; Au-gold

1, unstabilized ilmenite→rutile; 2, sometimes two-phase sulfides, unstabilized ilmenite; 3, highly unstabilized ilmenite; 4, highly unstabilized ilmenite→Ru+Hmt+Lex; 5, ilmenite highly unstabilized on borders; 6, wholly unstabilized ilmenite; 7, unstabilized ilmenite

For type rock and location see Table 3.

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Table

Modal composition of some Northern

No	Sample	Mineralogical composition %							
		Qz	Or	Pl	Hb	Bi	OM	Ap	Zr
1	9316	32.49	32.39	31.93		2.36	0.49	0.07	0.05
2	9319	28.84	31.67	31.90		7.09			
3	$9319'$	31.93	27.16	32.85		6.65			
4	$9319''$	28.64	22.84	43.13		5.00	0.33	0.02	
5	9335	20.91	27.18	44.57	3.79	1.86	0.73	0.33	
6	$9335'$	24.88	22.93	40.12	3.18	6.69	1.47	0.22	0.04
7	$9335''$	24.72	34.55	32.78	0.62	5.36	1.11	0.01	
8	$9335'''$	26.12	29.13	36.14	4.86	1.86	1.17	0.07	
9	9335^{IV}	16.16	27.98	40.22	7.80	5.31	1.99	0.01	0.01
10	9336	30.79	42.00	20.16	1.88	3.98	0.79	0.21	0.09
11	9339	22.19	21.39	46.67	2.66	2.90	0.44	0.05	
12	9320	27.39	13.30	39.20	7.19	12.20	0.45	0.14	0.09
13	$9320'$	22.68	17.63	38.20	6.51	14.17			
14	$9320''$	19.39	14.26	55.28	3.88	6.77			
15	$9320'''$	11.69	2.34	53.75	2.08	27.86	1.56	0.72	
16	9320^{IV}	21.74	16.11	48.79	5.05	7.82	0.16		
17	9321	25.98	25.38	43.98	2.04	1.54	0.53	0.06	0.07
18	9345	22.14	17.11	48.25	2.50	8.04	1.34	0.22	0.03
19	9444	22.14	19.95	43.35	7.49	6.53	0.30	0.04	
20	$9344'$	23.14	22.93	42.17	5.38	5.25			
21	$9344''$	21.54	22.08	47.80	0.13	6.08	0.87	0.36	0.01
22	9355	18.26	22.76	46.85	3.46	6.80	1.10	0.16	0.02
23	9327	27.50	16.53	41.50	1.67	12.06		0.31	0.04
24	$9339'$	14.58	20.58	48.05	4.08	0.28	3.05	0.23	0.02
25	9331	5.42	2.91	65.07		8.87	2.91	0.72	
26	9341	4.23	4.18	73.61	2.84	8.49	4.57	0.04	0.07

Rock type and location: 1, granite, Drăganului Valley; 2-4, microgranites, Drăganului Valley; 5-10, granites, Sebişel Valley; 11, granodiorite, Pietroasa; 12-16, granodiorites, Drăganului Valley; 17, porphyritic microgranodiorite, Valea Lungii; 18-22, granodiorites, Valea Seacă; 23, granodiorite, Ierţii Valley-Băişoara; 24, quartz microgranodiorite, Pietroasa; 25, quartz diorite, west Stîna de Vale; 26, diorite, Zăpozii Valley

The analysed andesite sample (19-9313, Tabs. 1, 2, 3) exhibits opacitized amphiboles, corroded quartz and albite and sericite pseudomorphosed plagioclases. The accessory minerals are represented by apatite and opaque minerals, partly resulting from amphibole alteration. The microcrystalline, sometimes micropoikilitic groundmass is crossed by fissures filled with calcite, chlorite and quartz.

Just like the other andesitic rocks, the hornblende and biotite quartz andesite collected from Cerhat quarry, at Dumbrava (27-9280) is not fresh enough.

The dacites (Lunca, Vişag and Mărgăuţa type) consist of phenocrysts (40-45 %) of plagioclase, quartz, hornblende and biotite, and sometimes the first two varieties also contain pyroxenes, which together with the accessory minerals (apatite, zircon and opaque minerals) are included in a vitreous to microcrystalline groundmass that often corrodes the phenocrysts.

The analysed Mărgăuţa dacite (22-9324) consists of twinned and very slightly zoned (slightly sericitized) plagioclase, quartz, green hornblende and partly chloritized and epidotized biotite.

The Vişag dacite (23-9326) is fresh and exhibits microcrystalline-grainy groundmass with phenocrysts of twinned and zoned plagioclases, corroded quartz, green hornblende, pyroxenes and biotite. The rhombic pyroxenes are surrounded by monoclinal pyroxene rims, in turn corroded by biotite. Apatite and oxides are accessory minerals (Tab. 1).

The Zărna rhyolites (3-9317), very poor in phenocrysts (ca. 5 %), of characteristic banded aspect, are recrystallized and biotite seems partly a secondary mineral.

The Ciripa rhyolites, which pass to granitic facies in places, with 5-10 % phenocrysts, are usually hydrothermalized (sericitized and pyrite impregnated). The sample collected from Moara Dracului Valley (6-9312)

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Apuseni banatitic rocks

Sph	Ep	Chl	Cc	Acc	Fl	Q	A	P	CI
0.17						33.59	33.45	32.98	3.15
		0.67		0.49		34.52	34.27	31.20	7.58
				0.71		34.72	29.54	35.72	8.03
						30.27	24.14	45.58	5.35
0.01	0.21	0.37				22.56	29.33	48.00	7.50
0.08	0.08	0.15	0.12			28.29	26.07	45.62	12.03
	0.07	0.73				26.85	37.53	35.61	7.90
0.11		0.51				28.58	31.87	39.54	8.58
0.05	0.13	0.02	0.29			19.16	33.15	47.68	15.60
0.06						3.12	45.18	21.68	7.01
0.31	0.57	2.69			0.07	24.58	23.70	51.71	9.69
						34.28	16.64	49.06	20.07
		0.61		0.15		28.88	22.45	48.65	21.44
				0.39		21.80	16.03	62.16	11.04
						17.25	3.42	79.32	32.22
						25.09	18.59	56.31	13.33
	0.06		0.32			27.24	26.62	46.12	4.62
					0.34	25.30	19.55	55.14	12.13
0.03		0.11				25.91	23.34	50.73	14.50
				1.09		26.22	25.98	47.79	11.72
0.05		0.65	0.12		0.24	23.56	24.15	52.28	8.15
0.18		0.36				20.78	25.90	53.31	12.08
0.07	0.08	0.11				32.15	19.32	48.52	14.42
2.00	1.71	5.35				17.52	24.73	54.74	15.01
0.63		6.42	0.60			7.38	3.97	88.65	20.15
		0.92				5.15	5.09	89.74	16.93

Abbreviations: Qz-quartz; Or-potash feldspar; Pl-plagioclase; Hb-hornblende; Bi-biotite; OM-opaque minerals; Ap-apatite; Zr-zircon; Sph-titanite; Ep-epidote; Chl-chlorite; Cc-calcite; Acc-unseparated accesories; Fl-fluorine; Q-quartz; A-alkali feldspar; P-plagioclase; CI-colour index.

consists of broken, corroded quartz phenocrysts, sericitized plagioclase feldspar, opacitized and/or chloritized biotite, opaque mineral grains (Tab. 1) especially at the expense of biotite, and on fissures. The quartz-feldspathic groundmass is microcrystalline or granophyric, being invaded by calcite or sericite.

The Vlădeasa rhyolites, exhibiting massive, fluidal and/or eutaxitic or vitroclastic structures and porphyritic textures have identical mineralogic and chemical composition, as well as similar physiographic features and size of crystals. The quartzite, sandstone, clay, limestone, andesite and dacite xenoliths often represent 5 % of the rock. The usually angular phenocrysts (5-25 %) are represented by quartz, plagioclase feldspar, orthoclase and very rarely by sanidine and biotite. Apatite, magnetite, zircon, orthite and rarely titanite are accessory minerals. Actinote, epidote, chlorite, calcite, albite, sericite and zeolites are abundant and result from metasomatic and hydrothermal processes. The usually micro- or cryptocrystalline, partly vitreous, groundmass exhibiting unhomogeneities due to the xenoliths and the crystallisation grade has characteristic features which define the Vlădeasa rhyolite facies. Out of the analysed Vlădeasa rhyolites, sample 7-9314 yields, besides the mentioned xenoliths, autoenclaves of the same eutaxitic rhyolite type or of pumice. Carbonate xenoliths are almost completely substituted by mineral neoformations of the epidote group.

The vitrophyric rhyolites from the west of the Vlădeasa massif (Istrate, 1978) contain high quartz, oligoclase and sanidine amounts and less biotite; hornblende and augite occurrences are sparse. Sample 11-9333 with slightly perthitic potash feldspar seems not to belong to the Vlădeasa rhyolites, as Istrate (1978) had assumed.

The banded biotite rhyolites (9-9332) in the Meziad Valley, often exhibiting fluidal structure, resemble the vitrophyric ones and contain fragments of Vlădeasa rhyolites (Istrate, 1978).

The diorites and quartz diorites (30-9331, 31-9341) are often strongly altered (Tabs. 1, 2). The quartz micromonzodiorite sample (26-9339-A), collected from the metric xenoliths contained by the granodiorites in the Pietroasa quarry, is also strongly altered hydrometasomatically (Tabs. 1, 2).

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The rhyodacites of the Valea Fagului body (Udubaşa et al., 1980) are leucocratic rocks consisting of finely crystallized groundmass and crystals of orthoclase, oligoclase and frequently corroded quartz. Biotite occurs sparsely, being partly chloritized (pennine) or epidotized. Opaque minerals, zircon, apatite and allanite are accessory minerals. Oligoclase is often illitized, illite is associated with calcite, epidote and adularia, while orthoclase is relatively fresh. Quartz, calcite, sparsely galena and epidote occur on fissures within the groundmass.

The samples collected by us (8-9342,15-9356) have yielded, besides the already mentioned minerals, amphibole crystals replaced by chlorite, while the groundmass exhibits granophyric textures.

The dacite-porphyritic microgranodiorite with porphyritic microdiorite xenoliths collected from the Valea Fetii quarry (28-9328) is relatively fresh, consisting of plagioclase, augite, hornblende and biotite phenocrysts contained by the groundmass built up of quartz, plagioclase feldspar and potash feldspar. Scarce calcite, albite, sericite, chlorite and argillaceous minerals occur subordinately. Apatite and opaque minerals are accessory (Tab. 1).

The porphyritic microgranodiorites in the Valea Lungii are usually fresh rocks with microgranular groundmass (Tabs. 1, 2; sample 21-9321).

The granodiorites occur in several places in the Northern Apuseni Mts and our samples have been collected from Băişoara, the Ierţii Valley (25-9327), the Drăganului Valley (20-9320), the Pietroasa quarry (18-9339) and Valea Seacă (24-9345). The different mineralogic composition of granodiorites is inferred from Tables 1 and 2 which include chemically analysed samples or only microscopically studied ones.

The rhyolites tending to micrographic rhyolites from Corniţel-Borod area (sample 2-9329) are characterized by a reduced amount of phenocrysts (quartz, oligoclase and biotite) and a groundmass with secondary quartz nests or radially arranged quartzine. Potash feldspar pierces oligoclase along the fissures. The opaque minerals constitute a fine powder within the rock (Tab. 1).

The granophyres (13-9318) collected from the Drăganului Valley have a typically granophyric groundmass including quartz, potash feldspar and sparse biotite. Among phenocrysts (amounting to ca. 5 % of the rock volume), plagioclase, potash feldspar and biotite are noted; hornblende occurs accidentally. The sample collected by us contains rhyolite and dacite xenoliths.

The microgranite in the Drăganului Valley (1-9319) consists of potash feldspar, plagioclase and quartz, associated with biotite and accessory minerals: zircon, apatite, opaque minerals and titanite, in places (Tabs. 1, 2). Hornfelses bearing andalusite, corundum, hercinite and biotite are formed at the expense of the xenoliths contained by these rocks.

The mineralogic composition of the granitic rocks assigned to the body located in the Drăganului Valley (4-9316) as well as to the granitic body in the Sebişel Valley (Pietroasa massif) (17-9335) is presented in Tables 1 and 2.

The rhyolites in the Cetăţeaua Hill (1-9281) near Dumbrava, which are late differentiates of granodiorite-granitic magmas, are massive rocks, often banded, with voids in places. The rock consists of a sometimes eutaxitic groundmass (90 %), uniformly cryptocrystalline or with spherulitic quartz-feldspathic aggregates other times. The fresh potash feldspar phenocrysts are often Karlsbad or seldom Baveno twinned. Plagioclase is usually wholly replaced by cryptocrystalline silica. Biotite is lacking and quartz is highly corroded. Lens-like or rounded voids are filled with opal or chalcedony, while on fissures and in the nests, like in the mentioned sample, the quartz is associated with secondary potash feldspar.

The megaporphyritic microgranitic rhyolites in the Colului Valley (5-9340) are characterized by finely grained quartz-feldspathic groundmass containing large (more than 1 cm) quartz, plagioclase and potash feldspar phenocrysts. Biotite is also present, the accessory minerals being zircon and opaque minerals (Tab. 1). Calcite, sericite and albite occur as secondary minerals.

The pyroxene and olivine andesites, south of the Cetăţeaua Hill (29-9282), consist of plagioclase phenocrysts, monoclinic, rhombic pyroxenes and olivine. Quartz, amphibole and biotite come from the volcanics previously emplaced, as well as some plagioclase crystals highly corroded by the groundmass.

The basalts which pierce the Senonian rocks and the ignimbritic rhyolites from the west of the Vlădeasa massif constitute NW-SE trending dykes, 10-20 m thick and hundreds of meters long. The breccious basalt sample collected from the Ilieşului Brook-Meziad (39-9334) exhibits intersertal texture, with prismatic crystals of albitized and calcitized plagioclase, the interstices of which contain highly chloritized pyroxenes or glass; grains of opaque minerals occur frequently.

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Geochemistry of Banatitic Magmatites in Northern Apuseni Mts

Distribution of Major Elements

The analysed samples (Tab. 3) point to wide range of petrographic types and large diversity of oxide contents with respect to the same petrotypes.

Table 4 present the variation range and the average SiO_2, CaO, Na_2O and K_2O contents characteristic of the Northern Apuseni volcanics by comparison to the data published by Le Maitre (1984) and Taylor (1969) on similar rocks.

If andesites usually yield lower CaO and higher K_2O and SiO_2 contents, the samples analysed by us (Tab. 3) being in fact not fresh enough, the dacites which were first defined from this area (Hauer, Stache, 1863) point to contents similar to those published by Le Maitre (1984) and Taylor (1969); SiO_2 and Na_2O are excepted due to their lower contents in the Northern Apuseni dacites and rhyolites, too.

Discussions on the nomenclature and chemical composition of the igneous rocks from Vlădeasa, Gilău and Mezeş were previously presented (Istrate, 1978; Ştefan, 1980; Ştefan et al., 1985, 1986). The representations on diagrams are illustrated by Figure 1, on which the fields inferred from previous analyses are also presented. In view of their plotting on the TAS diagram the samples were recalculated by eliminating CO_2 and H_2O. There is an appropriate connection between the nomenclature proposed by us (Tab. 3) and the TAS and QAP classifications, except for the andesite samples which are partly altered.

On the QAP-normative diagram (Fig. 1) the distribution field of dioritic rocks previously analysed and of those analysed by us is extremely large due to the high biotite amount contained. This is shown by the plotting of modal values on the diagrams of Figure 1 in view of comparison. The variation range of SiO_2 contents yielded by dioritic rocks is of 58.25-67.00 %, with the upper limit a little higher, also due to the alteration processes which affected these rocks.

Granodiorite rocks and the associated rhyodacites are plotted on the QAP diagram within the appropriate fields based on both chemical and modal analyses on grained rocks (Fig. 1). The variation ranges of oxides are the normal ones of banatitic rocks from Northern Apuseni Mts. By comparison to the average values presented by Taylor (1969), these rocks yield lower SiO_2 average contents (66.90 %/66.19 %) and Na_2O average contents (3.84/3.59 %).

The granitic rocks and their late alkaline differentiates show a wide chemical composition diversity (Tab. 3). The very different subvolcanic rocks (granophyres, microgranites, microgranitic megaporphyritic rhyolites) divided mainly acording to their texture have been only partly analysed chemically. Similarly to the case of Valea Seacă and Valea Fagului rhyodacites, somehow uncertain, to this rock family the more or less microgranitic rhyolites at Corniţel (2-9329) have been added.

On the QAP diagram of Figure 1 the rocks are plotted in the granite field, based on both normative and modal compositions. The Na_2O and K_2O distribution points to high values of these components ($Na_2O+K_2O=7.14$-8.58 %) and in the increase of K_2O content of the last differentiates which cross the previously consolidated granodiorites and granites (5-9340). The CaO contents range from 0.36 to 1.28 % except for sample 17-9335 from the Sebişel Valley yielding higher CaO (2.67 %) and Na_2O (3.74 %) contents and a lower K_2O one (3.80 %), related to the reduced SiO_2 amount (68.37 %).

Among basic rocks, the pyroxene and olivine andesites and the basalts are plotted on the QAP diagram (Fig. 1) within the field of these rocks figured according to previous data.

Minor Element Distribution

Ba, Rb, K (Tabs. 5, 6, 7) are mainly associated into K-feldspars and biotite minerals, well represented in most of the analysed rocks. Within the evolution of the first cycle rock sequence, Ba values increase from andesites (≈ 360 ppm) to dacites (≈ 1100 ppm) and rhyolites (≈ 1600 ppm), following the K increase (Fig. 2A). The average Ba contents of the andesites and dacites analysed by us are similar to the average values presented by Taylor (1969), except for the Mărgăuţa dacite, close to a hydrothermally circulated area. The investigated rhyolites yield a much higher Ba content than classic values.

Rb exhibits the same tendency of average values increase which accompanies the differentiation process from andesites (≈ 90 ppm) to dacites (≈ 120 ppm) and rhyolites (≈ 170 ppm). The average Rb contents of andesitic and dacitic rocks are three times higher than those presented by Taylor (1969) and those of rhyolites are 16 times higher, which accounts for Rb assimilation from the crust. The large distribution field of Ba and Rb contents (Fig. 2A, 2B) yielded mainly by rhyolitic rocks is determined by the diversity of petrographic

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Table

Chemical composition of some Northern

No	Sample	Rock type and location	SiO_2	TiO_2	Al_2O_3
1	9281	Silicified microgranite rhyolite, Cetăţeaua Hill	78.76	0.16	11.25
2	9329	Altered rhyolite, Cornişel	76.41	0.04	11.96
3	9317	Zărna rhyolite, Zărna Valley	75.85	0.16	12.40
4	9316	Granite, Drăganului Valley	75.14	0.26	12.50
5	9340	Megaporphyritic microgranitic rhyolite, Cohului Valley	75.09	0.24	12.20
6	9312	Ciripa rhyolite, Moara Dracului	74.94	0.12	12.94
7	9314	Vlădeasa rhyolite, Voiosului Brook	74.93	0.16	12.00
8	9342	Rhyodacite, Valea Seacă	74.90	0.16	12.45
9	9332	Biotite rhyolite, Meziad Valley	74.79	0.22	13.04
10	9319	Microgranite, Drăganului Valley	73.68	0.24	13.74
11	9333	Vitrophyric rhyolite, Meziad Valley	73.33	0.16	12.40
12	9315	Vlădeasa rhyolite, Drăganului Valley	72.80	0.28	11.95
13	9318	Granophyre, Drăganului Valley	72.76	0.38	13.07
14	9330	Vlădeasa rhyolite, Iadului Valley	72.43	0.20	11.85
15	9356	Rhyodacite, Fagului Valley	70.00	0.44	13.82
16	9323	Vlădeasa rhyolite, Răcadului Valley	69.72	0.32	13.74
17	9335	Granite, Sebişel Valley	68.37	0.60	14.85
18	9339	Granodiorite, Pietroasa	67.93	0.58	14.78
19	9313	Quartz andesite, Voiosului Brook	67.87	0.76	14.15
20	9320	Granodiorite, Drăganului Valley	67.56	0.76	14.55
21	9321	Porphyritic microgranodiorite, Valea Lungii	67.25	0.76	15.25
22	9324	Mărgăuţa dacite, Mărgăuţa Valley	66.90	0.48	14.80
23	9326	Vişag dacite, Vişag	65.90	0.62	15.70
24	9345	Porphyritic granodiorite, Valea Seacă	65.73	0.47	15.45
25	9327	Granodiorite, Băişoara	65.46	0.60	15.69
26	9339A	Hydrothermalized quartz micromonzodiorite, Pietroasa	63.78	0.94	14.95
27	9280	Hornblende and biotite quartz andesite, Dumbrava	63.30	0.48	15.58
28	9328	Porphyritic microgranodiorite-dacite, Valea Fetii	63.20	0.84	15.15
29	9282	Pyroxene and olivine andesite, west Dumbrava	60.99	0.64	16.44
30	9331	Quartz diorite, west Stîna de Vale	57.23	1.06	18.34
31	9341	Quartz diorite, Zăpozii Valley	56.89	1.24	17.99
32	9334	Breccious basalt, Ilieşului Valley	52.61	1.80	15.76

types and the unhomogeneity of rocks assigned to the Vlădeasa petrotype.

The second cycle magmatites point out the same increase of Ba and Rb contents concomitantly with the progress of the differentiation process. The higher Ba values yielded by the two quartz diorite samples are accounted for only by a low hydrothermal supply. The variation of Ba and Rb contents inferred from our analyses (Tabs. 5, 6) and literature data on Ba is presented on the diagrams of Figure 2.

The general characteristic feature pointed out by the Sr distribution is the impoverishment tendency of evolved magmas assigned to both differentiation cycles of banatitic magmatism in the Northern Apuseni Mts (Tab. 6, Fig. 2C).

The average Sr values are similar to those presented by Taylor (1969) for semblable rocks.

Excepting the very high zirconium contents of the two quartz diorite samples, the study of Zr-SiO_2 diagram (Fig. 2D) drawn up according to available data points out constant Zr contents for all the rock types distinguished with respect to the two magmatic cycles but for rhyolites. The wide dispersion of Zr contents reported for rhyolites is due to the assimilation of Zr-rich rocks by the rhyolitic magma. The average Zr contents are similar to those presented by Taylor (1969) for rocks of the same composition, excepting the rhyolites in the Apuseni Mts which yield higher contents.

The siderophyle elements Ni, Co, Cr, V and Sc, usually present in the ferromagnesian minerals, are placed within the normal limits of the calc-alkaline rock sequences, but for Co, V and Sc contents of andesites which are lower for our samples.

Although the U and Th distribution (Tab. 7), which marks the differentiation degree, shows a positive correlation between U and Th, the characteristic plots are widely spread, a feature also proving the assimilated crustal matter (Fig. 2E). The correlation and dispersion of U-K (Fig. 2F) and Th-K (Fig. 2G) shown by these

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Apuseni banatitic rocks

Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O	CO_2	S	Total
0.43	0.03	0.02	0.10	0.50	2.67	4.47	0.04	1.07		0.25	99.83
1.03	0.08	0.02	0.12	0.36	2.60	5.98	0.10	0.86		0.16	99.72
1.23	0.12	0.04	0.13	0.61	1.82	6.70	0.06	0.45		0.25	99.82
1.49	0.09	0.06	0.56	1.08	3.56	4.53	0.06	0.29		0.26	99.91
1.70		0.05	0.40	0.97	3.15	5.00	0.12	0.29	0.44	0.20	99.85
1.77	0.14	0.05	0.22	0.64	2.91	5.30	0.06	0.42		0.27	99.78
1.95	0.02	0.08	0.54	2.10	2.48	4.24		0.69		0.91	100.10
1.20		0.07	0.44	0.98	2.71	4.72	0.04	1.28	0.68	0.12	99.75
1.18	0.04	0.05	0.28	0.75	3.60	5.04	0.06	0.80		0.16	100.01
1.20	0.90	0.05	0.35	1.00	4.00	3.99	0.06	0.37		0.30	99.88
1.01		0.05	0.22	0.75	3.70	4.69	0.04	3.37		0.17	99.89
2.30	0.27	0.05	0.59	2.75	3.84	4.01	0.08	0.30	0.13	0.28	99.63
2.41	0.69	0.08	0.50	1.28	3.90	4.13	0.08	0.30		0.26	99.84
1.68	0.59	0.04	0.48	2.35	3.21	4.93	0.04	1.43	1.18	10.15	99.76
2.73	0.05	0.09	1.31	2.58	2.91	4.15	0.13	0.96	0.89	0.29	100.35
2.21	0.80	0.09	0.57	1.99	3.37	4.05	0.08	1.31	1.21	0.25	99.71
3.00	0.58	0.08	1.23	2.67	3.74	3.81	0.20	0.43		0.14	99.70
3.71	0.09	0.10	1.25	2.79	3.63	3.77	0.20	0.75		0.15	99.73
3.48	0.62	0.12	0.72	3.75	3.82	2.68	0.13	0.95	0.47	0.30	99.82
3.65	0.50	0.08	1.57	3.62	3.80	2.73	0.20	0.45		0.19	99.66
3.28	0.45	0.06	1.47	3.50	3.96	2.93	0.14	0.83		0.21	100.09
3.64	0.31	0.09	1.74	2.77	3.70	3.33	0.14	1.15	0.56	0.21	99.82
3.96	10.51	0.06	1.87	4.10	4.00	2.46	0.20	0.31		0.20	99.91
3.17	1.06	0.09	2.23	3.72	3.42	3.30	0.15	0.72		0.26	99.91
4.31	0.13	0.11	1.88	4.35	3.13	2.72	0.20	0.90		0.21	99.77
4.23	0.99	0.12	1.42	3.64	3.83	3.74	0.32	1.69		0.15	99.67
2.96	0.94	0.07	2.08	3.85	3.38	2.10	0.20	4.53	0.30	0.26	99.80
4.54	0.69	0.18	2.45	3.56	3.56	2.52	0.24	2.91		0.20	100.03
2.56	2.45	0.10	3.62	5.82	3.92	1.21	0.16	1.52		0.25	100.04
4.49	1.83	0.11	2.26	5.35	3.91	2.08	0.44	1.49	0.95	0.15	99.68
5.20	1.95	0.14	2.64	6.05	3.94	2.00	0.54	0.93		0.23	99.69
8.28	2.12	0.19	3.08	6.87	3.67	1.50	0.54	3.38		0.12	99.74

Table 4

Average SiO_2, CaO, Na_2O and K_2O contents of banatitic volcanics in Northern Apuseni Mts and classic values

Rock type	Oxides	Minimun value	Maximum value	Average value	Average values Le Maitre (1984)	Taylor (1969)
Andesites	SiO_2	58.50	67.87	60.70	59.94	59.50
	CaO	3.04	7.31	5.36	6.56	7.03
	Na_2O	1.68	4.40	3.43	3.46	3.68
	K_2O	0.60	3.20	2.05	1.69	1.60
Dacites	SiO_2	63.32	67.17	64.92	66.17	65.10
	CaO	2.77	5.04	4.12	4.19	4.75
	Na_2O	1.61	4.89	3.76	3.82	4.60
	K_2O	1.55	4.52	2.96	2.39	3.25
Rhyolites	SiO_2	64.40	76.46	68.40	74.26	75.30
	CaO	0.31	3.78	1.42	1.14	1.49
	Na_2O	0.60	5.21	2.93	3.71	4.12
	K_2O	2.19	6.70	3.86	4.26	3.39

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Table 5

Minor elements (ppm) in Northern Apuseni banatitic rocks

No	Sample	Pb	Cu	Zn	Sn	Ga	Ni	Co	Cr	V	Sc	Nb	Zr	Yb	Y	La	Sr	Ba
1	9281	17	5	<30	3.5	14	2.5	<2	8	6	3	<10	145	2.4	19	44	40	1400
2	9329	15	3	52	<2	8.5	<2	<2	2	<2	2.5	17	42	1.8	20	<30	[illegible]	1500
3	9317	24	<2	38	4	15	<2	<2	<2	<2	5.5	<10	120	3	29	38	47	1300
4	9316	18	<2	<30	<2	14	<2	<2	3	10	3.5	<10	140	3.4	31	38	125	530
5	9340	18	7	42	<2	11	<2	<2	2.5	5	3.5	<10	150	2	25	62	130	480
6	9312	48	5	87	3	15	<2	<2	2	<2	9	<10	150	4.5	46	53	67	1450
7	9314	20	5.5	55	5.5	16	3	<2	4	3	9.5	10	220	4.2	42	50	150	1400
8	9342	18	3	44	<2	11	4	<2	5	3	5	<10	145	3.4	42	53	145	1350
9	9332	22	<2	48	<2	11	3	<2	6.5	2	4.5	<10	140	3.2	38	53	58	1550
10	9319	25	<2	52	4	16	<2	<2	3	3	7	10	260	3.7	35	46	100	950
11	9333	24	6.5	49	3	13	<2	<2	3	<2	5	<10	145	3.3	35	53	53	1300
12	9315	32	4.5	93	4	15	3	<2	6	5	8	<10	270	3.1	35	53	75	1400
13	9318	17	3.5	70	2	16	4	<2	5	4.5	13	14	370	4.6	48	90	75	2100
14	9330	27	4	78	4.5	15	5	<2	7.5	5.5	9.5	<10	210	3.4	38	62	115	2100
15	9356	17	22	65	<2	13	3	<2	4.5	21	7	<10	280	3.1	34	40	300	1500
16	9323	25	17	64	5	15	6	3.5	10	16	13	<10	310	4	40	78	100	2100
17	9335	14	3	48	<2	13	5.5	5	10	53	11	<10	170	2.9	37	56	140	420
18	9339	13	3	56	<2	12	7.5	6	11	55	13	<10	220	3.4	42	63	170	530
19	9313	10	6	64	<2	12	9.5	6.5	44	62	12	<10	210	3	30	50	100	340
20	9320	5	2	46	<2	11	10	7.5	16	78	12	<10	230	3.2	31	44	190	430
21	9321	8.5	2	44	<2	13	9.5	6	14	57	10	<10	165	2.5	30	40	150	420
22	9324	15	13	57	<2	14	7	4	9.5	50	8	<10	155	2.5	26	38	800	2100
23	9326	12	10	64	<2	13	14	8.5	17	67	9.5	<10	130	2.4	29	35	700	1100
24	9345	6	9	56	<2	11	10	9.5	40	78	15	<10	180	2.1	28	34	200	410
25	9327	8.5	16	70	3.5	15	10	7	14	65	12	<10	115	2.1	21	50	270	500
26	9339A	14	7.5	60	<2	14	5.5	8	3.5	85	15	<10	270	2.3	56	65	210	440
27	9280	7.5	10	40	<2	10	10	9	8.5	105	9	18	135	1.7	18	37	210	380
28	9328	6	3	58	<2	12	11	8	11	85	12	<10	140	2.3	24	50	380	460
29	9282	7	16	70	<2	14	47	13	13.5	80	15	<10	110	1.9	22	33	220	340
30	9331	13	3	52	<2	25	7	9	16	115	15	<10	720	3.1	39	50	300	1300
31	9341	38	8	130	<2	16	7	8.5	13	110	17	<10	930	2.5	40	37	300	900
32	9334	18	5	63	<2	19	4	17	6	170	30	<10	260	3.2	53	40	260	300

Table 6

Minor elements (ppm) in Northern Apuseni banatitic rocks determined by non-dispersive XRF spectrometry

No	Sample	Rb	Sr	Y	Zr	Nb	Rb/Sr	No	Sample	Rb	Sr	Y	Zr	Nb	Rb/Sr
1	9281	167	46.2	21.8	142	20.5	3.61	17	9335	163	260	31.2	222	17.1	0.63
2	9329	217	55.7	26.0	59.7	25.6	3.89	18	9339	168	269	31.1	237	20.8	0.62
3	9317	263	54.2	36.1	116	19.6	4.85	19	9313	97	235	31.1	203	20.9	0.41
4	9316	241	109	35.1	127	21.2	2.21	20	9320	126	238	30.2	163	22.1	0.53
5	9340	234	94.3	21.4	143	19.1	2.48	21	9321	119	236	30.0	174	17.1	0.51
6	9312	196	74	45.4	142	18.8	2.65	22	9324	130	272	23.8	172	16.2	0.48
7	9314	182	126	48.9	175	24.5	1.43	23	9326	110	280	28.4	164	20.5	0.39
8	9342	191	120	46.3	158	18.7	1.59	24	9345	155	362	23.0	153	16.3	0.43
9	9332	224	58.6	37.5	166	22.8	3.82	25	9327	124	360	19.4	125	18.3	0.34
10	9319	192	95.2	40.1	228	21.3	2.02	26	9339A	171	266	32.9	229	17.2	0.64
11	9333	226	62.9	37.3	163	20.3	3.60	27	9280	83.8	322	19.0	148	24.4	0.26
12	9315	161	71.7	42.7	258	20.1	2.25	28	9328	95.5	422	17.2	156	23.1	0.23
13	9318	149	88	46.7	386	24.2	1.69	29	9282	81.3	311	20.4	136	18.5	0.26
14	9330	166	83.5	36.1	266	21.2	1.99	30	9331	73.1	487	29.3	543	15.2	0.15
15	9356	159	212	35.6	297	21.0	0.75	31	9341	103	599	28.9	730	15.2	0.17
16	9323	155	85	40.0	335	20.0	1.84	32	9334	43.9	318	40.5	237	23.5	0.14

diagrams are typical of rocks frequently containing K-feldspar and biotite. The increasing amounts of these minerals from basic to acid rocks influence the U contents which amount to the highest values in rhyolites. The highest Th contents are reported from granitic rocks. The average U contents yielded by the sample analysed

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are two or three times higher than Taylor's (1969) average values. The Th average contents are also higher, especially for andesites and dacites, by comparison to the same author's data.

The chalcophile elements Cu, Pb, Zn usually show lower values than those of banatitic rocks previously analysed.

The REE and other rare elements contents (Tabs. 5, 8) have been chondrite-normalized according to Henderson (1984).

Table 7

U, Th, K contents and Th/U ratio in Northern Apuseni banatitic rocks

No	Sample	U ppm	Th ppm	K %	Th/U	No	Sample	U ppm	Th ppm	K %	Th/U
1	9282	4.1	21.0	3.86	5.12	17	9535	3.2	13.0	3.50	4.06
2	9329	6.6	6.5	5.91	0.98	18	9339	3.9	13.6	3.41	3.49
3	9317	3.9	13.0	5.90	3.33	19	9313	3.3	11.3	2.79	3.42
4	9316	3.6	17.8	4.08	4.94	20	9320	3.7	12.9	2.64	3.49
5	9340	4.3	21.6	4.63	5.02	21	9321	4.7	13.4	2.73	2.85
6	9312	4.0	15.0	5.19	3.75	22	9324	4.1	12.4	2.86	3.02
7	9314	4.8	14.0	4.01	2.92	23	9326	3.7	10.8	2.18	2.93
8	9342	4.5	17.0	4.20	3.77	24	9345	3.6	11.7	3.00	3.25
9	9332	5.2	17.6	4.46	3.39	25	9327	5.4	10.9	2.63	2.02
10	9319	4.7	13.6	3.74	2.89	26	9339A	4.8	14.2	3.66	2.96
11	9333	6.4	15.1	4.12	2.35	27	9280	1.8	11.9	2.05	6.60
12	9315	4.5	14.2	3.68	3.16	28	9328	3.9	11.0	2.50	2.82
13	9318	3.4	16.0	3.73	4.71	29	9282	2.7	9.4	0.84	3.48
14	9330	4.9	14.4	3.77	2.94	30	9331	2.4	7.0	1.72	2.94
15	9356	5.6	12.8	3.51	2.28	31	9341	2.3	5.9	1.53	2.57
16	9323	4.0	13.8	3.83	3.45	32	9334	1.4	6.8	1.37	4.84

Table 8

Analyses of rare earths, Hf, and Th (ppm) in Northern Apuseni banatitic rocks

No	Sample	La	Ce	Sm	Eu	Tb	Hf	Th	No	Sample	La	Ce	Sm	Eu	Tb	Hf	Th
1	9281	26	71	3.5	0.50	1.40	5.5	24.0	16	9335	38	73	4.5	1.12	0.70	6.5	9.8
2	9329	11	28	2.5	0.47	0.91	2.4	11.1	17	9339	34	66	5.3	1.08	0.89	4.5	10.7
3	9317	35	85	9.2	0.93	0.43	5.7	15.3	18	9313	15	51	3.7	0.97	0.84	4.6	9.0
4	9316	33	89	5.4	0.85	1.37	6.1	22.4	19	9320	31	47	3.7	0.84	0.88	2.9	10.8
5	9340	40	75	5.4	0.52	1.40	3.6	21.6	20	9321	25	49	3.3	0.87	0.60	4.7	10.1
6	9312	34	80	8.4	1.07	1.12	4.8	14.2	21	9324	28	61	3.6	0.88	0.61	4.1	12.4
7	9314	29	53	9.4	1.06	0.75	4.4	9.7	22	9326	27	43	2.9	0.83	0.90	2.2	9.2
8	9342	43	69	7.9	1.00	1.02	3.1	15.1	23	9327	30	56	4.7	0.92	0.64	3.0	11.2
9	9332	31	74	6.4	0.95	1.05	4.5	16.4	24	9339A	37	56	6.0	1.34	0.55	5.9	10.2
10	9319	37	74	4.2	1.14	1.07	9.4	14.9	25	9280	25	60	3.0	0.91	0.25	4.6	10.2
11	9333	40	69	4.8	0.89	1.04	4.8	15.2	26	9328	28	34	3.6	0.90	0.43	3.9	9.1
12	9315	48	97	5.7	1.15	1.31	7.5	14.5	27	9282	11	28	3.0	0.63	0.36	1.5	5.2
13	9318	60	109	6.0	1.64	0.88	7.8	13.1	28	9331	22	30	4.9	1.48	0.89	7.8	3.7
14	9330	34	97	7.6	1.04	0.97	4.9	11.3	29	9341	26	32	4.5	1.58	0.81	11.0	4.3
15	9323	62	117	6.9	1.47	1.14	7.2	13.5	30	9334	27	25	5.6	1.28	0.40	2.6	4.4

The data contained by the table and represented on diagrams point to REE enrichment, especially of light ones, by comparison to chondrites; these values are higher than those of similar rocks presented by Taylor (1969). However, the granodiorites from the Apuseni Mts yield lower contents.

The Eu concentrations are three times higher in the case of Northern Apuseni volcanics and seven times lower for granodiorites, respectively, by comparison to Taylor's (1969) data.

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The REE distribution curves of banatitic rocks (Fig. 3) are most often similar, accounting for important light REE differentiation and enrichment and slight or inexistent heavy REE enrichment and differentiation. Usually, excepting the diorites, basalt andesites and basalts, a negative Eu anomaly is noted concomitantly with the decrease of plagioclase content, associated with the evolution of fractional magma crystallisation.

The REE concentration and distribution in andesites and dacites illustrated by the general shape of curves (Fig. 3A), but for the Tb content of sample 9280, are similar accounting for the common origin and the slight composition difference of the two petrographic types. Although the Eu anomaly is less obvious than for the succeeding rocks, it is difficult to account for its presence only by the fractionation of plagioclases, as banatitic rocks older than andesites are not known.

The REE distribution in rhyolitic rocks (Fig. 3B) shows, excepting Zărna petrotype (as regards Tb only), almost similar contents and features. On the whole, one notes a slight increase of REE contents by comparison to andesites and dacites; the LREE/HREE ratio is also slightly increased, maybe due to the fractionation of ferromagnesian minerals (D'Elio et al., 1987). The increase of LREE contents of the rhyolitic rocks glass by comparison to the contained minerals, can also be considered, in accordance with Mahood and Hildreth (1983), when speaking of the Vlădeasa rhyolites. The less obvious Eu anomalies exhibited by some rhyolites particularly assigned to the Vlădeasa main body might be due to the inclusion of plagioclases from penetrated rocks into the rhyolitic melt.

The quartz diorites and monzodiorites exhibit REE distribution only partly similar with that one of andesites or granodiorites. It is to note the lack of Eu anomaly as far as the diorites include most of the plagioclases from the melt. This time, too, LREE contents are manifestly differentiated and more than 50-100 times higher than those of chondrites, while HREE are less abundant and differentiated.

The REE differentiation curves of granodioritic rocks (Fig. 3D) show similar behaviour with those of corresponding volcanic rocks, thus proving the consanguineous character of banatitic magmatites. The plagioclase amount fractionated during diorite crystallisation accounts for the negative Eu anomaly. The LREE contents resemble those of quartz diorites of the same mineralogic composition as granodiorites. The absence of Eu anomaly from Valea Fetii porphyritic micrograndiorite sample (28-9328) suggests that the magma from which the melt was released, generating the mentioned rock, does not show an advanced fractionation of plagioclases. The REE distribution curve of rhyodacites resembles the one of granodiorites, pointing only to well-marked negative Eu anomaly.

The REE differentiation in granitic rocks pointed out by distribution curves (Fig. 3E) is more advanced than in granodiorites. Besides the higher LREE contents one notes well-marked Eu anomalies, generally as the evolution of rocks advances, accounted for by the removal of an increased amount of plagioclases by means of fractional crystallisation. As regards the Cornițel microgranitic rhyolites, this anomaly is caused by squeezing out the melt from the granitic magma during the crystallisation. The less obvious Eu anomaly of granophyres is difficult to account, but one could adopt Toth's (1987) opinion that it is brought about by the oxidation of Eu^{2+} to Eu^{3+}.

The REE distribution curves in basalts and basaltic andesites obviously point out different origins with respect to calc-alkaline banatitic rocks, and the lack of Eu-anomaly proves the absence of a preceding plagioclase fractionation process.

Genesis and Evolution of Magmas

The petrologic features of Upper Cretaceous-Paleogene magmatites and their subsequent position with respect to Alpine orogeny have been presented in several general (Giuşcă, 1950; Giuşcă et al., 1966; Ianovici et al., 1976; Cioflică et al., 1973) or minute studies.

In the past, the lithogenic origin of calc-alkaline banatitic magmas through sialic crust anatexis in the orogen area was sustained (Giuşcă, 1950). Later, by applying the plate tectonics concept to the Romanian territory, the hypothesis that banatitic rocks have been generated by subduction and fusion of oceanic-type crust was emitted (Boccaletti et al., 1973; Rădulescu, Săndulescu, 1973; Herz, Savu, 1974 etc.).

Although the arrangement of banatitic magmatites does not show a time and space succession corresponding to igneous evolution in the known classical subduction areas, the global tectonics cannot be contradicted with respect to the Apuseni Mts. Although the available data on Sr isotopes are not enough, the published ones regarding the granites assigned to the Bihor batholith (Pavelescu et al., 1985) show an initial $^{87}Sr/^{86}Sr$ ratio of 0.708, graphically calculated, which surely points to the contamination of the melt generated by the fusion of basic rocks. The values reported by the laboratories IFIN Bucharest for volcanic rocks samples collected from

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the Vlădeasa massif (Popescu, 1987, unpubl. report) are lower than those for granitic rocks used by Pavelescu et al. (1985) for drawing up the isochron, pointing however to magma contamination. Therefore, the isotope ratio of andesites ranges from 0.7058 to 0.7084, of dacites from 0.7053 to 0.7086 and of rhyolites from 0.7054 to 0.7090, symptomatically higher than the 0.704±0.002 value typical of the upper mantle. On the other hand, the isotope ratio $^{87}Sr/^{86}Sr$ is lower than the one of magmas generated by sialic crust anatexis.

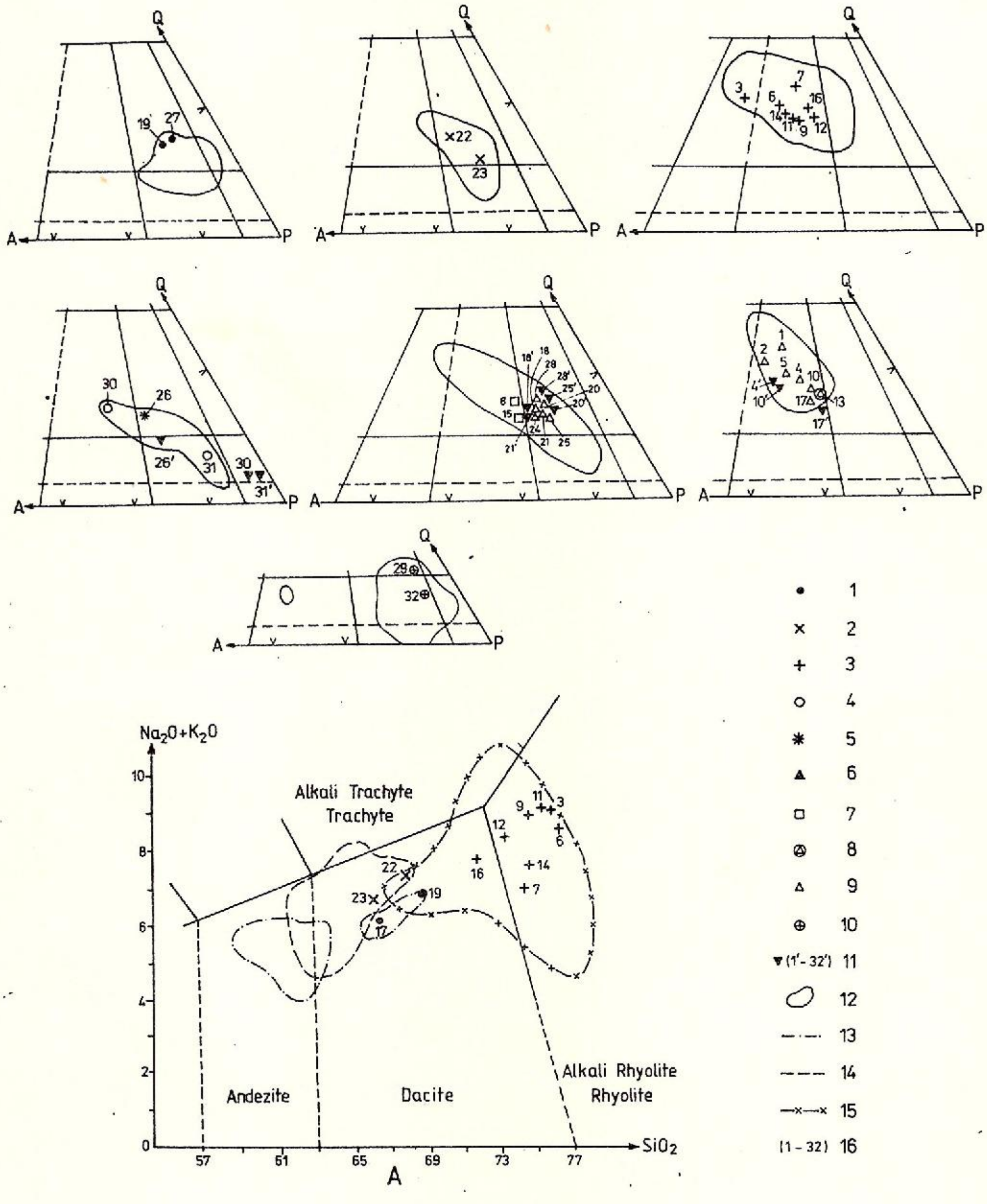

Fig. 1 – QAP and TAS diagrams
1, andesites; 2, dacites; 3, rhyolites; 4, quartz diorites; 5, quartz monzodiorites; 6, granodiorites and their porphyritic varieties; 7, rhyodacites; 8, granophyres; 9, granites and their porphyritic varieties; 10, basaltic andesites and basalts; 11, rocks plotted according to their modal composition (1′-32′); 12, plotting areas of QAP normative diagram: 1A, plotting areas of TAS diagram; 13, andesites; 14, dacites; 15, rhyolites; 16, number of analysed sample (1–32, Table 3).

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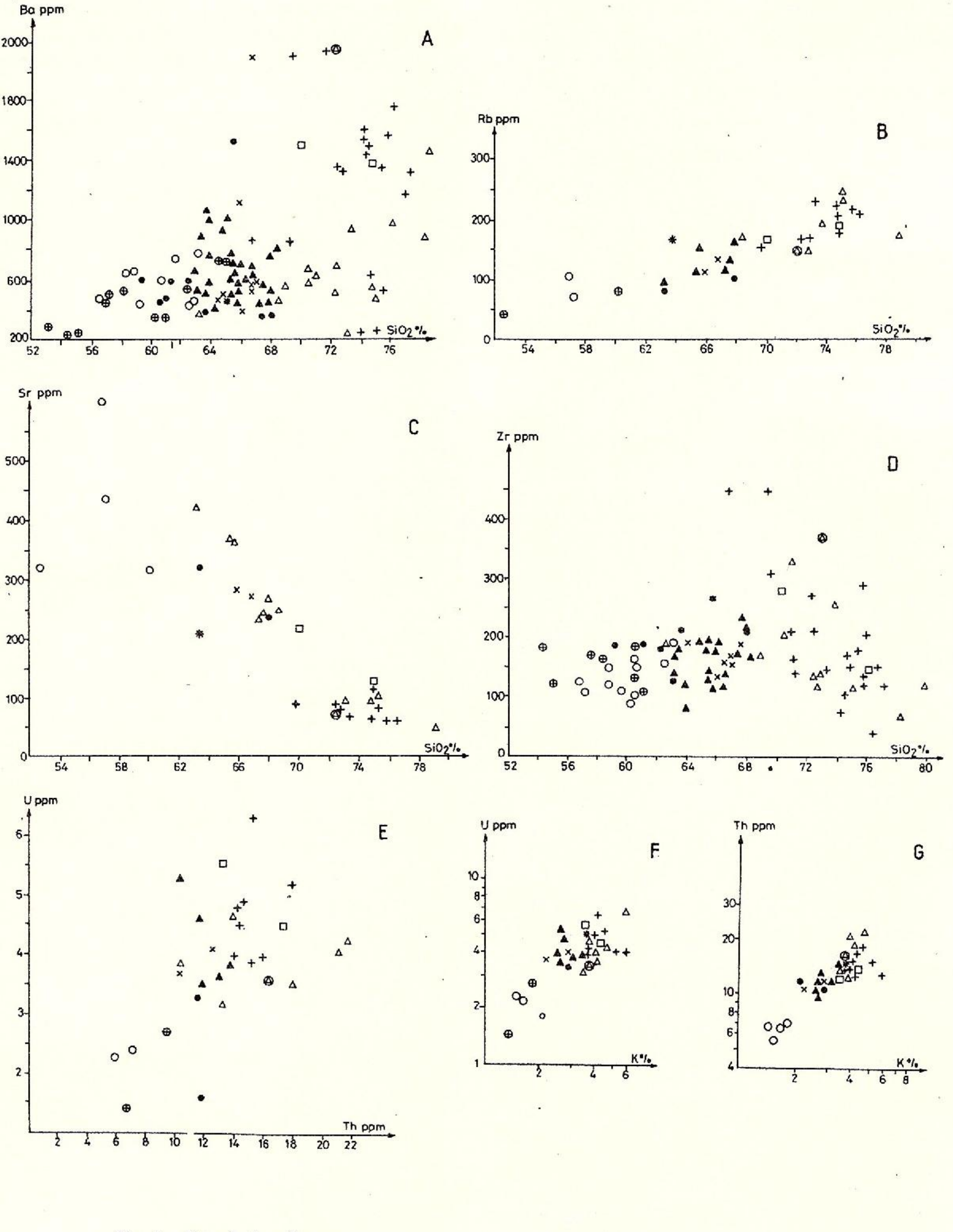

Fig. 2 – Correlation diagrams
A, Ba–SiO_2; B, Rb–SiO_2; C, Sr–SiO_2; D, Zr–SiO_2; E, U–Th; F, U–K; G, Th–K.
The same legend as for Fig. 1.

The data previously presented, corroborated with the Sr isotope ratios, suggest that the calc-alkaline magma generated by subduction had a complex subsequent evolution, by means of differentiation processes, sometimes in intermediate magmatic chambers, concomitantly with variable assimilation of country rocks.

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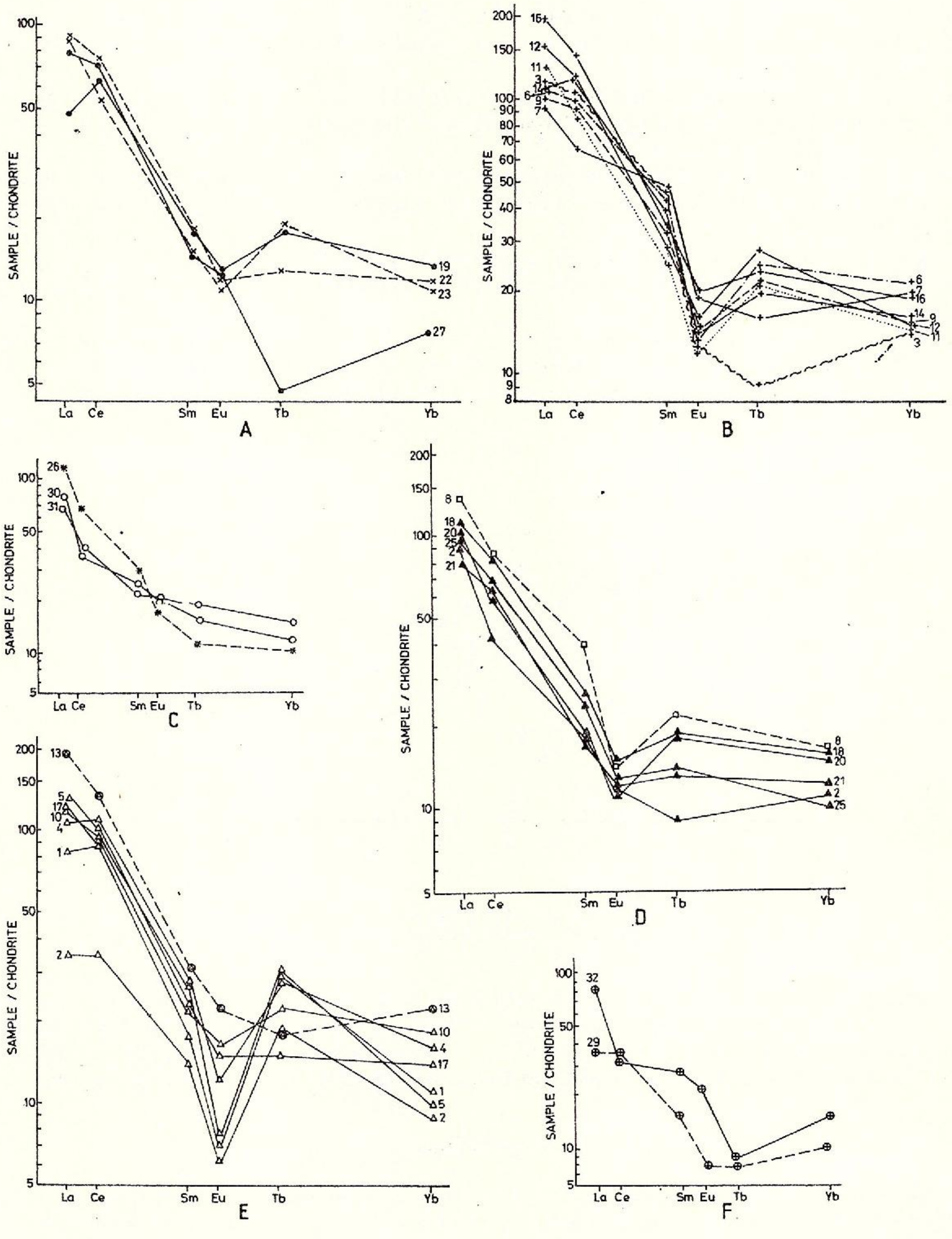

Fig. 3 – REE fractionation diagrams
A, andesites and dacites; B, rhyolites; C, diorites and quartz monzodiorites; D, granodiorites and their porphyritic varieties and rhyodacites; E, granites and their porphyritic varieties and granophyres; F, basaltic andesites and basalts.

Certain petrogenetic remarks on the Northern Apuseni banatitic magmatites can be inferred from the morphologic features of zircon crystals. Starting from the growth of prism (110), (100) or pyramid faces (211),

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(101) depending on the different chemistry of magmas (particularly the $Al_2O_3/(K_2O+Na_2O)$ ratio) and the temperature of zircon crystals constitution (Pupin, Turco, 1972; Pupin, 1980) the following are argued:

- each sample, particularly the ones collected from the first cycle volcanic rocks, exhibits zircon crystals assigned to two different groups, one showing the crustal origin and the other of deep origin, probably in the subducted oceanic crust;
- the zircon crystals supplied by the plutonic and subvolcanic rocks of the second cycle exhibit the same characteristics, most of the samples showing transition types from one group to another, generated by an advanced process of melt homogenization;
- the similar morphologic types of zircon crystals (characterized by a wide range of shapes for subvolcanic and plutonic magmatites) point to similar chemistry and temperature conditions, typical of a single source of the banatitic magma;
- the generation of zircon is conditioned by temperatures ranging from 600 to 900°C, the most frequent ones grouping round 850-900°C in the case of magmatites originating in the mantle and round 750-800°C for those associated with the crust.

The magmatic origin of the primary material which generated the banatitic calc-alkaline melt is also shown by the Na_2O/K_2O diagram (Fig. 4) on which most of the analysed samples (not only granites, but volcanics, too) are plotted in the field of I-type (igneous) granites according to Chappell and White (1974, fide Toth, 1987).

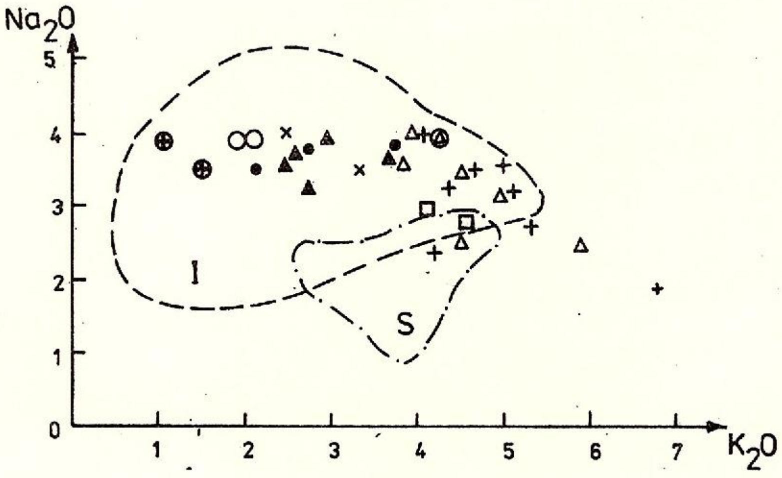

Fig. 4 - Na_2O-K_2O diagram.
The same legend as for Fig. 1.

There are also other features which point to I-type granites: the occurrence of hornblende and titanite; as well as of magnetite as accessory mineral with a wide composition range and linear correlation diagrams. The peraluminous character of rocks and certain deviations from the variation of Na_2O content or the increase of Sr isotope ratio probably reflect the same contamination processes of the initial calc-alkaline magma.

Conclusions

The distribution area of banatitic magmatites in Northern Apuseni Mts is a complex geostructural assemblage represented by metamorphic and/or sedimentary and igneous rocks, starting with crystalline schists (of different lithology and metamorphic degrees) and associated granitoids to sedimentary detrital and carbonate rocks and acid or basic magmatic rocks emplaced previously to banatites.

All these rocks are overlain by the Upper Cretaceous sedimentary cover consisting of carbonate and detrital rocks.

The banatitic magmatites pierce or lie over these formations, generating magmatic contact phenomena or mineralisations, and are transgressively overlain by the Tertiary sedimentary rocks of the Transylvanian Basin. The banatitic calc-alkaline magmatic activity in the Northern Apuseni Mts took place during two cycles: the former, mainly volcanic, is marked by andesites, dacites and rhyolites, while the latter, the main one, is

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represented by plutonic and subvolcanic rocks, of dioritic, granodioritic and granitic composition, associated with the known ore deposits. The second cycle ends up with alkaline differentiates dykes represented by aplites, microgranitic rhyolites, micropegmatitic rhyolites; no mineralizations associated with these magmatic products are known.

The consolidation of the granodiorite-granitic magma was followed by the emplacement on deep fractures of basaltic andesites, basalts and lamprophyres from a deeper magmatic source, partly concomitantly with the above mentioned alkaline differentiates and the post-magmatic solutions associated with the second cycle magmatites.

The geochemical study, implying our samples (Tab. 3) corroborated with the previously published data and the microscopic study of the analysed rocks, shows the distribution of major and minor elements considering the magmatism evolution.

According to $^{87}Sr/^{86}Sr$ isotope data and other geochemical data graphically represented, one accounts for the genesis of calc-alkaline magmas by oceanic crust subduction, followed by complex evolution of melt, within several magmatic chambers, associated with assimilation and differentiation processes.

The pointing out of two and respectively three zircon populations, which show different origins, agree with the petrogenetic concept according to which the calc-alkaline melt, generated by oceanic crust subduction, was contaminated with sialic matter during its ascension, the more obviously as the contact with the sialic crust had longer existed.

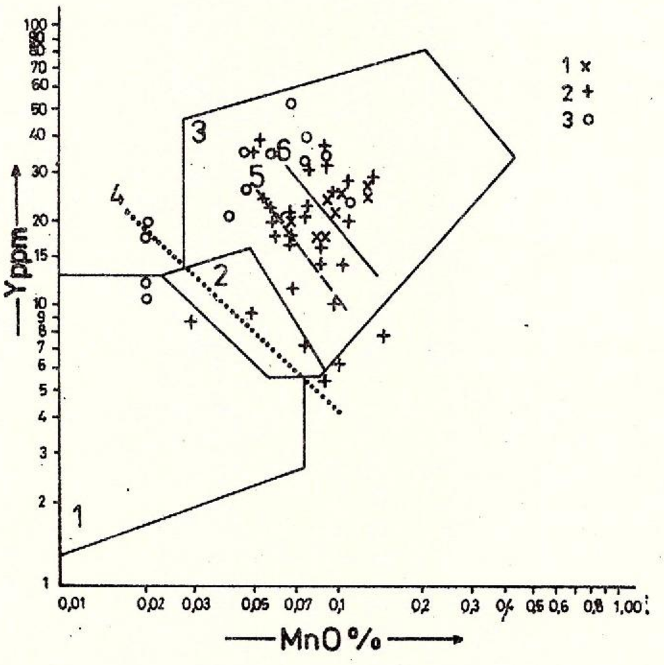

Fig. 5 - Y-MnO diagram
1, diorites; 2, granodiorites;
3, granites.

The geochemical data obtained for generally fresh rocks show the metallogenetic 'specialization' of certain intrusive banatitic magmatites. Therefore, for porphyry copper ores the Y/MnO covariation diagram, applied by Baldin and Pearce (1982) to distinguish the productive porphyritic intrusions of diorite-granitic type from the non-productive ones shows that most of the second cycle magmatic rocks occurring in the Gilău Mts, some of them accompanied by the known mineralizations, are plotted on the field of productive rocks (Fig. 5). The possible occurrence in this area of some porphyry copper ores is suggested in this paper, confirming a previous assumption (Andrei, Udubaşa, Ştefan in Borcoş et al., 1988, unpubl. report).

Institutul Geologic al României

References

Baldin A. J., Pearce A. J. (1982) Discrimination of productive and non productive porphyritic intrusions in the Chilean Andes. *Econom. Geol.*, 77, 3, p. 664–674, Lancester.

Boccaletti M., Manetti P., Peltz S. (1973) Evolution of the Upper Cretaceous and Cenozoic Magmatism in the Carpathians arc Geodynamic significance. *Mem. Soc. Geol. Ital.*, XII, p. 367–377.

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

Cioflica G., Savu H., Borcoş M., Ştefan A., Istrate G. (1973) Alpine Volcanism and Metallogenesis in the Apuseni Mountains. Guide to Excursion 2 Ab. Symp. Volc. and Metallogenesis, 1973, Geological Institut, Bucureşti.

Dragoş I. (1971) Fauna şi flora cretacică din regiunea Vlădeasa (Munţii Apuseni). Rezumatul tezei de doctorat, Univ. Bucureşti.

D'Elio G., Di Girolamo P., Guido M. (1987) Geological and petrological characters of some quaternary calc-alkaline tuffites of Cilento (southern Italy). *Boll. Soc. Geol., It.*, 106, p. 699–716.

Giuşcă D. (1950) Le massif éruptif de la Vlădeasa. *An. Com. Geol.*, XXIII, p. 199–251, Bucureşti.

– , Cioflica G., Savu H. (1966) Caracterizarea petrologică a provinciei banatitice. *An. Com. Geol.*, XXXV, p. 13–45.

– , Istrate G., Ştefan A. (1969) Le complex volcano-plutonique de la Vlădeasa (Roumanie). *Bull. Vol.*, XXXIII/4, p. 1118–1127, Napoli.

Hauer F., Stache G. (1863) Geologie Siebenbürgens. Wien.

Henderson P. (1984) Rare Earth element geochemistry. Ed. P. Henderson, Elsevier, Amsterdam.

Herz N., Savu H. (1974) Plate Tectonics History of Romania. *Geol. Soc. Amer. Bull.*, 85, p. 1429–1434.

Ianovici V., Borcoş M., Bleahu M., Patrulius D., Lupu M., Dimitrescu R., Savu H. (1976) Geologia Munţilor Apuseni. 631 p. Edit. Acad. R.S.R., Bucureşti.

Istrate G., Bratosin I. (1976) Caracterele geochimice ale banatitelor din partea vestică a masivului Vlădeasa şi sugestii pentru originea magmelor banatitice. *D. S. Inst. Geol. Geofiz.*, LXII/1, p. 99–142, Bucureşti.

– (1978) Petrologic study of the Vlădeasa Massif (Western part). *An. Inst. Geol. Geofiz.*, LIII, p. 177–298, Bucureşti.

– , Udubaşa G. (1981) Contribuţii la cunoaşterea metalogenezei masivului banatitic de la Budureasa (Munţii Apuseni). *D. S. Inst. Geol. Geofiz.*, LXV/2, p. 5–19, Bucureşti.

Lazăr C., Întorsureanu I., Popescu M. (1972) Studiul petrografic al rocilor banatitice din zona Maşca-Băişoara (Munţii Apuseni). *D. S. Inst. Geol. Geofiz.*, LVIII/1 (1971), p. 143–173, Bucureşti.

Le Maître R. W. (1984) A proposal by the IUGS subcommission on the systematics of igneous rocks for a chemical classification of volcanic rocks based on the total alkali silica (TAS) diagram. *Austr. J. of Earth Sci.*, 31, p. 243–255.

Mahood G., Hildreth W. (1983) Large partition coefficients for trace elements in high-silica rhyolites. *Geochimica et Cosmochimica Acta*, 47, p. 13–30.

Mureşan I. (1971) Studiul riolitelor din bazinele văilor Fetii şi Stolnei (satul Stolna, jud. Cluj). *Studia Univ. "Babeş-Bolyai", ser. Geol. Miner.*, 2, p. 17–26, Cluj.

Pavelescu L., Pop Gr., Weisz E., Popescu G. (1985) La nature et l'âge du batholite banatitique de Bihor. Proceeding reports of the XIII-th Congress of KBGA. Additionaly received reports, Poland-Cracow.

Pupin J. P., Turco G. (1972) Une typologie originale du zircon accessoire. *Bull. Soc. Mineral. Cristallogr.*, 95, p. 348–359, Paris.

– (1980) Zircon and granite Petrology. *Contrib. Mineral. Petrol.*, 73, p. 207–220, Berlin.

Rădulescu D. P., Săndulescu M. (1973) The plate-tectonics concept and the geological structure of Carpathians. *Tectonophysics*, 16, p. 155–161.

Săndulescu M. (1984) Geotectonica României. Ed. Tehnică, 334 p., Bucureşti.

Ştefan A. (1969) Structura geologică a părţii de est a masivului Vlădeasa. *St. cerc. geol. geofiz., geogr., Geol.*, XIV/2, p. 525–530, Bucureşti.

– (1980) Petrographic study of the eastern part of the Vlădeasa eruptive massif. *An. Inst. Geol. Geofiz.*, LV, p. 207–325, Bucureşti.

– , Lazăr C., Întorsureanu I., Horvath A., Gheorghiţă I., Bratosin I., Şerbănescu A., Călinescu E. (1985) Petrological study of the banatitic eruptive rocks in the eastern part of the Gilău Mountains. *D. S. Geol. Geofiz.*, LXIX/1, p. 215–246, Bucureşti.

– , Rusu A., Bratosin I., Colios E. (1986) Petrological study of the Alpine magmatites in the link zone between the Apuseni Mountains and Oaş-Gutîi-Ţibleş Chain. *D. S. Inst. Geol. Geofiz.*, 70-71/1, p. 243–262, Bucureşti.

– , Lazăr C., Berbeleac I., Udubaşa G. (1988) Evolution of banatitic magmatism in the Apuseni Mountains and associated metallogenesis. *D. S. Inst. Geol. Geofiz.*, 72-73/2, p. 195–213, Bucureşti.

Taylor S. R. (1969) Trace element chemistry of andesites associated calc-alkaline rocks. *Int. Upper Mantle Project Sci. Pop.*, 16, 65, p. 43–65. Proceed. Andesite Conf. Eugen. and Bend, Oregon, SUA.

Institutul Geologic al României

Toth M. J. (1987) Petrology and origin of the Bitterroot lobe of the Idaho batholith. *US Geol. Surv. Prof. Papper.*, 1436, p. 9–35.

Udubaşa G., Istrate G., Popa C. (1980) Date preliminare asupra mineralizaţiilor de la Juleşti-Valea Fagului, Munţii Apuseni. *D. S. Inst. Geol. Geofiz.*, LXIV/2, p. 185–211, Bucureşti.

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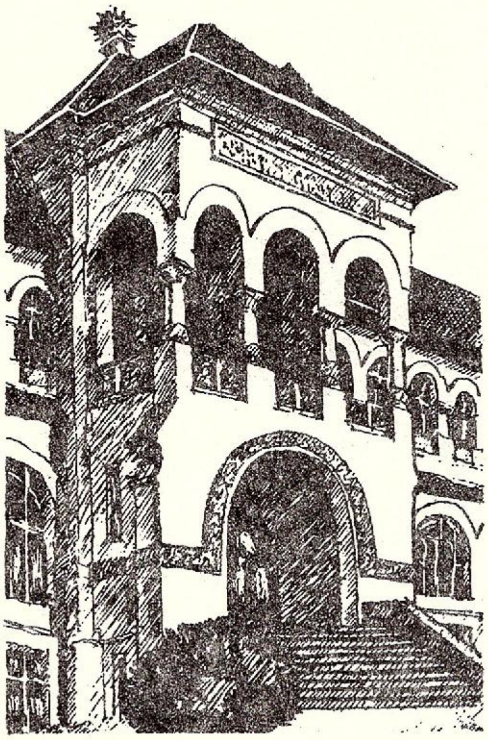

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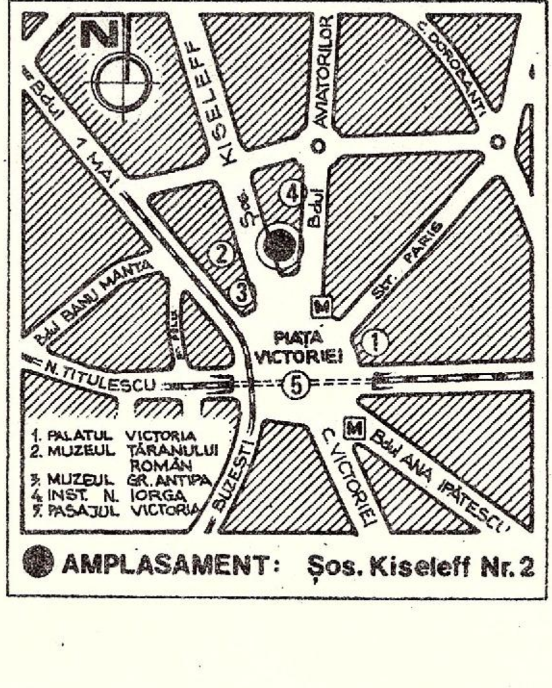

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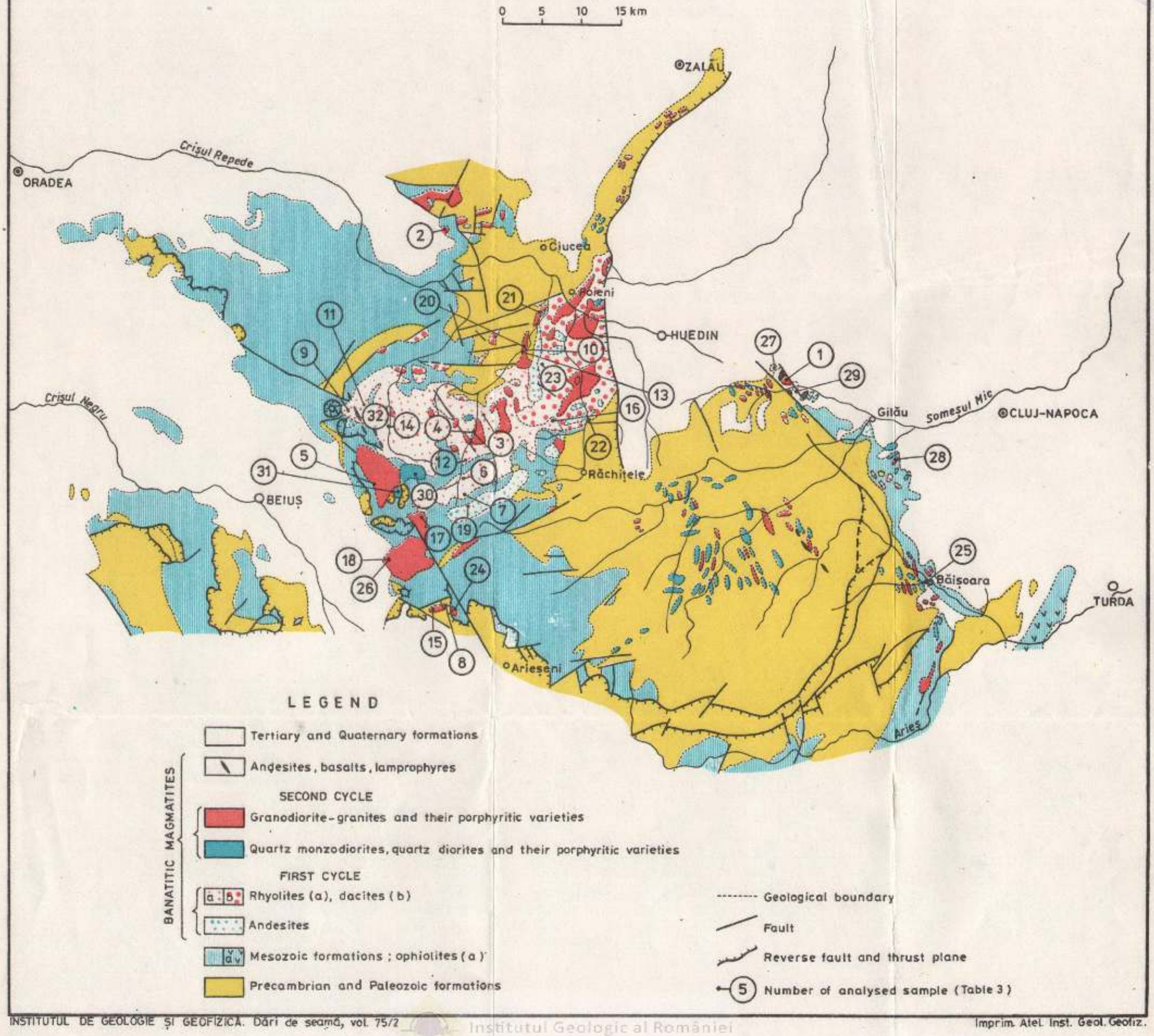
DISTRIBUTION OF BANATITIC ROCKS IN THE NORTHERN AREA OF THE APUSENI MTS

(acc. to A. Ștefan et al., 1988)



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PETROLOGY, GEOCHEMISTRY AND ORIGIN OF THE LARAMIAN VOLCANICS OF THE MUREŞ COULOIR, BETWEEN ZAM AND GURASADA

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Key words: Volcanic rocks. Island arcs. Magmas. Magmatic differentiation. Calc-alkalic composition. Subduction. Major elements. Minor elements. Apuseni Mountains – Laramian eruptive – Sîrbi-Lăpugiu-Bulza area.

Résumé: *Pétrologie, géochimie et origine des vulcanites laramiennes du Couloir du Mureş, entre Zam et Gurasada.* Les vulcanites laramiennes du Couloir du Mureş représentent le résultat d'un volcanisme d'arc continental interne des Carpathes Méridionales, déterminé par la subduction de la plaque moesienne au-dessous de la plaque transylvaine. Elles reposent sur un soubassement tectonisé polyphasiquement, appartenant à la partie sud de la Zone de Mureş. Les roches forment une série volcanique calco-alcaline à tendences alcalines où on sépare trois groupes pétrographiques: (1) basaltes à olivine et andésites basaltiques; (2) andésites et trachyandésites; (3) roches leucocrates acides et alcalines. Les roches présentent les caractères géochimiques des vulcanites d'arc insulaire. Les roches basiques et celles intermédiaires représentent "orogenic high-K andesites", aspect qui souligne la tendence alcaline de la série et les sépare des vulcanites de l'arc volcanique externe qui manifestent des caractéristiques de "medium-K andesites". La présence des roches alcalines, tant dans le groupe andésitique que dans celui leucocratique, souligne aussi la consanguinité des vulcanites. Le magma parental s'est formé à des profondeurs de 200 km, dans le manteau surjacent au plan de Benioff, par la fusion du substratum éclogitique métasomatisé par des solutions émanées de la croûte océanique enfoncée par subduction, en voie d'être transformée en éclogite. Le magma a eu une composition dioritique à 58 % SiO_2, mais il a été fortement différentié, en résultant andésites, basaltes à olivine et roches acides et alcalines.

Introduction

Along the Mureş Couloir, cropping out between Groşi and Sîrbi, on the northern margin of the Lăpugiu posttectonic basin, the products of a Laramian continental arc volcanism are to be found vis-à-vis the Pietroasa-Tomeşti volcanics on the southern margin of the basin. Up till 1983 the mentioned rocks were considered Neogene volcanics by different workers, e.g. Kadič (1906), Papiu (1954), Savu (1962), Peltz and Peltz (1965), Ghiţulescu, Borcoş (1966), Rădulescu, Borcoş (1968), Ianovici et al. (1969) etc., who studied them from different points of view.

In 1983–1984, Popescu (in Borcoş et al., 1986) established that the microfauna of the sedimentary deposits associated to the "Neogene" volcanics of the Mureş Couloir, at Sîrbi, are of Uppermost Cretaceous-Paleocene age. At the same time, Savu et al. (1984, unpubl. report) obtained the radiometric age of 59.6±2.4 Ma on a biotite tuff at Vica. On the basis of these data Savu (1984) presented the volcanics as products of a Laramian (banatitic) volcanism like those in the Rusca Montană Basin. In the years 1984-1985 Roşu (unpubl. reports) studied the Laramian volcanics in the outskirts of the locality of Gurasada with a view to elaborating the

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geological map scale 1:50000, sheet Gurasada (1986). In 1984 Borcoş et al. (unpubl. report) mentioned that in the Lăpugiu Basin both Paleogene and Neogene volcanic rocks are to be found.

Although the above-mentioned papers presented numerous petrographical and petrochemical data on the volcanics of the Mureş Couloir, no modern, complex study has been elaborated up till now. That is why the present paper gives a detailed description of the volcanic rocks lying between Zam, Boiu de Sus and Gurasada, north of the Mureş (Pl. I), which comprise a complete suite of the Laramian volcanics in the region. This study is based on the data obtained in the period 1983–1986 as a result of the surveys carried out for the elaboration of the geological map scale 1:50 000, sheet Vorţa (Savu et al., 1987, unpubl. report), and its correlation with the adjacent zones.

Prevolcanic Basement of the Region

The Laramian volcanics overlie a prevolcanic basement comprising the whole Lăpugiu Basin. The geological formations of this basement belong to two tectonic units: (1) Căpîlnaş-Techereu Unit and (2) Tisa Unit (Pl. I).

1. Căpîlnaş-Techereu Unit (Lupu, 1975) is post-Neocomian in age and belongs to the Mureş Zone. Its formations crop out from under the Laramian volcanics and Neozoic deposits in the northern part of the region. Their basement is represented by a Liassic ocean crust resulting from the splitting of the ophiolitic sheet (megaslab) of the Mureş, obducted at the end of the Jurassic from the Mureş Ocean over the convergent plates, north and south of it (Savu, 1983). The ocean crust basement is exposed along a narrow strip between Glodghileşti and Boiu de Sus. It represents a segment of the Glodghileşti-Săliştioara tectonic rise occurring along a sub-Hercynian or Laramian longitudinal fault trending E-W, which puts into tectonic contact the Liassic ophiolites of the nappe basement with the Barremian-Aptian deposits (Savu, 1983,1984; Savu et al.,1987). The ophiolitic basement consists of the ocean floor basalt complex (O1), usually in pillow lava facies, represented by hyalobasalts, anamesites, basalts – often amygdaloidal – and very rarely dolerites.

The ophiolitic basement is unconformably overlain by postophiolitic Late Kimmmerian island arc volcanics (J_3-Cr_1) in association with jaspers and red argillites which pass laterally and vertically to Stramberg reef limestones (J_3-Cr_1) and to Tithonian-Neocomian flysch formations. The erosion period which probably succeeded to the emplacement of the Căpîlnaş-Techereu Nappe is followed unconformably by Barremian-Albian wildflysch deposits. They form several folds, south of the tectonic rise area of the basement ophiolites, and an asymmetrical syncline consisting of Upper Aptian-Albian wildflysch formations (Lupu, in Savu et al., 1987, unpubl. report), occurring unconformably in the north of the region, approximately in the axial zone of the southern Late Kimmerian island arc of the Mureş Zone (Pl. I). It is to be noted that the Late Kimmerian volcanics unconformably overlie the ocean crust basement like the Laramian volcanics at Boiu de Sus (Pl. I) and the Neogene ones at Obîrşia, north of the area represented on the annexed map. As the latter ones, they do not belong to the ocean crust because during the manifestation of the Late Kimmerian island arc volcanism in the eugeosyncline the ocean crust was to be obducted, blocks being broken from it, which were included as olistoliths in the J_3-Cr_1 flysch (Drocea Mts), or even in the Neokimmerian volcanics in the Metaliferi Mountains (Savu, 1984). All the mentioned series of island arc volcanics are postophiolitic (Rocci et al., 1980). Provided that by definition the ophiolites represent obducted ocean crust (Penrose Conference, 1972; Mesorian et al., 1973; Moskow Conference, 1974) the Late Kimmerian volcanics, like the other mentioned volcanics, cannot be referred to this category.

The Căpîlnaş-Techereu Unit and its southern autochthon were affected by later tectonic movements and sectioned by longitudinal faults strongly dipping northwards – e.g. that in the south of thè Glodghileşti-Săliştioara tectonic rise – as well as by thrust planes, e.g. that which determined the thrust of the Tisa Unit over the Căpîlnaş – Techereu Unit.

2. Tisa Unit (Savu, in Andrei et al., 1987, unpubl. report) was probably determined by the sub-Hercynian Orogeny. It lies between Groşi, Tisa and Vărmaga, an alignment along which the crystalline schists (retromorphic phyllites and amphibolites) of the Poiana Ruscă Mountains thrust from the south to the north over the formations situated on the Căpîlnaş-Techereu Unit, the last thrust formation being the Barremian-Aptian wildflysch deposits at Tisa, south of the Mureş (Pl. I). The thrust plane was affected by the regional crustal fracture pointed out by Andrei and Cristescu (1966) between Cuieş on the Mureş and Recaş in the Banat, along about 45 km. The contact between the two tectonic units is mostly masked by Laramian volcanics and by recent sediments (Pl. I, geological section), from under which crop out in some places, like that on the Şendreasca Valley, at Tisa, strongly tectonized crystalline schists (Peltz, Peltz, 1965) thrust over Barremian-Aptian deposits (Savu, in Andrei et al., 1987, unpubl. report).

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This tectonic accident is marked by gravimetric anomalies on whose basis Andrei et al. (1975) referred the formations on this alignment to the "Mureş Cordillera". It is obvious that the Tisa Nappe began its rising in the Cretaceous Sea as a cordillera which at present is eroded. From this Cretaceous cordillera were probably torn off the crystalline schists olistolith on the Toma (Dobîrleşti) Brook and the microcline granite olistoliths insedimented in the Barremian-Aptian deposits south of Căprioara (Savu, 1966).

Distribution of Volcanics and Their Petrography

Like the Pietroasa eruptive rocks – on the southern border of the Lăpugiu Basin (Savu, 1962) – the Laramian volcanics in the Zam-Gurasada region belong to two eruption phases: (1) phase of mostly leucocrate rocks; (2) phase of intermediary and melanocrate rocks.

1. The eruptive rocks of the first phase are rarely found and are represented especially by stratified tuffs, some of them with an ignimbritic character, situated at the base of the Laramian volcanics. They occur at Gurasada, Tătărăşti and Vica (Pl. I) and are represented by leucocrate biotite tuffs with feldspar, rarely quartz, phenocrystalloclasts whose composition resembles that of rhyodacites, dacites, trachydacites, trachytes and biotite andesites, as shown in Table 1 and on TAS diagram (Le Bas et al., 1986) in Figure 1. It is worth mentioning that since Kadič (1906) the leucocrate rocks in the Mureş Couloir have been considered as biotite andesites and trachytes. Rhyodacites and dacites are quite similar with the acid rocks described by Savu (1962) at Pietroasa and Tomeşti. Trachytes are found as blocks within an agglomerate of biotite andesites on the Tătărăşti Valley. They consist of a groundmass with trachytic texture including phenocrysts of potash feldspar and biotite. It is to be noted that potash feldspar phenocrysts are rarely found in tuffs of alkaline leucocrate rocks; it indicates that K_2O is mostly contained in the glassy mass. Biotite andesites are grey leucocrate rocks consisting of a groundmass formed of oligoclase microcrysts, within which occur phenocrysts of zoned andesine-oligoclase plagioclase and biotite.

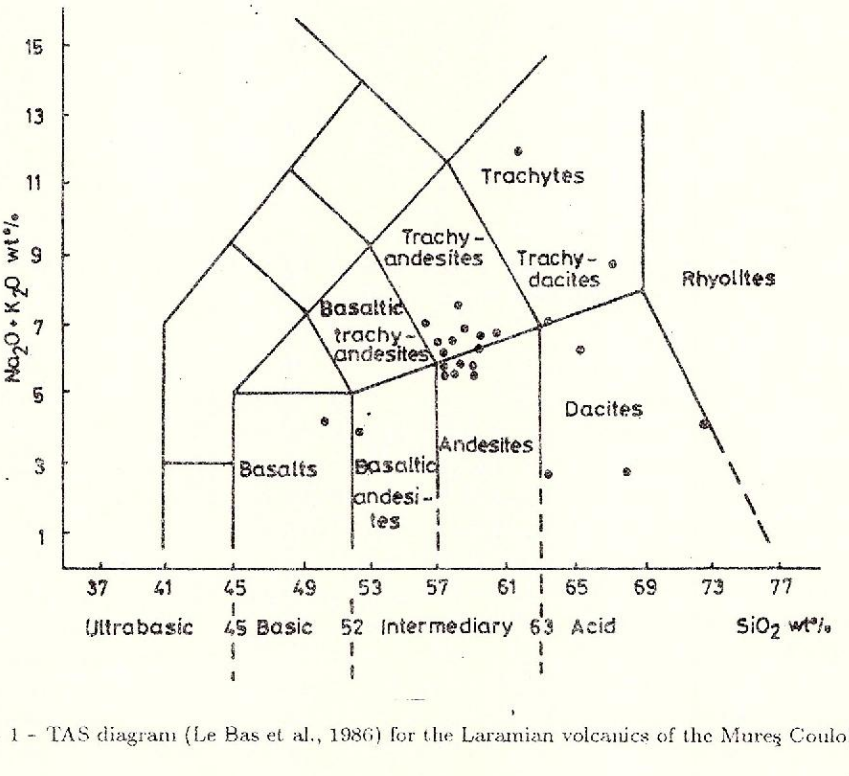

Fig. 1 - TAS diagram (Le Bas et al., 1986) for the Laramian volcanics of the Mureş Couloir

The biotite tuffs at Vica consist of a vitrocrystalloclastic mass which includes fragments of glassy rock with sinuous flow lines. The phenocrysts are represented by oligoclase (An_{28}) and biotite. At Gurasada the leucocrate rock tuffs, particularly the dacite ones, have been affected by bentonitization processes, at the expense of their groundmass secondary minerals being formed, such as montmorillonite and cristobalite, determined by Vanghelie (I.G.G.) by X-ray analyses. For this reason the tuffs are exploited in quarries.

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Table

Chemical composition of the Laramian

No	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	50.78	52.90	56.82	57.20	57.24	57.49	57.60	57.77	58.15	58.37	58.38	58.42
TiO_2	1.50	1.04	1.20	1.18	1.04	1.16	1.02	0.84	1.14	1.18	0.88	0.86
Al_2O_3	15.24	13.92	16.46	19.11	16.06	16.41	16.51	16.11	16.31	17.31	17.11	16.81
Fe_2O_3	4.33	2.63	3.16	3.63	3.05	2.61	2.30	4.28	3.63	6.34	2.71	3.99
FeO	5.37	5.67	3.49	3.39	3.57	4.02	4.44	2.42	3.36	0.40	3.76	2.15
MnO	0.17	0.15	0.15	0.17	0.17	0.16	0.17	0.17	0.16	0.08	0.14	0.01
MgO	6.18	9.76	3.58	2.90	4.66	4.08	4.17	3.69	2.88	2.47	2.47	3.70
CaO	11.21	9.78	6.96	7.13	7.41	7.24	7.13	6.97	6.73	5.98	6.89	6.69
Na_2O	2.43	2.55	3.27	3.61	2.71	2.80	2.87	2.67	2.82	3.25	3.28	3.12
K_2O	1.53	0.98	3.41	1.50	3.35	2.86	2.74	3.55	2.97	3.64	2.35	2.20
P_2O_5	0.23	0.11	0.20	0.08	0.31	0.10	0.16	0.33	0.16	0.32	0.15	0.10
S	0.07	0.27	0.23	0.18	0.22	0.29	0.19	0.25	0.25	0.25	0.23	0.27
CO_2												
H_2O^+	1.05	0.75	0.64	0.78	0.69	1.25	1.38	1.53	1.19	1.19	1.43	2.08
H_2O^-												
Total	100.11	100.51	99.57	100.86	100.48	100.47	100.68	100.58	99.75	100.78	99.78	100.40
Ni	17	21	19	7.5	24	17	13	18	12	15	8.5	24
Co	34	25	20	15	22	15	16	23	19	14	15	18
Cr	100	260	40	1.5	110	80	48	47	16	30	4.5	65
V	300	100	210	160	190	190	210	180	210	200	175	180
Sc	44	17	17	14	20	16	19	17	17	16	15	15
Y	21	12	26	19	23	20	23	23	29	22	20	24
Yb	2.8	1.3	2.1	2.3	2.0	2.2	2.3	1.8	2.8	2.1	1.8	3.1
La	<30	<30	44	<30	38	36	42	42	60	50	<30	<30
Zr	110	75	210	105	190	190	180	180	200	200	175	180
Ba	280	300	600	360	600	650	650	600	750	600	500	500
Sr	440	720	700	560	480	530	680	600	720	680	620	720
Pb	7.5	11	20	14	14	25	25	18	23	16	12	15
Cu	23	80	30	8.5	20	20	22	19	30	22	12	23
Ga	13	15	17	20	12	18	20	13	20	17	16	14
Sn	2	<2	3	<2	2.5	2.5	3	3	<2	3	<2	<2

* The analyses in this table represent: basalt, 1–Petreşti; basaltic andesite, 2–Runcşor; andesites, 6, 12, 14, 15–Runcşor; 5, 8–Glodghileşti; 11–Voia Valley; 17–Petreşti Valley; trachyandesites, 4, 7, 9, 16–Runcşor; 3–Chihu Summit; 10, 13–Măgura Summit, 18–Glodghileşti; 20, 24–Gurasada quarry; rhyodacite, 25–Pîrîul Bisericii.

2. After this short extrusive episode with leucocrate rocks, the volcanic activity continues with volcanic breccias, agglomerates and stratified tuffs (Glodghileşti), and rarely lava flows. The volcanic products are represented by basalts, basaltic andesites, andesites and trachyandesites (Fig. 1), rocks constituting the main mass of the Laramian volcanics in the region (Pl. I). According to the type of the phenocrysts, about seven petrographic varieties can be distinguished among the second phase volcanics: olivine basalts, basaltic andesites, plagioclase-pyroxene andesites, hornblende-pyroxene andesites, hornblende-pyroxene-plagioclase andesites, hornblende andesites, trachyandesites, and their tuffs.

Olivine basalts are rarely found. They are aphanitic rocks consisting of a hyalopilitic up to pilotaxitic groundmass, represented by microcrysts of plagioclase, rarely of augite and magnetite (Pl. II, Fig. 1), in which augite and olivine microphenocrysts, as well as idingsite pseudomorphoses after the latter, are floating. The augite microphenocrysts ($c \wedge Ng = 43^0$-48^0) are twinned after (100). They are grouped determining the glomeroporphiritic texture of the rock. Augite appears as idiomorphic crystals, locally resorbed by the magma. Olivine microphenocrysts are more rare; they are idiomorphic and are replaced by idingsite on fissures, or entirely. A fine magnetite powder is spread in the whole rock.

Basaltic andesites can be separated into augite andesites and augite hypersthene andesites. They are porphyritic rocks consisting of a hyalophilitic groundmass, represented by plagioclase microlites oriented parallel to the lava flow trend, within which augite and hypersthene phenocrysts are to be found. The first are polysinthetically twinned after (100) and are grouped forming a glomeroporphiritic texture. The extinction angle ($c \wedge Ng = 45^0$-55^0) shows that the mineral is a Ti-augite, with an hour glass structure. It is to be noted the

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the laramian volcanics of the Mureş Couloir*

13	14	15	16	17	18	19	20	21	22	23	24	25**
58.45	59.14	59.18	59.53	59.62	60.79	61.50	63.16	63.34	65.47	67.47	68.57	72.05
0.76	1.04	0.76	1.16	0.98	0.96	0.36	0.78	0.19	0.54	0.18	0.19	0.30
17.36	17.16	17.56	17.51	14.41	16.56	19.19	18.61	12.85	19.19	16.92	12.72	11.55
1.91	2.99	3.47	4.46	2.15	3.46		4.14	1.81	2.13	1.84	1.24	3.75
3.94	2.62	2.57	1.12	2.95	1.95	1.66	0.18	1.86	0.18	0.35	0.39	0.22
0.14	0.12	0.12	0.05	0.13	0.10	0.04	0.05	0.07	0.04	0.05	0.03	0.05
2.45	3.56	2.49	1.25	2.28	2.93	1.10	0.35	3.25	0.50	1.00	2.60	0.50
5.76	6.64	6.26	7.03	6.40	5.74	2.48	5.56	1.76	4.84	2.30	1.63	2.00
3.16	3.59	3.42	3.60	3.51	2.90	2.84	4.06	1.12	2.45	3.20	1.05	1.50
3.93	2.28	1.74	2.65	2.43	3.45	9.20	2.40	1.17	2.94	4.90	1.17	2.50
0.29	0.10	0.08	0.16	0.10	0.16	0.10	0.08	0.02	0.15	0.06	0.03	tr.
0.29	0.20	0.29	0.26	0.27	0.35	0.01	0.25	0.01	0.34	0.30	0.08	abs.
						0.43		0.62			0.19	
1.24	1.17	1.73	0.97	1.24	1.64	1.04	0.64	5.19	1.69	2.22	4.75	3.93
								6.59			5.66	1.50
99.68	100.61	99.67	99.75	99.47	100.89	99.98	100.26	99.85	100.46	100.79	100.20	99.85
7.5	21	5	16	7.5	21	9.5	6.5	2	14	5.5	2	
11	13	10	9	13	14	2	5	2	4.5	3.5	2	
6.5	44	1	54	4	75	17	2.5	2.5	95	2	4	
150	145	120	250	180	135	54	120	8.5	80	14	17	
12	13	12	20	13	14	8.5	8	5	8	3.5	7.5	
21	16	17	24	19	23	22	15.5	18	12	12	18	
2.0	2.3	2.2	2.3	2.3	1.8	1.5	1.7	1.5	1.2	1.5	1.5	
45	<30	<30	36	< 30	60	52	30	30	33	50	30	
150	145	100	115	180	135	96	120	87	80	140	130	
620	500	370	630	420	600	1100	530	440	380	920	380	
650	620	600	550	470	430	250	730	85	360	580	75	
23	20	9.5	21	15	22	21.5	6	24	18	23	20	
14	31	5	22	10	22	7.5	2	3.5	18	17	19	
16	20	14.5	20	17	17	16	9	30	17	18	10	
3	2	<2	<2	2	2	2	2	2	2	3	2	

** The analysis no. 25 was offered to us by E. Roşu, for which we are very grateful.

Remarks: Nb is lower than 10 ppm, excepting samples 20 (12 ppm) and 24 (20 ppm). Mo is present in the samples 20 (3 ppm) and 24 (57 ppm).

presence of "opacite" pseudomorphoses after a resorbed amphibole. Plagioclase phenocrysts (An_{50-54}) show a poorly zoned structure and polysynthetic twins; they often contain glassy inclusions. Hypersthene phenocrysts are light pleochroic: Ng=greenish; Nm=brown-greenish; Np=yellowish.

Plagioclase-pyroxene andesites can be divided, according to the character and amount of phenocrysts, into plagioclase-augite andesites and plagioclase-augite-hypersthene andesites. The first variety is frequently found. The rocks with a porphyritic texture consist of the same hyalophilitic groundmass and unoriented or oriented plagioclase phenocrysts, rarely resorbed, with a zonary structure and polysynthetically twinned. Locally they display glassy inclusions. The anorthitic content varies in different zones from An_{56} to An_{50}. The pyroxene phenocrysts are rare and they are represented by augite and more rarely hypersthene. In some cases the melanocrate minerals have been replaced by a semiopaque reddish material, formed of a powder of iron and titanium oxides, or by green-bluish or yellowish chlorite. In places the magnetite crystals, usually small-sized, are larger than the plagioclase microcrysts, looking like microphenocrysts. Within these andesite varieties pyroxene crystals are grouped, forming a glomeroporphyritic texture.

Hornblende-pyroxene andesites display a porphyritic texture and consist of a high-glass groundmass within which the plagioclase microcrystals are often parallel to the margins of the augite, hornblende and plagioclase phenocrysts. Hornblende phenocrysts are quite frequently found; they display the following pleochroism: Ng=green-brownish; Nm=brown-greenish; Np=yellow-greenish and are locally resorbed and usually opacitized on margins (Pl. II, Fig. 2). Augite phenocrysts ($c\wedge Ng=55^0$) are idiomorphic and they locally present a zonary

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or hour-glass structure. Plagioclase phenocrysts (An_{30-48}) are polysynthetically twinned and display glassy inclusions. Locally they are partly resorbed.

Hornblende-pyroxene-plagioclase andesites are rocks with a porphyritic or glomeroporphyritic texture, originating in a glassy groundmass with a fluidal structure in which the plagioclase microcrystals and the magnetite grains are oriented parallel to the lava flow trend, surrounding the phenocrysts of plagioclase, hornblende, augite and hypersthene. Plagioclase phenocrysts (An_{40-48}) present a poor zonary structure; they are polysynthetically twinned and contain glassy inclusions (Pl. II, Fig. 3). Rarely they are partly resorbed. In places they form crystal agglomerations in which the composition of different zones varies from An_{40} to An_{28}. Hornblende phenocrysts (usually 5 mm long) are usually idiomorphic and opacitized on margins, microphenocrysts being often entirely replaced by iron oxides. They display twins after (100) and include plagioclase and magnetite crystals. Locally the amphibole substitution is made (1) by opacitization and (2) by its alterations on margins into an aggregate of plagioclase, magnetite and augite microphenocrysts, in which plagioclase moulds the other minerals. Augite phenocrysts ($c \wedge Ng=46^0$) occur less frequently. They are idiomorphic or resorbed by the magma and present polysynthetic twins after (100). Hypersthene phenocrysts are idiomorphic and light pleochroic: Ng=greenish; Np=light brown-redish. In some rocks one can observe small aggregates of plagioclase, clinopyroxene, hypersthene, magnetite and, more rarely, hornblende crystals displaying a microgabbro-diorite aspect (Pl. II, Fig. 4), in which pyroxenes are moulded by plagioclase. They have been torn off from an intermediary magmatic chamber.

Hornblende andesites are widely porphyritic, with hornblende phenocrysts from 2 to 10 mm long, resorbed on margins and opacitized. They are parallel to the trending of the flow lines from the groundmass, which displays a hyalophilitic texture in which the plagioclase microlites tend to be parallel.

Andesite tuffs are rarely found. They usually occur in the base of the volcanics pile (Glodghileşti), pointing out that the volcanic activity began also in the second phase with eruptions whose pressure was very high. Plagioclase-pyroxene andesite tuffs appear at Glodghileşti. They are stratified, with a crystallovitroclastic structure, and consist of a glassy groundmass including plagioclase microphenocrystalloclasts, sedimented parallel to the tuff stratification. The different levels or bands occurring in tuffs vary as regards the amount of microphenocrystalloclasts.

Trachyandesites include rocks with a high K_2O content (> 3 %, Tab. 1), falling on the diagram in Figure 1 in the trachyandesite field. However, it is to be noted that no potash feldspar phenocrysts have been observed within them, so that in petrographic respect they do not differ much from andesites.

Geochemistry and Tectonic Setting

In the Zam-Gurasada area the rocks form a strongly differentiated volcanic series – from basalts to rhyodacites (Tab. 1). Their SiO_2 contents vary as follows: basalts – $SiO_2 < 52$ per cent, andesites and trachyandesites – $SiO_2 = 53–63$ per cent, leucocrate rocks – $SiO_2 > 63$ per cent, thus being included in Gill's (1981) and Ewart's (1982) classifications. Diagrams in Figures 1 and 2 show a calc-alkaline series, as in case of the andesite group where trachyandesites occur and in case of the leucocrate rocks, associated with trachyandesites and trachytes.

Basalts and andesites present high MgO contents ranging between 6.18 and 9.76 per cent, and CaO contents varying from 9.78 to 11.21 per cent. In trachyandesites the alkali contents are higher than in andesites, their amount being higher than 3 per cent. In rhyodacites SiO_2 is higher than 72 per cent, and in trachytes it is lower (61.50 per cent), but K_2O is much higher, reaching 9.20 per cent.

The SiO_2 variation from 50.78 per cent to 72.05 per cent, as well as the variation of other oxides, show that the parental magma underwent an intense differentiation process. It was a dioritic calc-alkaline magma, with a little higher alkali content. The presence of alkaline rocks, both in the andesites group and in the leucocrate rocks, points out the consanguinity of the volcanics.

Trace elements (Tab. 1) are characteristic of the calc-alkaline series. Table 1 shows that andesites and trachyandesites differ clearly from basalts and leucocrate rocks. Basalts display higher contents of Co, V and Sc and lower contents of Ba. Leucocrate rocks are poor in siderophile elements and are enriched in Ba. Due to their trace elements contents, the andesitic rocks belong to the suite of "high-K" orogenic andesites (Gill, 1981), as also indicated by the high Ba values (300–750 ppm, mean value 530 ppm) and high Sr values (430–720 ppm, mean value 600 ppm). These averages can be compared with those presented by Gill (1981) for this type of andesites. The value of the Ni/Co ratio ≤ 1, obtained in most of the rocks, is characteristic of the andesites. Chrome from andesites shows contents varying within a wide interval, as compared with Ni whose values range

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from 5 to 24 ppm. This aspect is also rendered evident on the diagram in Figure 3, on which the rocks fall into the island arc field. The appurtenance of the Laramian rocks to the island arc volcanics is also emphasized on the diagram in Figure 4, on which the leucocrate rocks plot, too.

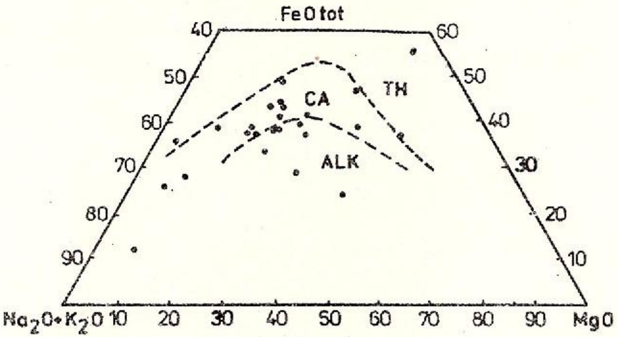

Fig. 2 – FeO_{tot} – MgO – Na_2O+K_2O diagram (Irvine, Baragar, 1971; Hutchinson, 1982).

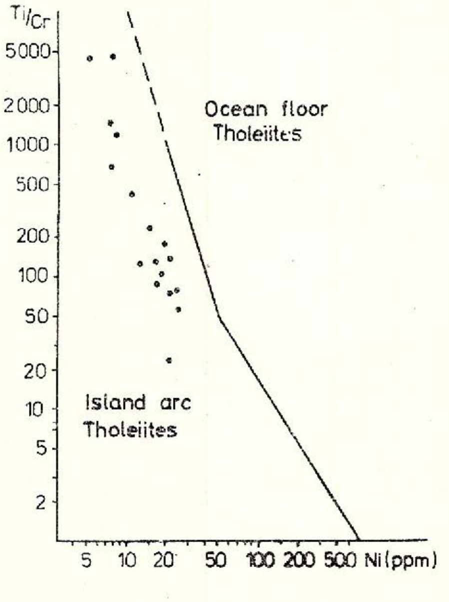

Fig. 3 – Ti/Cr – Ni diagram (Beccaluva et al., 1979).

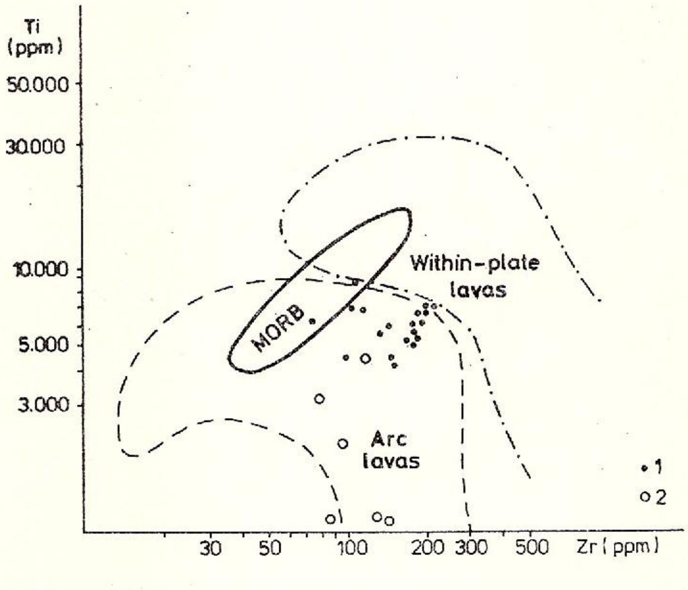

Fig. 4 – Ti–Zr diagram (Pearce, 1980). 1, basalts and andesites; 2, leucocrate rocks.

The values of the Rb/Sr ratio determined by Grabari (I.G.G.–Bucharest) on samples 2 and 23 (Tab. 1) are 0.051 for the basaltic andesite and 0.226 for trachyandesite. On the same rocks, Popescu (I.F.I.N.– Măgurele) determined for the $^{87}Sr/^{86}Sr$ ratio values of 0.710 for the former and 0.706 for the latter. All values are characteristic of the arc volcanics.

The REE contents presented in Table 2 are higher for LREE and lower for HREE, typical of the arc calc-alkaline series, as indicated also on the diagram in Figure 5. As Table 2, this diagram points out an increase of the REE contents from the melanocrate rocks to the leucocrates ones. The patterns of the three rock groups are approximately parallel and they are similar to those presented by Gill (1981) for the orogenic andesites of the "high-K" suite. An exception is the pattern of the leucocrate rocks, which points to a positive anomaly for Ce and to a negative one for Yb. There is no negative anomaly for Eu, which is typical of the rocks with an alkaline tendency, e.g. monzogranites and syenogranites (Cullers, Graf, 1984). The values of the Eu/Sm ratio (0.16–0.39) and of the La/Lu ratio (8.5–5.0) are within the limits indicated by Cullers and Graf (1984)

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for the continental orogenic andesites, showing that the Laramian magmatic rocks in the Mureş Couloir are arc volcanics formed in a continental subduction zone, as shown on the diagrams in Figures 3 and 4. This conclusion also results from the diagram in Figure 6, on which the rocks are situated both in the field of the plate margin basalts (62 per cent) and in that of the intraplate basalts (38 per cent), pointing out that some rocks from the continental volcanics arcs present higher Ti and Zr contents than those from intraocean island arcs. The average values of the Ba/La[1] ratio are of 17-24 and they are placed within the limits given by Gill (1981) for the arc volcanics in New Zealand (12-25).

Table 2

REE contents (ppm) in the volcanic rocks

No*	La	Ce	Sm	Eu	Tb	Yb	Lu
1	12	16	4.1	0.64	0.46	2.8	0.22
2	17	26	3.6	1.00	0.47	1.3	0.20
4	15	30	3.3	1.00	0.74	2.3	0.26
5	26	68	6.0	1.50	0.43	2.0	0.28
13	32	82	6.2	1.70	1.00	2.0	0.52
16	25	43	4.2	1.30	0.60	2.3	0.22
21	15	56	4.0	1.23	0.87	1.7	0.35
23	32	100	4.3	1.70	0.98	1.5	0.58

* The number of the analyses corresponds to that in Table 1.

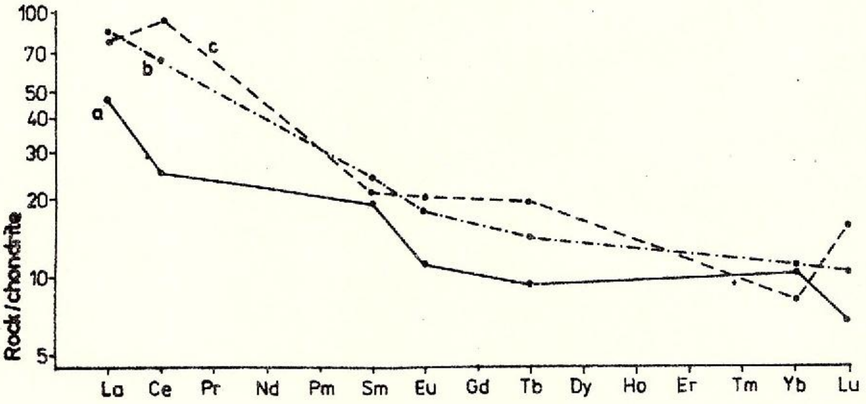

Fig. 5 – REE distribution, normalized by chondrite, in the volcanic rocks. a, basalts and basaltic andesites; b, andesites and trachyandesites; c, leucocrate rocks.

Origin of the Volcanics

The Laramian volcanics in the Mureş Couloir occur along an arc of gravimetric anomalies spread from the Laramian granitoid intrusions in Banat towards ENE, beyond Sîrbi (Mureş Couloir), thus indicating the extension of the volcanic arc which generated them. This volcanic arc is parallel to the Severin ophiolitic suture (Fig. 7), that is to a curve line of the collisional contact between the Moesian Plate and the Transilvanian

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[1] When calculating the Ba/La ratio, for La were used only the values presented in Table 2, obtained by neutron activation, because the contents determined by emission spectrography (Tab. 1), being close to the detection limit, can present greater errors of analysis.

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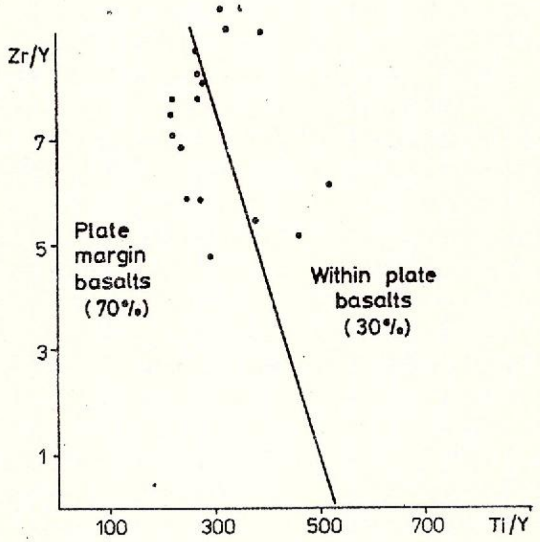

Fig. 6 – Zr/Y–Ti/Y diagram (Pearce, Gale, 1977).

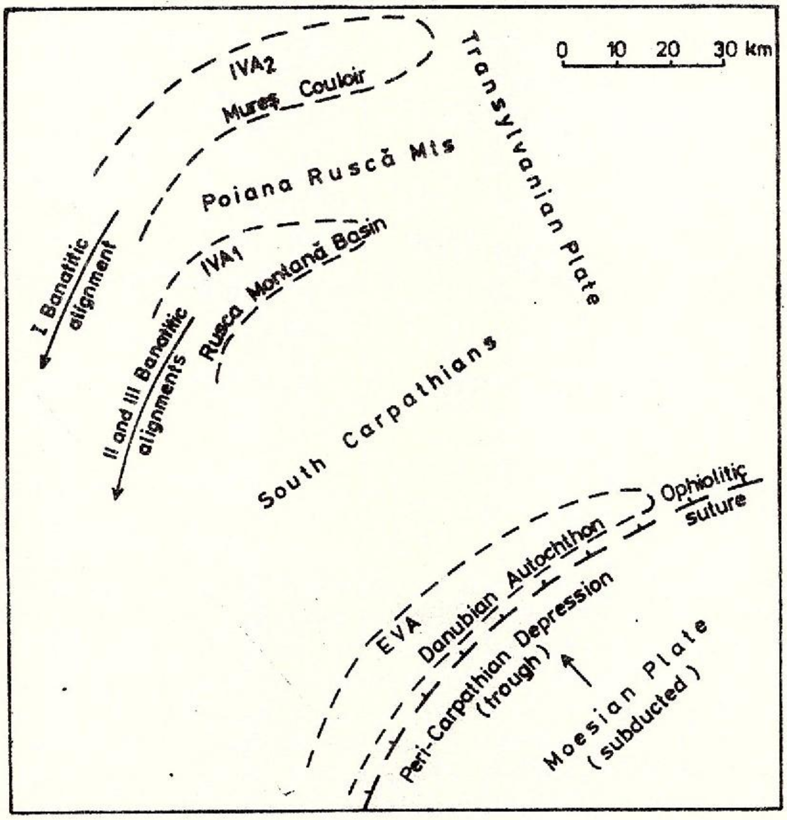

Fig. 7 – Position of the three volcanic arcs as against the Severin ophiolitic suture.
EVA, external volcanic arc; IVA$_1$, first internal volcanic arc; IVA$_2$, second internal volcanic arc.

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Plate (Savu et al., 1985), on whose external margin also occur Laramian volcanics belonging to an external volcanic arc (EVA), and to the axis of the South Carpathians a first internal volcanic arc (IVA_1), in the Rusca Montană Basin (Savu et al., 1987). The first internal volcanic arc (IVA_1) extends south-westwards with the Laramian intrusions from the second and the third banatitic (Laramian) alignment established by Giuşcă et al. (1966). The second internal continental volcanic arc (IVA_2) – which we are dealing with in this paper – extends south-westwards with Laramian intrusions from the first banatitic alignment (Fig. 7), generated by subduction (Rădulescu, Săndulescu, 1973). We mention that on this alignment, like in the Mureş Couloir, there are igneous rocks with a more alkaline tendency than in the other two volcanic arcs (EVA, IVA_1) or the two mentioned Laramian alignments in Banat. These volcanic arcs extend in Yugoslavia and Bulgaria.

The position of the three continental volcanic arcs as against the Severin ophiolitic suture, the margin of the peri-Carpathian Depression (Trough) respectively, shows that their formation is connected with the subduction of the Moesian Plate under the Transylvanian Plate (Savu et al., 1985, 1987); it reminds us of the volcanic arcs succession in the Andes described by Palacios (1984). It is a Chilean-type volcanism (Uyeda, 1981), more compressive and with porphyry copper mineralizations. The volcanic arcs occurred successively, concomitantly with the WNW advance of the subduction process; the external arc volcanics (EVA) erupted during the Upper Turonian-Senonian (Stan et al., 1979; Savu et al., 1987), the volcanics from the first internal volcanic arc (IVA_1) were formed during the Maastrichtian (Maier et al., 1975) and those of second volcanic arc (IVA_2) occurred during the Paleocene (Popescu in Borcoş et al., 1986; Savu, 1984). The volcanics from the external volcanic arc erupted in the Carpathian miogeosyncline in which the Upper Cretaceous wildflysch was being deposited and the rocks from the two internal volcanic arcs erupted in smaller epicontinental sedimentary basins.

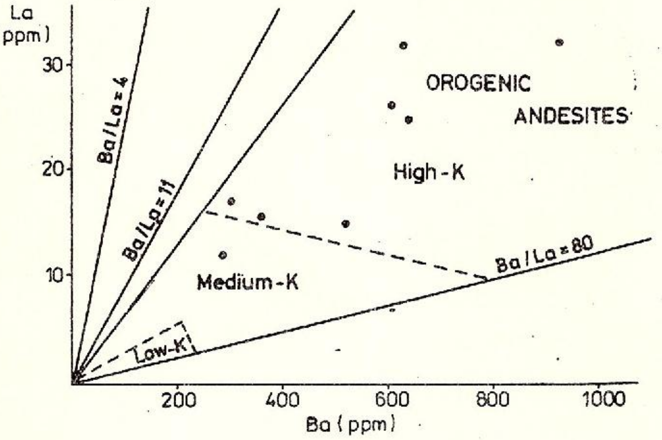

Fig. 8 – La-Ba diagram (Gill, 1981).
For La, the values presented in Table 2 were used.

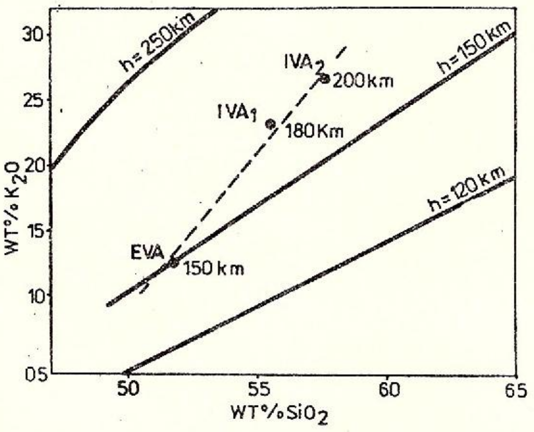

Fig. 9 – K_2O–SiO_2 diagram (Ninkovich, Hayes, 1971).

The mentioned succession in the formation of the continental volcanic arcs and the regional zonality as against the Severin ophiolitic suture line determined a petro- and geochemical polarity of the volcanics. Within the external volcanic arc (EVA), the rocks are more basic and display "medium-K andesites" characteristics (Savu et al., 1987), whereas those of the second internal volcanic arc (IVA_2) correspond to "high-K" andesites (Fig. 8). As regards the first internal arc (IVA_1) there are no geochemical data necessary for the establishing of these characteristics. However, the position of the rocks on the diagram in Figure 9 and the value of the K_2O/SiO_2 ratio, as we shall see further on, show that these rocks must have an intermediary position versus the rocks from the other two volcanic arcs. The K_2O and SiO_2 contents of the volcanics increase concomitantly with the distance from the margin of the Transylvanian Plate and the sinking of the Benioff plane, as shown on the diagram in Figure 9, on which occurs the plotting (average values) of the rocks from the three volcanic arcs. The value of the K_2O/SiO_2 ratio increases from 0.024 in the external volcanic arc to 0.041, in the first internal volcanic arc, and to 0.046 in the second internal volcanic arc.

In the Zam-Gurasada area the parental magma of the volcanic rocks, situated in the second internal volcanic arc (IVA_2), was a dioritic magma with about 58% SiO_2 formed at a depth of 200 km (Fig. 9). Similarly occurred the parental magmas of the other two volcanic arcs, the parental magma of external volcanic arc displaying a

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basaltic character with an average SiO_2 content of 51-52 per cent. Those magmas were formed at even smaller depths (180 km and 150 km, respectively) in the mantle, decreasing concomitantly with the lowering of the depth of the Benioff plane (Fig. 10).

If the formation depth of the magmas rendered on the diagram in Figure 9 is compared with the actual small distance between the three continental volcanic arcs and the Severin ophiolitic suture one must admit either that the dipping of the Benioff plane was greater than 45^0 or that the distance between the volcanic arcs and the subduction line was longer. In fact, the spatial relationships observed nowadays (see Fig. 7) represent the result of the evolution of a geotectonic system from an oceanic zone, by subduction processes, to the collision stage, with crust consumption, therefore those elements must have initially been farther one from another.

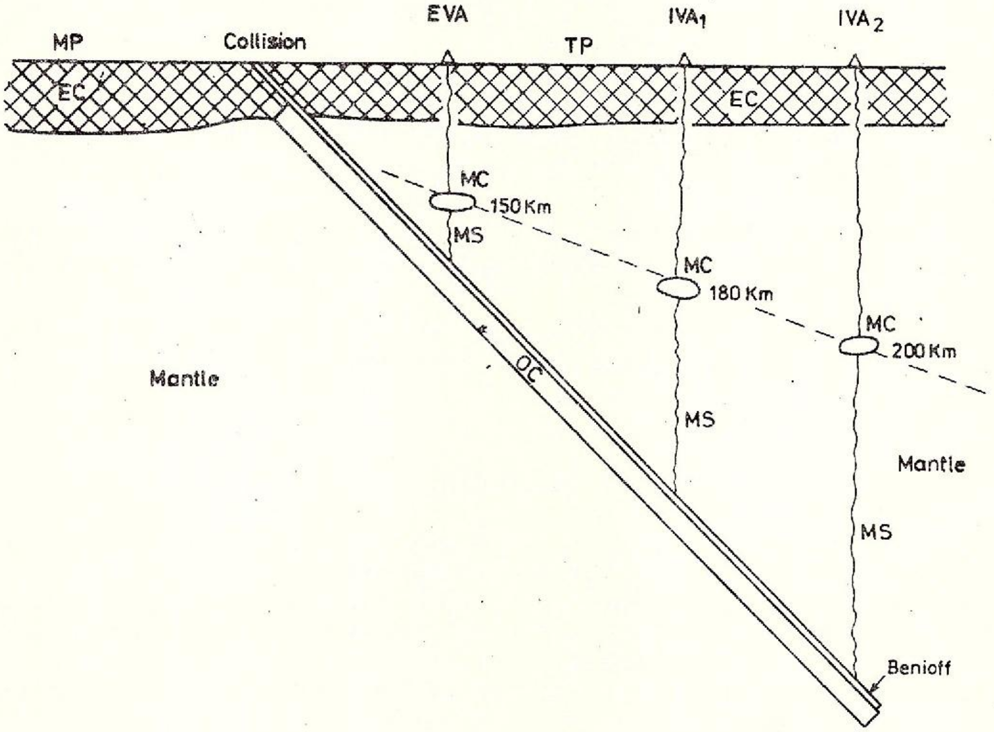

Fig. 10 – Formation of the calc-alkaline parental magmas in the three continental volcanic arcs of the South Carpathians by the melting of the metasomatized mantle under conditions of the subduction of the Moesian Plate under the Transylvanian Plate.
MP, Moesian Plate; TP, Transylvanian Plate; EC, Earth's crust; MC, magmatic chambers; MS, metasomatic solutions; OC, subducted oceanic crust which is being eclogitized.

There are several opinions as regards the magma formation in the subduction zones. Taylor (1968) considered that magma resulted from the melting of the subducted oceanic crust, but Gill (1981) stated that the mere melting of the oceanic crust could not generate a calc-alkaline magma. Mysen and Boettcher (1975) and Green (1976) showed that the andesites came from hydrated meltings of the peridotites at pressures of ca. 10 kbar. Considering the above-mentioned opinions, as well as those published by Minear and Toksöz (1970), Mysen (1979) and Roden and Murthy (1985), we came to the conclusion that the parental magma from the second internal volcanic arc (IVA_2) was formed in the mantle suprajacent to the Benioff plane – like the magma of the other two volcanic arcs – at a depth of about 200 km, as a result of the partial melting of the eclogitic source, that had been affected by metasomatic processes with substances emanated from the oceanic crust of the Moesian Plate subducted on the Benioff plane (Fig. 10). The average value of the (La/Ce)cn ratio (1.16) suggests that the parental magma was generated by a source that underwent a process of LREE depletion (Gill, 1981) or, according to us, it originates in the metasomatized mantle. This magma generated "high-K andesites" and trachyandesites. Small amounts of olivine basalts and basaltic andesites, as well as acid and

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alkaline leucocrate rocks were formed by partial melting and differentiation. The percentages of the three rock groups point to those from the Brokeoff volcano in California (USA) studied by Fountain (1979). The small amounts of basalts and leucocrate rocks in comparison with the great amount of andesitic rocks in the Mureş Couloir are also comparable with the percentages established by Sugimura (1968) for the island arc volcanics in Japan.

Conclusions

Laramian volcanics are products of a continental arc volcanism, formed in connection with the subduction of the Moesian Plate under the Transylvanian Plate.

The rocks constitute a calc-alkaline volcanic series with obvious alkaline tendencies, within which three petrographic groups can be distinguished : (1) olivine basalts and basaltic andesites; (2) andesites and trachyandesites; (3) acid and alkaline leucocrate rocks.

The rocks display geochemical features of island arc volcanics. The basic and intermediary rocks belong to the "high-K andesites" series; it points out their alkaline trend, differentiating them from the rocks of the external volcanic arc of the South Carpathians, which display "medium-K andesites" characteristics.

The presence of the alkaline rocks both in the andesite group and in leucocrate group emphasizes the volcanics consanguinity.

The parental magma was formed at a depth of 200 km in the mantle suprajacent to the Benioff plane by the partial melting of the eclogitic source metasomatized by substances emanated from the subducted and eclogitized oceanic crust.

Magma displayed a dioritic composition with 58 per cent SiO_2 but it strongly differentiated generating mainly andesites and sporadically olivine basalts and leucocrate rocks.

References

Andrei J., Cristescu T. (1966) Asupra prezenţei unei fracturi crustale pe rama nordică a bazinelor Lăpugiu şi Timiş-Bega. *Stud. cerc. geol. geofiz. geogr., Geofiz.*, 4, 3, p. 311–315, Bucureşti.

– , Calotă C., Scurtu F. (1975) Considérations structurales sur le sillon ophiolitique des Monts Métallifères à l'aide des données géophysiques et de certains éléments quantitatifs obtenus grâce au modelage. *Rev. Roum. Géol., Géophysique*, 19, p. 101–111, Bucureşti.

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

Borcoş M., Popescu Gh., Roşu E. (1986) Nouvelles données sur la stratigraphie et l'evolution du volcanisme tertiaire des Monts Métallifères. *D. S. Inst. Geol. Geofiz.*, 70-71/4 (1983; 1984) p. 245–259, Bucureşti.

Cullers R. L., Graf J. L. (1984) Rare earth elements in igneous rocks of the continental crust: intermediate and silicic rocks – their petrogenesis. In: P. Henderson(edit), Rare Earth Element Geochemistry, p. 275–316, Elsevier, Amsterdam.

Ewart A. (1982) The mineralogy and petrology of Tertiary – Recent orogenic volcanic rocks: with special reference to the andesitic-basaltic compositional range. In R. S. Thorpe, J. Wiley (edits) Andesites. p. 25–27, Chichester.

Fountain J. C. (1979) Geochemistry of Brokeoff volcano, California. *Geol. Soc. Am. Bull.*, 90, p. 294–300, Boulder.

Ghiţulescu T. P., Borcoş M. (1966) Încadrarea funcţională a magmatismului alpin din Munţii Metaliferi. *Stud. cerc. geol., geofiz., geogr., Geologie*, 10, 2, p. 267–279, Bucureşti.

Gill J. (1981) Orogenic Andesites and Plate Tectonics. Springer-Verl., 390 p., Berlin.

Giuşcă D., Cioflica G., Savu H. (1966) Caracterizarea petrologică a provinciei banatitice. *An. Com. Stat Geol.*, XXXV, p.14–40, Bucureşti.

Green D. H. (1976) Experimental testing of "equilibrium" partial melting of peridotites under water-saturated, high-pressure conditions. *Can. Mineral.*, 14, p. 255–268, Toronto.

Hutchinson C. S. (1982) Indonesia. In R. S. Thorpe, J. Wiley (edits) Andesites. p. 207–224, Chichester.

Ianovici V., Giuşcă D., Ghiţulescu T. P., Borcoş M., Lupu M., Bleahu M., Savu H. (1969) Evoluţia geologică a Munţilor Metaliferi. 743 p., Bucureşti.

Irvine T. N., Baragar W. R. A. (1971) A guide to the chemical classification of common volcanic rocks. *Canad. J. Earth Sci.*, 8, p. 523–548, Ottawa.

Kadič O. (1906) Die geologischen Verhältnisse des Berglandes am linken Ufer der Maros in der Umgebung von Czela, Bulza und Pozsoga. *Jahr. d. k. ung. geol. Anst.*, 1904, p. 148–165, Budapest.

Institutul Geologic al României

Le Bas M. J., Le Maître R. W., Streckeisen A., Zanettin B. (1986) A chemical classification of volcanic rocks based on the total alkali-silica diagram. *Jour. Petrol.*, 27 (3), p. 745–750, Oxford.

Lupu M. (1975) Einige Bemerkungen zur Tektonik des südlichen Apuseni Gebirges (Siebenbürgischen Erzgebirge). *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 19, p. 95–104, Bucureşti.

Maier O., Solomon I., Zimmermann P., Zimmermann V. (1975) Studiul geologic şi petrografic al cristalinului din partea sudică a Munţilor Poiana Ruscă. *An. Inst. Geol. Geofiz.*, XLIII, p. 65–189, Bucureşti.

Mesorian H., Juteau T., Lapierre H., Nicolas A., Parrot J. F., Ricou L. E., Rocci G., Rollet M. (1973) Idées actuelles sur la constitution, l'origine et l'evolution des assemblages ophiolitiques mésogeens. *Bull. Soc. Géol. France.*, 15, p. 478–493, Paris.

Minear J. W., Toksöz M. N. (1970) Thermal regime of a downgoing slab and new global tectonics. *J. Geöphys. Res.*, 75, 8, p. 1397–1419, Richmond.

Mysen B. O. (1979) Trace element partitition between garnet peridotite minerals and water-rich vapor: experimental data from 5 to 30 Kba. *Am. Mineral.*, 64, p. 274–287, Washington.

– , Boettcher A. I. (1975) Melting of a hydrous mantle: II. Geochemistry of crystals and liquids formed by anatexis of mantle peridotite at high pressure and high temperature as a function of controlled activities of water, hydrogen and carbon dioxide. *J. Petrol.*, 16, p. 459–593, Oxford.

Ninkovich D., Hayes J. D. (1971) Tectonic setting of Mediterranean volcanics. *Acta Int. Sci. Congr. Volcano Thera (Greece)*, 1, p. 111.

Palacios M. C. (1984) Considerations about the plate tectonic model, volcanism and continental crust in the southern part of the Central Andes. *Tectonophysics*, 106, p. 205–214, Amsterdam.

Papiu V. C. (1954) Géologie de la région Valea Mare-Căprioara-Bulza-Pojoga. *C. R. Inst. Géol. Roum.*, XXXVIII, p. 169–178, Bucureşti.

Pearce J. A. (1980) Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites. In A. Panayiotou. Ophiolites. *Proc. Intern. Ophiol. Symp. Cyprus, 1979*, p. 261–272, Nicosia.

– , Gale G. H. (1977) Identification of the deposition environment from trace-element geochemistry of associated igneous host rocks. In M. Jones Volcanic processes in ore genesis. *Inst. Mining and Metallurgy and Geol. Soc. Special Publ.*, 7, p. 14–24, London.

Peltz S., Peltz M. (1965) Notă asupra unor iviri de şisturi cristaline în regiunea Tisa-Ioneasa. *D. S. Com. Geol.*, LI/1, p. 109–116, Bucureşti.

Rădulescu D., Borcoş M. (1968) Aperçu général sur le déroulement du volcanisme néogène en Roumanie. *An. Com. Stat Geol.*, XXXVI, p. 177–193, Bucureşti.

– , Săndulescu M. (1973) The plate-tectonics concept and the geological structure of the Carpathians. *Tectonophysics*, 16, p. 155–161, Amsterdam.

Rocci G., Baroz F., Bebien J., Desmet A., Lapierre H., Ohnenstetter D., Ohnenstetter M., Parrot J. F. (1980) The Mediterranean ophiolites and their related Mesozoic volcano-sedimentary sequences. In A. Panayiotou (edit.) Ophiolites. *Proc. Intern. Ophiolite Symp. Cyprus, 1979*, p. 273–286, Nicosia.

Roden M. F., Murthy V. R. (1985) Mantle metasomatism. *Ann. Rev. Earth Planet. Sci.*, 13, p. 269–296, Palo Alto.

Saunders A. D. (1984) The rare earth element characteristics of igneous rocks from the ocean basins. In P. Henderson (edit.). *Rare Earth Element Geochemistry*, p. 205–236, Amsterdam.

Savu H. (1962) Asupra erupţiilor neogene din partea de nord a masivului Poiana Ruscă. *D. S. Com. Geol.*, XLIII, p. 113–130, Bucureşti.

– (1966) Metamorfismul calcarelor la contact cu ofiolitele de la Căpîlnaş (Valea Mureşului). *D. S. Com. Geol.*, LII/1, p. 123–140, Bucureşti.

– (1983) Geotectonic and magmatic evolution of the Mureş Zone (Apuseni Mountains). Carp.-Balk. Assoc. 12th Congr. Bucharest, 1981, *An. Inst. Géol. Géophys.*, LXI, p. 253–262, Bucureşti.

– (1984) Mélange-ul cu matrice piroclastică asociat arcului insular sudic al Zonei Mureş. *Stud. cerc. geol., geofiz., geogr., Geologie*, 29, p. 36–43, Bucureşti.

– , Udrescu C., Neacşu V., Bratosin I., Stoian M. (1985) Origin, geochemistry and tectonic position of the Alpine ophiolites in the Severin Nappe (Mehedinţi Plateau, Romania). *Ofioliti*, 10 (2/3), p. 423–440, Bologna.

– , Udrescu C., Lemne M., Romanescu O., Neacşu V. (1988) Petrology, geochemistry and tectonics of the ophiolites and Late Kimmerian island arc volcanics from the Glodghileşti-Săliştioara tectonic rise (Mureş Zone). *D.S. Inst. Geol. Geofiz.*, 72-73/5, p. 259–281, Bucureşti.

– , Udrescu C., Lemne M., Romanescu O., Stoian M., Neacşu V. (1987) Island arc volcanics related to the wildflysch on the outer margin of the Danubian Autochthon (South Carpathians) and their geotectonic implications. *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 31, p. 19–27, Bucureşti.

Stan N., Stănoiu I., Năstăseanu S., Moisescu V., Seghedi A., Pop Gr. (1979) Harta geologică a R.S.R., scara 1:50 000, foaia Cîmpu lui Neag, Bucureşti.

Sugimura A. (1968) Spatial relations of basaltic magmas in island arcs. In A. Poldervaart (edit.) Basalts, p. 537–571.

Taylor S. R. (1968) Geochemistry of andesites. In L.H. Ahrens (edit.), Origin and Distribution of the Elements. p. 559–583, Oxford.

Institutul Geologic al României

Uyeda S. (1981) Some thoughts on geodynamics of Asia. *Reports Geol. Surv. Jap.*, 261, p. 1–6, Higashi.

*** Penrose Field Conference on Ophiolites. *Geotimes*, 17, p. 24–25, 1972, Alexandria, Virginia.

*** Conference on the scientific program of the project "Ophiolites of Continents and Comparable Oceanic Rocks". Moscow, 1974.

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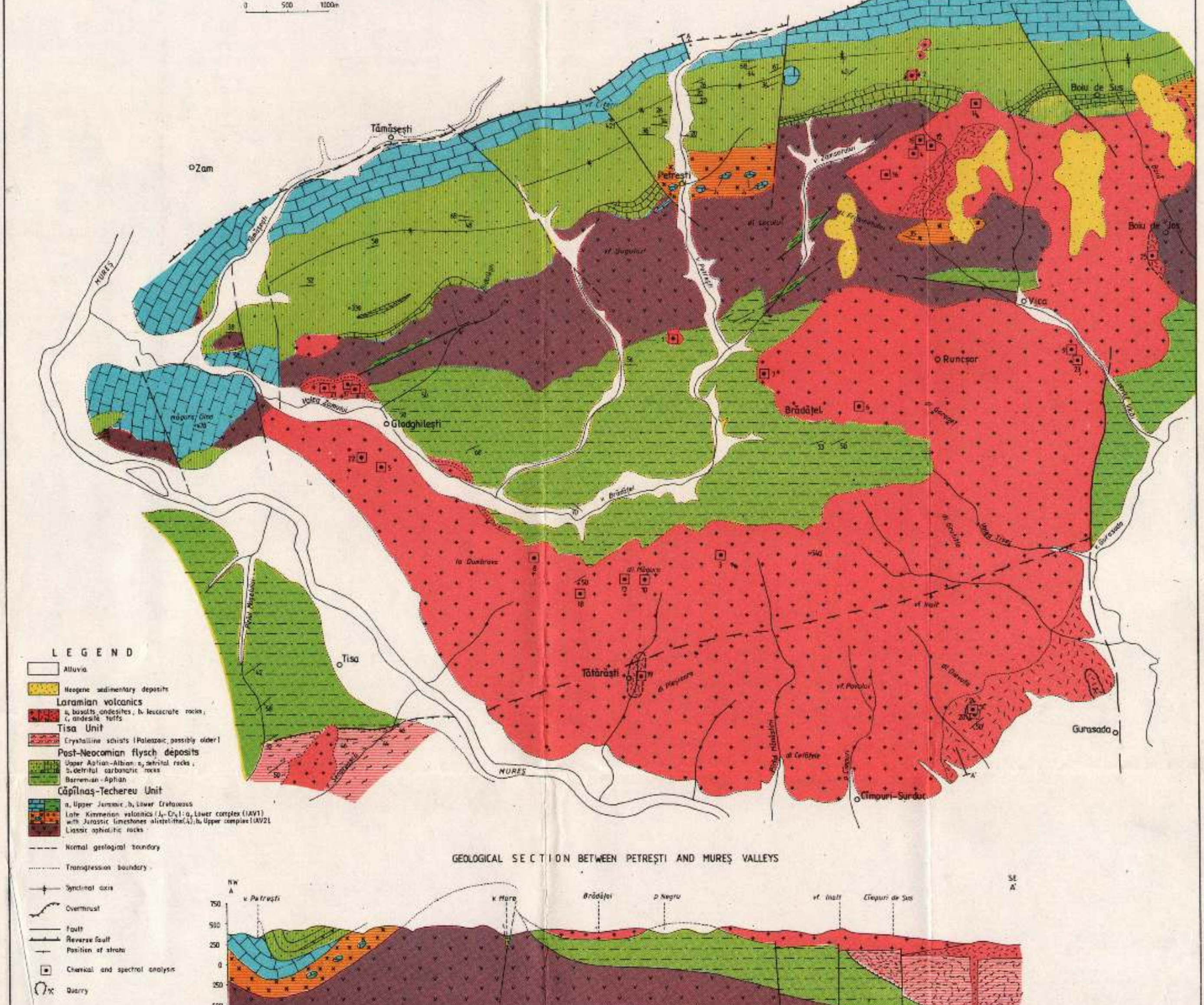
GEOLOGICAL MAP OF THE ZAM-BOIU DE SUS-GURASADA AREA



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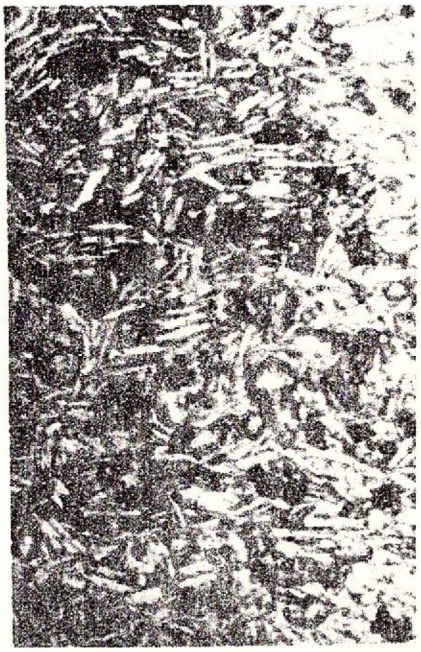

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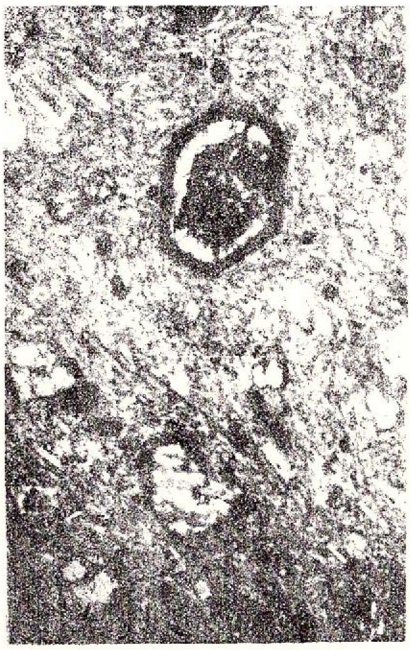

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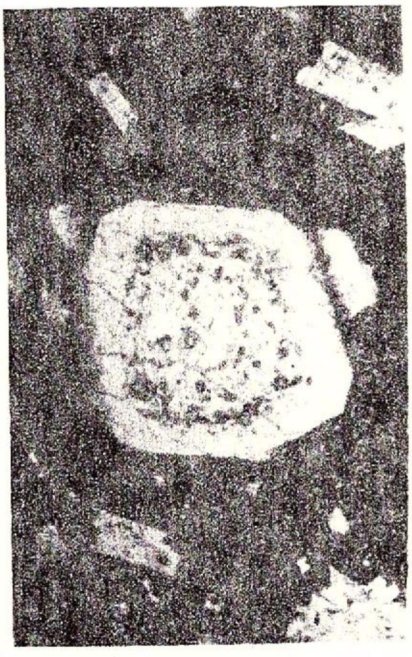

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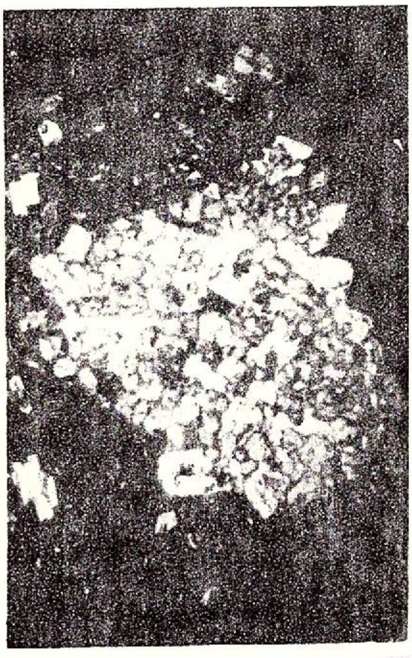

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Plate II

Fig 1 – Structure of the olivine basalt groundmass. Petreşti. Nic. II, x 44.

Fig. 2 – Pyroxene hornblende andesite within which the amphibole phenocrysts are opacitized on margins. Glodghileşti. Nic. II, x 34.

Fig. 3 – Hornblende pyroxene plagioclase andesite within which the plagioclase phenocrysts display glassy inclusions with a zonary disposition. Pîrîul Negru. Nic. II, x 50.

Fig. 4 – Microgabbrodiorite xenolith within a hornblende pyroxene plagioclase andesite. Petreşti. Nic. II, x 20.

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DONNÉES PRÉLIMINAIRES CONCERNANT LA CONSTITUTION GÉOLOGIQUE DE LA ZONE DE SĂPÎNŢA–VALEA BRAZILOR (MONTS DE IGNIŞ)

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Key words: Igneous rocks. Andesites. Magmas. Calc–alkalic composition. Hydrothermal alteration. Haloes. Major elements. East Carpathians – Neogene eruptive–Quaternary – Igniş Mountains.

Abstract: *Some Problems of the Geological Structure of the Săpînţa – Valea Brazilor Area (the Igniş Mts) – Preliminary Data.* In the central part of the northern slope of the Igniş Mts there is a complex intrusive body made up of pyroxene andesites ± quartz, quartz andesites passing to porphyry quartz diorites, clinopyroxene bearing andesites. It is supposed that the intrusions are newer than the lava flows made up of more pyroxene andesites varieties and that both the hydrothermal transformations and the mineralization are associated to the emplacement stage of the intrusions. A second complex igneous body is deduced from the geophysical data.

La zone en discution a une surface de 13 km^2 et se situe dans la partie centrale du versant septentrional des monts de Igniş. Des contributions importantes à la connaissance géologique de la région ont été apportées par: Gheorghiţă, Dofescu (1962); Gheorghiţă (1964); Lang (1972, 1973); Rădulescu, Lang (1973); Edelstein et al. (1978–1980), dans des études publiées et Cristescu, Ştefănciuc (1967); Stan et al. (1968); Andrei et al. (1970); Borcoş et al. (1971); Cojocea et al. (1971); Edelstein et al. (1971); Edelstein et al. (1983); Maran, Mihale (1971); Pătraşcu (1976); Rusu (1971); Vâjdea et al. (1976); Demetrescu et al. (1980); Veliciu, Diaconu (1980); Vlaicu, Vlaicu (1982); Scurtu et al. (1983), dans des études non-publiées.

La présente note est fondée spécialement sur les données fournies par les prospections effectuées par l'équipe de prospections de Baia Mare, en 1981 (Edelstein et al., 1983).

Données géologiques générales

Par analogie avec les manifestations magmatiques du versant septentrional des monts de Igniş, dans la zone de Săpînţa–Valea Brazilor, on peut séparer, plus ou moins clairement, trois étapes:

– la première étape a eu un caractère explosif et subordonnément effusif, engendrant une formation vulcano-sédimentaire;

– pendant la deuxième étape a eu lieu la mise en place d'une succession de laves, généralement andésites pyroxéniques, qui a déterminé la réalisation de l'édifice volcanique soumis ultérieurement à l'érosion. Vers la fin de l'étape ont été mises en place des dacites, probablement sous forme de petites extrusions;

– la troisième étape vise le corps à caractère intrusif.

En absence des données certes, paléontologiques ou d'âge K–Ar, on considère que toutes ces étapes peuvent être encadrées au Pliocène. La flore prélevée de la partie supérieure de la formation vulcano-sédimentaire, qui affleure dans le versant gauche de la vallée de Runcu–Săpînţa (au nord de la zone analysée dans cette ouvrage), comportant *Populus cf. tremula* L. et *Castanea Kubinyii*, KOV., est considérée par Givulescu (1977, données non-publiées) d'âge pontien–dacien.

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La formation vulcano-sédimentaire n'affleure pas dans la zone de Săpînţa–Valea Brazilor, mais elle a été interceptée par le forage B-vallée de Săpîncioara, entre 257,6 et 290,3 m (le niveau inférieur du forage étant situé à la cote 700), étant représentée par des épiclastites.

Dans la zone de Valea Runcului–Săpînţa, la formation affleure entre les cotes 630 et 800; dans le forage situé dans la vallée du ruisseau de Ghiurchi elle est interceptée entre les cotes 520 et 300. Ce forage nous a offert, jusqu'à présent, l'image la plus complète sur la formation vulcano-sédimentaire et c'est ici qu'elle a été le mieux étudiée (Lang, 1972; Edelstein et al., 1972–1973, données non-publiées). La formation comporte des écoulements de laves de andésites pyroxéniques, pyroclastites andésitiques (brèches et microbrèches pyroclastiques, tufs fins et grossiers où les éléments ont la même nature que les intercalations de laves), ainsi que des roches épiclastiques (conglomérats, grès andésitiques et roches pélitiques). La formation a une épaisseur de presque 850 m, les laves andésitiques représentant seulement 23 %; la puissance maximum d'un écoulement de laves ne dépasse pas 60 m. Y prédominent les andésites hypersthénique où le rapport hypersthène/augite est de 10:5. Les andésites pyroxéniques intercallées représentent les roches les plus basiques qui apparaissent dans le versant septentrional des monts de Igniş; on observe une tendence de croissance de l'acidité, des écoulements inférieurs vers ceux supérieurs.

Le complexe effusif, bien qu'il comporte surtout des andésites pyroxéniques et ait un aspect homogène et unitaire, on constate, à une investigation minutieuse, qu'il présente des écoulements de laves successives, déterminés par des caractéristiques minéralogiques, structurales, texturales ou de composition chimique. Pétrographiquement, il comporte des andésites et dacites.

Quant aux andésites, on a séparé des andésites exclussivement pyroxéniques (la majorité), des andésites pyroxéniques à hornblènde (dans la zone de la crête de Brazi), ainsi que des andésites pyroxéniques à quartz, hornblende et biotite (dans la zone de la crête de Braga). Selon la teneur en pyroxènes dans la masse de la roche (notamment le rapport entre ortho et clinopyroxènes) et en verre et tenant compte de certaines particularités de la composition, on a séparé, dans le cadre des andésites exclusivement pyroxéniques, plusieurs types et sous-types, tels: andésites à pyroxènes de Valea Brazilor, où les pyroxènes ne représentent jamais plus de 50 % de la masse de la roche et renferment de fréquentes séparations primaires à caractère gloméroporphyrique; andésites pyroxéniques à verre et andésites vitreuses qui comportent du verre representant 15 à 50 % de la masse de la roche; andésites pyroxéniques à séparations feldspathiques sous forme de rubans leucocrates qui impriment à la roche un aspect rubané (séparations constituées d'oligoclase±quartz et des quantités réduites de biotite); andésites pyroxéniques microcristallinnes de la zone de la crête de Rotunzi et d'au-dessous de la crête de Brazi, dont les fénocristaux ne dépassent pas 1,5 mm, la dimension moyenne étant inférieure à 1 mm; il y a aussi d'autres variétés moins individualisées cartographiquement.

Tableau 1

Succession presumée, simplifiée des produits éruptifs de la zone de Săpînţa–Valea Brazilor

Complexe intrusif	Andésite pyroxénique, augitique surtout
	Diorite quartzifère porphyrique et andésite quartzifère à pyroxénes, biotite et hornblende
	Andésite pyroxénique ± quartz
Complexe effusif	Dacite à biotite de Pleşca Mare
	Andésite pyroxénique à quartz, amphiboles et biotite de Braga
	Andésite pyroxénique à amphiboles du sommet de Brazi
	Andésite pyroxénique microcristalline de Rotunzi
	Andésite pyroxénique de Vîrful Stînilor
	Andésite pyroxénique à verre et andésite pyroxénique vitreuse
	Andésite pyroxénique à séparations feldspathiques et andésite pyroxénique±quartz
	Andésite à pyroxénes de Valea Brazilor
	Andésite pyroxénique hydrothermalisée et andésites pyroxéniques de Mlăci
Formation vulcano-sedimentaire	

Dans le cadre des dacites à biotite on a séparé une variété hyaline, moins acide, considérée comme faciès marginal d'une petite extrusion, localisée aux sources de la Valea Brazilor de Dreapta.

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Le tableau 1 présente la succession presumée des produits eruptifs de la zone de Săpînţa–Valea Brazilor; cette succession a un caractère hypothétique et sera confirmée ou infirmée par les travaux qu'on va exécuter. La figure 1 présente la carte géologique simplifiée.

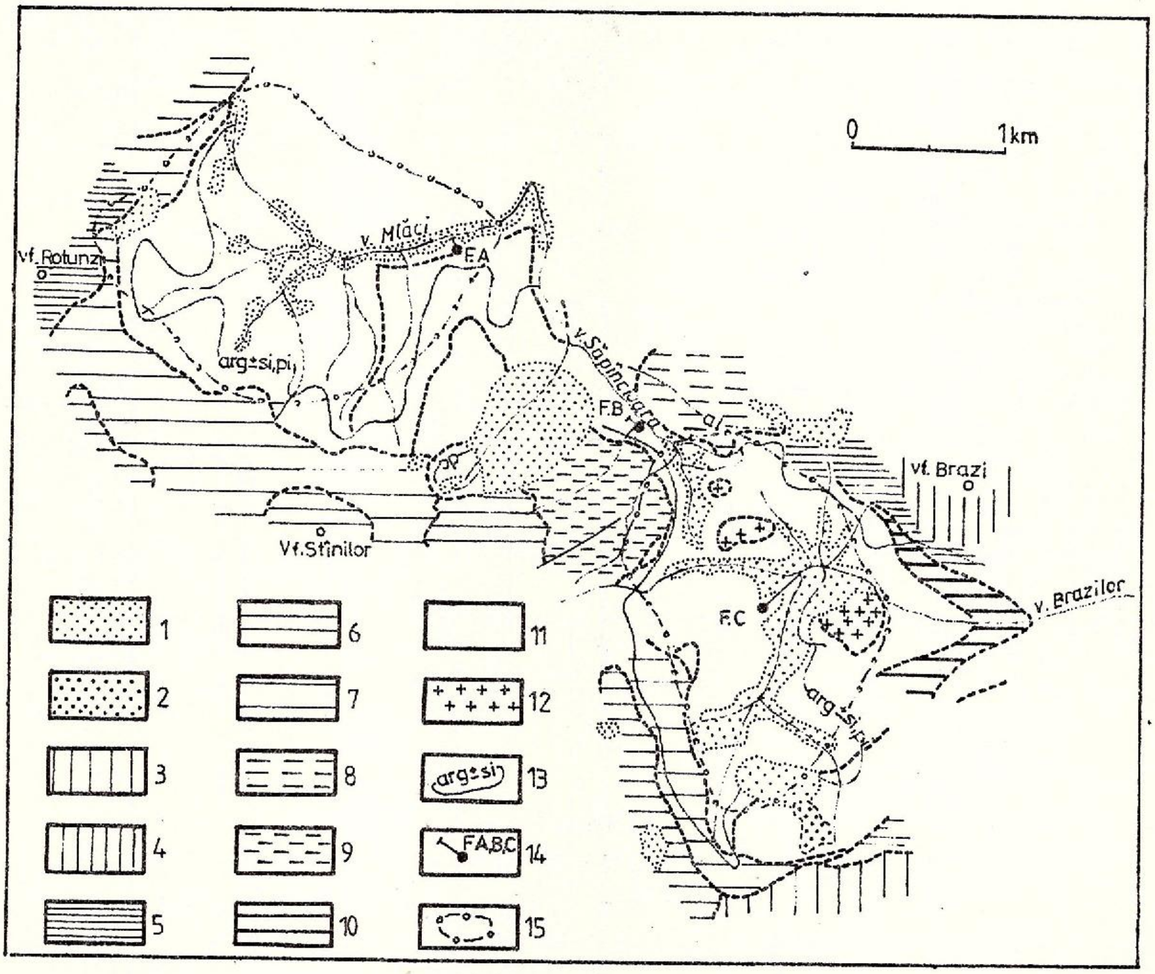

Fig. 1 - Carte géologique de la zone de Săpînţa–Valea Brazilor.
1, dépôts quaternaires; 2, dacites à biotit; 3, andésite pyroxénique à quartz, amphiboles et biotite, de Braga; 4, andésite pyroxénique à amphiboles du sommet de Brazi; 5, andésite pyroxénique microcristalline; 6, andésite pyroxénique de Vîrful Stînilor; 7, andésite pyroxénique au verre et andésite pyroxénique vitreuse; 8, andésite pyroxénique ± quartz; 9, andésite pyroxénique à séparations feldspathiques; 10, andésite à pyroxénes de Valea Brazilor; 11, andésite pyroxénique hydrothermalisée et andésite pyroxéniques de Mlăci; 12, corps intrusifs présumés; 13, zones à des roches transformées du point de vue hydrothermal; arg, argilisation; si, silicifications; pi, pyritisations; op, opalicifications; al, alunitisations; 14, forage éxécuté; 15, le contour supposé des corps éruptifs majeurs.

La composition minéralogique des principales variétés de roches, qui participent à la formation du complexe effusif, est présentée dans le tableau 2.

Le long d'un affluent de droite de la vallée de Săpînţa on a observé des pyroclastites sous forme de fragments dissipés, tufs vitreux cristallo-lithoclastiques à opale et alunite; dans le bassin de Valea Brazilor il y a tuffs cristallo-clastiques à cristaux de quartz silicifiés, argilisés.

Le complexe intrusif comporte le corps intercepté par le forage C-Valea Brazilor. On lui attribue aussi, sous réserve, d'autres roches à aspects structuraux et texturaux propres pour certains produits intrusifs. La figure 2 présente la composition du corps intercepté par le forage C et le tableau 3 les compositions minéralogiques des roches qui forment le complexe intrusif.

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Tableau 2

Composition minéralogique des principaux types de roches du complexe effusif

Nr. crt.	Types de roches	Phénocristaux							Masse fondamentale	
		Plagioclases		Quartz	Pyroxènes		Amphiboles	Biotite	%	Structure, texture
		%	% An		opx	cpx				
1.	Dacite à biotite de Pleşca Mare	16–20	26–38	4–10	0,5–2*		2–4	4–5	66–68	phellsophyrique non orientée
2.	Andésite pyroxénique à quartz, hornblende et biotite du sommet de Braga	20	36–52	1	10*		1	sp.	68	hyalopilitique et eutectophyrique
3.	Andésite pyroxénique à amphiboles du sommet de Brazi	16–20			6–8*		2–3		68–74	pilotaxitique orientée
4.	Andésite pyroxénique microcristallin de Rotunzi	26	35–60		5	2			67	finement intergrenue orientée
5.	Andésite pyroxénique de Vîrful Stînilor	22–27	36–53		5–10	2–7			56–71	microlitique, fluidale
6.	a) Andésite pyroxénique à verre	21–26	36–62		6–8	3–5			63–69	hyalopilitique orientée
	b) Andésite pyroxénique vitreuse	23–30	29–54		6–8	2–5			59–66	vitrophyrique
7.	Andésite pyroxénique à séparations feldspathiques	23–28	24–67		4–7	2–4			64–67	microlitique fluidale
8.	Andésite à pyroxénes de Valea Brazilor	28–32	50–56		3–4	0,5–1			65–68	finement intergrenue non orientée

* Pyroxénes

Tableau 3

Composition minéralogique des principales roches intrusives du forage C–Valea Brazilor

No.	Type de roche	Phénocristaux						Masse fondamentale (mésostase)
		Plagioclases	Q	Pyroxénes		Hornblende	Bi	
				Opx	Cpx			
1.	Andésite piroxénique, augitique surtout	12–23		4–6	8–14			65–75
2.	Andésite pyroxénique ± quartz	26–29	sp*	4–10**				62–68
3.	Andésite-diorite quartzifère porphyrique	26–35	1–3	5–13**		0,5–3	sp–1	51–63

* sp = sporadique

** Pyroxénes

Le corps intrusif intercepté par le forage C comporte plusieurs types pétrographiques à nombreuses variétés. En général on peut parler de trois types principaux (fig. 2, tab. 3): andésite pyroxénique à augite dans l'intervalle de 49,2 à 55,6 m; 148,5 à 155,5 m et 172,2 à 255,0 m; andésite pyroxénique ± quartz dans l'intervalle de 255,0 à 459,0 m, auquel on rapporte aussi, sous réserve, andésite pyroxénique compris entre 652 et 1066 m; andésite quartzifère avec des passages à diorite quartzifère porphyrique à pyroxènes, hornblende et biotite, amplement cristallisé dans l'intervalle de 459 à 625 m et présentant des caractères andésitiques dans l'intervalle de 1066 à 1260 m. Dans le cadre des autres deux types principaux on a évidentié des aspects structo-texturaux, soit à caractère de transition, soit différentiés entre eux par le degré de cristallisation, granulation – en tant que dimensions absolues et relatives – ainsi que par le degré de l'idéomorphisme. On a observé aussi des aspects qui suggèrent des roches de type aplitique et lamprophyrique.

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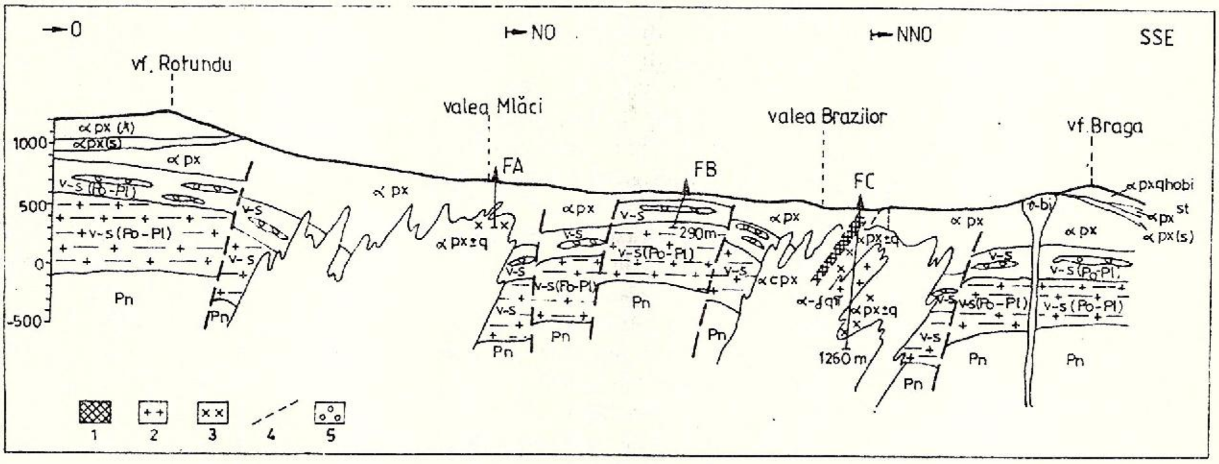

Fig. 2 – Coupe géologique par les forages exécutés dans la zone de Săpînţa–Valea Brazilor.
Complexe intrusif: 1, acpx, andésite augitique; 2, a–dqP, andésite quartzifère–diorite quartzifère porphyrique; 3, apx±q, andésite pyroxénique ± quartz; 4, fractures; 5, pyroclastites. Complexe effusif: Dbi, dacite à biotite de Pleşca Mare; apxqhobi, andésite pyroxénique à quartz, amphiboles et biotite de Braga; apx(m), andésite pyroxénique microcristallin de Rotunzi; apx St, andésite pyroxénique de Vîrful Stînilor; apx(s), andésite pyroxénique à verre et andésite pyroxénique vitreuse; apx, andésite pyroxénique de Mlăci et andésite pyroxénique hydrothermalisée. Complexe vulcano-sédimentaire: V–S, la partie supérieure à des fréquents niveaux de lave; V–S (Po–Pl), prédominament épiclastique; Pn, sédimentaire pannonien.

Chimisme des roches

La composition chimique des roches éruptives est discutée sur des analyses effectuées. Des données présentées dans le tableau 4 résulte:

- par rapport à la composition moyenne des andésites du versant septentrional (Lang, 1976), les andésites de la zone de Săpînţa–Valea Brazilor sont plus acides; la même différence est plus marquante vis-à-vis des andésites pyroxéniques de Seini (Stan et al., 1983, données non publiées), de Ilba (Stan et al., 1983, données non publiées), de Jereapăn (Borcoş et al., 1979), de l'ensemble d'andésites pyroxéniques des Carpathes Orientales (Peltz et al., 1982), ainsi que des andésites basaltiques des Transcarpathes de l'Ukraine;
- la composition moyenne des andésites pyroxéniques de la zone de Săpînţa–Valea Brazilor est proche des valeurs moyennes des andésites (tab. 5);
- à remarquer la teneur plus réduite en Na_2O par rapport aux autres andésites des monts Igniş, des Carpathes Orientales et notamment de la moyenne des andésites (plus de 1 %); probablement, la teneur en K_2O est plus élévée que pour toutes les autres roches comparées, mais sans noter de grandes différences.

L'indice Kuno évidencie le caractère basique des roches, les andésites pyroxéniques ayant des valeurs entre 29 et 20.

Ayant en vue la participation massive de la masse fondamentale dans la constitution de la roche (souvent plus de 65 %), ainsi que l'impossibilité de determiner avec précision la composition minéralogique, le nom des roches est établi sur base des diagrammes de la composition chimique: TAS (fig. 3) et Peccerillo, Taylor (1976)(fig. 4).

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Tableau

Composition chimique des roches éruptives

No.	Echant.	Localisation	SiO_2	Al_2O_3	Fe_2O_3
1	1235 BE	Valea Brazilor de Stînga	53,45	23,29	5,26
2	1361 E	Valea Brazilor	54,05	18,90	3,48
3	26771-1	Source Valea Brazilor	56,30	16,02	3,00
4	1732 E	Au nord du sommet de Rotunzi	56,75	17,72	4,88
5	1747 F	Source de la vallée de Mlăci	57,60	19,69	1,42
6	6464 F	Versant gauche de la vallée de Săpîncioara	58,35	18,09	1,00
7	7862 L	Vallée de Mlăci	58,80	17,94	1,40
8	26642-1	Valea Brazilor de Stînga	58,85	13,55	5,83
9	1600 E	Sommet de Braga	58,95	19,57	1,88
10	1406 E	Valea Coconului	59,55	15,32	4,71
11	7886 L	Versant droit de la vallée de Mlăci	59,60	17,96	1,53
12	27055-1	Source de la vallée de Mlăci	60,00	16,02	2,84
13	26579-1	Sommet de Brazi	58,30	21,03	1,03
14	26660-1	Valea Brazilor de Stînga	61,75	14,88	2,55
15	1557-AE	Afluent de droite de Valea Brazilor	62,75	17,94	2,08
16	1572-AE	Source de Valea Brazilor	65,00	18,00	2,22
17	1592-E	Sommet de Pleşca Mare	68,00	14,86	2,07
18	F.C-471	Forage C à 471,5 m	60,30	15,64	3,60
19	F.C-626	Forage C à 629 m	61,00	14,85	3,76
20	F.C-484	Forage C à 484,5 m	61,20	13,62	4,24
21	F.C-605	Forage C à 605 m	61,30	13,65	6,00

1, andésite à pyroxènes de Valea Brazilor; 2, andésite pyroxénique à quartz; 3, andésite pyroxénique de Vîrful Stînilor; 4, andésite pyroxénique microcristallin; 5, andésite pyroxénique; 6,7,11,12, andésite pyroxénique de Mlăci; 8, andésite pyroxénique à verre; 9, andésite pyroxénique à quartz, amphiboles, biotite; 10, andésite pyroxénique à séparations feldspathiques;

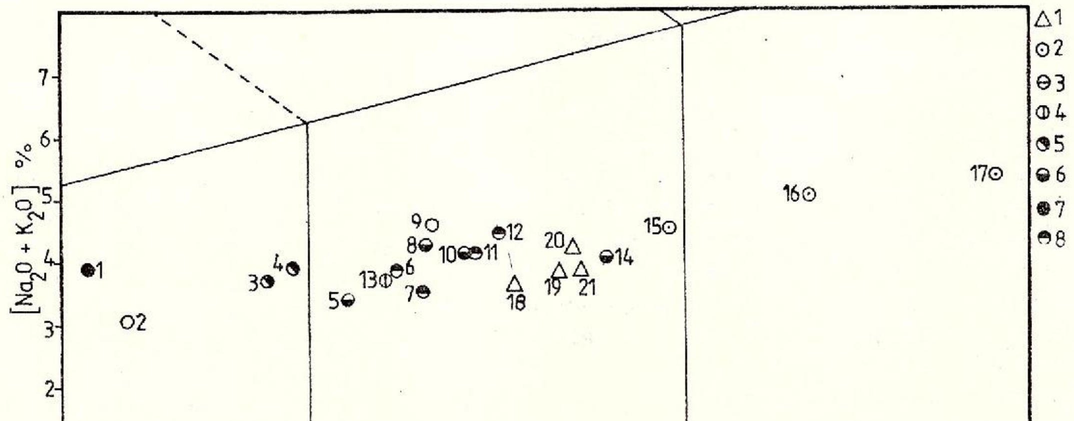

Fig. 3 – La position des roches éruptives de la zone de Săpinţa–Valea Brazilor dans le diagramme TAS.

1, andésites quartzifères-diorites quartzifères du forage C; 2, dacites de Pleşca Mare; 3, andésites pyroxéniques à quartz, amphiboles et biotite de Braga; 4, andésite pyroxénique à amphiboles du sommet de Brazi; 5, andésite pyroxénique microcristallin de Rotunzi et andésite pyroxénique de Vîrful Stînilor; 6, andésite pyroxénique à séparations feldspathiques, andésite pyroxénique à verre, andésite pyroxénique vitreuse et andésite pyroxénique comportant deux générations de phénocristaux; 7, andésite à pyroxénes de Valea Brazilor; 8, andésites pyroxéniques de Mlăci.

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4
de la zone de Săpînţa-Valea Brazilor

FeO	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	CO_2	H_2O^+	H_2O^-
3,29	0,16	2,75	8,12	2,64	1,26	0,83	0,04		0,70	0,65
5,85	0,18	3,55	7,72	1,88	1,20	0,93	0,09		1,05	0,60
6,01	0,16	4,55	8,35	1,97	1,75	0,78	0,04		1,00	0,20
4,72	0,11	3,80	7,00	2,37	1,52	0,90	0,07		0,30	0,40
5,58	0,15	3,85	6,44	2,10	1,32	0,88	0,03		0,85	0,10
5,98	0,18	3,85	7,02	2,02	1,85	0,75	0,04		1,15	0,30
4,86	0,13	3,55	6,30	2,05	1,47	0,83	0,05		1,10	1,15
5,37	0,28	4,50	4,90	2,34	1,90	0,87	0,06		0,35	0,40
4,30	0,15	3,50	5,32	2,65	1,97	0,83	0,06		0,35	0,85
4,41	0,28	3,50	5,75	2,47	1,68	0,95	0,16		0,45	0,65
4,29	0,10	3,50	5,46	2,16	1,97	0,80	0,04		1,00	1,15
5,32	0,21	1,70	5,16	2,35	2,07	0,73	0,04		3,10	0,80
4,29	0,10	3,90	5,95	1,89	1,89	0,68	0,05		0,20	0,45
4,00	0,13	3,30	6,02	2,02	2,03	0,70	0,04		2,10	0,20
3,71	0,10	2,90	3,38	2,35	2,17	0,63	0,03		1,80	0,20
1,85	0,06	2,75	3,08	2,50	2,54	0,53	0,02		0,90	0,85
1,57	0,08	1,85	2,95	2,60	2,73	0,40	0,02		0,70	0,50
4,22	0,08	3,05	6,30	2,20	1,49	0,70	0,10		1,10	0,45
3,45	0,15	3,20	5,90	2,09	1,79	0,55	0,10	0,20	2,07	0,30
3,75	0,13	2,80	6,59	2,23	1,97	0,75	0,10		1,55	0,40
3,60	0,13	2,80	5,25	2,23	1,85	0.85	0.10	0.25	1.60	0.25

13, andésite pyroxénique à amphiboles; 14, andésite vitreuse; 15, dacite vitreuse à biotite; 16, dacite à biotite; 17, dacite à biotite de Pleşca; 18, 20, diorite quartzique porphyrique; 19, 21, andésite quartzique. Analystes: V. Mercheş, (1-15, 18, 19); I. Apostolescu, (17, 21); R. Cîrciumaru (16, 20); I. P. G. G. , Bucureşti.

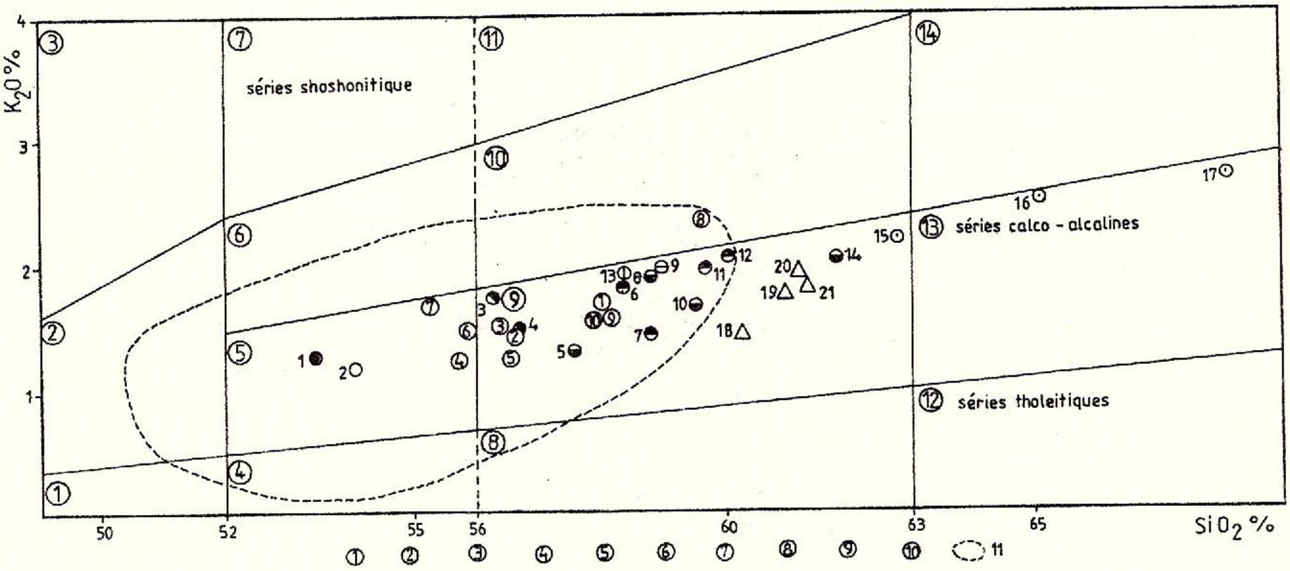

Fig. 4 La position des roches de la zone de Săpînţa Valea Brazilor dans le diagramme Peccerillo et Taylor (1976). Valeurs moyennes: 1, andésites pyroxéniques de la zone de Săpînţa-Valea Brazilor; 2, andésites pyroxéniques du versant septentrional des monts d'Igniş (Lang, 1976); 3, andésites pyroxéniques des Carpathes Orientales (Peltz et al., 1982); 4, andésites pyroxéniques de Seini (Stan et al., 1983); 3, andésites pyroxéniques d'Ilba (Stan et al., 1983); 6, andésites pyroxéniques ± hornblende de Jereapăn (Borcoş et al., 1979); 7, andésites basaltiques de la zone transcarpathique; 8, andésites de la zone transcarpathique; 9, andésites cénozoïques; 10, andésites; 11, zone des andésites pyroxéniques du versant septentrional des monts d'Igniş (Lang, 1976).

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Tableau 5

Les teneurs moyennes en oxydes des andésites pyroxéniques

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	58,02	56,79	56,49	55,64	55,55	55,97	55,25	59,71	58,17	57,92	56,93
Al_2O_3	17,85	18,25	18,70	16,85	17,50	17,90	18,67	17,66	16,26	16,88	17,33
Fe_2O_3	2,88	4,12	3,60	4,19	2,80	7,88	3,09	3,16	3,07	3,30	3,88
FeO	4,87	4,19	2,95	4,56	4,75		4,35	3,06	4,17	4,13	4,14
MnO	0,16	0,17	0,16	0,17	0,20		0,08	0,08		0,14	0,16
MgO	3,55	3,43	3,68	3,57	3,85	3,65	3,71	2,48	3,23	3,41	3,43
CaO	6,39	7,21	7,25	7,28	7,47	7,55	7,74	6,03	6,93	6,84	7,07
Na_2O	2,20	2,41	3,55	2,24	2,22	2,41	2,48	2,75	3,21	3,42	2,54
K_2O	1,70	1,55	1,59	1,25	1,27	1,49	1,66	2,35	1,61	1,60	1,54
TiO_2	0,81	0,68	0,82	1,03	1,03		0,74	0,65	0,80	0,86	0,79
P_2O_5	0,05	0,14	0,16	0,12	0,16		0,20	0,20	0,20	0,21	0,22

1, andésite pyroxénique – Săpînţa–Valea Brazilor (y compris l'andésite pyroxénique à quartz et celle à amphiboles), (14 analyses); 2, andésites pyroxéniques de Mara et Săpînţa (18 analyses) (Lang, 1976); 3, andésites pyroxéniques des Carpathes Orientales (53 analyses) (Peltz et al., 1972 in Lang, 1976); 4, andésites pyroxéniques sarmatiennes de Seini – la zone Ilba–Seini, (26 analyses) (Stan et al., 1983); 5, andésites pyroxéniques pannoniennes d'Ilba (9 analyses) (Stan et al., 1983); 6, andésites pyroxéniques pontiennes de Jereapăn (Borcoş et al., 1979); 7, andésites basaltiques de la zone transcarpathique; 8, andésites de la zone transcarpathique; 9, andésites cénozoïques (1775 analyses); 10, andésites (1981 analyses); 11, andésites – Oaş–Văratec (206 analyses) (Kovacs et al., 1986).

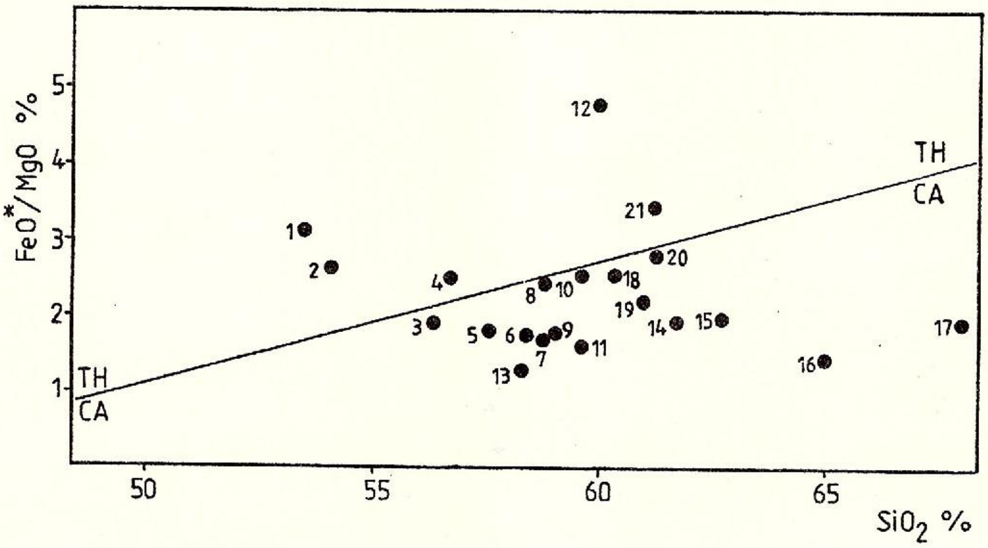

Fig. 5 – Diagramme FeO/MgO–SiO_2 (avec la position des séries magmatiques. TH, série tholéiitique; CA, série calco-alcaline.

Les roches de la zone de Săpînţa–Valea Brazilor appartiennent à l'association des roches calco-alcalines. La valeur élevée de l'indice Peacock (60–62), similaire aux valeurs calculées pour les roches d'autres secteurs des monts d'Oaş–Văratec et Ţibleş, relève leur caractère calcique. L'indice Rittmann, à des valeurs entre 1,12 et 1,50 et la moyenne 1,34, relève le caractère fortement pacifique. Le caractère calco-alcalin des magmas qui ont engendré les roches de la zone de Săpînţa–Valea Brazilor résulte des diagrammes FeO/MgO–SiO_2 (fig. 5) et AFM (fig. 6). (Les diagrammes qui renferment aussi les valeurs Na_2O, qui se situent, selon on a déjà vu, avec 1 % sous la moyenne des roches similaires d'autres zones, manifestent une faible tendence tholéiitique).

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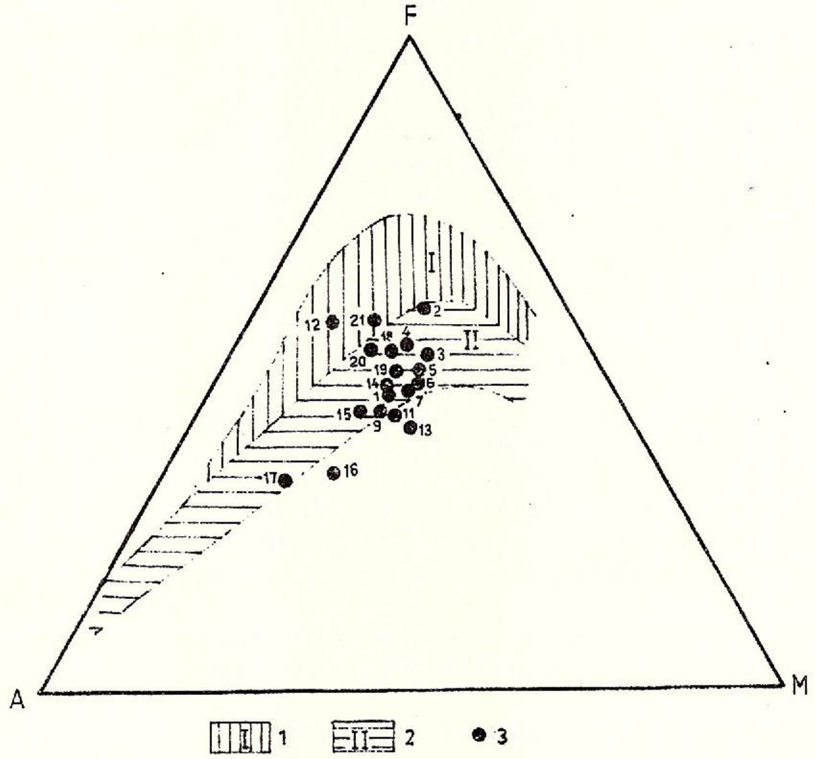

Fig. 6 – Diagramme AFM (selon Kuno 1968, in Girod, 1978).
I, domaine de la série tholéiitique (pigéonitique); II, domaine de la série calco-alcaline (hypersthénique).

Métamorphisme hydrothermal

Les roches hydrothermalisées occupent approximativement 40 % de la zone. On a individualisé, à la surface, trois auréoles hydrothermales: les auréoles de Valea Brazilor et de Valea Mlăci, montrant des roches fortement argilisées (à kaolinite et montmorillonite calcique ou sodique, auxquelles on ajoute des hydroxides de fer et l'illite et la ferrihalloysite) et l'auréole du versant droit de la vallée de Săpîncioara, caractérisée par une association de néominéraux renfermant: silice criptocristalline (opale et cristobalite) ± alunite et minéraux argileux. Les roches sont blanc-jaunâtre, grisâtres, roses, à éclat gras, satinées, dures, cassantes; quelques échantillons présentent un aspect faiblement bréchifié.

Au microscope on observe la structure porphyrique relique, où les phénocristaux de plagioclases et pyroxènes conservent parfaitement leur contour, bien qu'il soient entièrement substitués avec l'opale, qui, à son tour se transforme en quartz largement développé + alunite; au-dessus de l'alunite se forme la sidérite. Aussi, on a mis en évidence, mais rarement, la présence de la pyrite, la calcite et la tridimite. L'alunite est présente en proportion de jusqu'à 59 % dans la roche (l'apréciation est fondée sur l'analyse thermo-différentielle; pour huit échantillons analysés le contenu a varié entre 24 et 59 %). Les roches du complexe intrusif du forage C présentent des transformations propilitiques de type autométamorphique, avec la formation de la chlorite ± carbonates ± minéraux opaques secondaires. Par endroits, au-dessus du fond propilitique, fait son apparition un apport de quartz et carbonates ± minéraux argileux. Dans l'intervalle de 299 à 394 m, par exemple, l'andésite pyroxénique ± quartz présente du quartz de néoformation, en quantité et largement développé, recouvert par des carbonates ± minéraux opaques. Dans la diorite quartzifère, l'association caractéristique est quartz + chlorite + carbonates + épidote, parfois sous forme de nids, par endroits avec un apport massif de quartz, carbonates et rarement séricite (515 à 583 m et 619 à 620 m). L'andésite pyroxénique de l'intervalle de 643 à

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1066 m est transformée la paragenèse spécifique étant: chlorite + actinote + biotite + uralite + carbonates. Les transformations hydrothermales qui ont lieu sont: silicification + carbonatation tardive + pyritisation; dans des zones plus restreintes a lieu une transformation phyllique.

Le nombre réduit d'affleurements, travaux miniers et forages, rend difficile la compréhension totale des processus hydrothermaux. Certains rapports temporels se déduisent du fait que les apophyses d'andésite à clinopyroxènes situées à 55,6 m et dans l'intervalle de 148 à 155 m, dans le forage C, attribuées au complexe intrusif, sont presque fraiches, pendant que l'andésite percé est fortement argilisé et silicifié. On déduit donc l'existence d'une première phase hydrothermale, associée à la mise en place des laves (qui a engendré des argilisations et des silicifications), ainsi que d'une seconde, associée, en temps et espace, au corps intrusif (engendrant aussi des argilisations et silicifications + carbonatations sur un fond propilitique comportant parfois actinote et uralite). On accepte l'idée de l'existence d'une unique étape hydrothermale, ultérieure à la formation de la pile d'écoulements et à l'intrusion du corps éruptif.

Indices de minéralisation

Les indices de minéralisation sont peu nombreux et relativement pauvres en comparaison avec les aires du versant méridional des monts d'Igniş, mais ils peuvent être considérés suggestifs.

À la surface, excepté la pyrite, les autres minéraux ont été rencontrés dans une seule zone (galene – sur le versant droit de la vallée de Mlăci à 720 m en amont de la vallée de Săpincioara). Dans tous les trois forages exécutés jusqu'à présent, on a mis en évidence des occurrences de blende et plus rarement de galene et de chalcopyrite. Les indices de minéralisation se trouvent tant dans les laves, aussi bien que dans les corps intrusifs.

En vue d'enrichir les informations métallogéniques on a recouru aussi aux données lithogéochimiques: on a examiné presque tous les affleurements des roches hydrothermalisées et les trois forages ont été testés de 10 en 10 metres.

Les tableaux 6, 7 et 8 présentent, synthétisées, les principales données[1].

Tableau 6

Les teneurs moyennes en Pb, Zn et Cu des andésites pyroxéniques et des dacites de Săpînța–Valea Brazilor et d'Igniş

Type de roche	N	Pb ppm	Zn ppm	Cu ppm
andésite*	14	31	136	21
dacite*	3	12	64	15
andésites pyrox. Săpînța**	26	18	64	41
andésites pyrox. Mara**	10	10	58	27
andésites augitiques**	8	13	79	43
corps subvolcaniques**	16	15	68	31

N, no de l'analyses; * analyses effectuées par absorbtion atomique (Lab. de IPGG de Bucarest en 1983); ** d'après Lang (1976).

Tableau 7

Les valeurs de fond de quelques aires à vulcanites andésitiques des monts Igniş et Gurghiu

Zone	Pb	Zn	Cu	Pb/Cu	Zn/Cu
Bassin Valea Brazilor*	172	263	123	1,3	2,1
Bassin Valea Mlăci*		186	98	1,8	1,8
Forage 604 Vallée de Mlăci*	76	89			
Forage 605 Vallée de Săpîncioara*	61	109	33	3,3	3,3
Forage 607 Valea Brazilor*	173	155	66	2,3	2,3
Cratère de Seaca–Tătarca**	8	35	35	1,0	1,0

* analyses effectuées dans la laboratoire de l'IPGG de Bucarest;
** d'après Peltz et al. (1982) (Lab. I.M. Bălan)

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[1] Etant donné le caractère informatif des analyses, les interprétations doivent rester encore sous le signe de la réserve.

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Tableau 8

La valeur du rapport $V_f/\bar{x}$ pour les éléments: Pb, Zn et Cu

Zone	$V_f/\bar{x}$(Pb)	$V_f/\bar{x}$(Zn)	$V_f/\bar{x}$(Cu)
Bassin Valea Brazilor	11	4	3
Bassin Valea Mlăci		3	3
Forage 604 – Vallée de Mlăci	5	1	
Forage 605 – Vallée de Săpîncioara	4	2	1
Forage 607 – Valea Brazilor	11	2	2

V_f = la valeur du fond géochimique (vulcanites andésitiques – Săpînţa-Valea Brazilor, tab. 7); $\bar{x}$ = les teneurs moyennes en Pb, Zn, et Cu des andésites pyroxéniques d'Igniş (Lang, 1976, tab. 6)

Des données présentées résulte:

– dans les auréoles de Valea Brazilor et la vallée de Mlăci, les composants géochimiques sont quelquefois plus grands que dans les andésites pyroxéniques fraîches, les valeurs plus élevées étant enregistrées pour le plomb;

– pour les éléments analysés les valeurs du fond lithogéochimique sont plus élevées (jusqu'a approximativement 30 %) dans la zone de Valea Brazilor, par rapport à la zone de Mlăci;

– les valeurs du fond sont plus réduites dans les forages qu'à la surface, indiquant que les forages ont traversé aussi des intervalles à roches fraîches;

– les fonds lithogéochimiques pour les roches hydrothermalisées de la zone de Săpînţa-Valea Brazilor sont beaucoup plus élevés que ceux des roches du cratère de Seaca-Tătarca; on doit remarquer aussi le caractère plombo-zincifère pour la zone de Săpînţa-Valea Brazilor, par rapport à celui nettement cuprifère de la zone de Seaca-Tătarca.

Eléments d'ordre structural

Les nouvelles données concernant les variétés pétrographiques existantes, leur disposition aréale et la succession, ainsi que la confirmation, par le forage C-Valea Brazilor, du corps intrusif complexe, dont la présence avait été supposée sur base des données géochimiques, permettent une meilleure interprétation des structures éruptives. Les données gravimétriques, magnétiques et géomorphologiques, ont, même dans le stade actuel de la connaissance, un rôle très important.

On sépare clairement deux structures majeures: Valea Brazilor et Mlăci. Elles correspondent à deux maximums gravimétriques pregnants, à aspect isométrique, les mieux individualisés de tout le versant septentrional. On doit souligner l'existence d'un superposition remarquable du contour de l'aire à roches hydrothermalisées avec les maximums gravimétriques. Etant donné que le forage C-Valea Brazilor a confirmé l'existence, dans le cadre de la structure de Valea Brazilor, du corps intrusif complexe (andésites pyroxéniques ± quartz, andésites quartzifères passant à des diorites quartzifères porphyriques et andésites à clinopyroxènes) on devrait accepter que le spécifique de ces structures (Valea Brazilor et Mlăci) consiste même dans la présence des corps éruptifs. Si on a vu aussi les observations géomorphologiques, ainsi que certaines données concernant la disposition des roches effusives on peut invoquer un appareil volcanique central, au moins le long du cours supérieur de la vallée de Mlăci. Il est très difficile de former une image claire, de détail, de ces structures quand on n'a pas déterminé encore la nature exacte de celles-ci. Si les limites à l'horizontale de ces structures correspondraient aux gradients des deux principales anomalies gravimétriques de maximum, la structure de Mlăci et de Valea Brazilor aurait un diamètre majeur de 2500 m environ et celui mineur de 1500 à 2000 m. La structure de Mlăci est orientée approximativement E-W et celle de Valea Brazilor, N-S. La colonne du forage C indique qu'à présent l'apex de ces structures se situe très proche de la surface (170 m), en profondeur elles sont développées sur une distance de plus de 1000 m. Elles sont des structures intrusives à caractère polystadial, comportant une gamme assez variée de magmatites, quelques-unes nettement différentes, par leurs caractèristiques minéralogiques et chimiques, des vulcanites qui forment la masse de la région.

Les observations microscopiques, géomorphologiques et géophysiques suggèrent l'existence, à la surface, de quelques petites intrusions (fig. 1).

Le sommet de Brazi représente probablement un petit centre d'émission, autour duquel se dispose un écoulement restreint de lave d'andésite pyroxénique à amphiboles. Des dacites y forment de petites extrusions.

Il est très difficile d'établir les centres d'émission pour chaque variété pétrographique mais on peut considérer que celui-ci se situe à l'intérieur de la surface où a été cartographié le respectif pétrotype.

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Dans le stade actuel il est très difficile aussi de reconstituer les grands appareils volcaniques supposés comme générateurs de la pile de volcanites du versant septentrional des monts d'Igniş. Rădulescu et Lang (1973) ont présenté les critères geomorphologiques, hydrographiques et pétrographiques qui ont suggéré l'existence des caldéiras de Mara et Săpînţa et Edelstein et al. (1971, données non-publiées) ont supposé l'existence d'un cratère à l'origine de la vallée de Mlăci. Les données actuelles ne nient pas cette idee, mais les observations géomorphologiques seront plus utiles pour déterminer les édifices volcaniques anciens, en présence d'une meilleure image géologique. En l'absence d'une telle image, à la suite des interprétations vulcanologiques des données géomorphologiques (Rădulescu, Lang, 1973; Coteţ, 1972, données non-publiées) on ajoute une autre: nos investigations éffectuées dans la zone des sommets de Pleşca et Rotunzi ont relevé que l'élément géomorphologique principal, avec une grande continuité, le représente la crête qui délimite la zone volcanique à l'est, nord-est et le nord et qui passe par les sommets de Ticeri, Frăsinei, Piatra Neagră, Ţiganul, Fetei, Piatra Borcutului et Piatra Săpînţei. À l'intérieur de cette aire on individualise morphologiquement les sommets de Pleşca Mare, Măgura Mică et Brazi, mais représentant d'autres entités. Les données géologiques indiquent que le sens des écoulements de laves est de l'exterieur de la zone volcanique vers l'intérieur de celle-ci. Ces faits nous conduisent à la conclusion qu'il s'agit d'un grand ensemble volcanique avec une zone centrale dépressionnaire, ou de l'apparition d'un colaps dans les phases finales de l'activité éruptive.

Quant aux rapports entre les roches éruptives et les minéralisations on peut considérer plusieurs variantes:

a) les minéralisations sont plus jeunes que toutes les écoulements de lave, ou

b) les roches plus jeunes que l'andésite pyroxénique hydrothermalisé et que l'andésite de Mlăci sont postérieures à la minéralisation. Dans ce cas, une large aire du versant septentrional des monts d'Igniş présenterait les caractères d'une structure masquée. Certaines anomalies géochimiques mises en évidence dans des aires où il n'y a pas des roches hydrothermalisées à la surface, sont en faveur de cette interprétation.

On peut conclure donc que:

– les andésites en faciès d'écoulement sont plus variées que l'on connaissait antérieurement, du point de vue minéralogique, chimique et structo-textural;

– il y a deux corps intrusifs complexes plus jeunes que les écoulements de lave;

– il y a des indices directs de minéralisation tant à la surface, aussi bien qu'en profondeur, jusqu'a moins 1200 m;

– les anomalies litho-géochimiques ont un caractère plombo-zincifère.

Bibliographie

Borcoş M., Peltz S., Stan N., Udrescu C., Vasiliu C. (1979) Consideraţii petrochimice şi geochimice asupra vulcanitelor neogene din munţii Gutîi (IV. Andezite piroxenice ± hornblendă, ponţiene). *St. tehn. econ.*, I, 16, p. 81–107, Bucureşti.

Edelstein O., Istvan D., Weisz G., Cojocea C., Bernad A., Stan D., Kovacs M. (1978–1980) Harta geologică a Munţilor Oaş–Ţibleş, scara 1:25.000, arhiva I.P.E.G. "Maramureş", Baia Mare.

Gheorghiţă I. (1964) Contribuţii la cunoaşterea eruptivului neogen din regiunea Remeţi–Săpînţa–Sarasău. *D. S. Inst. Geol.*, I, Bucureşti.

– , Dofescu M. (1962) Cercetări geologice în regiunea Firiza–Izvoare–Crăceşti. *D. S. Com. Geol.*, XLVI (1958–1959), Bucureşti.

Girod M. (1978) Les roches volcaniques. Pétrologie et cadre structural. Ed. Doin, 239 p., Paris.

Kovacs M., Edelstein O., Istvan D. (1986) Andezitele din Munţii Oaş–Ţibleş; consideraţii privind definirea şi clasificarea lor pe baza datelor petrochimice. *St. cerc. geol., geofiz., geogr., Geol.*, 32, p. 12–24, Bucureşti.

Lang B. (1972) Date noi privind chimismul andezitelor cu piroxeni din nordul munţilor Gutîi. *St. tehn. econ.*, I, 6, Bucureşti.

– (1973) Formaţiunea vulcano–sedimentară ponţiană din partea nordică a munţilor Gutîi. *St. cerc. geol., geofiz., geogr., Geol.*, 18, 2, Bucureşti.

– (1976) Mineralogy and Geochemistry of the Neogene Pyroxene Andesites from the Northern Part of the Gutîi Mountains (Romania). *An. Inst. Geol. Geofiz.*, XLIX, Bucureşti.

Peccerillo A., Taylor (1976) Geochemistry of Eocene Calcalkaline Volcanic Rocks from Kastamom Ares, Northern Turkey. *Contr. Mineral, Petrol.*, 68.

Peltz S., Peltz M., Botar N. (1982) Observaţii litochimice şi implicaţii metalogenetice în aria vulcanică Găineasa (craterul Seaca–Tătarca, munţii Gurghiu). *D. S. Inst. Geol. Geofiz.*, LXVII/2, p. 85–112, Bucureşti.

– , Stanciu C., Balla Z., Gheorghiu A., Niţulescu I., Pomârleanu V., Udrescu C., Anastase S. (1982) Date noi privind mineralizaţia hidrotermală de la Stînceni (Munţii Călimani de Sud). *D. S. Inst. Geol. Geofiz.*, LXVII/2, p. 113–158, Bucureşti.

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Rădulescu D., Lang B. (1973) Sugestii privind interpretarea structurii geologice a părţii nordice a munţilor Gutîi. *D. S. Inst. Geol. Geofiz.*, LIX/5, Bucureşti.

Seghedi I. (1982) Contribuţii la studiul petrologic al calderei Călimani. *D. S. Inst. Geol. Geofiz.*, LXVII/1 (1979-1980), p. 87-126, Bucureşti.

Stanciu C. (1982) Structura eruptivă de la Mădăraşul Mare din partea centrală a Munţilor Harghita. *D. S. Inst. Geol. Geofiz.*, LXVII/1, p. 127-146, Bucureşti.

– (1976) Transformări hidrotermale în craterul Östörös din munţii Harghita. *D. S. Inst. Geol. Geofiz.*, LXII/1, p. 199-213, Bucureşti.

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DISTRIBUTION OF REE, K, Rb, Sr AND OF THE $^{87}Sr/^{86}Sr$ RATIOS IN THE NEOGENE ANDESITES OF THE IGNIŞ-VĂRATEC MOUNTAINS (GUTÎI)

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Key words: Andesites. Neogene. Minor elements. Rare earths. $^{87}Sr/^{86}Sr$. East Carpathians – Neogene–Quaternary eruptive – Gutin Mts.

Résumé: *Distribution des terres rares, du K, Rb, Sr et des rapports $^{87}Sr/^{86}Sr$ dans les andésites néogènes des monts de Igniş – Văratec (Gutîi).* On présente la distribution des terres rares (23 analyses), de Rb et Sr (41 analyses), K (plus de 200 analyses) et des rapports $^{87}Sr/^{86}Sr$ (14 déterminations) dans les principaux types d'andésites des monts de Igniş-Văratec (d'âge sarmatien-pliocène), par rapport aux andésites des zones géotectoniques semblables (à volcanisme calco-alcalin engendré par des processus de subduction). La variation des terres rares, normalisée aux chondrites, atteste l'homogénéité des magmas dans l'intervalle sarmatien-pannonien et une transformation relativement importante de leur caractère pendant le Pontien, tout comme la participation considérable des processus de cristallisation fractionnée à la formation des andésites en discussion. Selon la teneur en Sr et la variation de celui-ci par rapport au Rb, les andésites des monts Igniş-Văratec présentent des affinités sûres avec les vulcanites calco-alcalines du type arc insulaire. Pour la teneur en K, proche de celles présentées par Taylor et Gill pour les andésites calco-alcalines, la teneur en Rb est élevée (la plupart des valeurs supérieures à 50 ppm), ainsi que le rapport Rb/Sr (0,24 la moyenne). Les valeurs des rapports $^{87}Sr/^{86}Sr$ sont, elles aussi, très grandes (0,7063-0,7101). Toutes ces valeurs font la preuve de l'existence d'une contamination de la croûte importante des magmas parentaux, déterminée aussi par l'existence d'une croûte sialique traversée relativement épaisse (la valeur calculée selon la relation de Condie – respectivement 25 à 40 km, est en accord avec les données sismométriques). Cette contamination s'est effectuée tant pendant l'ascension des magmas parentaux à travers la croûte sialique, aussi bien que pendant leur évolution dans les chambres magmatiques secondaires localisées à des différents niveaux dans la croûte.

Introduction

The available volcanological and petrochemical data concerning the Igniş-Văratec (Gutîi) Mts area, as well as the whole area of the Oaş-Ţibleş Mts, make possible a relatively different interpretation of the evolution of the magmatism in the study region in relation to the geotectonic evolution of the northwestern part of the Carpathian Orogen.

The Igniş-Văratec (Gutîi) Mts are best studied. Andesites represent more than 85 per cent of the surface covered nowadays by volcanics. They were the object of recent petrochemical considerations concerning their definition and classification according to the recommendations of the IUGS Subcommission on the volcanic rocks systematics and to the criteria used by Gill (1981) for the denomination of the andesites generated by subduction processes (Kovacs et al., 1987). Starting from this and from the fact that there are no significant analyses on the other rock types (rhyodacites and dacites) – the present paper deals with the distribution of REE, K, Rb, Sr and $^{86}Sr/^{87}Sr$ isotopic ratios only in andesites.

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There are 23 analyses on REE – La, Ce, Sm, Eu, Tb, Lu – determined by thermic neutron activation, nondestructive variant, and Yb by emission spectrometry; 41 analyses on Rb, Sr and K – the first two determined by XRF spectrometry and K values computed from chemical analyses by K_2O transformation; 14 determinations of the $^{87}Sr/^{86}Sr$ isotopic ratios effectuated in laboratories of IFIN Măgurele.

Peltz (in Udubaşa et al., 1986, unpubl. report) gave a first processing of the data as well as a petrogenetic interpretation – including also Badenian rhyodacites and Pannonian-Pliocene dacites. The author reached several interesting conclusions on the diverse differentiations of the volcanics according to the variation of Rb, Sr and REE, the supply of sialic material in magma generation, as well as the advanced unhomogeneity of the parental magma.

General Data on the Volcanic Evolution in the Igniş-Văratec Mountains

In the Igniş-Văratec Mts the volcanic activity started in the Badenian with explosions that generated a rhyodacitic volcanoclastic complex.

During the Sarmatian-Pliocene time occurred a succession of andesites and dacites within which andesites, particularly the pyroxene ones, are predominating. Intrusive manifestations occur as well.

From the rocks more widespread and more significant in the developement of the petrogenetic processes are the following petrotypes (Fig. 1): Sarmatian pyroxene andesites (Seini), Pannonian pyroxene andesites (Ilba), Pannonian quartz andesites (Piscuiatu-Şuior), Pontian pyroxene andesites and amphibole-pyroxene andesites (Jereapăn), Pontian-Pliocene (?) hornblende-pyroxene andesites (Breze) and Pliocene pyroxene andesites – Igniş andesites and Mogoşa pyroxene andesites.

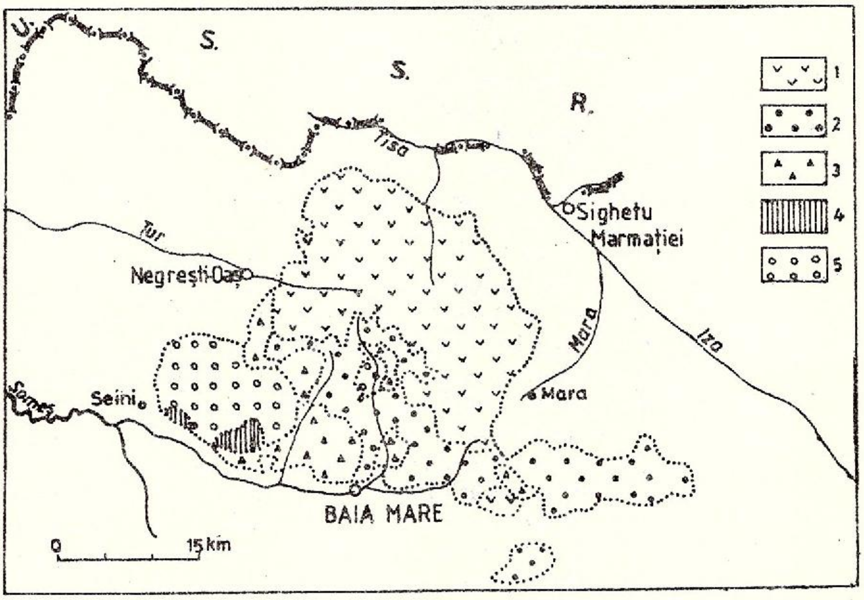

Fig. 1 – Disposition of the main types of andesites in the Igniş-Văratec Mts.
1, Pliocene pyroxene andesites; 2, Pontian pyroxene andesites and hornblende pyroxene andesites; 3, Pannonian quartz andesites; 4, Pannonian pyroxene andesites (Ilba); 5, Sarmatian pyroxene andesites (Seini).

The calc-alkaline character is pointed out by the distribution of the average of the main types of andesites on the AFM and $MgO/FeO+Fe_2O_3$ diagrams (Fig. 2).

REE Distribution

23 REE analyses are presented, according to ages and composition, in Table 1.

The SiO_2 content of the study rocks ranges between 50.92 and 62.52, with one exception, when the rocks represent andesites according to TAS classification.

La and Ce increase generally concomitantly with the content of SiO_2. Most of the La values are higher than 15 ppm; only 7 samples display lower values, corresponding to SiO_2 values ranging between 50.92 and 57.20. Sm, Eu and HREE (Tb, Yb, Lu) do not vary significantly with SiO_2.

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TABLE 1

REE distribution in andesites from the Igniș-Văratec (Gutâi) Mountains

No.	Sample	Mineralogical type	TAS classification	Age	SiO_2	La	Ce	Sm	Eu	Tb	Yb	Lu	ΣLa, Ce, Sm	La/Sm
1	16.500	αpx	αβ	Sm	55,10	18	14	5,2	1,10	0,94	2,9	0,33	37,2	3,46
2	16.501	αpx	αβ	Sm	53,20	13	18	4,6	1,16	0,93	3,1	0,26	35,9	2,65
3	5.260	αpx	αβ	Sm	53,66	20	12	5,0	1,01	0,99	3,1	0,23	37,0	4,00
4	6.248	αpx	α	Sm - Pn ?	57,46	15	19	5,3	1,13	0,76	3,3	—	39,3	2,83
5	6.702	αpx	αβ	Sm - Pn ?	55,61	17	8	3,7	0,99	0,95	3,3	—	28,7	4,59
6	27.235	αpx	αβ	Pn	56.97	17	16	4,0	0,98	0.77	2,9	0,37	37,0	4,25
7	8.772	αqpxho	α	Pn	60.72	15	12	3,1	0,78	0,68	1,9	0,24	30,1	4,83
8	5.680	αpx	α	Po	58,02	15	22	4,2	1,02	0,61	2,4	0,29	41,2	3,57
9	27.823	αpx	α	Po	59,20	25	17	4,2	0,93	0,50	2,7	0,28	46,2	6,16
10	2.291 *	αpx	αβ	Po	54,85	12	14	4,6	0,95	0,60	2,8	0,29	30,6	1,17
11	13.351 *	αpx	αβ	Po	55,81	13	12	4,2	0,79	—	2,2	0,25	29,2	3,57
12	27.820 *	αpx + ho	α	Po	57,20	5	7	4,2	0,84	—	2,4	0,26	16,2	1,19
13	14.560	αpx ± ho	α	Po - Pl ?	62,52	16	17	3,6	0,85	0,51	2,3	0,37	36,6	4,44
14	27.021	αpx	αβ	Po - Pl ?	56,47	12	19	4,4	0,90	0,49	2,5	0,38	35,4	2,72
15	9.274	αpxho	α	Po - Pl ?	60,06	16	23	3,2	1,00	0,61	2,3	0,22	42,2	5,00
16	6.713	αpx	α	Pl	60,32	20	29	4,9	1,03	0,72	2,3	0,28	53,9	4,00
17	14.595	αpx (±ho)	αβ	Pl	54.58	12	7	4,3	0,97	0,50	2,2	0,28	23,3	2,79
18	9.185 *	αpx	β	Pl	50,92	12	8	4,0	0,90	0,64	2,0	0,31	24,0	3,00
19	1.747	αpx	α	Pl	57,60	18	19	4,3	1,02	0,76	3,4	0,26	41,3	4,18
20	26.579	αpx + ho	α	Pl	58,30	26	27	4,5	1,05	0,98	1,8	0,20	57,5	5,77
21	1.600	α(q)pxhobi	α	Pl	58,95	17	29	5,4	1.07	0,67	2,0	0.32	51,4	3,14
22	607 *	αqpxhobi	α	Pl	60,30	23	30	4,6	0,89	0,72	1.9	0,23	57,6	5,00
23	8.372 *	αqpxhobi	α	Pl	61,17	35	38	4,2	1,02	1,00	1,7	0,31	78,2	8,33

1. * - intrusive bodies; 2. α - andesite, αβ - basaltic andesites, β - basalts; 3. minerals of the porphyritic phase: px - pyroxenes, ho - hornblende, bi - biotite, q - quartz

The study of the REE distribution in the andesites from the Igniş-Văratec Mts normalized versus chondrites (Haskin et al., 1968) was made firstly by classifying them according to the age.

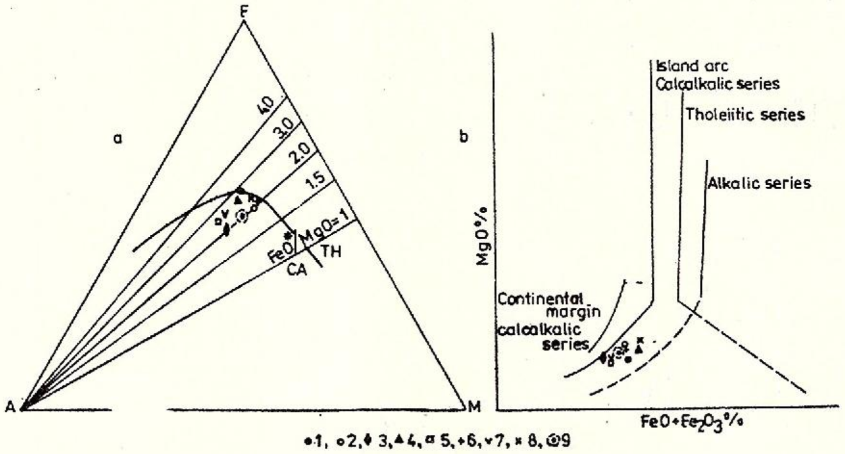

Fig. 2 – Distribution of the main types of andesites (average values) in the Igniş-Văratec (Gutii) Mts on the AFM (a) and MgO-FeO+Fe_2O_3 (b) diagrams.
1, Sarmatian pyroxene andesites (Seini); 2, Pannonian pyroxene andesites (Ilba); 3, Pannonian quartz andesites (Piscuiatu); 4, Pontian pyroxene andesites (+hornblende) (Jereapăn); 5, Pontian-Pliocene ? hornblende pyroxene andesites (Breze); 6, Pliocene pyroxene andesites (Igniş); 8, Pliocene pyroxene andesites (Mogoşa); 9, mean value for the Igniş-Văratec (Gutîi) andesites.

Figure 3 presents the REE distribution for the Seini Sarmatian pyroxene andesites (16500, 16501, 5260), for the Sarmatian-Pannonian (?) pyroxene andesites – 6248, Pleşa Summit and 6702, Comşa Summit (Ilba zone) and for the Ilba Pannonian pyroxene andesite – 27235, Valea Porcului. Several particularities are observed, as follows :

– parallelism of the variation diagrams in all andesites, pointing out the common origin and the same petrogenetic mechanism;

– strong fractionation, especially in the LREE field, pointing to an advanced fractional crystallization;

– Eu negative anomaly, indicating an early fractionation of plagioclase;

– Ilba pyroxene andesite occurs less fractionated both in the LREE field and in the HREE field.

Figure 3b presents Pannonian andesites, Ilba pyroxene andesites (27235) and Poiana Cremenii quartz andesites (8772, Baia Sprie E – Şuior zone), respectively. It is worth mentioning the similarity, up to identity, of the fractionation way of the two andesites, suggesting an identical petrogenetic mechanism.

Figure 3c presents andesites from lava flows of Pontian (5680, Higea Summit, 27823, Roata-Cavnic) or uncertain Pontian-Pliocene(?) age (27021, Piatra Prislop Summit – Băiuţ, 14560, Hust-Negreia zone) as well as andesites from Pontian intrusive bodies (2291, Baia Sprie, 13351, Cavnic, 27820, Bloaja). It is to be noted the parallelism of the variation diagrams for lavas (even for andesites with an uncertain position), on the one hand, and lavas and intrusive bodies, on the other hand, pointing to a common origin and a similar petrogenetic mechanism. At the same time, one can notice a stronger fractionation in the LREE field for bodies attesting a process of fractional crystallization more advanced as compared with lavas.

Figure 3d represents the hornblende pyroxene andesite from Blidari Summit (Chiuzbaia), considered as an equivalent of the Breze andesites, of Pontian-Pliocene(?) age. As compared with the Pontian andesites (Jereapăn phase), the variation diagram displays a different feature, suggesting a quite different petrogenetic mechanism.

As regards Pliocene andesites, the REE distribution normalized to chondrites is shown in Figures 3e and 3f. Figure 3e presents the Mogoşa pyroxene andesites (14595, Mogoşa Summit and 9185, intrusive body in Şuior zone) and the Igniş pyroxene andesites (6713). Sensible differences are observed between the fractionation way of the two types – especially in the field of the LREE. Igniş andesite shows higher values both for LREE and for HREE.

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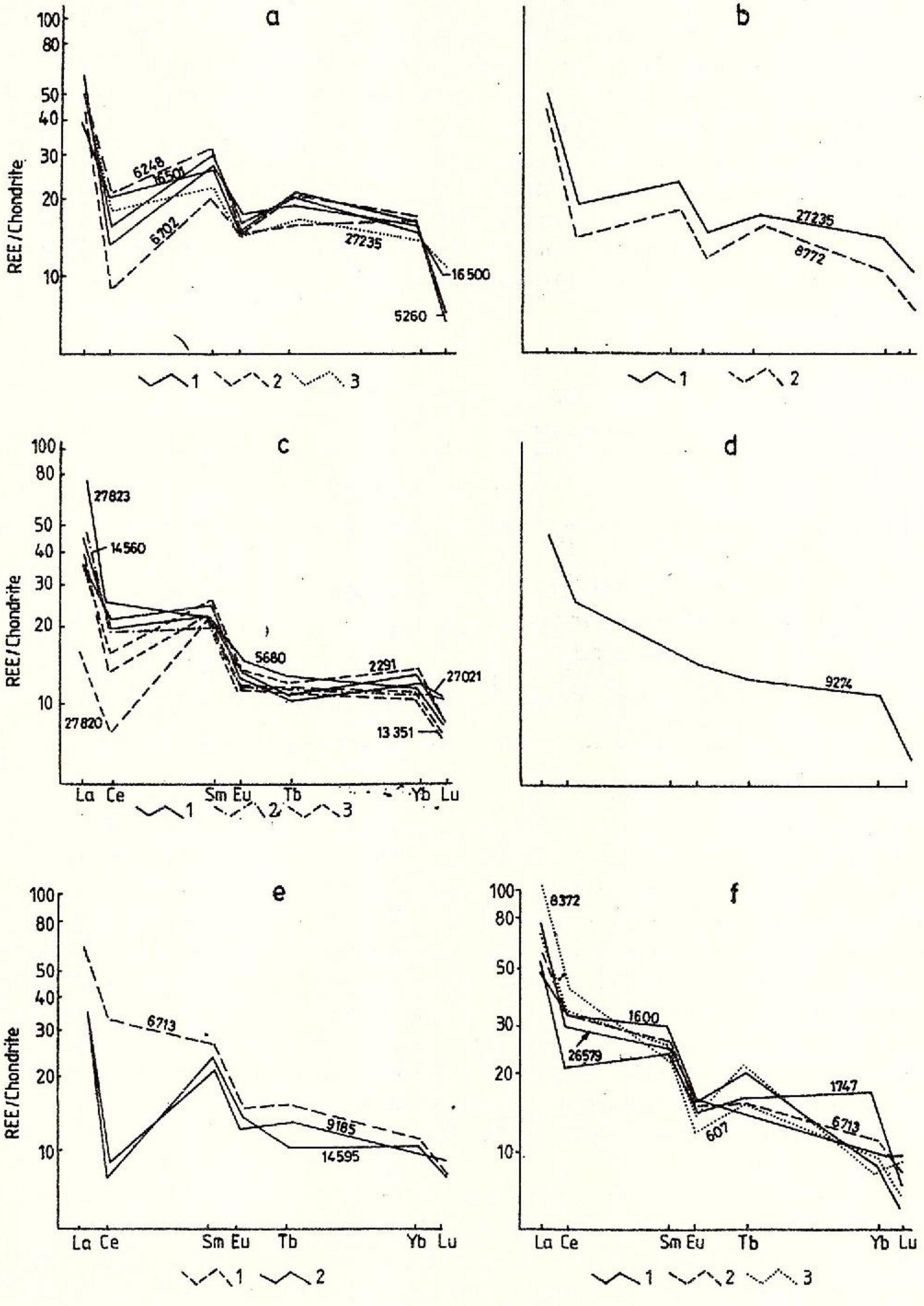

Fig. 3 – REE distribution normalized to chondrites.
a) Sarmatian (1), Sarmatian-Pannonian ? (2) and Pannonian (3) pyroxene andesites.
b) Pannonian andesites: 1, Ilba pyroxene andesite; 2, Piscuiatu-Şuior quartz andesite.
c) Pontian andesites: 1, pyroxene andesites (lavas); 2, pyroxene±hornblende andesite Hust-Mogoşa zone; 3, pyroxene andesites (2291, 13351) and hornblende pyroxene andesites (27820) in intrusive bodies. d) Pontian-Pliocene (?) hornblende pyroxene andesite (Breze). e) Pliocene pyroxene andesites of Igniş (1) and Mogoşa (2). f) Pliocene andesites: 1, lavas; 2. Igniş andesites; 3, intrusive bodies.

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Figure 3f displays andesites from the northern side of the Igniş Mts, representing lavas (1747, pyroxene andesite, Rotunzi Summit-Mlăci, 26579, hornblende-pyroxene andesites, Brazi Summit and 1600, hornblende-biotite-quartz pyroxene andesites, Braga Summit) and intrusive bodies (607, quartz andesite (diorite), Brazi Valley and 8372, quartz andesite with pyroxene, hornblende and biotite in Laleaua Albă dykes). The quite different aspect of the diagrams points to relatively different sources and petrogenetic processes. The rocks from bodies display a similar fractionation way; however, the andesites from the Brazi Valley-Mlăci and that from Igniş display a different fractionation, particularly in the field of HREE.

Table 2 presents the average values of LREE and Lu for the andesites from the Igniş-Văratec; as compared with those from the Călimani-Harghita andesites (Peltz et al., 1985), as well as with Taylor's averages (1969) for calc-alkaline andesites. It is to be noted the similar value of La in Igniş-Văratec and Călimani-Harghita andesites, and higher than Taylor's average, and the different values, in case of Igniş-Văratec andesites, of Sm (higher value) and Ce (lower value). All this is determined by a stronger fractionation of the Igniş-Văratec andesites in the LREE field.

TABLE 2

Averages of LREE

		La	Ce	Sm	Eu	Lu
1	Igniş - Văratec (Gutâi) andesites (23 samples)	16.2	17.2	4.3	0,97	0,28
2	Călimani - Harghita andesites (Peltz, 1985 ~ 20 samples)	16.6	23.3	2.9	0,96	0,27
3	Calcalkaline andesites (Taylor, 1969)	11.9	24	2,9	1.0	-

Several conclusions can be inferred from the REE distribution in Igniş-Văratec andesites:

- in spite of the similar behaviour within the Sarmatian-Pontian andesites, on the one hand, and the Pontian ones, on the other hand, which consequently implies similar petrogenetic mechanisms, there is a clear differentiation between the two main phases on the REE diagrams. Generally, the Sarmatian-Pannonian andesites display a higher degree of fractionation, with marked Eu negative anomalies (due also to the Tb content about twice higher than the one of the Pontian andesites). As for lavas, the Sarmatian-Pontian ones are more fractionated in the LREE field as compared with the Pontian lavas, suggesting that the process of fractional crystallization played a greater role in their formation.
- Mogoşa basaltic andesites occur as the most fractionated volcanics from the Igniş-Văratec Mts in the LREE field, only the Sarmatian pyroxene andesites displaying affinities with them;
- Pliocene andesites occur less fractionated in the LREE field, as compared with the other andesites, partly because of the higher content of Ce.

The relatively small number of analyses and the unrepresentativity of some of the study andesites (especially the Pliocene ones) make difficult the whole reconstitution of the evolution of parental magmas versus temporal evolution of the volcanism. Magma homogeneity in the Sarmatian-Pannonian time-interval and a relatively significant change of their character in the Pontian phase can be preliminarily stated. At the level of the Pliocene volcanism, the different fractionation of the HREE field suggests parental magmas different from the Sarmatian-Pontian ones, on the one hand, and different from one another, on the other hand. All this could be somehow connected with the evolution of the magmas in secondary magma chambers with different localization in the sialic crust.

K, Rb and Sr Distribution

As regards K distribution, more than 200 chemical analyses have been published by Kovacs et al. (1987). More than 90 per cent of the Igniş-Văratec andesites are medium-K andesites and low-K andesites are lacking (Fig. 4). It is to be noted the grouping of quartz andesites and of pyroxene hornblende andesites at the top of the andesite field, where among pyroxene andesites only Igniş andesites are to be found. In the basalts and

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low-silica andesites field only pyroxene andesites occur. Only one sample from the Igniş andesite and Laleaua Albă quartz andesite can be referred to the high-K andesites field.

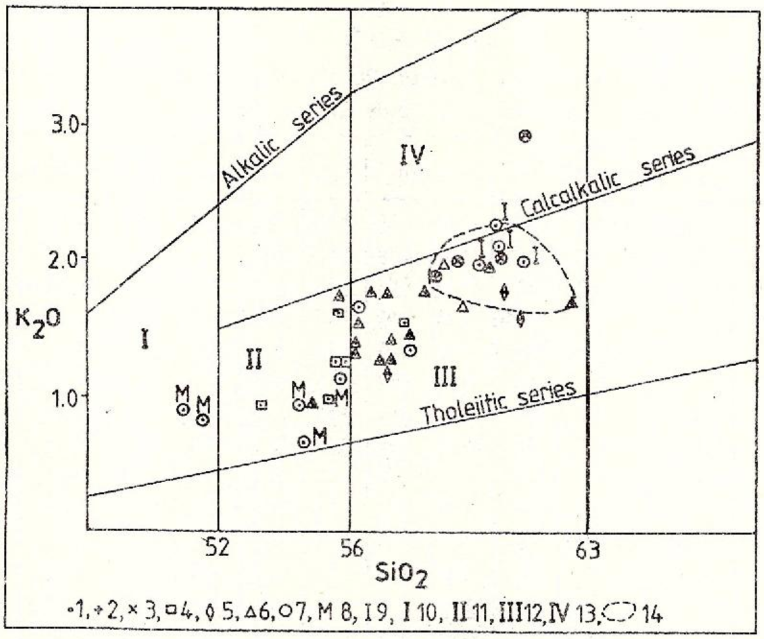

Fig. 4 – K_2O SiO_2 diagram (Peccerillo, Taylor, 1976).
1, pyroxene andesites; 2, hornblende pyroxene andesites; 3, quartz andesites; 4, Sarmatian and Sarmatian–Pannonian (?) andesites; 5, Pannonian andesites; 6, Pontian and Pontian–Pliocene (?) andesites; 7, Pliocene andesites; 8, Mogoşa andesites; 9, Igniş andesites. Fields of the diagram (I, II, III, IV); 10, basalts; 11, low-Si andesites; 12, andesites; 13, high-K andesites; 14, field of the Igniş pyroxene andesites (Pl), quartz andesites (Pn) and hornblende pyroxene andesites (Po–Pl ?).

The distribution of Rb (ppm), Sr (ppm), K (%) as well as of the Rb/Sr and K/Rb ratios for the study andesites is presented in Table 3.

Rb generally displays high values, most andesites including more than 50 ppm. No significant differences can be observed in Rb distribution between the main phases. In general, Rb variation can be correlated with the variation of the K_2O and SiO_2 contents, as shown on the Rb-K/Rb diagram (Fig. 5). It is to be noted the clear individualization of the variation domains of the Mogoşa and Igniş andesites.

As regards Sr distribution, it is worth mentioning the homogeneity and the low values of the Sr content. It ranges between 213 and 340 ppm, except for three samples which exceed 300 ppm.

There are no significant differences concerning the Sr content in rocks varying in age, mineralogic-petrographic and chemical respect. For instance, Table 4 shows Sarmatian-Pliocene andesites with a SiO_2 content varying between 51.4 and 61.1 per cent and a K content varying between 0.98 and 2.23 per cent, with relatively different mineralogical compositions and a highly homogeneous but very low Sr content.

No Sr increase with K_2O is observed for andesites with different features (low-K, medium-K and high-K, Gill, 1981). Thus, the Mogoşa pyroxene andesites have a K_2O average of 0.98 and a medium-towards low-K character and the Igniş pyroxene andesites an average of 2.06 and a transitional character towards high-K andesites. The Sr contents of the former (6 samples) are, on the whole, even higher than those of the Igniş andesites (3 samples).

The Rb-Sr diagram (Fig. 6) presents the variation domains for different geotectonic zones with orogenic andesites (Gill, 1981). Taking into account the Sr content, the andesites from the Igniş-Văratec Mts present affinities with the island arc andesites – the variation domain being mostly superposed over that of the andesites from New Zealand arc. The localization of the Igniş-Văratec andesites has to be regarded as a proof of the existence of a low content of primary Sr, probably connected with the implication of a thinned oceanic crust into the melting process of subducted plate.

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TABLE 3

Distribution of SiO_2, Na_2O, K_2O, and of Rb, Sr, K in andesites from the Igniș-Văratec Mountains. Symbols as in Table 1

No.	Sample	Mineralogical type	Peccerillo, Taylor (1976)	Age	SiO_2	Na_2O	K_2O	Rb	Sr	Rb/Sr	K(%)	K/Rb
1	16 500	αcpx	αc lowSi	Sm	55.10	2.26	0.98	56.2	231	0.243	0.81	144
2	16 501	αcpx	αc lowSi	Sm	53.20	2.40	0.98	50	256	0.197	0.81	162
3	26 078	αcpx	αc lowSi	Sm	55.54	2.82	1.61	48	280	0.173	1.34	300
4	6 248	αcpx	αc	Sm-Pn?	57.46	2.42	1.52	60.4	253	0.238	1.26	208
5	6702	αcpx	αc lowSi	Sm-Pn?	55.60	2.40	1.24	74.6	254	0.293	1.03	136
6	27 235	αcpx	αc	Pn	56.99	2.42	1.16	78.9	225	0.327	0.96	121
7	521	αcpx	αc	Pn	61.11	3.14	1.52	64.2	260	0.246	1.26	121
8	8772	αcqpxho	αc	Pn	60.79	2.99	1.74	81.4	222	0.366	1.44	176
9	5680	αcpx	αc	Po	58.02	2.51	1.76	63.2	234	0.270	1.46	231
10	27823	αcpx	αc	Po	59.20	2.56	1.64	52.4	252	0.207	1.36	259
11	2291*	αcpx	αc lowSi	Po	54.85	2.63	0.97	51.4	266	0.193	0.80	155
12	8186	αcpx	αc	Po	56.98	3.17	1.23	61.3	230	0.265	1.02	166
13	13 351	αcpx	αc lowSi	Po	55.81	2.48	1.75	58.3	250	0.233	1.45	248
14	27 820	αcpxho	αc lowSi	Po	57.20	3.02	1.40	54.8	232	0.235	1.16	211
15	7 316	αcpx	αc lowSi	Po-Pl?	56.05	2.81	1.33	53.1	274	0.193	1.10	207
16	8 294	αcpx	αc	Po-Pl?	57.04	2.60	1.74	92.1	240	0.383	1.44	156
17	8 574	αcpx	αc	Po-Pl?	56.84	2.94	1.23	40.4	257	0.157	1.02	252
18	14 014	αcpx	αc	Po-Pl?	57.63	2.87	1.46	98.9	240	0.244	1.21	122
19	14 039	αcpx	αc	Po-Pl?	56.19	2.87	1.36	50.7	239	0.211	1.13	220
20	14 560	αcpx (±ho)	αc	Po-Pl?	62.52	3.26	1.65	77.7	217	0.356	1.37	175
21	17 004	αcpx	αc	Po-Pl?	56.07	2.00	1.52	97.2	278	0.205	1.26	129
22	27 021	αcpx	αc	Po-Pl?	56.47	2.10	1.76	65.7	253	0.258	1.46	223
23	9 274	αcpxho	αc	Po-Pl?	60.06	2.89	1.94	86.9	239	0.363	1.61	276
24	27 460	αcpxho	αc	Po-Pl?	58.63	2.72	1.94	71.4	255	0.279	1.61	225
25	4 325	αcpx(I)	αc	Pl	60.98	2.66	1.94	92.9	232	0.399	1.61	173
26	6 713	αcpx(I)	αc	Pl	60.32	2.72	2.13	84.5	234	0.360	1.77	208
27	6 714	αcpx(I)	αc	Pl	59.80	2.55	1.94	78.1	234	0.332	1.61	205
28	27 459	αcpx(I)	αc highK	Pl	60.46	2.58	2.23	59	261	0.186	1.85	357
29	14 595	αcpx(M)	αc lowSi	Pl	54.58	2.84	0.68	22.9	229	0.097	0.56	244
30	26 786	αcpx(M)	β	Pl	51.46	2.77	0.88	27.6	257	0.107	0.73	264
31	9 185*	αcpx(M)	β	Pl	50.92	2.55	0.92	34.6	253	0.136	0.76	191
32	14 532	αcpx(M)	αc lowSi	Pl	54.26	2.84	0.98	34.9	231	0.151	0.81	232
33	9 399	αcpx(M)	αc lowSi	Pl	55.50	2.89	1.23	43.4	273	0.158	1.02	235
34	7 788	αcpx(M)	αc lowSi	Pl	55.85	2.87	1.12	45.5	236	0.192	0.92	202
35	1 747	αcpx	αc	Pl	57.60	2.10	1.32	52	213	0.128	1.09	209
36	9 047	αcpx	αc	Pl	56.10	3.41	1.64	55.1	267	0.206	1.36	246
37	1 774	αcpx	αc	Pl				58	235	0.247	1.30	228
38	26579	αcpx+ho	αc	Pl	58.30	1.89	1.89	87	312	0.278	1.57	179
39	1 600	αc(q)pxhobi	αc	Pl	58.95	2.65	1.97	79.5	305	0.260	1.64	205
40	607*	αc(δ)qpxhobi	αc	Pl	60.30	2.20	1.49	93.9	251	0.370	1.65	130
41	8 372*	αcqpxhobi	αc highK	Pl	61.17	2.83	2.97	108	342	0.317	2.47	227

(I)= Igniș andesites; (M)= Mogoșa andesites

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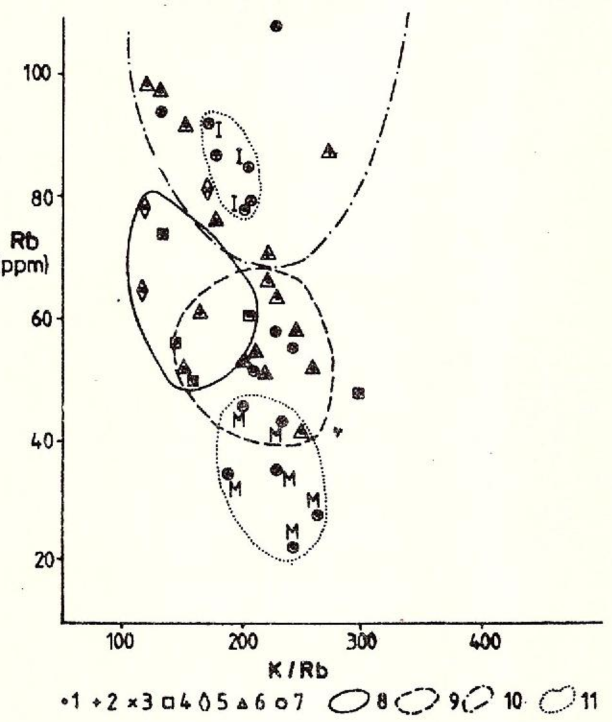

Fig. 5 – Rb–K/Rb diagram
Symbols 1-7 the same as in Figure 4; 8, variation domain of the Sarmatian–Pannonian pyroxene andesites; 9, variation domain of the Pontian pyroxene andesites; 10, the same as 14 in Figure 4; 11, variation domain of Mogoşa and Igniş pyroxene andesites.

TABLE 4

Distribution of ^{87}Sr/ ^{86}Sr and of Rb/Sr in andesites from the Igniş-Vãratec Mountains. Symbols as in Tables 1 and 3

	Sample	Mineralogical type	Age	SiO$_2$	K	Sr
1	16 501	αpx	Sm	53.2	0.81	256
2	6 702	αpx	Sm-Pn?	55.6	1.03	254
3	521	αpx	Pn	61.1	1.26	260
4	27 823	αpx	Po	59,2	1.36	252
5	13 351	αpx±ho	Po	55.8	1.45	250
6	27 460	αpxho	Po-Pl?	58.6	1.61	255
7	27 459	αpx	Pl	60.4	1.85	261
8	26 786	αpx	Pl ?	51.4	0.73	257
9	9047	αpx	Pl	56.1	1.36	267
10	607	αgpxhobi	Pl	60.3	1.65	251

1. samples taken from Hertz et al. (1974) and Peltz et al. (1985).

Starting from the high-Rb and the low-Sr values, the Rb/Sr ratio shows high values – between 0.10 and 0.39, an average of 0.24. The lowest values occur in andesites wih the lowest Rb contents where the Sr contents are relatively constant.

Considering the higher values of the Rb/Sr ratio in differentiated products, the Rb/Sr-SiO$_2$ variation diagram (Fig. 7) was drawn up for Pliocene andesites (15 samples), which present the largest variation domains for SiO$_2$, K$_2$O and Rb. The andesites under discussion differ also, mineralogically and petrographically, as well as regarding the age. There is a good correlation of the Rb/Sr ratio with SiO$_2$; except for the Igniş pyroxene

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andesites (which display a high K_2O and Rb content), the highest value occur in highly differentiated rocks, in mineralogical respect, too.

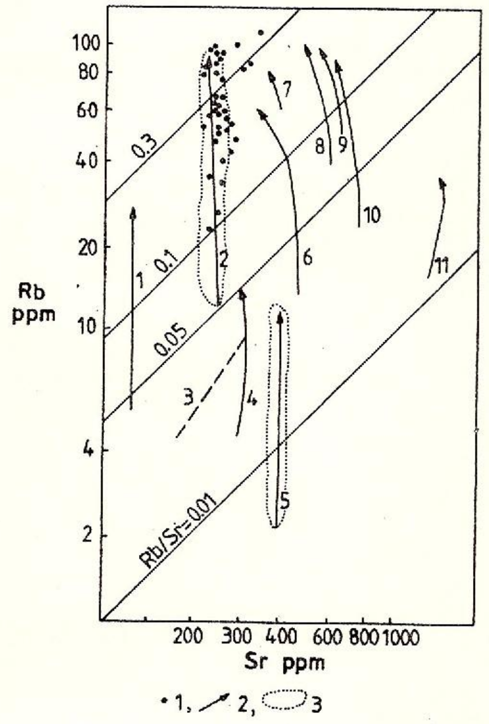

Fig. 6 – Rb–Sr diagram.
1, andesites in the Igniş–Văratec Mts; 2, volcanics from: 1, South Sandwich; 2, New Zealand; 3, Tonga; 4, Antilles; 5, New Britain; 6, Aleutians; 7, Sumatra; 8, Chile; 9, Peru; 10, Eolian; 11, Cascade; 3, variation field of the data with example for two of the cases (fide Gill, 1981).

It is worth mentioning that the Rb/Sr ratio increases concomitantly with the younger age of the andesites (in agreement with the increase of the differentiation degree).

Distribution of $^{87}Sr/^{86}Sr$ Isotopic Ratios

Table 5 presents the values of the $^{87}Sr/^{86}Sr$ ratio determined for the Igniş-Văratec andesites. Several conclusions can be inferred as follows:

– on the whole, the values of the isotopic ratios are high, especially in comparison with the value of the Racoş basalt, considered to have originated in a parental magma without crustal contamination (Peltz et al., 1985);

– the pyroxene andesite in the Pleşa-Ilba summit (sample 6248), whose age is differently interpreted by various authors, shows the same value of the isotopic ratio as the pyroxene andesite in the Zugău Valley-Seini (sample 26078), unanimously accepted as Sarmatian in age;

– the noticeable differences between the Mogoşa andesites and the other Pliocene andesites, especially Igniş andesites, which display the highest value of the isotopic ratio;

– the quartz andesite of the intrusive bodies from Laleaua Albă is a special case, due to the value of the isotopic ratio – the lowest value of the Rb/Sr ratio versus that of the Igniş andesites;

– the Pliocene andesites, except for the Mogoşa ones, show higher values as compared with the Sarmatian-Pontian ones which, from this point of view, are highly homogeneous (the values range between 0.7080 and 0.7085).

Petrogenetic Considerations

For a synthetical approach of the andesites from the Igniş-Văratec Mts (Gutîi) as regards the Rb, Sr and K distribution, the average values, calculated in comparison with values from the relevant literature of andesites from similar geotectonic areas, are presented in Table 6. It is to be noted that the andesites under discussion include all the main types occurring in the Igniş-Văratec Mts, displaying averages for SiO_2 – 57.2 per cent and

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K_2O – 1.53 per cent, similar up to identity, on the one hand, with the averages for all the andesites from the Oaş-Ţibleş volcanic chain – 56.93 and 1.54 %, respectively (Kovacs et al., 1987) and, on the other hand, with the averages for orogenic andesites – 57.6 and 1.5 %, respectively (Gill, 1981). The following conclusions can be drawn:

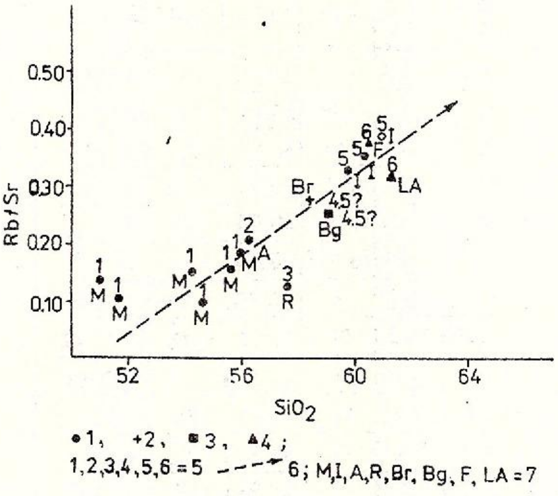

Fig. 7 – Rb/Sr–SiO_2 diagram for Pliocene andesites. 1, pyroxene andesites; 2, hornblende pyroxene andesites; 3, quartz+biotite+hornblende pyroxene andesites; 4, biotite+hornblende+pyroxene quartz andesites (intrusive bodies); 5, age relationship from old to recent; 6, variation direction of the Rb/Sr ratio with SiO_2; 7, andesites and their location: M – Mogoşa andesites; I – Igniş andesites; A – Arsura andesites (east of Gutin Pass); R – Rotunzi Summit andesites (Săpînţa); Br – Brazi Summit andesites (Valea Brazilor-Săpînţa); Bg – Braga Summit andesites (Valea Brazilor-Săpînţa); F – Valea Brazilor intrusive body; L.A. – Laleaua Albă intrusive body (Serpentinele Gutinului – Gutin winding road).

– high Rb value – twice higher than the averages given by Taylor (1969) and Gill (1981);

– the content of K is identical with that of the andesitic magma of the Călimani-Harghita Mts (Peltz et al., 1985) and it corresponds to the values given for calc-alkaline andesites (Taylor, 1969; Gill, 1981);

– the low Sr value is close to that of the New Zealand volcanics as it is shown on the Rb–Sr diagram;

– the K/Rb ratio is very small (205) in comparison with the other andesites under discussion.

Several petrogenetic particularities can be inferred from the K, Rb, Sr distribution and the isotopic ratios $^{87}Sr/^{86}Sr$ in the Igniş-Văratec andesites, as well as from the comparison with volcanics from similar geotectonic areas (convergent plate margins with a calc-alkaline magmatism generated by subduction processes).

The high Rb content suggests the enrichment of parental magma in this element from the crossed continental crust. The low K/Rb ratio points out the magma contamination with high-Rb lithospheric material during the crossing of the crust. The bigger the thickness of the crossed crust, the higher the crustal contamination is (Gill, 1981). The K/Rb-K_2O variation diagram (Fig. 8) presents the variation domain of the Igniş-Văratec andesites versus those of the orogenic andesites from other zones (fide Gill, 1981). The lower values of the ratio in comparison with continental margin zones, as well as with many island arc zones, and the localization of the K/Rb ratios versus the values of the crust thickness under the respective arcs, could be in favour of a relatively thick crossed crust.

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TABLE 5

No.	Sample	Mineralogical type	TAS	Age	SiO_2	K_2O	Sr^{87}/Sr^{86}	Rb/Sr
1	26078	αpx	αβ	Sm	55,54	1,61	0,7085	0,17
2	6248	αpx	α	SmPn?	57,46	1,52	0,7085	0,23
3	8772	αqpxho	α	Pn	60,72	1,74	0,708	0,37
4	5680	αpx	α	Po	58,02	1,76	0,7083	0,27
5	8574	αpx	αβ	Po-Pl?	56,08	1,23	0,708	0,16
6	27021	αpx	αβ	Po-Pl?	56,47	1,76	0,709	0,26
7	14595	αpx±ho(M)	αβ	Pl	54,58	0,68	0,7065	0,09
8	14532	αpx(M)	αβ	Pl	54,26	0,98	0,7072	0,15
9	H 74 ●	αpx	α	Pl			0,7084	0,24
10	26579	αpx±ho	α	Pl	58,30	1,89	0,709	0,28
11	1747	αpx	α	Pl	57,60	1,32	0,710	0,18
12	1600	αpxhobiq	α	Pl	58,95	1,97	0,7092	0,26
13	6713	αpx(l)	α	Pl	60,32	2,13	0,7101	0,36
14	8372	α(δ)qpxhobi	α	Pl	61,17	2,97	0,7063	0,31
15	R 74 ●	βpx	β	Q			0,7043	0,05

● 1. Samples taken from Hertz et al. (1974) and Peltz et al. (1985)

TABLE 6

No			Rb	Sr	Rb/Sr	K(%)	K/Rb
1.	Igniș - Văratec (Gutâi) andesites		62,5	249	0,24	1,2	205
2	Călimani-Harghita andesitic magma (Peltz, 1985)		43,3	297	0,15	1,2	292
3	Calcalkaline andesites (Taylor, 1969)		31	385	0,08	1,3	430
4	Orogenic andesites (Gill,1981)	Medium K basic andesites	25	501	0,05	0,9	365
5		Medium K acid andesites	20	490	0,04	1,1	544
6	New Zealand volcanics (Siegers,1969 fide Peltz,1985)		40	271	0,17	–	325
7	Chile volcanics (Pichler, Zeil,1972 fide Peltz, 1985)		42	743	0,14	–	397
8	The Perşani basalts (Peltz, 1985)		378	777	0,05	1,2	317

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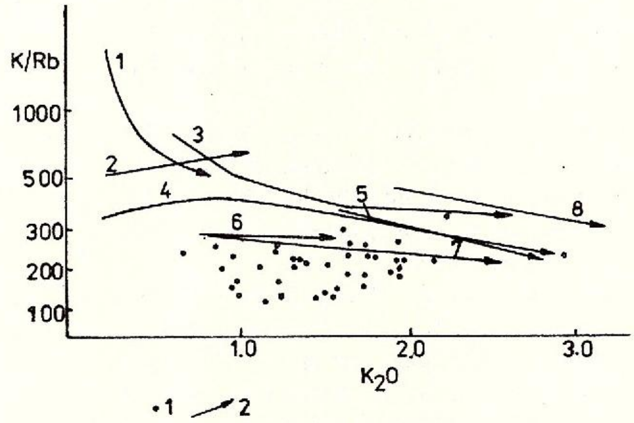

Fig. 8 – K/Rb–K_2O diagram.
1, Igniş-Vărătec andesites; 2, variation of K/Rb with K_2O for the volcanics from other zones (in parantheses the crust thickness, fide Gill, 1981). 1, New Britain (20-40 km); 2, Tonga (13 km); 3, Aleutians (18-25 km); 4, Jawa (25-30 km); 5, New Guinea (30-35 km); 6, Grenada (30-35 km); 7, New Zealand (36 km); 8, Peru (40-70 km).

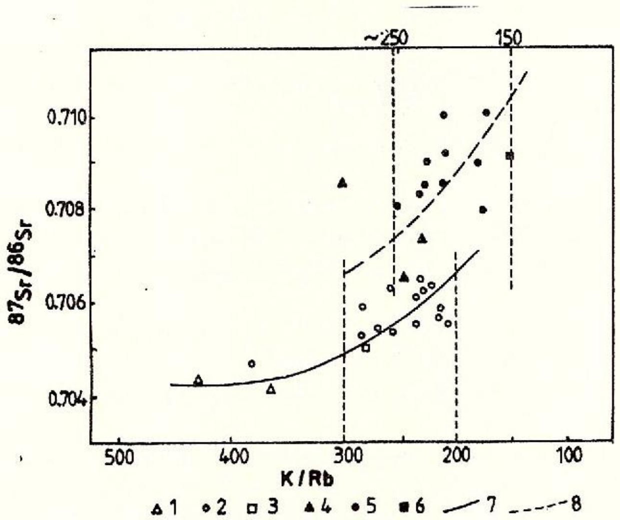

Fig. 9 – $^{87}Sr/^{86}Sr$–K/Rb diagram.
1, basalts; 2, andesites; dacites (New Zealand); 4, basaltic andesites; 5, andesites; 6, dacites (Igniş-Vărătec); 7, "mixing" curve interpreted as contamination direction with sialic material for New Zealand volcanics (Ewart, Stipp, 1968, fide Gill, 1981); 8, contamination curve for andesites of the Igniş-Vărătec (Gutii) Mts.

The crustal contamination can be also inferred from the assigning of the values of the $^{87}Sr/^{86}Sr$ to K/Rb ratio on such a diagram. Ewart and Stipp (1968, fide Gill, 1981) observed the variation of the $^{87}Sr/^{86}Sr$ and K/Rb ratios for the volcanics in New Zealand, starting from basalts, considered least contaminated. In

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comparison with them, the Igniş-Văratec volcanics (Fig. 9), although displaying the same variation direction – the decrease of the K/Rb ratio concomitantly with the increase of the isotopic ratios – possess one particularity: for smaller values of the K/Rb ratio (150–250) the values of the $^{87}Sr/^{86}Sr$ isotopic ratio are much higher. This can be explained, on the one hand, by the high Rb (^{87}Rb passes to ^{87}Sr and, consequently, the values of the $^{87}Sr/^{86}Sr$ increase, being higher in the lithosphere than in the mantle (Faure, Powell, 1972)) and, on the other hand, by the lower content of primary Sr.

Values of the isotopic ratios higher than 0.704–0.708 are typical of the arcs whose crust is thicker than 30 km (Gill, 1981). Starting from these data, the thickness of the crust for the Igniş-Văratec andesites was calculated according to Condie's relation (1973, fide Bleahu, 1983). More than 200 analyses, representing all the main types, were considered. The values occur generally within a quite large interval (15–42 km), most of them being assigned to the interval of 25–35 km.

The punctual seismic drilling (Socolescu et al., 1972) indicate in the Negreşti-Oaş area a depth of 28–31 km for the Moho and SE of Baia Mare the depth of 33 km, increasing under the Carpathians.

The interpretation of the K, Rb, Sr distribution and of the $^{87}Sr/^{86}Sr$ ratios in the Igniş-Văratec andesites points out a significant crustal contamination of the parental magma, both during the crossing of the sialic crust and during their evolution in secondary magma chambers, where the predominant fractional crystallization processes were associated with assimilation.

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References

Borcoş M., Peltz S., Stan N., Udrescu C., Vasiliu C. (1979) Consideraţii petrochimice şi geochimice asupra vulcanitelor neogene din Munţii Gutîi (II, III, IV). *St. tehn. econ.*, ser. I/16, Bucureşti.

Bleahu M. (1983) Tectonica globală. Edit. ştiint. encicl., Bucureşti.

– , Boccaletti M., Manetti P., Peltz S. (1973) Neogene Carpathian Arc; a continental arc displaying the features of an "island arc". *J. Geophys.*, 78, 23, p. 5025–5032.

Edelstein O., Istvan D., Cojocea C., Weisz G., Bernad A., Stan D., Kovacs M. (1980) Harta geologică a munţilor Oaş-Ţibleş, scara 1:25000.

Faure G., Powell J. L. (1972) Strontium Izotope Geology. Springer-Verlag, Berlin, Heidelberg-New York.

Gill J. B. (1981) Orogenic andesites and plate tectonic. Springer-Verlag, Berlin, Heidelberg-New York.

Hanson N. G. (1980) Rare earth elements in petrogenetic studies of igneous systems. *Ann. Rev. Earth Planet. Sci.*, p. 371–406.

Hart S. R., Brooks C., Krogh F. E., Davis G. L., Nava D. (1970) Ancient and modern volcanic rocks: a trace element model. *Earth Planet. Sci. Lett.*, 10, p. 17–18, Amsterdam.

Irvine T. N., Baragar W. R. A. (1971) A guide of the chemical classification of the common volcanic rocks. *Can. J. Earth Sci.*, 8, Ottawa.

Kovacs M., Edelstein O., Istvan D. (1987) Andezitele din Munţii Oaş-Ţibleş; consideraţii privind definirea şi clasificarea lor pe baza datelor petrochimice. *Stud. cerc. geol. geofiz. geogr., Geologie*, 32, p. 12–24, Bucureşti.

Peccerillo A., Taylor S. R. (1976 a) Geochemistry of Eocene Calc-Alkaline volcanic rocks from Kastamonu area, Northern Turkey. *Contr. Mineral. Petrol.*, 58, 1, Heidelberg.

– (1976 b) Rare earth elements in East Carpathians volcanic rocks. *Earth Planet. Sci. Lett.*, 32, p. 121-126, Amsterdam.

Peltz S., Grabari G., Tănăsescu A., Vâjdea E. (1985) Rb, Sr and K distribution in young volcanics from the Călimani-Harghita and Perşani Mountains. Petrogenetic implications. *D. S. Inst. Geol. Geofiz.*, LXIX/1, p. 323–338, Bucureşti.

– , Stoian M. (1985) REE distribution in young volcanics from the Călimani-Harghita and Perşani Mountains. *D. S. Inst. Geol. Geofiz.*, LXIX/1 (1982), p. 339–349, Bucureşti.

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

Seghedi I., Grabari G., Ianc R., Tănăsescu A., Vâjdea E. (1986) Rb, Sr, Th, U, K distribution in the neogene volcanics of the South Harghita mountains. *D. S. Inst. Geol. Geofiz.*, 70-71/1, p. 453–473, Bucureşti.

Socolescu I., Airinei S., Ciocîrdel R., Popescu N. (1972) Fizica şi structura scoarţei terestre din România. Edit. tehn. ştiinţ., Bucureşti.

Institutul Geologic al României

Taylor S. R. (1969) Trace element chemistry of andesites and associated calc-alkaline rocks. Proc. Andesite Conference Oregon Dept. *Geol. Min. Res. Bull*, 65, Oregon.

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FORMAŢIUNILE METAMORFICE DIN MASIVUL GÎRBOVA (MUNŢII PERŞANI)

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Key words: Metamorphic rocks. Cambrian. Stratigraphic units. Metamorphism. Palynomorphs. Foliation. East Carpathians - Perşani Mountains.

Abstract: *The metamorphic rocks of the Gîrbova Massif (Perşani Mts.)* It is proved that the metamorphics of the Perşani Mts belong to the Cambrian Tulgheş Group. In the structural framework of the East Carpathians, they represent the southernmost outcrop of the crystalline basement of the Bucovinian Nappe. The exposed sequence may be correlated with the formations Tg_3 and Tg_4 of the stratotype from the Bistriţa Mts and the Sîndominic–Gheorghieni region.
Due to the low grade metamorphism (Q+ab+chl+bt+alm-spess+ilm), relict structures and minerals (quartz and feldspar) from the primary volcanic and clastic rocks are preserved. The metamorphhic history includes three stages of deformation with mineral growth, a postkinematic porphyroblasic development of albite, muscovite, biotite, spessartinic almandine and a late retrograde chloritisation of the mentioned biotite and garnet (Fig. 9). The main synkinematic mineral formation is linked to an early caledonian event (sardic phase). The later development may be Variscan.

1. MASIVUL GÎRBOVA ÎN ANSAMBLUL STRUCTURAL DE LA CURBURA CARPAŢILOR

Cristalinul masivului Gîrbova a constituit vreme îndelungată o dilemă în zona de curbură a Carpaţilor. Acest lucru se datora faptului că, în timp ce, din punctul de vedere al faciesurilor depozitelor sedimentare premezocretacice, munţii Perşani reprezintă evident extremitatea sudică a structurilor din Carpaţii Orientali (pînze transilvane şi pînza bucovinică – Patrulius et al., 1966; Săndulescu, 1976), formaţiunile metamorfice erau considerate fie un element exotic, fie un element al Carpaţilor Meridionali (Dessila-Codarcea et al., 1969). După recunoaşterea de către Vodă, Vodă (1985) a apartenenţei acestui cristalin la grupul Tulgheş, a devenit evident că masivul cristalin al Gîrbovei reprezintă continuarea sudică a cristalinului din pînza Bucovinică a Carpaţilor Orientali. Relaţiile faţă de cristalinul Carpaţilor Meridionali (din zona Şinca-Făgăraş) sînt considerate a fi de ordin tectonic, delimitarea fiind realizată printr-o fractură transcrustală – falia Sud-Transilvană – situată sub culoarul Vlădeni şi reluată local în mişcări tinere (falia Dealul Mare; Săndulescu, 1984).

În munţii Perşani cristalinul de tip Tulgheş este acoperit transgresiv de seria depozitelor sedimentare bucovinice, care la rîndul lor suportă tectonic unităţile transilvane, şi anume pînza de Perşani, constituită preponderent din ofiolite, şi pînza de Hăghimaş formată din calcare jurasice şi eocretacice. După Patrulius et al. (1966) între cuvertura sedimentară bucovinică a cristalinului Gîrbovei şi unităţile Transilvane se interpune o unitate alohtonă constituită din depozitele de wildflysch.

Unităţi ale pînzelor central-est carpatice situate sub pînza Bucovinică nu se cunosc în munţii Perşani. De altfel zona frontală a acesteia şi relaţiile cu unităţile flişului (pînza de Baraolt şi pînza de Ceahlău) sînt ascunse sub cuvertura sedimentară post mezocretacică. Prin acest fapt structura munţilor Perşani se deosebeşte de segmentul situat la nord de Miercurea-Ciuc, în care ansamblul pînzelor central-est carpatice (zona cristalino-mezozoică) a fost reşariată, după finele cretacicului, peste dacidele externe (zona flişului).

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2. CERCETĂRI GEOLOGICE ANTERIOARE

Existenţa formaţiunilor metamorfoce în munţii Perşani a fost semnalată de Hauer şi Stache (1863) în urma unui profil parcurs pe valea Comana. O delimitare cartografică a acestui cristalin, inclusiv a ivirii din valea Hamaradia, se datoreşte lui Wachner (1918), care defineşte masivul cristalin al Gîrbovei drept "boltirea fundamentului cristalin de la Veneţia". Din punct de vedere petrografic, el atrage atenţia asupra diferenţelor litologice faţă de cristalinul Făgăraşului, bogat în intercalaţii de amfibolite, care lipsesc în munţii Perşani.

În urma progresului realizat în studierea rocilor sedimentare mezozoice, Ilie, Preda (1940) recunosc structura în pînză a munţilor Perşani. În perioada 1960–1970, Patrulius, Popa, Popescu (1963, 1965, 1966, 1967, rapoarte nepublicate), Murgeanu, Contescu, Jipa, Mihăilescu, Panin (1963, raport nepublicat), Dumitriu, Dumitriu (1964), Kusko et al. (1970, raport nepublicat) aduc contribuţii stratigrafice şi paleontologice la cunoaşterea mai detaliată a depozitelor sedimentare. Patrulius et al. (1966) confirmă pînza munţilor Perşani, recunoscînd încadrarea ei în sistemul pînzelor Transilvane, generate prin decolare şi deplasare gravitaţională în timpul Apţianului. Autorii menţionaţi consideră că această tectogeneză a fost însoţită de acumularea unor depozite de wildflisch, de vîrstă barremiană şi – eventual – bedouliană, care la rîndul lor au fost şariate peste depozitele mezozoice subjacente ("pînza wildflyschului" – Patrulius, 1973).

Spre deosebire de datele referitoare la formaţiunile sedimentare mezozoice, care indicau din ce în ce mai argumentat încadrarea munţilor Perşani în structura Carpaţilor Orientali, cercetările privind formaţiunile metamorfice tindeau spre corelări cu cristalinul Carpaţilor Meridionali. Astfel, Ilie (1953, 1954) grupează şisturile cristaline din zona Comana-Veneţia în "seria epizonală de Veneţia", pe care o echivalează cu seria de Ciuta (Schmidt, 1930) din estul munţilor Făgăraş, fără a exclude însă şi o posibilă paralelizare cu seria de Tulgheş din Carpaţii Orientali.

Prima tratare modernă a formaţiunilor metamorfice din masivul cristalin al Gîrbovei a fost realizată de Dessila Codarcea et al. (1969), care separă o "serie epimetamorfică de Gîrbova" în poziţie transgresivă peste un fundament mai vechi, format din gnaise. In seria de Gîrbova delimitează trei unităţi litostratigrafice: orizontul metagraywackelor (în poziţie inferioară), orizontul şisturilor cuarţo-sericitoase, orizontul filitelor (în poziţie superioară).

Această imagine litostratigrafică şi structurală a fost preluată în harta geologică 1:50 000, foaia Perşani (Popescu, 1970). După Dessila Codarcea et al. (1969) "seria de Gîrbova" reprezintă un flişmetamorfozat în faciesul şisturilor verzi, la finele ciclului baikalian. Autorii citaţi menţionează, de asemenea, procese metamorfice ulterioare, de grad mai scăzut, manifestate prin cloritizarea parţială sau totală a biotitului şi prin recristalizarea muscovitului în condiţii statice.

Dessila Codarcea, Iliescu (1969) corelează seria de Gîrbova cu "complexul metagraywackelor" din versantul nordic al munţilor Făgăraş şi subasmentul ei mezometamorfic cu seria de Măgura din munţii Cibin. În această viziune încadrează "seria de Gîrbova" (împreună cu cristalinul din nordul munţilor Făgăraş) unui "metafliş intracarpatic" pe care-l atribuie Proterozoicului superior, pe baza unor date microfloristice care însă nu exclud posibilitatea încadrării în Cambrianul inferior.

Preciziuni noi referitoare la încadrarea cristalinului din munţii Perşani în edificiul structural al Carpaţilor Orientali, şi anume în pînza Bucovinică, au fost aduse de Săndulescu (1976, 1984) pe baza cuverturii sedimentare mezozoice a cristalinului Gîrbovei. La nivelul formaţiunilor cristaline această încadrare structurală a fost confirmată de Vodă, Vodă (1983, 1984, rapoarte nepublicate, 1985) prin recunoaşterea faptului că metamorfitele masivului Gîrbovei reprezintă un echivalent al grupului Tulgheş. Autorii menţionaţi recunosc în munţii Perşani formaţiunea Tg_3 cu metavulcanite riolitice şi formaţiunea Tg_4 cu roci blastodetritice şi filite cu "paramorfoze de rutil după brookit". Suprafeţe relativ întinse sînt atribuite grupului Bretila, considerat într-o poziţie echivalentă pînzei de Rarău. În linii foarte generale rocile atribuite de Vodă, Vodă (1983, 1984, rapoarte nepublicate, 1985) formaţiunii Tg_3 ar corespunde, după Dessila Codarcea et al. (1969), subasmentului mezometamorfic al "seriei de Gîrbova", în timp ce aceasta din urmă s-ar suprapune atît peste formaţiunea Tg_4, cît şi peste grupul Bretila din imaginea cartografică dată de Vodă, Vodă (1983, 1984, rapoarte nepublicate, 1985).

Recent, sectoare din bazinul văilor Comana, Veneţia şi Arinoasa au fost acoperite prin lucrări de prospecţiune geologică detaliată (Vodă, Velio, 1985, raport nepublicat), prospecţiuni geochimice şi geofizice (Ciobanu et al., 1987, raport nepublicat).

3. LITOSTRATIGRAFIA GRUPULUI TULGHEŞ DIN MASIVUL GÎRBOVA

Modelul litostratigrafic elaborat pentru masivul Gîrbova (fig. 1) se bazează pe admiterea unei boltiri anticlinale pe aliniamentul Valea Dabijului-Valea Secăturii, în care aflorează metavulcanitele riolitice din formaţiunea

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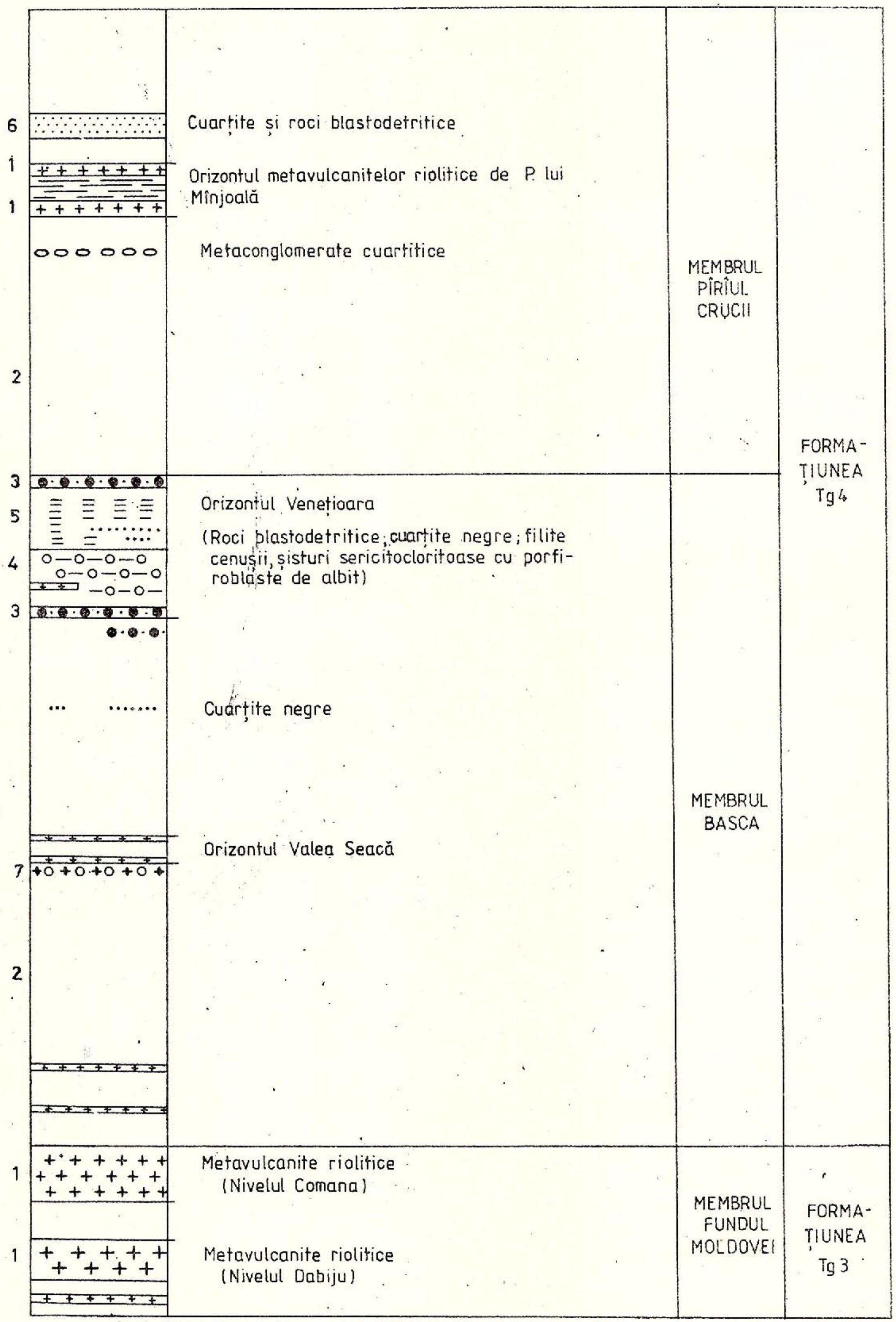

Fig. 1 – Succesiunea litostratigrafică în grupul Tulgheş din masivul Gîrbova.
1, metavulcanite riolitice; 2, şisturi sericito-cloritoase în alternanţă cu şisturi sericitoase cenuşii şi intercalaţii de şisturi cu porfiroblaste de albit; 3, roci blastodetritice; 4, şisturi sericito-cloritoase cu porfiroblaste de albit; 5, şisturi sericitoase grafitoase; 6, cuarţite sericitoase şi roci blastodetritice; 7, şisturi sericito-cloritoase cu porfiroblaste de albit şi diseminări de pirită±calcopirită.

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Tg$_3$. O simetrie a succesiunilor litologice pe cele două flancuri ale structurii anticlinale este evidentă pentru partea inferioară a coloanei litostratigrafice. Pentru nivelele superioare simetria litologică nu poate să fie examinată, deoarece cuvertura sedimentară se dispune asimetric pe structura plicativă a cristalinului. Din această cauză modelul litostratigrafic propus se bazează, în special, pe succesiunea de pe flancul vestic al boltirii anticlinale Comana, expus în bazinele văii Comana, Valea Largă, Valea Gîrbovei, Valea Seacă, valea Veneţioara, pîrîul Răşinari, valea Veneţia şi valea Arinoasa.

Elementul litostratigrafic cel mai evident constă în superpoziţia geometrică a două asociaţii litologice distincte, reprezentate printr-o asociaţie vulcano-sedimentară riolitică, în partea inferioară, şi o secvenţă metapelitică cu intercalaţii de roci metapsamitice, în partea superioară. Detalii de ordin petrografic şi litostratigrafic nu lasă nici un dubiu asupra echivalenţei celor două asociaţii litologice menţionate cu formaţiunile Tg$_3$ şi Tg$_4$ ale grupului Tulgheş din munţii Bistriţei şi din regiunea Sîndominic-Gheorgheni (Carpaţii Orientali)

Formaţiunea Tg$_3$

Limita dintre formaţiunea Tg$_3$ şi Tg$_4$ a fost trasată deasupra stivei principale de metavulcanite şi metaepiclastite riolitice. În această situaţie, ultimele vestigii ale vulcanismului riolitic, şi anume 2–3 nivele de roci cuarţo-feldspatice, în general subţiri, rămîn incluse în formaţiunea Tg$_4$.

Delimitarea menţionată asigură limitei un isocronism mai bun decît luarea în considerare a ultimelor produse riolitice cu răspîndire sporadică şi discontinuă. Ea marchează schimbarea în regimul de sedimentare după paroxismul vulcanic; deasupra limitei respective încep să apară şisturi cu porfiroblaste de albit, şisturi sericitoase-cloritoase cenuşii cu pseudomorfoze de rutil după ilmenit, reprezentînd elemente litologice caracteristice pentru formaţiunea Tg$_4$.

De asemenea, ea corespunde limitei dintre Tg$_3$ şi Tg$_4$ admisă în munţii Bistriţei, unde echivalenţe ale nivelelor subţiri de metavulcanite riolitice în baza formaţiunii Tg$_4$ se regăsesc, de exemplu, în metariolitele de Praşca.

În formaţiunea Tg$_3$ din zona anticlinală Comana au fost distinse două orizonturi principale de roci cuarţo-feldspatice, separate printr-un nivel de şisturi sericito-cloritoase ± biotit.

Orizontul Comana (superior) apare cu grosimi relativ constante la partea superioară a formaţiunii Tg$_3$, pe întreaga arie de extindere a acesteia. Excepţie face partea de sud-vest, unde a fost retezat de o falie cu extindere regională.

Orizontul Dabiju este separat de orizontul superior printr-o bandă continuă de şisturi sericito(muscovito)-cloritoase ± cuarţoase. Din punct de vedere litologic are o constituţie identică cu aceea a nivelului Comana. Spre partea inferioară se intercalează mai multe nivele de şisturi sericito-cloritoase-cuarţoase.

Pentru orizontul superior este evident că reprezintă un echivalent, cel puţin parţial, al membrului Fundu Moldovei din Carpaţii Orientali. Pentru secvenţa situată sub acest orizont poate fi luată în considerare o echivalare fie tot cu membrul Fundu Moldovei, fie cu membrul Leşu Ursului, în care caz banda de şisturi interpusă ar corespunde membrului Moroşan.

Formaţiunea Tg$_4$

Şisturile atribuite formaţiunii Tg$_4$ constituie masa principală a cristalinului din masivul Gîrbova. Sînt dispuse pe cele două flancuri ale boltirii anticlinale Comana. Succesiunea cea mai completă (cca 2000 m) este expusă pe flancul vestic al structurii plicative menţionate; ea poate fi considerată drept stratotip pentru grupul de Tulgheş din munţii Perşani.

Identificarea mai multor nivele reper cu asociaţii litologice specifice a permis subdivizarea litostratigrafică a formaţiunii Tg$_4$. Semnificativ în acest sens este în special nivelul constituit dintr-o asociaţie de roci cuarţo-feldspatice blastodetritice, uneori cu cuarţ detritogen violaceu, remarcat şi de Vodă, Vodă (1985). Avînd în vedere atît asemănarea petrografică foarte evidentă, cît şi poziţia litostratigrafică asemănătoare faţă de orizontul Arşiţa Rea din munţii Bistriţei şi din regiunea Gheorghieni-Sîndominic, reperul litostratigrafic respectiv a fost folosit pentru separarea, în cristalinul din masivul Gîrbova, a două unităţi cu rang de membru, echivalabile membrilor Basca şi Pîrîul Crucii din sectoarele mai nordice ale Carpaţilor Orientali.

Membrul Basca. Este constituit dintr-o alternanţă de şisturi sericito-cloritoase, şisturi sericitoase cenuşii cu pseudomorfoze de rutil după ilmenit, şisturi sericito-cloritoase cu porfiroblaste de albit, şisturi cuarţitice şi cuarţite cu sericit şi clorit. Local apar şisturi sericito-cloritoase cu magnetit.

În partea inferioară a secvenţei predomină şisturile sericito-cloritoase-cuarţoase cu rare intercalaţii de şisturi cu porfiroblaste de albit. Aproape de baza succesiunii au fost întîlnite local nivele discontinue de cuarţite negre (± granat spessartinic), echivalabile ca poziţie litostratigrafică cu cuarţitele negre de Sîndominic. Tot în partea inferioară a stivei se individualizează 2-3 nivele subţiri de metavulcanite şi metaepicalstite riolitice, care amintesc atît prin constituţia litologică, cît şi prin poziţia litostratigrafică, nivelul metariolitelor de Praşca din

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munţii Bistriţei.

În partea mediană a stivei se remarcă un orizont relativ subţire – *orizontul Valea Seacă* – constituit din 1-2 nivele de cuarţite albe sericitoase feldspatice (metaepiclastite şi metavulcanite acide), uneori limonitizate datorită prezenţei piritei. Local (pe Valea Seacă) se asociază şisturi albitice clorito-sericitoase cu diseminări stratiforme de pirită (± calcopirită). Roci asemănătoare au fost întîlnite pe Valea Secăturii, în derocările de lîngă galeria 1. Sub orizontul Valea Seacă, în versantul drept al văii Comana, apare primul nivel de roci blastodetritice cu extindere locală. Poziţia acestor roci blastodetritice faţă de nivelele subţiri de roci albe cuarţo-feldspatice din orizontul Valea Seacă aminteşte orizontul Cianod separat în regiunea Gheorghieni, într-o poziţie litostratigrafică similară şi orizontul Tonigăreşti de la Fundu Moldovei.

Deasupra orizontului Valea Seacă se individualizează un nivel subţire şi discontinuu de cuarţite negre asociate cu şisturi sericitoase, slab grafitoase.

Orizontul Veneţioara constituie termenul superior al membrului Basca. Este constituit atît la partea superioară, cît şi la partea inferioară din cîte un nivel continuu de roci blastodetritice de tip Arşiţa Rea. Între aceste două repere apare, în partea inferioară, o alternanţă de şisturi sericito-cloritoase şi şisturi cu porfiroblaste de albit, în timp ce în partea superioară predomină şisturi sericitoase cenuşii (± grafit), asociate local cu nivele decimetrice şi discontinue de cuarţite negre grafitoase. Orizontul Veneţioara este echivalent cu orizontul Arşiţa Rea din munţii Bistriţei, fie prin unul din cele două nivele de roci blasodetritice, fie prin ambele.

Membrul Pîrîul Crucii. Este format dintr-o asociaţie de şisturi sericito-cloritoase, şisturi sericitoase cenuşii cu pseudomorfoze de rutil după ilmenit şi şisturi cuarţitice sericito-cloritoase situate deasupra orizontului Veneţioara.

În jumătatea inferioară a stivei se conturează un nivel subţire de metamicroconglomerate cuarţitice, cu elemente relicte de cuarţ rotunjit, bine conservate.

În partea mediană a succesiunii se intercalează un orizont de metavulcanite riolitice – *orizontul Pîrîul lui Mînjoală* – constituit din nivele de roci albe cuarţo-feldspatice, între care se intercalează şisturi sericito-cloritoase. Unele din rocile cuarţo-feldspatice au structuri relicte porfirice foarte bine conservate. După poziţia lor litostratigrafică aceste metavulcanite acide ar corespunde metariolitelor de Dealul Fagi din munţii Bistriţei.

Spre partea superioară a succesiunii se individualizează un orizont constituit predominant din şisturi cuarţitice cu clorit, sericit, biotit la care se asociază subordonat şi local roci blastodetritice. Această asociaţie de roci predominant cuarţoase aminteşte dezvoltarea detritogenă-cuarţoasă a membrului Afinet din munţii Bistriţei.

4. VÎRSTA CRISTALINULUI DIN MASIVUL GÎRBOVA

Întrucît cristalinul din masivul Gîrbovei se corelează, atît litologic, cît şi pînă la detalii litostratigrafice, cu grupul Tulgheş din Carpaţii Orientali, vîrsta cambriană a sedimentării şi metamorfismul ordovician (caledonian timpuriu), stabilite prin date palinologice (Iliescu, Mureşan, 1972; Iliescu et al., 1983) şi radiometrice (Mînzatu et al., 1975; Kräutner et al., 1976) în munţii Bistriţei şi în regiunea Bălan, sînt evidente şi pentru metamorfitele din munţii Perşani.

Primele datări palinologice pentru cristalinul Gîrbovei au fost efectuate de Dessila Codarcea şi de Iliescu (1969). Palinomorfele identificate se încadrează în intervalul Pt–Cb_1, cu excepţia unei forme (*Stenozonoligotriletum validum* Tim.) pe baza căreia autorii menţionaţi au încadrat "seria de Gîrbova" în partea terminală a Proterozoicului superior, avînd în vedere şi unele corelări cu şisturile cristaline din Carpaţii Meridionali. Cercetările palinologice au fost continuate de Iliescu şi Mureşan (date nepublicate), precizînd vîrsta cambriană a cristalinului Gîrbovei. Pentru studiul de faţă au fost investigate patru probe (din care numai una a fost sterilă), repartizate pe întregul interval litostratigrafic al formaţiunii Tg_4 (tab. 1, fig. 2).

Asociaţia obţinută este săracă şi cuprinde următoarele palinomorfe: *Protospheridium sp.*, *P. densum.* Tim., *P. tuberculiferum* Tim., *Gleocapsomorpha sp.*, *Symplassosphaeridium sp.*, *Synsphaeridium sorediforme* Tim., *Favosphaeridium favosum* Tim.

Probele colectate anterior de Dessila Codarcea, Mureşan şi studiate palinologic de către Iliescu au furnizat următoarea asociaţie: *Zonosphaeridium sp.*, *Protosphaeridium sp.*, *P. conglutinatum* Tim., *Stenozonoligotriletum sokolovi* Tim., *S. validum* Tim., *Psophosphaera obscura* Pisch., *Leioligotriletum bistrovi* Tim., *Uniporta sp.*, *Protosphaeridium cf. flexuosum* Tim., *P. minutissimum* Tim., *P. cf. clarum* Tim.

Distribuţia litostratigrafică şi stratigrafică a întregului material palinologic existent la ora actuală din cristalinul Gîrbovei este sistematizată în tabelul 1.

Din analiza acestor date se conturează trei categorii de elemente microfloristice (tab. 1): a) elemente de largă circulaţie, din Proterozoic pînă în Ordovician; b) elemente microfloristice cu interval de evoluţie mai

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scurt, cuprinzînd Proterozoicul superior şi Cambrianul inferior; c) elemente microfloristice cu interval de evoluţie delimitat la Cambrianul inferior (*Psophosphaera obscura* Pisch., *Uniporta sp.*). Prezenţa unor genuri a căror evoluţie se încheie în Cambrianul inferior, în asociaţie cu forme caracteristice pentru Cambrianul inferior ne conduc la concluzia că formaţiunea investigată corespunde Cambrianului inferior, cu posibilitatea de a trece în Cambrianul mediu.

Tabelul 1

Distribuţia stratigrafică a palinomorfelor identificate în grupul Tulgheş din Munţii Perşani

PRECAMBRIAN		CAMBRIAN			ORDOVICIAN	Unităţi taxonomice	Nr. probă
1	2	1	2	3			
	—	—	—	—	—	*Zonospaeridium sp.*	18^{x}
—	—	—	—	—	—	*Protosphaeridium sp.*	18^{x} G-2
—	—	—	—	—	—	*Protosphaeridium conglutinatum* Tim.	10^{xx}
—	—	—	—	—	—	*Stenozonoligotriletum sokolovi* Tim.	13^{xx}
		—				*Uniporata sp.*	18^{x}
		—				*Psophosphaera obscura* Pisch.	18^{x}.
—	—	—				*Leioligotriletum bistrovi* Tim.	20^{x}
	—	—				*Stenozonoligotriletum validum* Tim.	15^{xx}
	—	—				*Symplassosphaeridium sp.*	G-87
	—	—				*Synsphaeridium sorediforme* Tim.	G-87
	—	—				*Protosphaeridium cf. flexuosum* Tim.	17^{x}
	—	—				*Protosphaeridium clarum* Tim.	20^{x}
	—	—				*Protosphaeridium minutissimum* Tim.	10^{xx}
—	—	—				*Protosphaeridium tuberculiferum* Tim.	G-2
—	—	—				*Gleocapsomorpha sp.*	G-87
—	—					*Favosphaeridium favosum* Tim.	G-83
—	—					*Protosphaeridium densum* Tim.	G-87

Sursa probelor:

G- Probe colectate în 1988

x Probe colectate de M. Mureşan şi analizate de V. Iliescu (1970, raport nepublicat)

xx Probe după M. Dessila Codarcea şi V. Iliescu (1969)

Din distribuţia litostratigrafică a acestor categorii de asociaţii microfloristice (fig. 2) rezultă că formele reprezentative pentru Cambrianul inferior se situează în partea inferioară a formaţiunii Tg_4. Probele situate spre partea superioară a succesiunii din munţii Perşani (fig. 2) au furnizat o asociaţie săracă, de tipul celor întîlnite în Proterozoic. Nu este exclus ca formele respective să reprezinte elemente remaniate. Datarea pentru Cambrianul inferior în munţii Perşani se referă, deci, la partea inferioară a formaţiunii Tg_3 (Membrul Basca). În Carpaţii Orientali au fost identificate asociaţii caracteristice pentru Cambrianul inferior în formaţiunile Tg_2 şi Tg_3 (Iliescu et al., 1983). Palinomorfele care urcă mai sus în Cambrian, în parte pînă în Ordovician, se află intercalate în formaţiunea Tg_4 la nivele cu poziţie superioară faţă de acelea în care se află asociaţia atribuită Cambrianului inferior din munţii Perşani.

5. PRINCIPALELE TIPURI PETROGRAFICE DIN MASIVUL CRISTALIN AL GÎRBOVEI

5.1 Roci de origine sedimentară

5.1.1 Roci blastodetritice. Rocile blastodetritice reprezintă un element litologic caracteristic pentru formaţiunea Tg_4 din grupul Tulgheş. Ele formează strate cu grosimi de ordinul metrilor, mai rar pînă la zeci de metri, dispuse la patru nivele principale: un nivel discontinuu sub orizontul Valea Seacă (în versantul drept al văii Comana), două nivele continue care delimitează orizontul Veneţioara (în culmea de la nord de valea Comana şi în bazinul Veneţioara), un nivel discontinuu în asociaţie cu roci cuarţitice la partea superioară a succesiunii.

Rocile au un aspect masiv în bancurile groase şi sînt şistoase în intercalaţii subţiri. Structura blastodetritică relictă este vizibilă cu ochiul liber, datorită granulelor de feldspat alb şi de cuarţ (uneori de culoare violacee), dispuse într-o masă sericito-cloritoasă. Local se observă tranziţii la varietăţi de culoare verde, datorită îmbogăţirii

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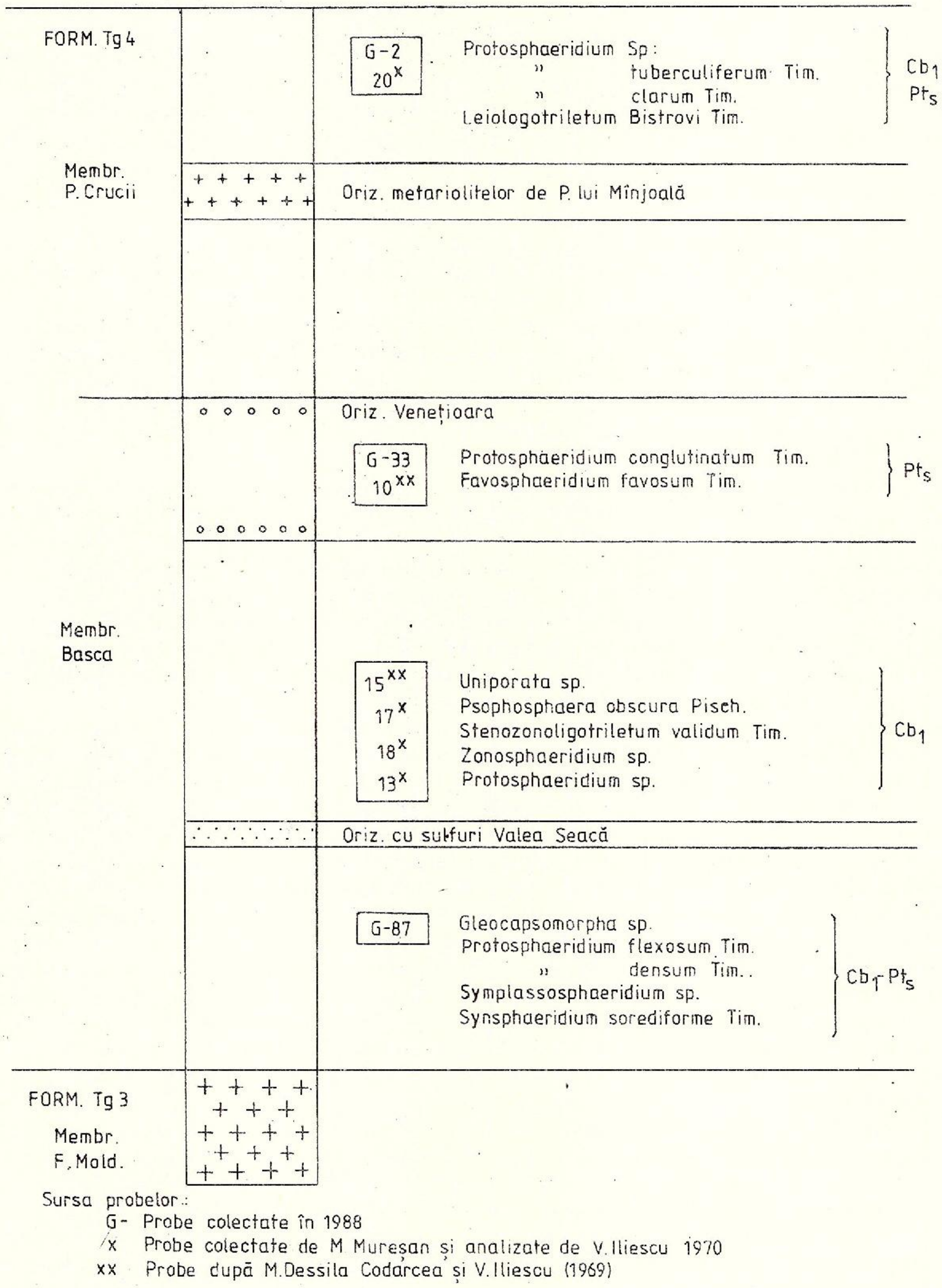

Fig. 2 – Distribuţia litostratigrafică a palinomorfelor identificate în cristalinul din munţii Perşani.

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rocii în clorit. Sub microscop se remarcă o structură granoblastică, marcată de cristalele de cuarţ şi albit de dimensiuni mai mari, situate într-o masă mediu granulară, care îmbracă aspecte lepidogranoblastice.

Parageneza minerală include cuarţ+albit+muscovit+clorit±biotit±granat. Dintre aceste minerale cuarţul şi albitul reprezintă în parte minerale relicte din roca psamitică primară. În cazul cristalelor de cuarţ se remarcă tranziţii treptate spre agregate poligonale de cuarţ mărunt, rezultate în urma deformărilor şi recristalizărilor metamorfice a granulelor de cuarţ detritice. Natura relictă a cristalelor de albit este sugerată uneori de prezenţa unor zone centrale slab sericitizate (± calcit), care marchează plagioclazul primar reorganizat, înconjurate de aureole cu aspect limpede, rezultate prin supracreştere metamorfică (pl. II, fig. 1, 2).

Micele (muscovitul, cloritul±biotit) sînt dispuse preponderent pe planele de şistozitate în varietăţile şistoase şi apar cu dispoziţii divergente în varietăţile masive.

5.1.2. Cuarţite. Cuarţite apar atît în formaţiunea Tg_3, cît şi în formaţiunea Tg_4, sub forma unor nivele cu grosimi ce variază de la cîţiva metri pînă la zeci de metri. În funcţie de compoziţia mineralogică, se pot distinge:

- *Cuarţite cu sericit şi clorit.* Sînt frecvent întîlnite în tot perimetrul, fie asociate cuarţitelor cu feldspat din formaţiunea Tg_3, fie în alternanţă cu şisturile şi filitele sericito-cloritoase din formaţiunea Tg_4. Sînt roci slab şistoase sau masive, dure, compacte, de culoare albă-cenuşie sau verzuie. Au o structură granolepidoblastică, orientarea fiind dată în special de dispunerea sericitului. Cuarţul apare în granule cu contur neregulat, de dimensiuni reduse, cu extincţie ondulatorie. Sericitul şi cloritul apar dispuse, fie divergent între cristalele de cuarţ, fie pe planele de şistozitate.
- *Cuarţite grafitoase.* Apar în mod subordonat, în strate subţiri şi discontinue, intercalate la cîteva nivele în formaţiunea Tg_4. Sînt roci masive sau şistoase, dure, de culoare negricioasă, uneori cu aspect rubanat, datorat alternanţei unor benzi mai mult sau mai puţin bogate în grafit.
- *Cuarţite cu granaţi.* Au fost întîlnite în extremitatea estică a zonei pe valea Comana. Sînt situate într-o poziţie litostratigrafică asemănătoare cu cuarţitele negre de Sîndominic din regiunea Sîndominic–Gheorghieni. Rocile au o culoare negricioasă şi sînt compuse în special din cuarţ, care alternează cu benzi subţiri alcătuite din clorit, sericit şi muscovit. Granaţii apar în porfiroblaste, sînt de dimensiuni mari (pînă la 5 mm) şi sînt frecvent cloritizaţi marginal sau pe fisuri. Porfiroblastele de granaţi conţin incluziuni de cuarţ şi minerale opace. Uneori se remarcă efecte de anizotropie, care trădează structuri zonare.
- *Cuarţite cu feldspat.* Apar între nivelele de metariolite şi metaepiclastite riolitice din formaţiunea Tg_3 şi din orizonturile Valea Seacă şi Pîrîul lui Mînjoală din formaţiunea Tg_4. Sînt greu de delimitat cartografic, deoarece există treceri gradate spre epiclastite. Compoziţia mineralogică cuprinde pe lîngă cuarţ şi albit în mod subordonat ortoză-microclin, sericit şi muscovit. Rocile sînt de regulă de culoare albă, cu structură masivă, rubanată sau slab şistoasă. Rubanarea este dată de alternanţa unor benzi de cuarţ cu benzi cuarţo-feldspatice, sericit, ± clorit. Aceste varietăţi rubanate trec lateral în metaepiclastite riolitice. Local cuarţitele feldspatice conţin diseminări de pirită, care prin alterare dau un aspect caracteristic rocii.

5.1.3 Metamicroconglomerate. Metamicroconglomeratele cuarţoase au fost întîlnite în versantul stîng al văii Comana şi în valea Veneţioara, sub forma unui nivel cu grosime redusă, intercalat în şisturile situate sub orizontul metavulcanitelor riolitice de Pîrîul lui Mînjoală. Macroscopic se disting granule de cuarţ într-o masă cuarţitică fină cu structură masivă. Sub microscop se recunosc relicte de granule rotunjite, distribuite neomogen într-o masă cuarţoasă mărunt cristalizată (pl. I, fig. 5). În jurul elementelor relicte de cuarţ se remarcă concreşteri metamorfice cu cuarţul micrograunlar din matrice (pl. I, fig. 6), care local se pot uni în zone de supracreştere subţiri şi discontinue. Subordonat apar paiete de sericit a căror orientare trădează poziţia şistozităţii metamorfice.

5.1.4 Şisturi sericito-cloritoase, mai mult sau mai puţin cuarţoase, formează fondul litologic al formaţiunii Tg_4. În funcţie de participarea cantitativă şi de granulaţia relictă a cuarţului şi a feldspatului se disting varietăţi metapelitice şi metapsamitice. Dispunerea alternativă a acestor două tipuri petrografice sugerează o sedimentare ritmică în cadrul căreia îşi au locul şi intercalaţiile de cuarţite, roci blastodetritice şi metaconglomerate. Frecvent se disting două foliaţii metamorfice dintre care prima este antrenată în cute cu alunecare concentrică, iar a doua urmăreşte planele axiale ale unor cute de forfecare mai tinere, în general coaxiale cu primele elemente plicative.

Metapelitele au, de obicei, aspect filitic şi sînt alcătuite preponderent din sericit şi clorit dispus în benzi cu aspect laminat, conforme cu şistozitatea principală şi afectate de microcute. Între aceste benzi filosilicatice se intercalează benzi subţiri de cuarţ şi albit micrograunlar.

Metapsamitele se caracterizează prin participarea substanţială a cuarţului şi uneori a albitului, blasteză mai avansată a filosilicaţilor şi o orientare mai puţin riguroasă a acestora. Cuarţul apare în cristale mai mari

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şi poate avea extincţie ondulatorie. Asociaţia minerală cuprinde cuarţ + clorit + sericit + albit (2–5 % An) ± biotit ± granat ± ilmenit. Atît în metapelite, cît şi în metapsamite sînt abundente pseudomorfozele de rutil după ilmenit, de regulă conforme cu foliaţia şi deformate atît pe S_1, cît şi pe S_2. Frecvent apar cristale de muscovit ($2V_x=35^0$) larg dezvoltate, crescute oblic sau transversal pe şistozitatea marcată prin sericit şi clorit. De asemenea, biotitul şi granatul apar, de obicei, porfiroblastic.

5.1.5 Şisturile sericito-cloritoase cu porfiroblaste de albit reprezintă un element petrografic caracteristic pentru cristalinul din masivul Gîrbova. Apar, sub formă de intercalaţii decimetrice sau metrice, în formaţiunea Tg$_4$. Sînt şisturi compacte verzui, în care porfiroblastele de albit sînt vizibile cu ochiul liber. Frecvent se remarcă două foliaţii metamorfice. Fondul rocilor constă dintr-o masă metapelitică sau metapsamitică, similară cu cea a şisturilor sericito-cloritoase comune. Porfiroblastele de albit (2-5 An %) conţin frecvent incluziuni de cuarţ, zircon, turmalină, muscovit sau ilmenit. Uneori incluziunile sînt dispuse în şiruri paralele, atît conforme, cît şi neconforme cu foliaţia rocii, indicînd evoluţii polistadiale cu rotiri în unele faze de creştere (pl. II, fig. 3). Uneori se remarcă coroane de supracreştere (pl. II, fig. 4, 6). Muscovitul apare frecvent porfiroblastic, crescut oblic pe S_1 şi S_2.

Tabelul 2

Analize chimice şi spectrale pentru principalele tipuri petrografice din cristalinul Gîrbovei

	Metavulcanite riolitice					Roci blastodetritice		Şisturi	
	G–6	G–41	G–83	G–11	G–13	2022	4344P	4339P	2P
SiO_2	74,84	77,42	75,11	77,10	70,21	64,53	71,95	49,81	49,57
TiO_2	0,13	0,22	0,12	0,30	0,75	0,65	0,44	1,32	0,92
Al_2O_3	12,40	11,45	11,95	11,30	12,70	17,15	12,90	22,94	23,29
Fe_2O_3	0,98	0,88	0,44	0,93	1,21	1,57	0,87	3,48	3,10
FeO	1,10	0,94	1,46	0,41	2,54	2,27	3,39	6,57	6,98
MnO	0,06	0,03	0,04	0,03	0,06	0,03	0,10	0,35	0,06
MgO	0,32	0,66	0,50	0,27	2,00	1,64	1,50	2,78	3,40
CaO	0,21	0,95	0,36	0,26	1,87	2,77	0,88	0,75	0,56
K_2O	7,28	1,00	7,40	6,15	5,76	2,62	2,01	3,80	4,24
Na_2O	1,21	5,02	1,84	2,07	1,71	4,15	2,66	2,71	1,48
P_2O_5	0,03	0,04	0,02	0,11	0,10	0,19	0,09	0,30	0,35
CO_2							0,51		
H_2O	0,72	0,89	0,31	0,53	0.98	1,97	2,19	4,57	5,42
S	0,24	0,14	0,14	0,16	0,15	0,14	0,15	0,19	0,20
Total	99,52	99,64	99,69	99,53	99,86	99,68	99,64	99,57	99,57
Pb	4	2	18	9,5	7,5	4	18	45	20
Cu	2	5,5	2	8	9	21	13	50	26
Zn	45	45	65	50	50	75	60	80	46
Ga	7,5	10	6	5	9.5	15	23	38	36
Sn	2	2	2	2	2	3	5,5	5.5	4
Ni	2	2	7	2	9	15	5	50	44
Co	2	2	2	2	8	7	3	15	11
Cr	6	5,5	23	9	26	54	9,5	100	100
V	8	9	6,5	13	52	70	18	120	125
Sc	3,5	4	4	4	10	15	13	20	20
Zr	36	55	37	110	115	330	580	220	155
Y	20	22	18	13	24	42	28	28	33
Yb	1,6	1,7	2,1	1	2,3	3	3,2	2.9	3,6
La	30	30	30	30	30	65	55	68	60
Ba	800	370	750	1500	480	19	700	550	700
Sr	44	100	850	52	70	150	46	62	140

Metariolite din Tg$_4$: G-6, V. Comana; G–41, V. Largă; G–83, S. Seacă

Metariolite din Tg$_3$: G-11 şi G-13, V. Dabijului

Roci blastodetritice: 2022, izvoarele Văii Largi; 4344P, afluent drept Valea Veneţia;

4339P, şist sericito-cloritos cu porfiroblaste de albit;

2P, filit sericitos cu pseudomorfoze de rutil după ilmenit.

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5.1.6 Şisturi sericito-cloritoase cu porfiroblaste de granat, biotit şi albit. Varietăţi cu granat şi/sau biotit porfiroblastic au fost întîlnite în asociaţie cu şisturile sericito-cloritoase cu porfiroblaste de albit, cu precădere în partea sud-estică a masivului, dar ele nu lipsesc în restul regiunii. Granatul este reprezentat printr-o varietate de almandin spessartinic şi apare în cristale adesea idiomorfe, de mărimi pînă la 5 mm. Aceste cristale conţin incluziuni de cuarţ şi ilmenit şi sînt afectate local de procese de cloritizare. Biotitul creşte oblic pe foliaţia principală. De obicei este proaspăt, dar local se remarcă cloritizări care, în cazuri izolate, pot avansa pînă la substituţia totală. Roca conţine de regulă şi un muscovit nou, crescut porfiroblastic oblic pe şistozitate. Ilmenitul este omniprezent, atît în masa rocii, cît şi sub formă de incluziuni în porfiroblastele de albit, granat, biotit şi muscovit.

5.2 Roci de origine vulcanică

Metavulcanitele riolitice apar cu precădere în formaţiunea Tg$_3$, din anticlinalul Comana, unde formează două stive (orizontul Comana şi orizontul Dabiju) cu grosimi de cca. 100 m şi cu intercalaţii de şisturi sericito-cloritoase. Metavulcanitele din formaţiunea Tg$_4$ apar în strate cu grosimi reduse, în cadrul orizontului metavulcanitelor riolitice de Pîrîul lui Mînjoală şi în orizontul Valea Seacă.

După aspectele structurale relicte şi compoziţia mineralogică pot fi distinse varietăţi metaeruptive (metariolite) şi metaepiclastite.

5.2.1 Metariolite. Metariolitele apar de regulă asociate cu rocile de origine epiclastică. Se prezintă în varietăţi masive, albe sau cenuşii şi în varietăţi cu rubanare metamorfică. Sub microscop se observă o structură blasto-porfirică (porfirică relictă) evidentă: într-un fond microgranoblastic alcătuit din cristale de cuarţ, feldspat, sericit, apar fenocristale relicte de cuarţ, feldspat potasic, albit, cu dimensiuni de pînă la 3 mm. Fenocristalele sînt dispuse uniform în masa rocii şi prezintă adesea conture cristalografice relicte (cuarţ bipiramidat, feldspat idiomorf) (pl. I, fig. 1).

În fenocristalele de cuarţ se recunosc figuri de coroziune magmatică şi tendinţe de recristalizare în aglomerări de cuarţ metamorfic sub efectul forfecării pe şistozitatea metamorfică. Feldspatul relict este reprezentat prin concreşteri de ortoză şi albit, formate prin descompunerea metamorfică a feldspaţilor sodo-potasici de temperatură înaltă din roca vulcanică (pl. I, fig. 2). Rar se remarcă fenocristale relicte de albit.

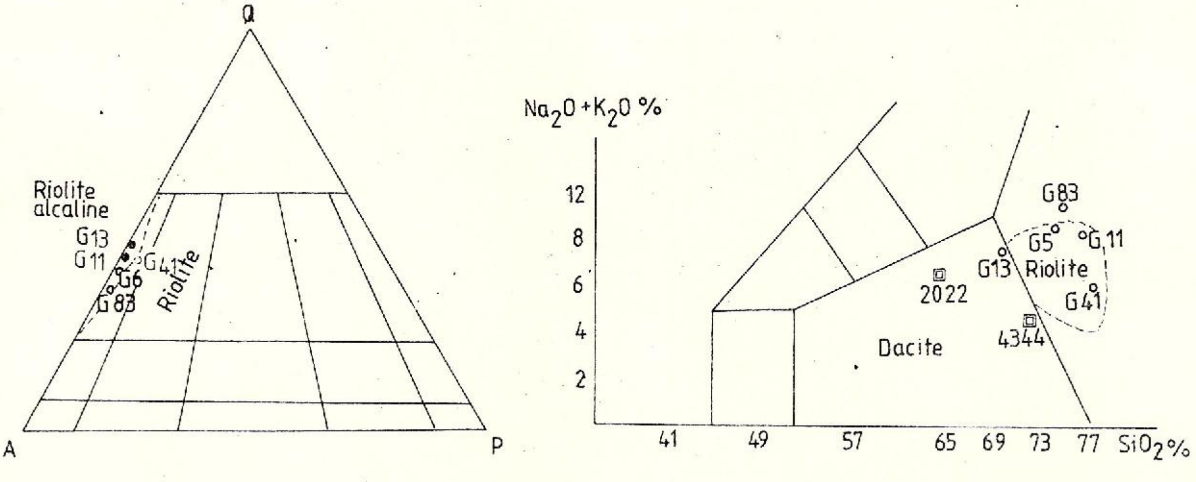

Fig. 3 – Diagramele QAP (Streckeisen) şi Na$_2$O+K$_2$O/SiO$_2$ (TAS) pentru metavulcanitele riolitice din masivul Gîrbova.

Din punct de vedere chimic (tab. 2) rocile corespund riolitelor în diagrama TAS şi riolitelor cu feldspaţi alcalini în diagrama Streckeisen (fig. 3). Este de subliniat faptul că analizele din munţii Perşani se încadrează în cîmpul metariolitelor grupului Tulgheş din sectoarele Fundu Moldovei–Leşul Ursului şi Gheorghieni–Bălan. Elementele urmă se înscriu în spectrul riolitelor. Semnificativ ar putea fi conţinutul mai ridicat în zircon al rocilor din formaţiunea Tg$_3$.

5.2.2 Metaepiclastite şi metatufite riolitice. Drept metaepiclastite şi metatufite riolitice au fost definite roci cuarţo-feldspatice albicioase masive sau rubanate cu structură granoblastică şi aspecte blastodetritice relicte.

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Sînt alcătuite din cuarţ, ortoză, albit, muscovit, biotit, ± granaţi, ± clorit. Unele aspecte structurale, cum ar fi de exemplu figuri de coroziune în cuarţ sau structuri de dezamestec albit-ortoză în cristale de feldspat, indică originea materialului epiclastic din roci de natură riolitică (pl. I, fig. 4). Prin aceasta metaepiclastitele se deosebesc de rocile blastodetritice cu cuarţ şi albit de provenienţă detritogenă. Alte elemente distinctive constau în prezenţa frecventă a cloritului în cele din urmă şi în compoziţia chimică neomogenă, în general mai bazică (echivalent-dacitică) a rocilor blastodetritice (fig. 3; tab. 2). Trecerea de la metaepiclastitele riolitice la cuarţitele feldspatice şi cuarţite se face treptat prin creşterea conţinutului de cuarţ în detrimentul feldspatului. Cuarţitele feldspatice se deosebesc prin gradul de sortare mai avansat al materialului resedimentat.

6. METAMORFISMUL GRUPULUI TULGHEŞ DIN MASIVUL GÎRBOVA

Metamorfismul grupului Tulgheş, atribuit iniţial ciclului baicalian (Giuşcă et al., 1969; Bercia et al., 1976), a fost plasat după ultimele date palinologice (Iliescu et al., 1983) şi radiometrice (Kräutner et al., 1976) într-o fază caledoniană timpurie – faza sardă. El se echivalează, deci, cu evenimentul metamorfic intra-Ordovician cu extindere regională foarte largă în aria vestică a lanţului alpin mediteranean (Becker et al., 1987) şi în vorlandul acestuia din Europa Centrală.

În zona cristalină a Carpaţilor Orientali centrali şi de nord metamorfismul grupului Tulgheş depăşeşte izogradul biotitului. Local se remarcă blasteza unui granat mărunt cristalizat, care probabil este bogat în componenta spessartinică. Evenimentul varistic a conferit cristalinului de tip Tulgheş un caracter polimetamorfic (Balintoni, Chiţimuş, 1973; Kräutner et al., 1975) şi a determinat regenerarea vîrstelor izotopice (Kräutner et al., 1976).

6.1 Feldspaţi

După aspectele morfologice ale feldspaţilor şi structura rocilor respective, în cristalinul masivului Gîrbova se pot distinge, în mod evident, feldspaţi relicţi, reorganizaţi în timpul metamorfismului şi feldspaţi metamorfici, formaţi exclusiv prin procese de blasteză.

Feldspaţii relicţi apar în metariolite, metaepiclastite riolitice şi în rocile blastodetritice (fig. 4) (pl. I, fig. 3, 4).

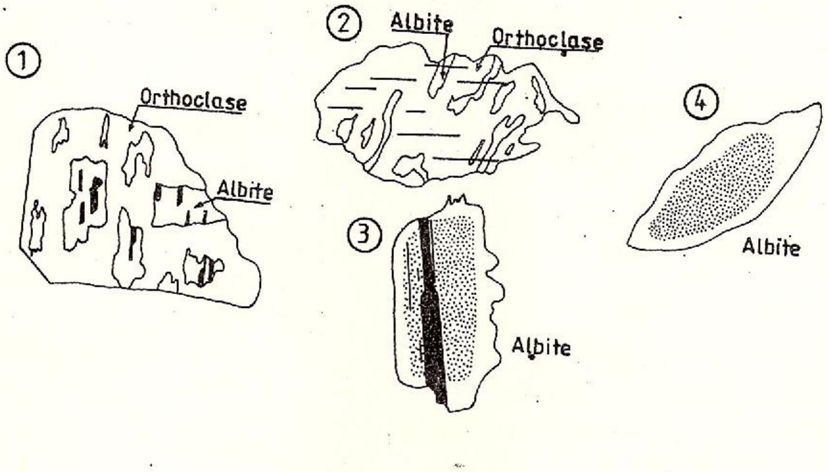

Fig. 4 – Feldspaţi relicţi reorganizaţi metamorfic.
1, feldspat idiomorf sodo-potasic de temperatură ridicată descompus într-o concreştere de albit+ortoză (metariolit); 2, feldspat xenomorf sodo-potasic de temperatură înaltă descompus într-o concreştere de albit+ortoză (metaepiclastit); 3, albit cu incluziuni de sericit şi zonă de supracreştere (metaepiclastit); 4, albit cu incluziuni de sericit şi zonă de supracreştere (rocă blastodetritică.

⁂

În metariolite şi în metaepiclastitele riolitice se întîlnesc fenocristale relicte de albit şi de feldspat sodo-potasic, uneori cu conture idiomorfe parţial conservate (fig. 4). Caracteristic este dezamestecul de ortoză +

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albit (pl. I, fig. 2) format pe seama feldspaţilor sodo-potasici de temperatură înaltă (anortoză, Na-Sanidin) din roca vulcanică primară, în urma adaptării la fazele stabile în metamorfismul regional (Kräutner et al., 1978, raport nepublicat). Realizarea acestui dezamestec implică temperaturi de ordinul a 500^0 C. Gradul de triclinicitate al feldspatului potasic determinat pe faţa 010 ($\Delta = 0{,}5–0{,}7$) este puţin mai ridicat, în comparaţie cu cel din metavulcanitele riolitice din regiunea Bălan, fapt ce ar putea fi corelat cu intensitatea mai ridicată a blastezei postcinematice din munţii Perşani. Albitul din epiclastite (0-5 % An) prezintă uneori zone centrale cu lamele fine de muscovit, înconjurate de aureole de supracreştere limpezi (fig. 4) (pl. II, fig. 1). Zonele centrale sericitizate reprezintă probabil albitul magmatic primar, resedimentat şi reorganizat prin expulzarea potasiului în condiţiile presiunii parţiale a H_2O de la începutul metamorfismului. Într-o altă ipoteză, s-ar putea admite sericitizarea metamorfică a albitului primar, fie datorită reorganizării unor produse de alterare supergenă incipientă din decursul sedimentării, fie sub influenţa fazelor fluide potasice supraîncălzite din stadiul incipient al metamorfismului.

Tabelul 3

Constituţia chimică şi moleculară a granaţilor din şisturile sericito-cloritoase cu porfiroblaste de albit

ANALIZA CHIMICĂ				COMPOZIŢIA CHIMICĂ (Calculată fără incluziuni)			
	2123	2165	2174		2123	2165	2174
SiO_2	37,42	42,16	39,48		36,7	36,7	37,1
TiO_2	0,68	0,89	0,79				
Al_2O_3	20,30	19,95	20,35		20,3	20,5	20,7
$FeO_{(total)}$	23,50	21,01	23,41	Fe_2O_3	0,7	0,3	0,6
				FeO	25,2	25,3	25,8
MnO	10,81	10,80	7,53		12,6	14,1	8,9
MgO	0,84	0,92	0,88		0,5	0,7	1,0
CaO	3,75	1,93	5,55		3,8	2,3	5,9
K_2O	0,36	0,58	0,62				
Na_2O	0,16	0,30	0,18				
P_2O_5	0,40	0,10	0,12				
H_2O	0,81	0,71	0,51				
Total	99,03	99,35	99,42		99,8	99,9	99,9

CATIONI DIN FORMULA STRUCTURALĂ*			
	2123	2165	2174
Si^{4+}	6,00	6,00	6,00
Al^{3+}	3,91	3,96	3,93
Fe^{3+}	0,09	0,04	0,07
Fe^{2+}	3,45	3,46	3,49
Mn^{2+}	1,75	1,95	1,23
Mg^{2+}	0,13	0,18	0,25
Ca^{2+}	0,67	0,41	1,03
	16,00	16,00	16,00

* Valori corectate prin eliminare normativă a incluziunilor de cuarţ, ilmenit, apatit, muscovit, paragonit/albit, clorit.

CONSTITUENŢI MOLECULARI (%)			
	2123	2165	2174
Almandin	57,5	57,7	58,2
Spessartin	29,2	32,5	20,5
Pirop	2,2		4,2
Grosular	8,9	5,8	15,4
Andradit	2,2	1,0	1,7
	100,0	100,0	100,0

Localizarea şi litologia probelor:

2123 Granat din şist sericito-cloritos cu porfiroblaste de albit şi de granat (Bazinul superior al v. Veneţia, afluent sting amonte de galeria IMR)

2165 Granat din şist sericito-cloritos cu porfiroblaste de albit şi de granat (Izvoarele văii Arinoasa)

2174 Granat din şist sericito-cloritos cu porfiroblaste de albit şi de granat (Bazinul superior al văii Arinoasa, afluent drept)

În rocile blastodetritice, feldspaţii vizibili cu ochiul liber reprezintă elemente clastice reorganizate şi sînt reprezentaţi aproape exclusiv prin albit (2-5 % An). Ca şi în plagioclazul epiclastitelor se disting frecvent

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zone periferice cu aspect limpede şi zone centrale cu grade diferite de sericitizare (fig. 4) la care se adaugă local epidot sau zoizit. Probabil şi în acest caz structura internă menţionată reflectă relicte de plagioclaz premetamorfic reorganizate şi supracrescute în timpul metamorfismului. Transformările mineralogice din zona centrală pot fi explicate, deci, fie prin reorganizarea fazelor rezultate din alterarea supergenă, fie prin substituţii în faza incipientă a metamorfismului. Avînd în vedere evoluţia transformării plagioclazilor din gnaise retromorfe (Kräutner, 1972), preferăm modelul din a doua ipoteză.

Feldspaţii metamorfici, reprezentaţi prin albit de temperatură joasă (2-8 % An), cresc în timpul metamorfismului, fie porfiroblastic, fie în benzi cu structuri echigranulare în asociaţie cu cuarţul. Evoluţia porfiroblastelor şi relaţiile faţă de cele trei elemente deformaţionale, S_1, S_2, şi S_3, permit recunoaşterea unor faze succesive de blasteză (fig. 9). Astfel se remarcă supracreşteri peste S_1 şi rotire după S_2 (pl. II, fig. 3); supracreşteri peste S_3 şi creşteri postcinematice după deformarea S_3 (pl. II, fig. 5).

6.2 Granaţi

Au fost identificate două tipuri (generaţii) de granaţi pentru care aspectele structurale sugerează o creştere sincinematică şi, respectiv, postcinematică.

Granaţii sincinematici se întîlnesc rar, în special în şisturi sericito-cloritoase-cuarţoase (în cursul superior al văii Veneţia). Apar în cristale mărunte, neidiomorfe, scheletice şi au multe incluziuni dispuse în benzi contorsionate, indicînd o rotire în timpul creşterii (pl. III, fig. 1).

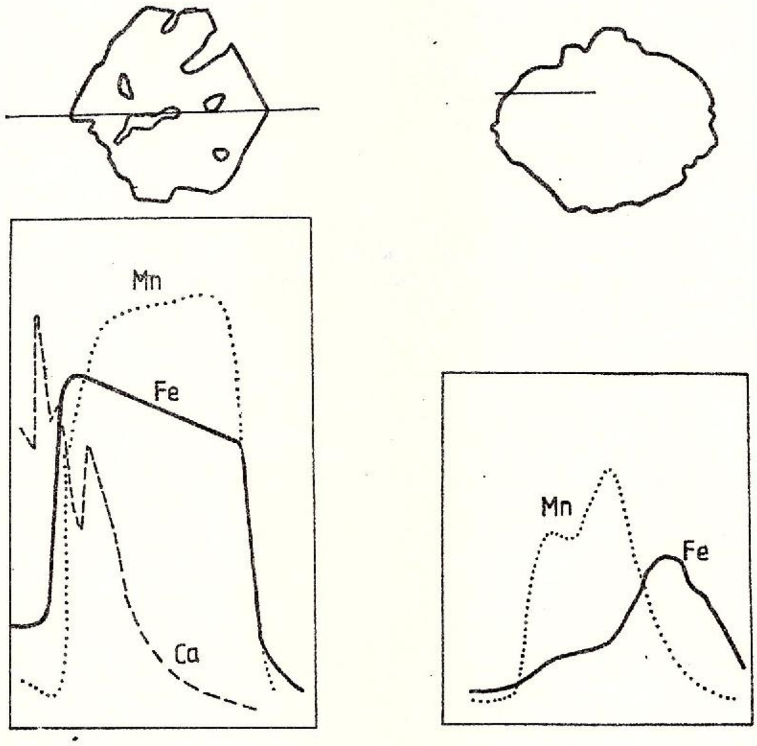

Fig. 5 – Variaţia conţinutului în Fe, Mn şi Ca în porfiroblaste de almandin spessartinic postcinematic. Profile investigate cu microsonda electronică.

Granaţii postcinematici se întîlnesc mai des, în special în şisturile cu porfiroblaste de albit şi biotit şi în cuarţitele din estul regiunii, dar nu lipsesc nici în partea centrală şi vestică. În comparaţie cu primul tip au dimensiuni mai mari, prezintă forme cristalografice bine individualizate şi au incluziuni mai puţine (în special cuarţ şi ilmenit). Observaţiile microscopice sugerează formarea prin blasteză statică (postcinematică) (pl. III, fig. 2). Uneori pe fisuri şi marginal apar cloritizări. Trei analize chimice efectuate pe probe monominerale alese sub microscopul binocular şi recalculate prin eliminarea normativă a incluziunilor de cuarţ, ilmenit, apatit,

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muscovit, albit şi clorit au indicat un almandin spessartinic cu următoarea compoziţie şi formulă moleculară (tab. 3):

2123 $Fe_{3.45}Mn_{1.75}Mg_{0.13}Ca_{0.67}$ $Al_{3.91}Fe_{0.09}$ $(SiO_4)_6$;
2165 $Fe_{3.46}Mn_{1.95}Mg_{0.18}Ca_{0.41}$ $Al_{3.96}Fe_{0.04}$ $(SiO_4)_6$;
2174 $Fe_{3.49}Mn_{1.23}Mg_{0.25}Ca_{1.03}$ $Al_{3.93}Fe_{0.07}$ $(SiO_4)_6$.

Caracterul spessartinic al granaţilor a fost confirmat şi prin analize în IR (632, 633, 634 cm^{-1}) şi Rx (a_0 = 11.606; 11.580; 11.626 Å).

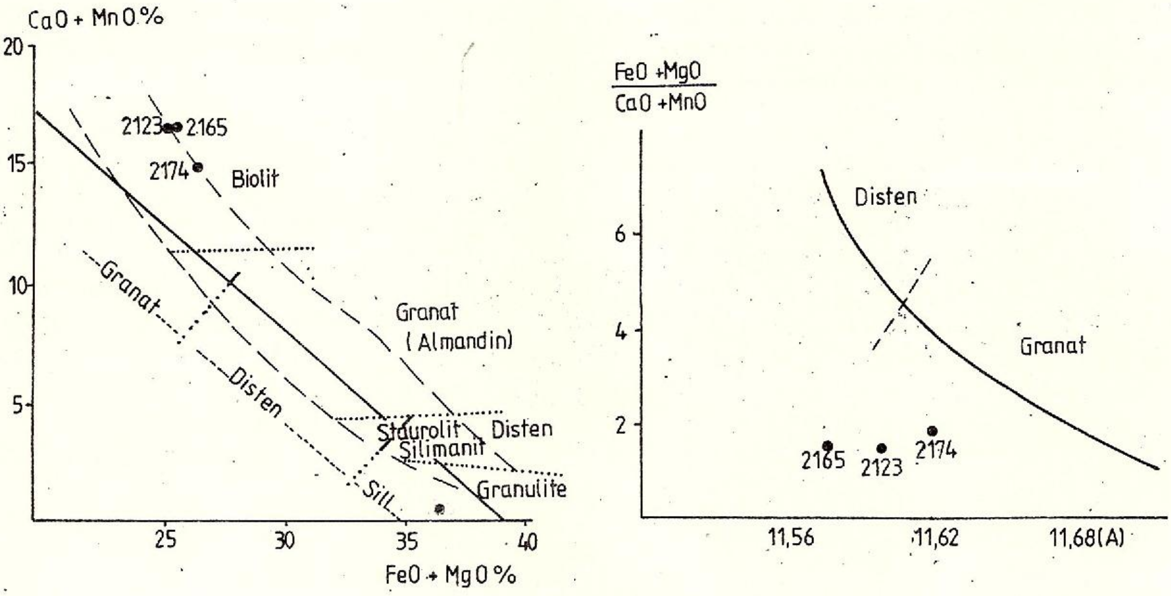

Fig. 6 – Poziţia almandinului spessartinic postcinematic din metapelitele masivului Gîrbova în trendul valorilor a_0 şi a constituţiei granaţilor din roci metapelitice în funcţie de intensitatea metamorfismului (zone cu minerale index). Cîmpul de variaţie după Sturt (1962) şi dreptele de regresie după Nandi (1967).

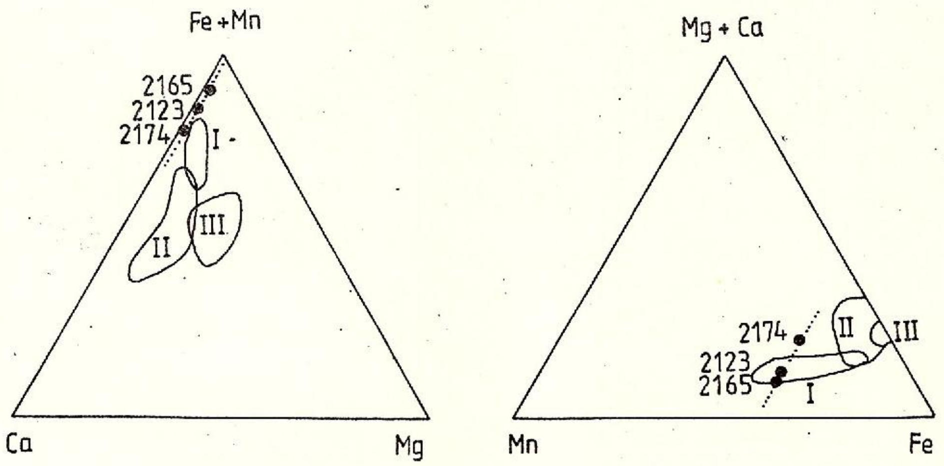

Fig. 7 – Granaţi postcinematici din metapelitele masivului Gîrbova în comparaţie cu granaţii din rocile metapelitice din regiunea Stavranger (după Müller, Schneider, 1971). Proporţii atomice reduse la 24 O.
I, zona cu parageneza Q+ab+mu+chl; II, zona cu parageneza Q+ab+ep+bt+alm; III, zona cu parageneza Di+alm+mu+Q.

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Investigări cu microsonda electronică au arătat că principalii cationi nu au o distribuţie omogenă în cadrul porfiroblastelor de almandin spessartinic (fig. 5). Astfel se remarcă atît o variaţie inversă a Fe şi Mn, cît şi o distribuţie neomogenă a acestor elemente în volumul cristalului. Este probabil ca neomogenităţile menţionate să reflecte disponibilitatea elementelor respective în timpul creşterii porfiroblastice. Acest lucru este sugerat şi de variaţia Ca, care se îmbogăţeşte spre zonele marginale care se află în contact cu minerale purtătoare de Ca (fig. 5). Constituţia moleculară a granaţilor analizaţi chimic (tab. 3) arată în acest sens o substituţie a Mn prin Ca, după cum indică şi variaţiile înregistrate cu microsonda (fig. 5).

În privinţa corelaţiei sugerată în literatură între constituţia granaţilor din rocile metapelitice şi intensitatea metamorfismului (Sturt, 1962; Nandi, 1967; Müller, Schneider, 1971), se constată că almandinul spessartinic din munţii Perşani se situează în domeniul compoziţional al granaţilor frecvent întîlniţi în roci cu grad de metamorfism scăzut, în zona biotitului şi a granatului (fig. 6, 7). Tot din acest punct de vedere, granaţii din cristalinul masivului Gîrbova se deosebesc de granaţii din rocile metapelitice ale secvenţelor cu metamorfism de grad mediu (grupurile Bretila, Rebra, Cumpăna) din Carpaţii Orientali şi din munţii Făgăraş (fig. 8, după Kasper, 1973).

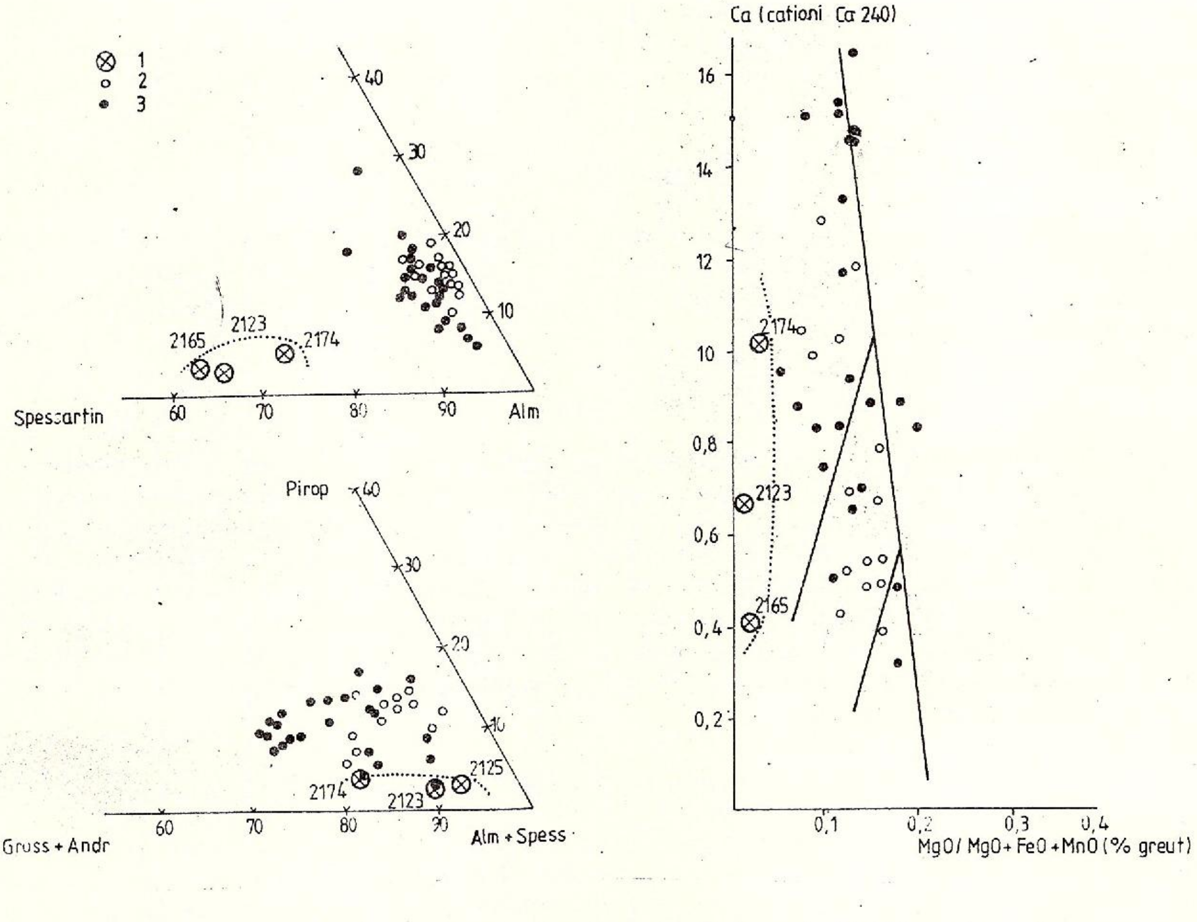

Fig. 8 – Constituţia granaţilor din metapelitele masivului Gîrbova în comparaţie cu granaţii din metapelitele de grad mediu de metamorfism din Carpaţii Orientali şi munţii Făgăraş (după Kasper, 1973).
1, almandin spessartinic postcinematic din cristalinul Gîrbovei; 2, granaţi din grupul Rebra şi grupul Bretila din Carpaţii Orientali; 3, granaţi din grupul Cumpăna din Carpaţii Meridionali.

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6.3 Biotit

Se recunosc cel puţin două generaţii de biotit în cristalinul Gîrbovei: un biotit sincinematic crescut pe foliaţia S_2 (pl. III, fig. 3) şi un biotit postcinematic crescut divergent faţă de elementele orientate din rocă (pl. III, fig. 4). Nu avem informaţii asupra constituţiei chimice a celor două generaţii de biotit.

6.4 Muscovit

Datorită omniprezenţei în toate tipurile principale de roci, muscovitul este filosilicatul a cărui evoluţie în timpul metamorfismului a putut fi cel mai bine urmărită. S-au decelat astfel patru generaţii de muscovit, corelabile cu cele patru stadii succesive din evoluţia metamorfică a rocilor: trei generaţii de muscovit sincinematic cu foliaţiile S_1, S_2 (pl. III, fig. 6) şi S_3 şi un muscovit mai nou, postcinematic, crescut porfiroblastic divergent faţă de foliaţiile amintite (pl. III, fig. 5). După valoarea $2V_x<25^0$ pentru muscovitele sincinematice, uneori verzui şi slab pleocroice şi valoarea $2V_x\sim35^0$ a muscovitului postcinematic, de regulă incolor şi fără pleocroism sesizabil, se pare că primele trei generaţii sînt în parte fengitice, în timp ce ultima generaţie este preponderent muscovitică. Această observaţie concordă cu stabilitatea preferenţială în condiţii izochimice a fengitului în condiţii de presiune mai ridicată şi a muscovitului în condiţii de presiune scăzută şi temperatură ridicată (Sassi, Scolari, 1974; Sassi et al., 1976). Prezenţa a două generaţii succesive de fengit şi muscovit în grupul Tulgheş a fost recunoscută de Kräutner et al. (1975) pe baza analizei statistice a valorilor b_o ale micelor albe potasice.

6.5 Clorit

Se pare că cloritul se comportă în mare parte similar cu muscovitul. Au fost recunoscute două generaţii sincinematice pe foliaţiile S_1 şi S_2 şi două generaţii postcinematice dintre care una probabil sincronă cu blasteza statică a muscovitului şi biotitului. O generaţie tardivă se datoreşte proceselor retrograde care au dus la cloritizarea biotitului şi granatului.

6.6 Cuarţ

Ca şi în cazul feldspaţilor, aspectele structurale permit recunoaşterea pe lîngă cuarţul metamorfic a unui cuarţ conservat relict din rocile premetamorfice. Astfel, de exemplu, se disting în metariolite fenocristale relicte de cuarţ bipiramidal cu figuri de coroziune magmatică, în metamicroconglomerate relicte de elemente clastice rotunjite (microgaleţi) şi în rocile blastodetritice claste relicte de cuarţ, uneori de culoare violacee. Sub microscop se disting frecvent stadii intermediare de sfărîmare şi forfecare ale acestor granule de cuarţ, însoţite de recristalizare şi reorganizare metamorfică. În acest sens pot fi menţionate concreşteri ale granulelor relicte de cuarţ cu cristale mărunte de cuarţ metamorfic din jur. Iau naştere, astfel, zone înguste de supracreştere metamorfică marcate frecvent prin incluziuni de cuarţ, feldspat sau muscovit, care trădează caracterul blastic al marginilor de supracreştere.

6.7 Pseudomorfozele de rutil după ilmenit

Pseudomorfoze de rutil după ilmenit, vizibile cu ochiul liber, sînt foarte frecvente în cristalinul grupului Tulgheş, în special în formaţiunea Tg_4 şi în partea superioară a formaţiunii Tg_3. Apar cu precădere în rocile de origine pelitică (şisturi şi filite sericitoase cenuşii, sericito-cloritoase) sau pelito-psamitice (şisturi sericito-cloritoase, şisturi sericito-cloritoase cu porfiroblaste de albit, biotit, granat). Descrise pentru prima dată drept "paramorfoze de rutil după brookit" de Balintoni, Chiţimuş (1973), studiul lor a fost aprofundat de Nedelcu (1986) şi Voicu et al. (1992).

Ambianţa structural-texturală a pseudomorfozelor, aspectele morfologice (plăcuţe pseudohexagonale) şi, în special, prezenţa ilmenitului relict indică provenienţa pseudomorfozelor din criastale de ilmenit crescute în timpul metamorfismului regional. Dispunerea cristalelor de ilmenit după prima şistozitate metamorfică (S_1), deformarea lor prin şistozitatea S_2 şi includerea în porfiroblaste, atestă formarea ilmenitului în primul stadiu al metamorfismului regional. Prezenţa ilmenitului în metamorfite ale faciesului şisturilor verzi, zona cu biotit a fost semnalată de Miyashiro (1973).

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Cristalele de ilmenit din masa sericito-cloritoasă a şisturilor sînt, de regulă, parţial sau total transformate într-un agregat de minerale de titan care constituie pseudomorfozele menţionate. Investigaţiile microscopice cu raze X, microsonda electronică şi cu microscopul electronic cu baleiaj au pus în evidenţă următoarele asociaţii de minerale formate prin descompunerea ilmenitului (Voicu et al., 1992): *a*) ilmenit → (± pseudorutil ?) rutil(1) ± anatas ± hematit ± hidroxizi de fier; *b*) ilmenit → rutil(1) + titanit; *c*) rutil(1) ± anatas → rutil(2).

Predomină transformarea conform relaţiei *(a)*. Titanitul (relaţia *(b)*) apare rar, probabil atît datorită lipsei calciului în mineralele asociate cît şi instabilităţii titanitului în prezenţa asociaţiei ilmenit-rutil (Force, 1976a, b). Relaţia *(c)* a fost presupusă de Balintoni, Chiţimuş (1973), Voicu et al. (1992) pe considerente morfologice; o reţea fină de ace de rutil(2) crescută în ochiurile reţelei constituită din cristale mari de rutil(1).

Formarea pseudomorfozelor a fost probabil amorsată de fluide în condiţii de creştere a fugacităţii O şi S, care determină instabilitatea ilmenitului (Force, 1976a, b; Udubaşa, 1982) şi facilitează transformarea paramorfă rutil(1)→anatas. Prezenţa uneori a calcopiritei şi bornitului în pseudomorfoză confirmă influenţa fugacităţii S.

Formarea pseudomorfozelor este greu de plasat în istoria metamorfismului, datorită lipsei unor date concludente. Se pare că descompunerea ilmenitului precede formarea porfiroblastelor de albit rotite de S_2, întrucît incluziunile de ilmenit din porfiroblastele menţionate prezintă acelaşi grad de transformare ca cele din masa şistoasă a rocii. Transformarea ilmenitului ar putea fi legată, deci, de momentul deformaţional S_2 sau intradeformaţional $S_1 - S_2$. Avînd însă în vedere faptul că momentul postcinematic după S_3 (fig. 9) reprezintă schimbarea cea mai evidentă a regimului de PT din timpul metamorfismului, nu este exclus ca transformarea ilmenitului să fi fost legată de această fază. În această ipoteză trebuie însă admis că transformarea s-a produs concomitent atît în cristalele de ilmenit din masa şistoasă a rocii, cît şi în cele incluse în porfiroblaste.

6.8 Structuri şi texturi

Rocile din cristalinul Gîrbovei provin dintr-un material variat, de provenienţă atît sedimentară, cît şi vulcanică. Datorită metamorfismului slab, unele aspecte structurale primare s-au păstrat sub formă relictă.

Structurile vulcanice relicte evidenţiază aspectul porfiric al materialului riolitic din care provin. Astfel, se disting fenocristale relicte de cuarţ bipiramidat şi cu coroziuni magmatice şi fenocristale relicte de feldspat sodopotasic, situate într-o matrice cuarţo-feldspatică microgranulară rezultată din recristalizarea pastei.

Structurile detritogene relicte indică caractere psamitice pînă la microconglomeratice, cele din urmă sesizabile prin elemente relicte de cuarţ rulat, prinse într-o masă cuarţoasă granulară săracă în filosilicaţi. Structuri blastodetritice au fost întîlnite atît în roci cuarţo-feldspatice, în care se păstrează fragmente de fenocristale de cuarţ cu figuri de coroziune, sau de feldspat sodo-potasic readaptat (concreşteri de albit+ortoză), indicînd o sursă vulcanică a materialului detritic (metaepiclastite), cît şi în roci sericito-cloritoase feldspatice provenite dintr-un material sedimentogen grezos (roci blastodetritice – metagresii feldspatice).

Structurile metamorfice indică recristalizări *sincinematice* succesive care au afectat întregul volum al rocilor şi au fost succedate de o blasteză postcinematică, în condiţii statice, în care structura rocilor a fost reorganizată numai parţial. Se pot distinge trei seturi succesive de foliaţii (S_1, S_2, S_3) dintre care numai primele două sînt însoţite de reorientare penetrativă a filosilicaţilor. Porfiroblastele de albit şi de granat conservă indicii asupra evoluţiei sincinematice polistadiale a rocilor (fig. 9).

Structurile postcinematice se datoresc creşterii neorientate a unor minerale. Este cazul cristalelor de muscovit şi de biotit larg dezvoltate, nedeformate şi crescute oblic pe foliaţiile S_1, S_2, S_3. Granatul, cu conture poligonale şi porfiroblastele de albit cresc, de asemenea, peste structurile sincinematice anterioare, fapt indicat de continuarea nederanjantă în interiorul acestor cristale a elementelor relicte din şistozităţile S_1, S_2, S_3, reprezentate prin incluziuni de cuarţ, muscovit, ilmenit. Unele porfiroblaste de albit poartă amprenta creşterii succesive în etape sincinematice şi postcinematice; nucleul sincinematic, cu incluziuni paralele, fiind înconjurat de o aureolă de supracreştere, nedeformată.

6.9 Evoluţia metamorfică

Relaţiile dintre fazele de deformare şi creşterea mineralelor sugerează o evoluţie polistadială a rocilor în timpul metamorfismului (fig. 9).

1. *Foliaţia S_1*, în general conformă cu limitele litologice, este slab conservată la scara microscopică datorită obliterării, uneori totale, de către foliaţia S_2. Este sesizabilă în special în şarnierele cutelor, pe flancuri confundîndu-se cu S_2 datorită unghiului mic dintre cele două foliaţii coaxiale. De aceea se observă doar sporadic o primă generaţie de muscovit (fengitic) clorit, cuarţ, albit şi ilmenit crescute conform cu S_1 şi deformate de S_2.

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2. *Foliaţia* S_2, foarte penetrativă în masa rocilor este însoţită de blasteza unei a doua generaţii de cuarţ, albit, muscovit (fengitic), clorit şi de o neoformaţie de biotit şi granat. Albitul se prezintă preponderent porfiroblastic. Mineralele din prima generaţie sînt în parte reorientate sau deformate. La porfiroblastele de albit şi de granat se constată, într-un stadiu incipient, creşteri cu rotiri sincinematice, urmate de desprinderea porfiroblastelor prin forfecare, datorită avansării mişcării pe S_2 şi a blastezei filosilicaţilor (fig. 9). În cazul în care într-adevăr nu s-au format granat şi biotit sincron cu foliaţia S_1, momentul deformaţional S_2 coincide cu o creştere progradă a intensităţii metamorfismului.

3. *Foliaţia* S_3 este slab penetrativă, fiind dezvoltată local sau reprezentată numai prin planele axiale ale unui sistem de kinkuri, mai mult sau mai puţin paralele. Această foliaţie deformează filosilicaţii din generaţiile anterioare şi lamelele de ilmenit, şi este însoţită local de o nouă generaţie de sericit ± clorit ± cuarţ.

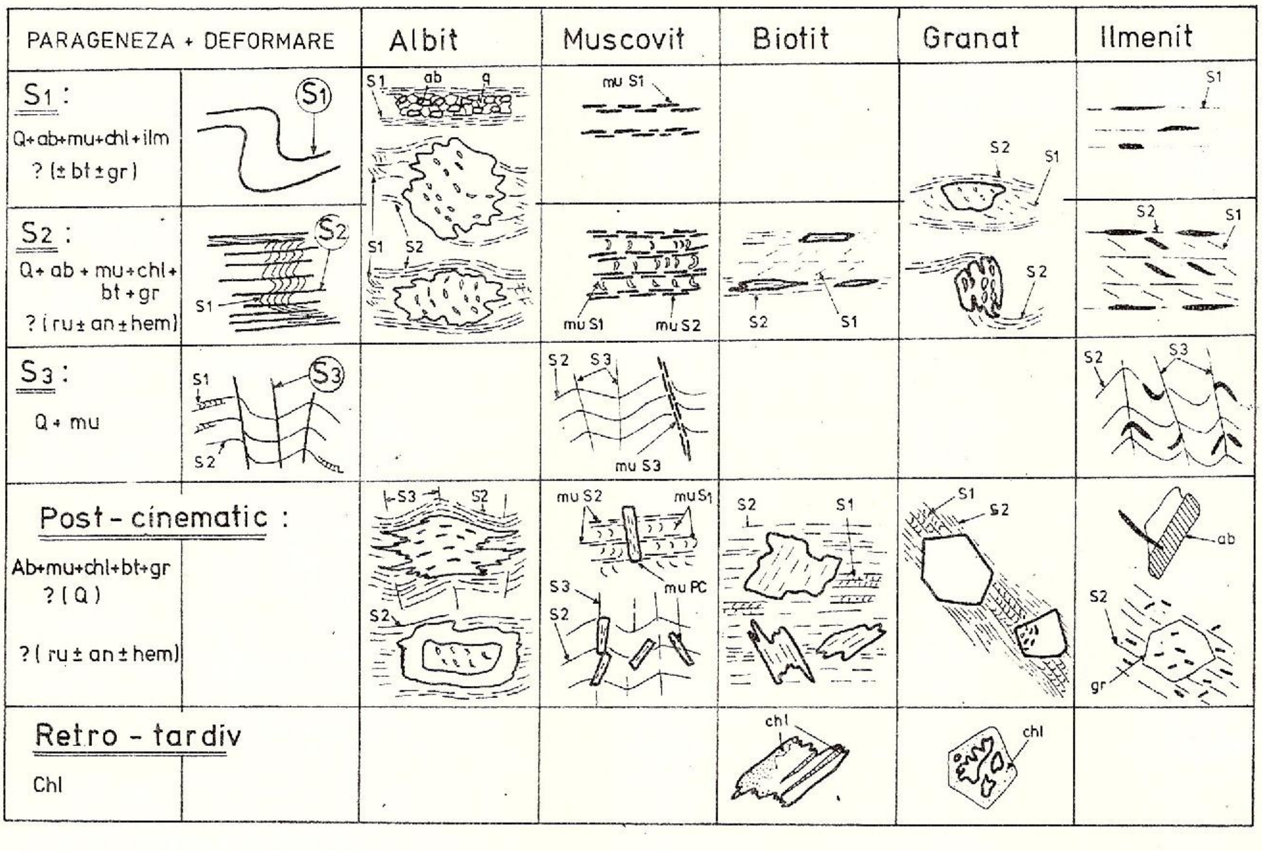

Fig. 9 – Reprezentarea schematică a relaţiilor dintre stadiile de deformare şi de creştere a mineralelor metamorfice din cristalinul de tip Tulgheş din masivul Gîrbovei.

4. Ulterior deformărilor S_1, S_2 şi cel puţin în parte S_3, apare o *blasteză în condiţii statice*, marcată prin creşterea porfiroblastică neorientată a unei generaţii de muscovit, biotit, clorit, albit şi almandin spessartinic. Spre deosebire de generaţiile anterioare muscovitul este sărac în fengit ($2V_x=35^0$). Toate mineralele menţionate cresc oblic peste foliaţiile S_1 şi S_2. Faţă de foliaţia S_3 muscovitul se dezvoltă oblic, deşi, uneori creşte local şi în lungul acestei foliaţii (fig. 9). Albitul creşte peste kinkurile S_3, care se recunosc uneori prin dispunerea incluziunilor din interiorul porfiroblastelor, fără a fi deformat de kinkurile respective. În alte cazuri au fost observate supracreşteri postcinematice de albit peste porfiroblastele de albit din stadiul sincinematic S_2. Biotitul şi granatul sînt probabil ulterioare deformaţiei S_3. Se pare, deci, că blasteza postcinematică menţionată reprezintă un stadiu prograd în evoluţia metamorfismului, deşi nu depăşeşte faciesul şisturilor verzi.

5. *O nouă reorganizare mineralogică* se manifestă prin cloritizarea locală, parţială, rareori totală, a biotitului şi granatului postcinematic. Este prima formare de minerale *cu caracter retrograd* din evoluţia metamorfitelor, sesizabilă în cadrul silicaţilor şi nu este legată de deformări sesizabile.

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6. Tot o *reorganizare mineralogică tardivă se datorește metamorfismului dinamic* din zonele de laminare care însoţesc deformările legate de punerea în loc a pînzelor alpine şi prealpine.

Deformările şi blastezele legate de S_1 şi S_2 pot fi atribuite evenimentului metamorfic caledonian timpuriu. Stadiul S_3 ar putea marca fie stingerea deformărilor caledoniene, fie deformări varistice. În funcţie de încadrarea deformării S_3, blasteza postcinematică ar putea, deci, să corespundă fie unui apex termic final al evenimentului caledonian, fie regimului termic din evenimentul varistic. O încălzire varistică este sugerată de regenerarea vîrstelor K/Ar din grupul Tulgheş (Kräutner et al., 1976). Dezvoltarea neuniformă a blastezei postcinematice, concentrarea ei preferenţială în areale uneori izolate pledează, de asemenea, în favoarea unui eveniment termic ulterior metamorfismului regional caledonian. Indiferent de încadrarea temporală a blastezei postcinematice, ea încheie o evoluţie metamorfică în general progradă, dar în limitele faciesului şisturilor verzi. Ea marchează schimbarea cea mai evidentă a condiţiilor de metamorfism în decursul istoriei metamorfice prograde. În unele privinţe aminteşte situaţii structurale întîlnite în aureola de contact a masivului Ditrău, faţă de care se deosebeşte însă prin modul de răspîndire, parageneză, şi prin unele aspecte structurale legate de evoluţia feldspaţilor şi a biotitului. Cloritizarea granatului şi biotitului postcinematic marchează un nou eveniment, retrograd, care ar putea corespunde atît unui eveniment varistic cît şi/sau unuia alpin.

Ca şi în alte zone din Carpaţii Orientali, metamorfismul grupului Tulgheş din munţii Perşani are un caracter polifazic şi totodată polimetamorfic, luînd în considerare transformările minerale care pot fi atribuite evenimetului varistic.

O problemă neclarificată la ora actuală este modul în care transformările mineralelor de titan se încadrează în istoria metamorfică menţionată. Transformarea principală ilmenit-oxizi de titan ar putea fi legată de stadiul S_2, sau de interstadiul S_1-S_2, avînd în vedere generarea ilmenitului în stadiul S_1 şi includerea pseudomorfozelor de rutil după ilmenit în porfiroblastele de albit ale stadiului S_2. Dacă admitem că aceste porfiroblaste nu au constituit o "structură armată", atunci generarea oxizilor de titan ar putea fi legată de evenimentul termic postcinematic, eventual varistic. O astfel de interpretare ar fi sprijinită de faptul că instalarea unui regim metamorfic diferit faţă de cel din evoluţia sincinematică progradă a putut implica o modificare în condiţiile de fugacitate a oxigenului, sulfului şi în regimul fluidelor intergranulare, modificare care să fi dus la instabilitatea ilmenitului şi formarea pseudomorfozelor cu oxizi de titan.

7. CRISTALINUL DE HAMARADIA

Pe flancul estic al structurii anticlinale Comana, în valea Veneţia, aflorează o stivă de şisturi sericito-cloritoase cu biotit şi granat sincinematic şi granat postcinematic. Avînd în vedere monotonia litologică a acestei secvenţe, lipsa intercalaţiilor de şisturi cu porfiroblaste de albit, este posibil ca stiva respectivă să nu aparţină grupului Tulgheş.

O limită clară faţă de asociaţii litologice tipice pentru grupul Tulgheş nu a putut fi sesizată pe teren.

Într-o astfel de interpretare se poate presupune că stiva respectivă reprezintă un cristalin în relaţie tectonică cu grupul Tulgheş, de tip Chiril, Rebra sau Rarău (Vodă, Vodă, 1985), acoperit transgresiv de acelaşi triasic care mai la nord (în bazinul Comana) se află dispus normal peste grupul Tulgheş. Ar fi deci vorba de relaţii tectonice prealpine cu un cristalin, probabil retromorf, care ulterior a fost supus împreună cu grupul Tulgheş blastezei poscinematice caledoniene sau varistice.

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BIBLIOGRAFIE

Balintoni I., Chiţimuş V. (1973) Prezenţa paramorfozei de rutil după brookit în cristalinul seriei de Tulgheş (Carpaţii Orientali). *St. cerc. geol., geofiz., geogr., Geologie*, 18, 2, p. 329–334, Bucureşti.

Becker L. P., Frank W., Höck V., Kleinschmidt G., Neubauer F., Sassi F. P., Schramm I. M. (1987) Outline of pre-Alpine metamorphic events in the Austrian Alps. Pre-Variscan and Variscan events in the Alpine–Mediteranean mountain belts. Ed. by Flügel, Sassi, Grecula, *Mineralia Slovaca Monography*, p. 69–106, Alfa Bratislava.

Bercia I., Kräutner H. G., Mureşan M. (1976) Premesozoic metamorphites of the East Carpathians. *An. Inst. Geol. Geofiz.*, L, p. 37–70, Bucureşti.

Codarcea, Dessila- M., Borcoş M., Drăgulescu Arghir A. (1969) Masivul cristalofilian al Gîrbovei. *D. S. Inst. Geol.*, LIV/3, p. 5–22, Bucureşti.

Institutul Geologic al României

– , Iliescu V. (1969) Asupra vîrstei metaflişului intracarpatic. *St. cerc. geol., geofiz., geogr., Geologie*, 14/1, p. 13–20, Bucureşti.

Dumitriu M., Dumitriu C. (1964) Contribuţii la geologia munţilor Perşani (regiunea Comana-Căciulata–Lupşa). *St. cerc. geol., geofiz., geogr., Geologie*, 9/1, p. 169–174, Bucureşti.

Force E. R. (1976a) Titanium minerals in deposits of others minerals. In Geology and resources of titanium. *Geol. Surv. Prof. Pap.*, 959 F_1, Washington.

– (1976b) Metamorphic source rocks of titanium placer deposits – a geochemical cycle. In Geology and resources of titanium. *Geol. Surv. Prof. Pap.*, 959 B_1, Washington.

Giuşcă D., Savu H., Bercia I., Kräutner H. G. (1969) Sequence of tectonomagmatic pre-alpine cycles on the territory of Romania. *Acta Geol. Acad. Sci. Hung.*, 13, p. 221–234, Budapest.

Hauer Fr., Stache G. (1863) Geologie Siebenburgens, Wien.

Ilie D. M. (1953) Structura geologică a munţilor Perşani. I Regiunea Căciulata–Lupşa–Comana–Veneţia. *An. Com. Geol.*, XXVI, p. 265–329, Bucureşti.

– (1954) Structura geologică a munţilor Perşani. II Defileul Oltului. *An. Com. Geol.*, XXVII, p. 175–258, Bucureşti.

– , Preda D. (1940) Présence des calcaires à Megalodus dans les Monts Perşani (Roumanie). *C.'R. Inst. Sci. Roum. (Acad. Sci. Roum.)*, IV/3 – 4, p. 335–337, Bucureşti.

Iliescu V., Mureşan M. (1972) Asupra prezenţei Cambrianului inferior în Carpaţii Orientali–Seria epimetamorfică de Tulgheş. *D. S. Inst. Geol.*, LVIII/4 (1971), p. 23–38, Bucureşti.

– , Kräutner H. G., Kräutner Fl., Horst H. (1983) New palynological proofs on the Cambrian age of the Tulgheş Group (East Carpathians). *An. Inst. Geol. Geofiz.*, LIX, p. 7–17, Bucureşti.

Kasper U. (1973) Studiul mineralogic şi geochimic al rocilor granatifere din cristalinul Carpaţilor Orientali şi Meridionali. Dr. Thesis, Univ. "Al. I. Cuza" Iaşi.

Kräutner H. G. (1972) Hercynische Regionalretromorphose im präkambrischen Kristallin der Ostkarpathen. *Rev. Roum. Géol., Géophys., Géogr., Géologie*, 16, 2, p. 121–129, Bucureşti.

– , Sassi F. P., Zirpoli G., Zulian T. (1975) The pressure characters of the pre-Alpine metamorphisms in the East Carpathians (Romania). *N. Jb. Miner. Abh.*, 124, 3, p. 278–296, Stuttgart.

– , Kräutner Fl., Tănăsescu A., Neacşu V. (1976) Interpretations des âges radiometriques K/Ar pour les roches metamorphiques régénerées. Un exemple – les Carpathes Orientales. *An. Inst. Geol. Geofiz.*, p. 167–229, Bucureşti.

Miyashiro A. (1973) Metamorphism and Metamorphic Belts. 491 p., London.

Mînzatu S., Lemne M., Vâjdea E., Tănăsescu A., Ioncică M., Tiepac I. (1975) Date geocronologice obţinute pentru formaţiuni cristalofiliene şi masive eruptive din România. *D. S. Inst. Geol. Geofiz.*, LXI, 5, p. 85–111, Bucureşti.

Murgeanu G., Patrulius D., Contescu L., Jipa D., Mihăilescu N., Panin N. (1963) Stratigraphie et sédimentogénèse des terraines crétacée de la partie interne de la courbure des Carpates. Asoc. Geol. Carpato–Balcanică, Congr. al V-lea, p. 31–54, Bucureşti.

Müller G., Schneider A. (1971) Chemistry and genesis of garnets in metamorphic rocks. *Contr. Min. Petrol.*, 31, 3, p. 178–200, Heidelberg.

Nandi K. (1967) Garnets as indices of progresive regional metamorphism. *Min. Mag.*, 36, p. 89–93, London.

Nedelcu L. (1986) The significance of titanium minerals in some crystalline schists from Romania. In *Mineral Paragenesis*, p. 623–646, Teophrastus Publ. S. A., Athens.

Patrulius D., Popa-Dimian E., Dimitrescu-Popescu I. (1966) Seriile mezozoice autohtone şi pînza de decolare transilvană în împrejurimile Comanei (munţii Perşani). *An. Com. Geol.*, XXXV, p. 397–432, Bucureşti.

– (1973) Le wildflysch et les olistolithes des Monts Perşani. Bull. du VIe Congrès de l'Association Géologique Carpatho–Balkanique, Vol. I, Stratigraphie, fasc. 2, p. 209–218.

Popescu I. (1970) Harta geologică a R.S.R. sc. 1:50.000, foaia Perşani.

Sassi F. P., Kräutner H. G., Zirpoli G. (1976) Recognition of the Pressure Character in Greenschist Facies Metamorphism. *Schweiz. Miner. Petr. Mitt.*, 56, p. 427–434, Zürich.

– , Scolari A. (1974) The b_0 values of the potassic white micas as a barometric indicator in lowgrade metamorphism of pelitic schists. *Contr. Miner. Petrol.*, 45, p. 143–152, Heidelberg.

Săndulescu M. (1976) La corrélation structurale du tronçon oriental avec celui méridional des Carpathes Roumains. *D. S. Inst. Geol. Geofiz.*, LXII/5 (1974–1975), p. 177–194, Bucureşti.

– (1984) Geotectonica României. Edit. tehn., Bucureşti.

Schmidt O. (1930) Cercetări geologice în ramificaţiile nord-estice ale munţilor Făgăraş. *D. S. Inst. Geol.*,XV, 1926–1937, p. 15–20, Bucureşti.

Sturt B. H. (1962) The composition of garnets from pelitic schists in relation to the grade of regional metamorphism. *Jour. Petr.*, 3, 181, Oxford.

Udubaşa G. (1982) Rutile of postmagmatic mineral formation. In Ore Genesis – The State of the Art. Pergamon Press, Oxford.

Institutul Geologic al României

Vodă Al., Vodă D. (1985) Masivul cristalin al Gîrbovei (Munţii Perşani). Noi date stratigrafice şi structurale. *St. cerc. geol., geofiz., geogr., Geologie*, 30, p. 70–75,Bucureşti.

Voicu G., Voicu C., Runceanu M. (1992) Asupra prezenţei pseudorutilului în cristalinul grupului Tulgheş din Munţii Perşani. *Rom. J. Mineralogy*,75, p. 101–106, Bucureşti.

Wachner H. (1918) Bericht über die im Sommer 1916 im Persányer Gebirge ausgeführten geologischen Aufnahmen. *Jahresbericht der königlich ungarischen Geologischen Reichsanstalt für 1916*, p. 259–284, Budapest.

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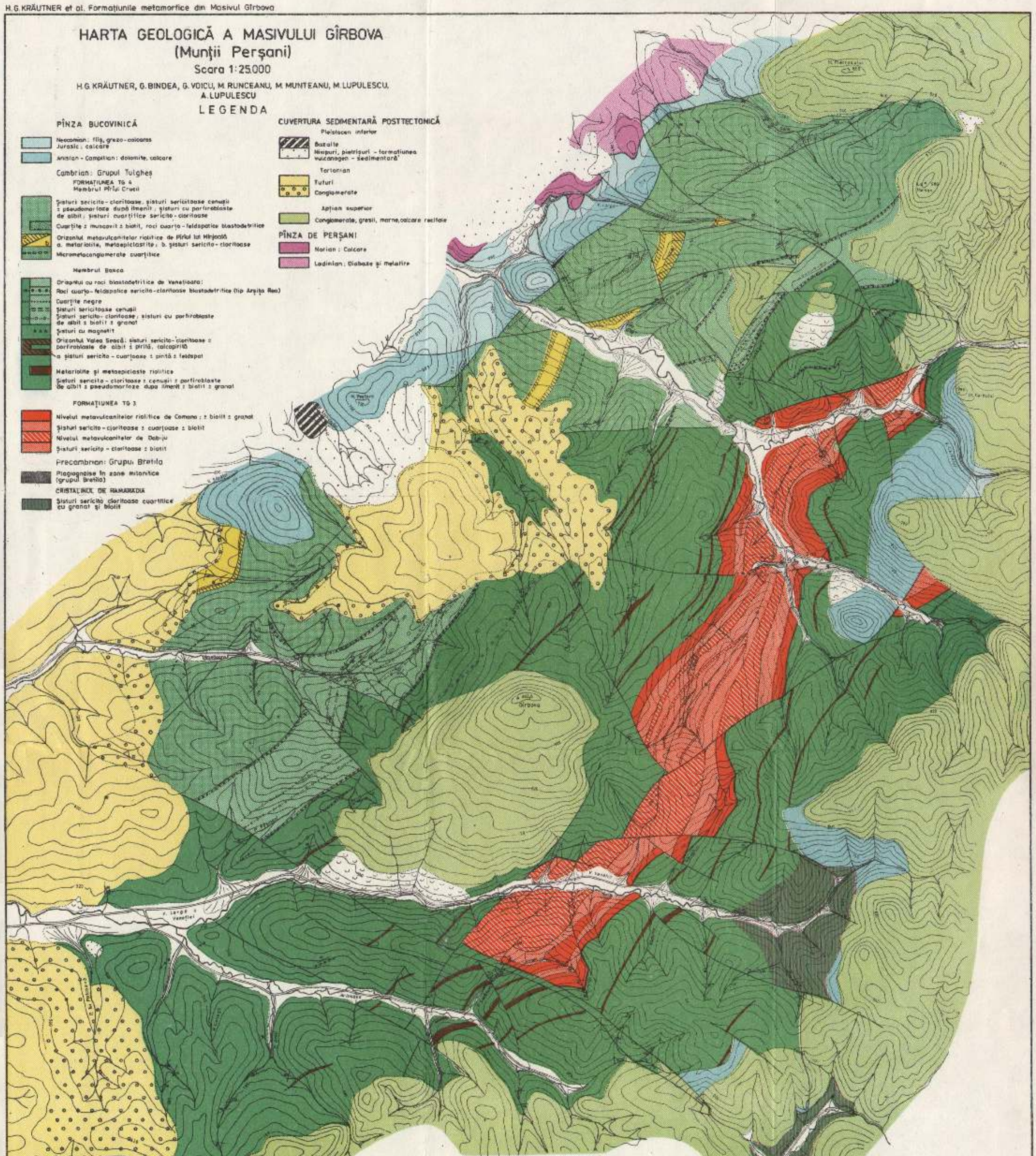

HARTA GEOLOGICĂ A MASIVULUI GÎRBOVA
(Munţii Perşani)

Scara 1:25.000

H.G.KRÄUTNER, G.BINDEA, G.VOICU, M.RUNCEANU, M.MUNTEANU, M.LUPULESCU,
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LEGENDA

PÎNZA BUCOVINICĂ

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Jurasic: calcare
Anisian - Campilian: dolomite, calcare

Cambrian: Grupul Tulgheş

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Membrul Pîrîul Crucii

Şisturi sericito-cloritoase, şisturi sericitoase cenuşii ± pseudomorfoze după ilmenit; şisturi cu porfiroblaste de albit; şisturi cuarţitice sericito-cloritoase
Cuarţite ± muscovit ± biotit, roci cuarţo-feldspatice blastodetritice
Orizontul metavulcanitelor riolitice de Pîrîul lui Mihjoacă
a. metariolite, metaepiclastite; b. şisturi sericito-cloritoase
Micrometaconglomerate cuarţitice

Membrul Barca

Orizontul cu roci blastodetritice de Venetioara:
Roci cuarţo-feldspatice sericito-cloritoase blastodetritice (tip Arşiţa Rea)
Cuarţite negre
Şisturi sericitoase cenuşii
Şisturi sericito-cloritoase; şisturi cu porfiroblaste de albit ± biotit ± granat
Şisturi cu magnetit
Orizontul Valea Seacă: şisturi sericito-cloritoase ± porfiroblaste de albit ± pirită, calcopirită
a şisturi sericito - cuarţoase ± pirită ± feldspat
Metariolite şi metaepiclaste riolitice
Şisturi sericito - cloritoase ± cenuşii ± porfiroblaste de albit ± pseudomorfoze după ilmenit ± biotit ± granat

FORMAŢIUNEA TG 3

Nivelul metavulcanitelor riolitice de Comana: ± biotit ± granat
Şisturi sericito - cloritoase ± cuarţoase ± biotit
Nivelul metavulcanitelor de Dobiju
Şisturi sericito - cloritoase ± biotit

Precambrian: Grupul Bretila

Plagiogneise în zone milonitice (grupul Bretila)

CRISTALINUL DE HAMARADIA

Şisturi sericito cloritoase cuarţitice cu granat şi biotit

CUVERTURA SEDIMENTARĂ POSTTECTONICĂ

Pleistocen inferior

Bazalte
Nisipuri, pietrişuri - formaţiunea vulcanogen - sedimentară

Tortonian

Tufuri
Conglomerate

Aptian superior

Conglomerate, gresii, marne, calcare recifale

PÎNZA DE PERŞANI

Norian: Calcare
Ladinian: Diabaze şi melafire

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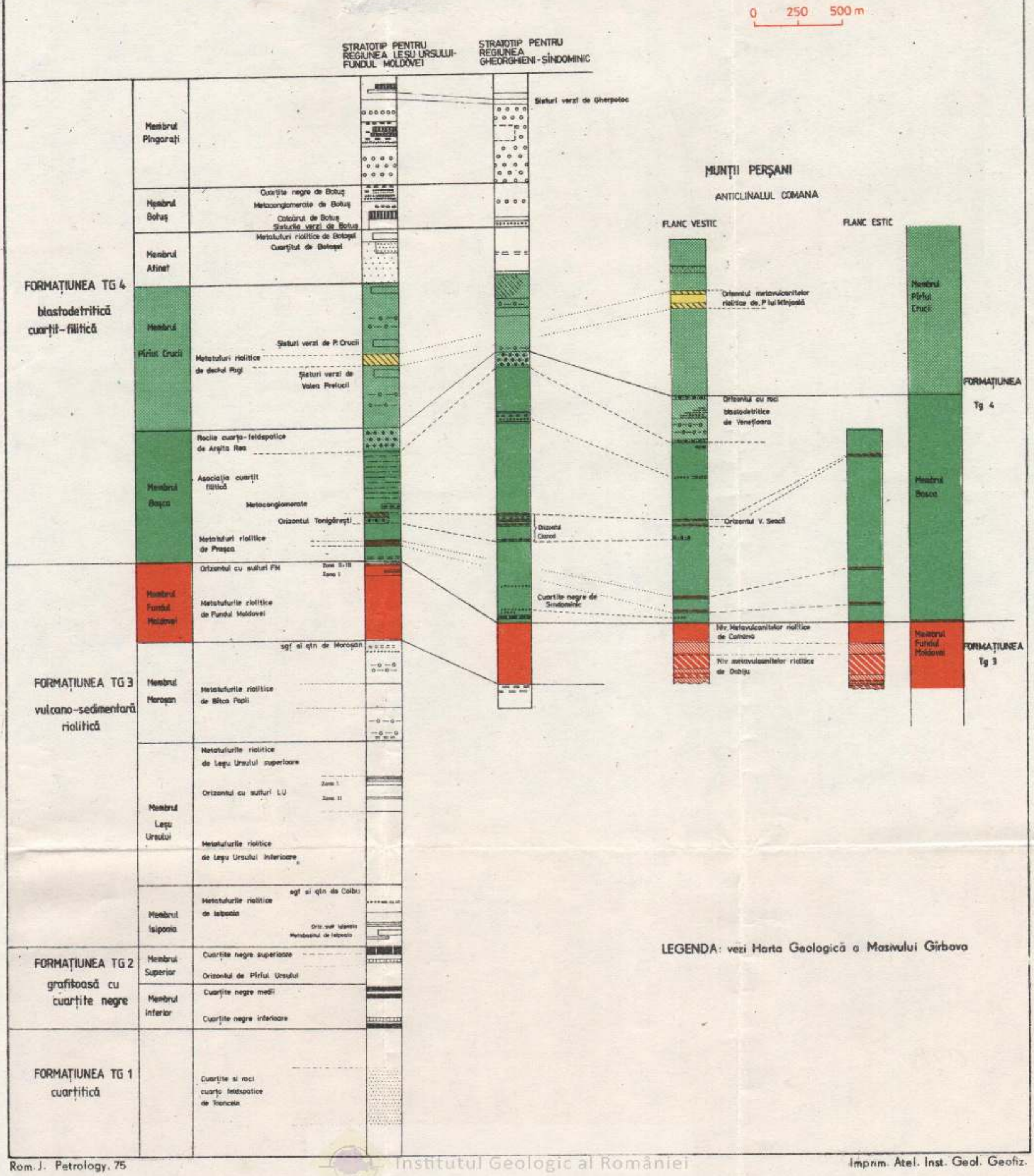
Imprim. Atel. Inst. Geol. Geofiz.

CORELAREA LITOSTRATIGRAFICĂ A GRUPULUI TULGHEŞ DIN PÎNZA BUCOVINICĂ ÎN CARPAŢII ORIENTALI CENTRALI ŞI MUNŢII PERŞANI



LEGENDA: vezi Harta Geologică a Masivului Gîrbova

Rom. J. Petrology, 75

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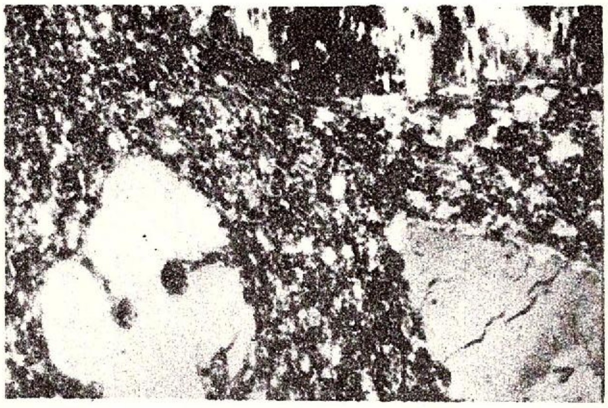

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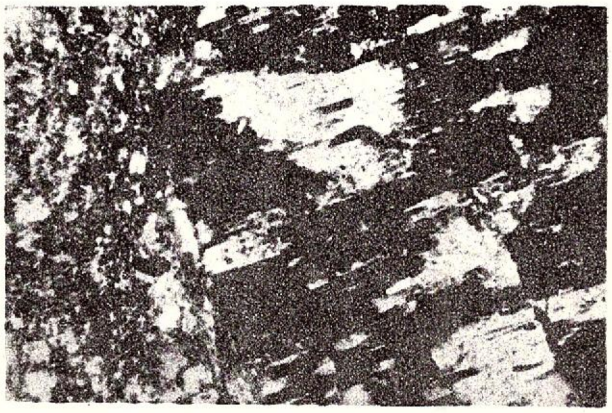

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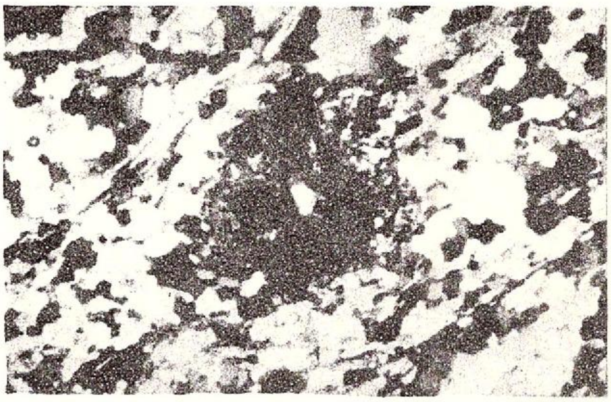

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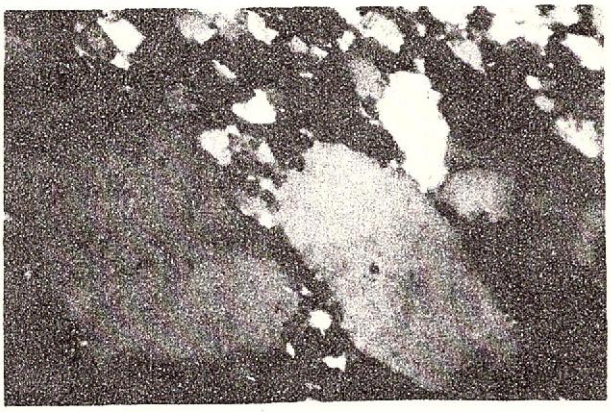

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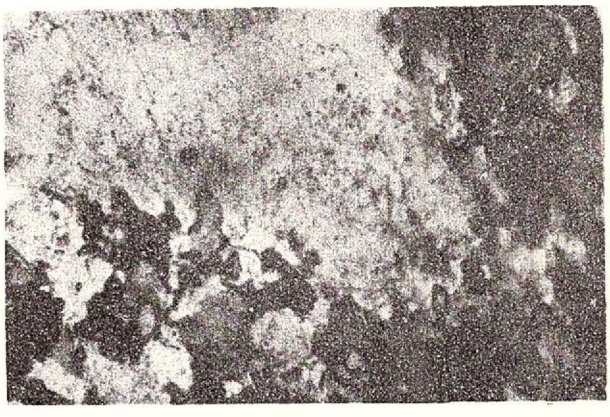

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Planşa I

Fig. 1 – Metariolit, structură porfirică relictă: fenocristale relicte de cuarţ bipiramidat cu coroziune magmatică şi de feldspat sodo-potasic de temperatură înaltă descopmus în ortoclaz+albit.

Fig. 2 – Metariolit: fenocristal relict de feldspat sodo-potasic descompus în ortocalz+albit.

Fig. 3 – Metariolit: fenocristal relict de feldspat şi pastă parţial recristalizată.

Fig. 4 – Metaepiclastit riolitic, structură blastodetritică: clast relict de feldspat sodo-potasic descompus în ortoclaz+albit.

Fig. 5 – Microconglomerat: Galeţi rotunjiţi de cuarţ relict într-o masă cuarţoasă microgranulară.

Fig. 6 – Idem: Detaliu pentru modul de concreştere a galeţilor de cuarţ relict cu matricea cuarţoasă recristalizată.

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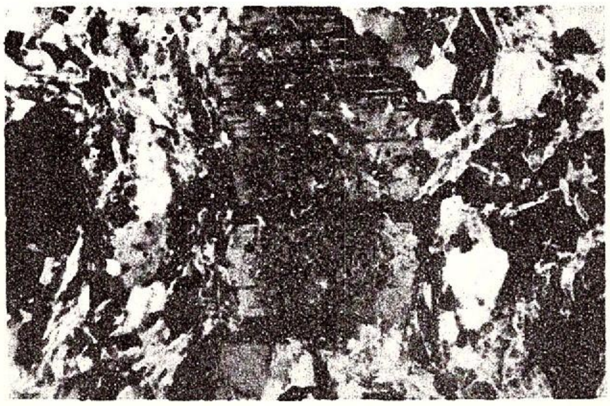

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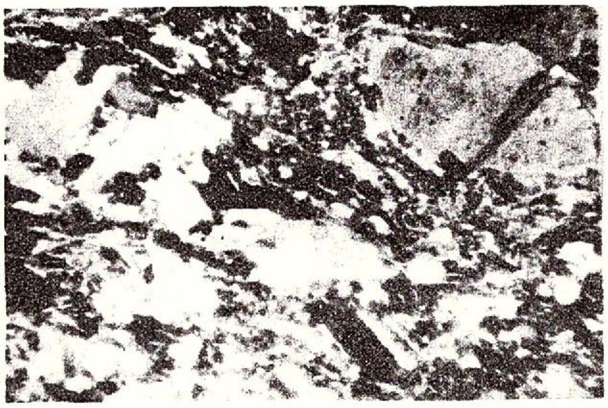

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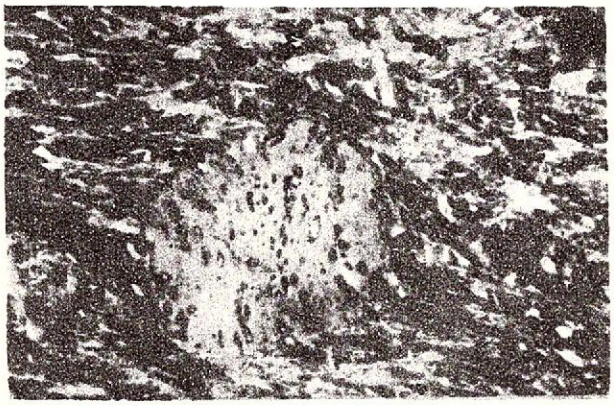

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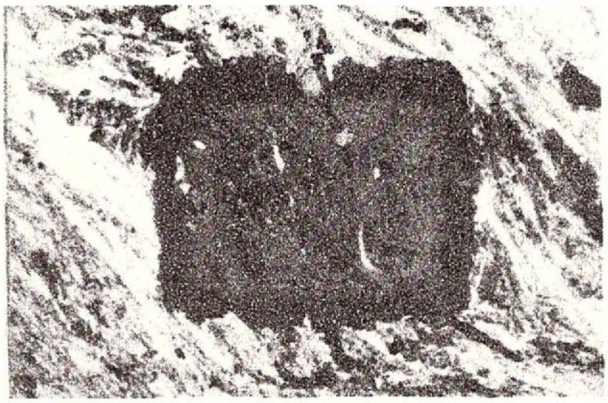

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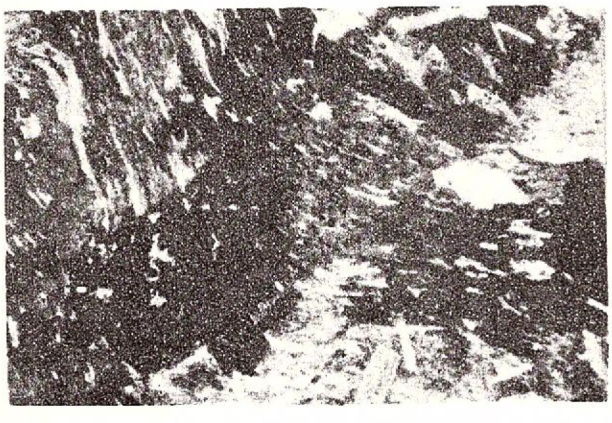

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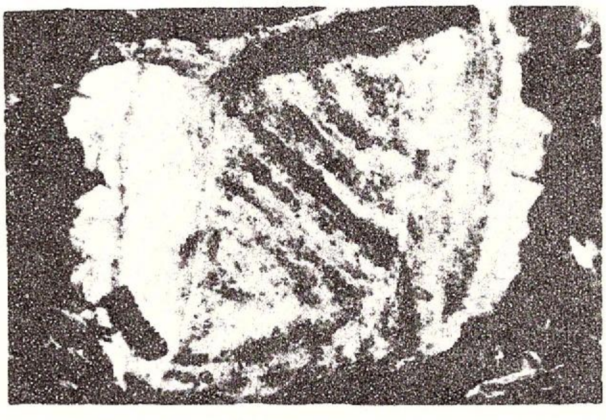

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Planşa II

Fig. 1 – Rocă blastodetritică cuarţo-feldspatică de tip Arşiţa Rea (metagresie feldspatică): clast relict de plagioclaz reorganizat în albit (2–5 % An) cu zonă centrală sericitizată şi supracreştere metamorfică.

Fig. 2 – Idem, inclusiv granul relict de cuarţ parţial sfărîmat şi vindecat prin cuarţ metamorfic.

Fig. 3 – Porfiroblast de albit rotit faţă de S_2.

Fig. 4 – Idem cu zonă de supracreştere mai săracă în anortit.

Fig. 5 – Porfiroblast de albit crescut peste kinkuri S_3.

Fig. 6 – Porfiroblast de albit în filit grafitos. Se remarcă creşteri succesive peste S_1 (cristal rotit), peste S_2 şi coroană de supracreştere probabil statică.

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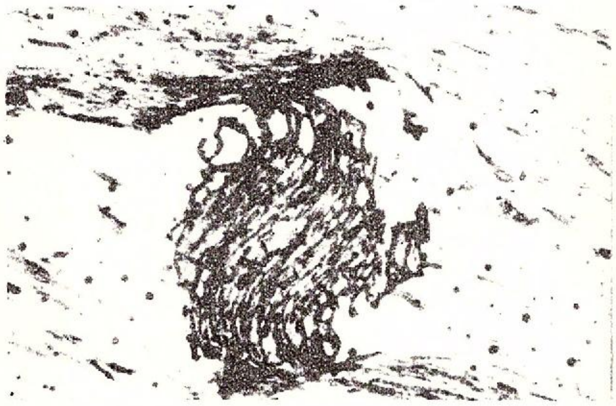

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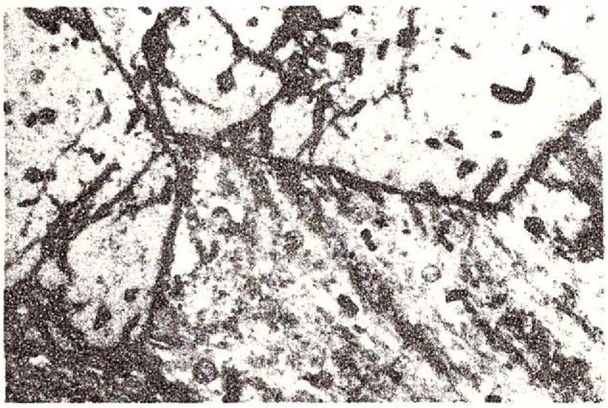

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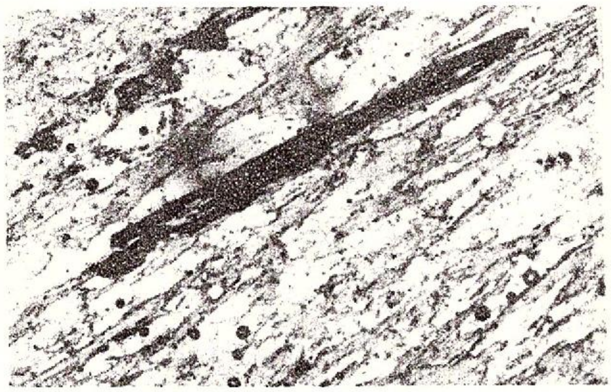

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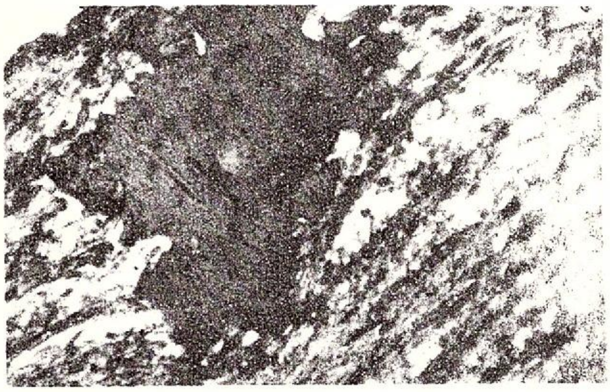

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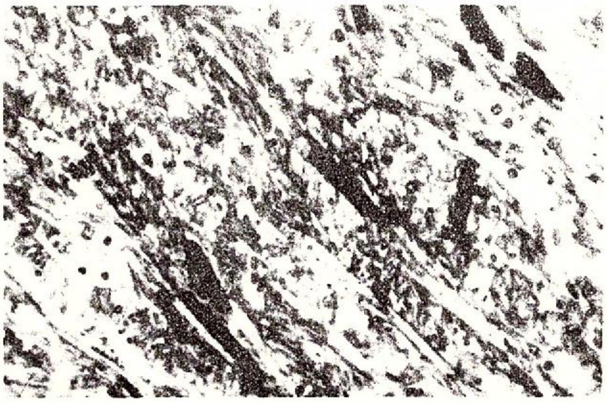

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Planşa III

Fig. 1 – Granat sincinematic rotit faţă de S_2 (incluziuni de cuarţ).
Fig. 2 – Granat postcinematic crescut peste S_1 şi S_2 (incluziuni de ilmenit).
Fig. 3 – Biotit sincinematic crescut pe S_2.
Fig. 4 – Biotit postcinematic crescut peste S_1 şi S_2.
Fig. 5 – Muscovit postcinematic crescut neorientat peste S_2.
Fig. 6 – O nouăgeneraţie de sericit-fengitic şi clorit pe S_2 crecută peste asociaţia mai mărunt cristalizată de sericit-fengitic, clorit, cuarţ şi albit conformă cu S_1. In negru ilmenit şi pseudomorfoze de rutil după ilmenit orientate după S_1 şi reorientate după S_2.

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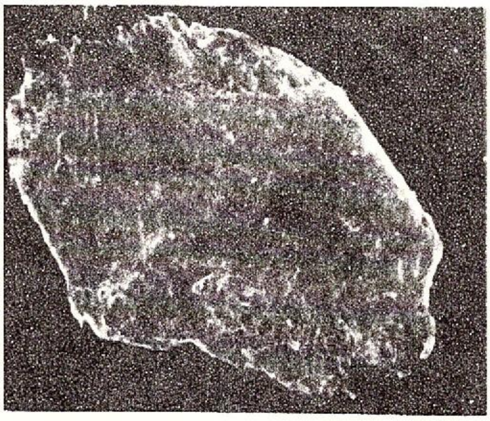
2

3



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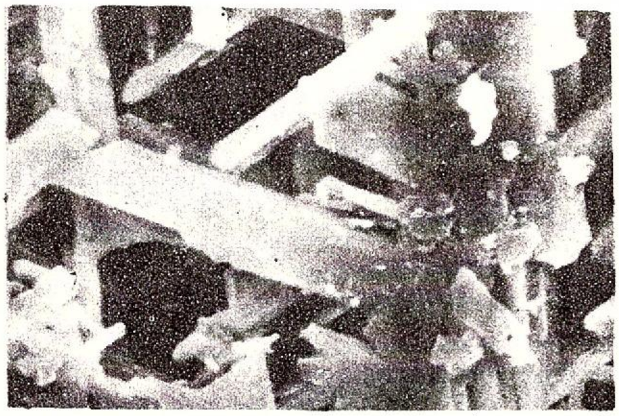

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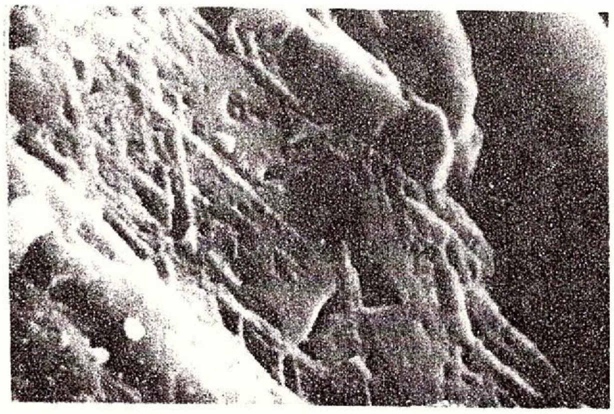

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Planşa IV

Fig. 1 – S_1 şi S_2 deformate de kinkuri S_3. O nouă generaţie de muscovit porfiroblastic crescută oblic (postcinematic) peste fengitul şi cloritul sincinematic faţă de S_1 şi S_2.

Fig. 2 – Deformarea sericitului, cloritului, ilmenitului şi pseudomorfozelor de rutil prin kinkurile S_3; regenerarea parţială a filosilicaţilor pe foliaţia S_3.

Fig. 3 – Aglomerarea de cristale de albit postcinematic, crescute neorientat. Incluziunile de cuarţ şi de pseudomorfoze de rutil după ilmenit păstrează relict orientarea structurii S_2.

Fig. 4 – Pseudomorfoză de rutil după ilmenit – faţa (0001). Microscop electronic, x 50.

Fig. 5 – Reţea de rutil (dintr-o pseudomorfoză după ilmenit), orientată după 3 direcţii cristalografice. Microscop electronic, x 6600.

Fig. 6 – Reţea de rutil 1 în interstiţiile căreia apare rutilul 2. Microscop electronic, x 12000.

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Rom. J. Petrology, 1992, 75, p. 183–188

CONSIDERATIONS UPON THE LITHOSTRATIGRAPHY AND METAMORPHISM OF THE FĂGĂRAŞ SERIES

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Key words: Metamorphic rocks. Stratigraphic units. Paragenesis. Silicates. Sillimanite. Polymetamorphism. Retrograde metamorphism. South Carpathians – Getic and Supragetic Crystalline Domain – Făgăraş Mountains.

Abstract: The Făgăraş series includes two formations of very different lithologies: the Şerbota Formation, made up of micaschists and paragneisses, and the Suru Formation, mostly consisting of amphibolites and crystalline limestones. The relationships among them are tectonic. On the background of a lithological homogeneity, the various petrographical aspects generated by retromorphism cannot sustain a lithostratigraphical subdividing of the two formations. The Făgăraş series metapelites contain five mineral paragenesis, resulted from four main metamorphic events. The metamorphic evolution is the same for the two Făgăraş series formations.

Introduction

The year 1983 marks the beginning of a new stage in the geological research of the Făgăraş Mts. The papers written during this period (Balintoni et al., 1984, 1985, unpubl. reports; Balintoni et al., 1986; Gridan, Dumitraşcu, 1990; Pană, Ricman, 1989; Gheuca et al., 1987, unpubl. report; Dimitrescu, 1987; Pană, 1990; Gheuca, 1988) stress out a quick evolution of the ideas concerning the geological knowledge about this mountain.

As the mentioned papers tackle mainly problems of structure and lithostratigraphy, we have considered it necessary to present some data about the metamorphism of the Făgăraş series. In order to define the action field of the metamorphism, we also deal here with certain aspects linked to the lithostratigraphy of the crystalline schists in this series.

The paper is based on many field and microscopic studies carried out on the entire outcropping area of the Făgăraş series. The most useful data come from the western part of the northern flank, an area for which we present a geological map achieved on the basis of the mapping carried out in 1984 and 1986.

1. Lithostratigraphy of the Făgăraş series

The Făgăraş series means the pile of metamorphic rocks included by Balintoni et al. (1984, unpubl. report; 1986) and Gheuca et al. (1987, unpubl. report) in the Făgăraş group.

Excepting the upper part, where the lithological association of amphibolites – crystalline limestones is widely developed, the Făgăraş series presents an unitary petrographic character, made up of paragneisses and micashists unhomogeneously retrogressed, regionally and dynamically. The lithological bodies, contrastive in comparison with this petrographic character (crystalline limestones, feldspathic paragneisses ± augen structures, amphibolites) are lenticularly developed and are disposed at all the levels of the series, therefore they cannot be used for dividing it lithostratigraphically unless one accepts a great number of subdivisions.

If the upper part of the series clearly stands out by the different composition of the premetamorphic material, the further subdivision of the Făgăraş series can only be done on the basis of the partially overlapped effects of the regional and dynamic retromorphism. Putting together the effects of the two retrogressions in the northern half of the Transylvanian slope and in the crest area gives to the sequences in these areas specific petrographic characters, easily recognizable in the field. But, besides, being recurrent, their intensity is graduated, and, therefore, it does not offer clear-cut limits that can be surely mapped. Even if we accepted conventional limits, these could not have lithostratigraphical significance, anyway. Besides, the limits established by retrogression

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are newer than the ones given by the difference in composition of the premetamorphic material and we consider that a correct parting of a metamorphite pile, irrespective of criteria, should use limits of the same age.

Within the Făgăraş series two formations of very different lithologies can be separated:

- the Suru Formation (upper) and
- the Şerbota Formation (lower).

This parting was recognized in the north-western part of the Făgăraş Mts and temporarily used by Balintoni et al. (1985, unpubl. report) and by Pană, Ricman (1989), with the difference that the mentioned authors included in the Suru Formation also the northernmost sequence west of the Avrigului Valley, strongly retrogressed and containing crystalline limestones, unassociated with amphibolites. Afterwards, this concept was modified (the northern sequence, bearing limestones unassociated with amphibolites was excluded from the Suru Formation) and detailed by Gheuca et al. (1987, unpubl. report) and Pană (1990). The detailings were carried out in the sequence named by us the Şerbota Formation, especially on the basis of retrogressive aspects. In the north of Făgăraş, the last retrogression is Alpine and any boundary always attributed to it should cross the boundary between the Suru and Şerbota Formations.

1.1. The Suru Formation

It is characterized by the presence of amphibolites – crystalline limestones lithological association, widely developed, over paragneisses and retromorphic quartzose paragneisses. The Suru Formation overlies the Şerbota Formation, the contact plane is tectonic, outcropping as tightly welded black ultramylonites. Its age could be pre-Alpine, as the Alpine laminations and shearing planes cross it. Examining the plane configuration in detail, Pană (1990) also admits a pre-Alpine age.

In the Lupu Valley basin, among the Suru Formation rocks, there also occur augen gneisses bearing potassic feldspar. Gridan, Dumitraşcu (1990) have presented this sequence, as an outlier of Cumpăna type crystalline. With this exception, all the researchers who work at present in the Făgăraş Mts agree with the unitary nature of this formation.

Eastwards, the Suru Formation retreats at the crest and tends to thin out. Isolately, the Suru type lithological associations occur still on the northern slope. Such an association (crystalline limestones, green schists, quartzose paragneisses, carbonate schists, veins of basic rocks), which is strongly tectonised, was found more eastwards, on the right slope of the Brescioara Valley.

1.2. The Şerbota Formation

It is mostly made up of paragneisses and micaschists bearing lenticular intercalations of crystalline limestones, feldspathic paragneisses ± augen structures and, very rarely, amphibolites. The Şerbota Formation limestones are not associated with amphibolites. Feldspathic paragneisses frequently have augen structures resulted from the porphyroblastic remobilization and recrystallization of plagioclase feldspar, relatively rich in these rocks. The spaced ocelli are subcentimetric and are made up of 15–20 % An oligoclase.

In the north-west of Făgăraş, the Şerbota Formation occurs in an antiform structure. Crossing this structure, but unrelated to its axis, we notice the following zonality of retrograde effects:

- unretrograded rocks (staurolite- and kyanite-bearing phaneroblastic micaschists and paragneisses);
- only regionally retrograded rocks (garnet-bearing chlorite micaschists, muscovite- and chlorite-bearing mica paragneisses; feldspathic paragneisses ± augen structures);
- both regionally and dynamically retrograded rocks (chlorite schists, sericite-chlorite schists, quartz-muscovite schists, quartz-feldspar schists, mylonites) bearing garnet and biotite relics.

The outcropping area of the Şerbota Formation widens eastwards, comprising the whole northern slope. West of Valea Caselor, it is perianticlinally covered by the Suru Formation.

2. Metamorphism of the Făgăraş Series

The microscopic examination of the Făgăraş series metamorphites and above all, of the Şerbota Formation ones, allowed the discovery of five parageneses, corresponding to four metamorphic events (M_1, M_2, M_3, M_4), each of them being retrograde as against the previous event. Against the background of these succesive retrograde transformations, the M_2 metamorphism has a complex evolution, and was first retrograde as against the M_1, and finally prograde.

Although we did not carry out detailed microscopic examinations on the Suru Formation, the observations we possess point out that the two Făgăraş series Formations had a common metamorphic evolution.

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2.1 The M_1 metamorphism

The oldest paragenesis, corresponding to the initial metamorphism, M_1, has given biotite, garnet, staurolite and kyanite; the paragenesis also contained plagioclase and quartz, minerals renewed by latter remobilization and recrystallization. It is well preserved in the north-west of Făgăraş, in the phaneroblastic micaschists and paragneisses on the southern flank of the antiform. This sequence reaches the main crest, in the Şerbota–Negoi peaks region. In the retrogressed rocks only garnet and biotite are preserved. On the Transfăgărăşan Road, at the southern mouth of the tunnel under the crest, Giuşcă et al. (1977), then Pană (1990) had found kyanite in phyllite micaschists belonging to the retrogressed Şerbota Formation. In the north-east of the massif the paragenesis is complete in the Şinca Nouă micaschists. In the Suru Formation, besides biotite and garnet, also microblastic staurolite occurs in micaschists (Pană, Ricman, 1989).

Minerals of the first paragenesis are coarse-grained. They are displaced and partially transposed on the S_2 foliation plane, by gliding and rolling. Nothing points to the forming of these minerals against other preexistent ones. They prove a medium temperature and pressure initial metamorphism, within staurolite and kyanite zone. The M_1 metamorphism affected the entire Făgăraş series and its effects are obliterated almost completely by the following metamorphic events.

The initial paragenesis represents the starting point of the succesive development of the other paragenesis (Fig.).

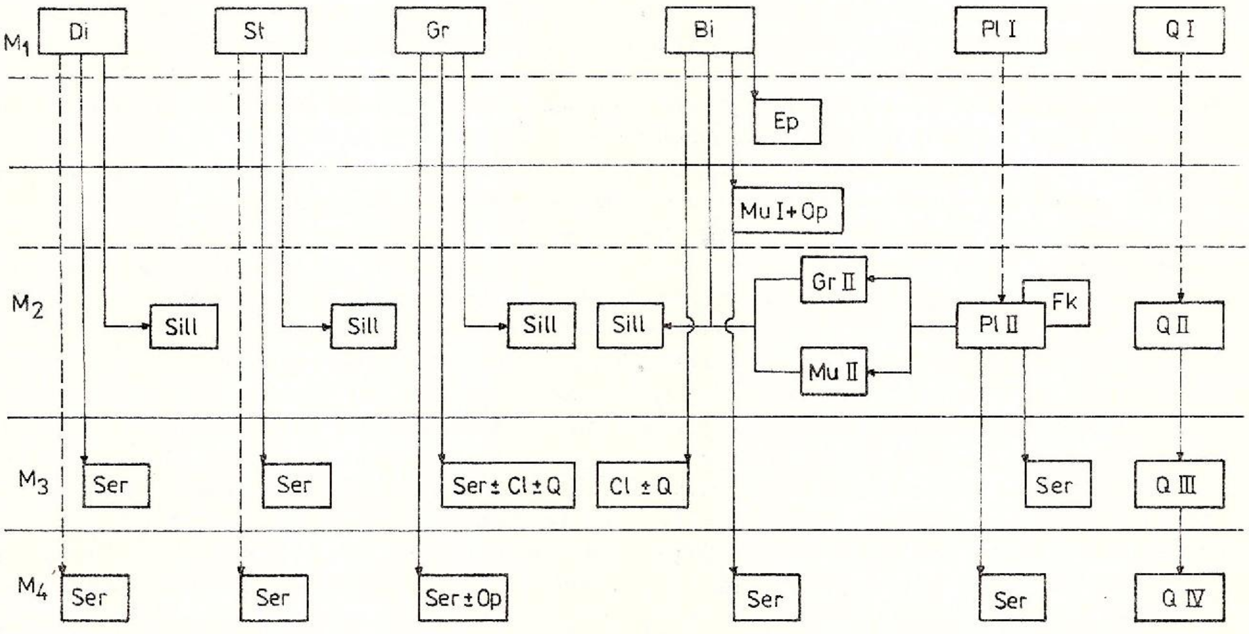

Fig. Succesion of metamorphic parageneses in the Făgăraş Series.
Di, kyanite; St, staurolite; Gr, garnet; Bi, biotite; Pl, plagioclase; Q, quartz; ser, sericite; ep, epidote; mu, muscovite; sill, fibrolitic sillimanite; cl, chlorite; op, opaque minerals.

2.2 The M_2 metamorphism

The M_2 metamorphism corresponds to two parageneses, a synkinematic one as against the S_2, which is retrograde (A paragenesis), and a postkynematic one, resulted from prograde transformations (B paragenesis).

A Paragenesis. It is made up of phengitic muscovite, bearing lamellar or pulverulent fine opaque minerals, disposed over the cleavage. Both minerals result from the retrograde transformation of biotite, a phenomenon regionally developing, with a constant intensity. The fusiform muscovite beds contain fresh biotite, with a reddish-colourless banded alternation, which gives these beds a banded microscopic aspect; the biotite/muscovite limit is sometimes more complicatedly drawn, crossing the cleavage planes. Thus formed muscovite has the 2V angle between 15^0–30^0 and pale green, pink or yellow shades. This paragenesis is synchronous to the S_2 regional foliation, that it renders real.

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B Paragenesis. It contains oligoclase, quartz, fibrolitic sillimanite, reaction garnet and muscovite. Excepting oligoclase and quartz, is a discreet paragenesis, that can be seen only under the microscope. In no thin section could it be seen in its completeness. Without garnet, it is present everywhere in the staurolite- and kyanite-bearing unretrograded sequence. Without reaction sillimanite and muscovite, but with reaction garnet it is present in the central area of the northern flank. In the north-eastern Făgăraş Mts, it contains both reaction sillimanite and garnet, but reaction muscovite was not noticed. Sporadically, fibrolitic sillimanite also occurs in the Suru Formation (Balintoni et al., 1984, unpubl. report). Plagioclase and quartz are remobilized at the M_2 moment in the whole Făgăraş series.

Plagioclase displays a constant composition (oligoclase 15–20 % An) and together with quartz it is postkynematically recrystallized. The plagioclase tendency to recrystallize porphyroblastically is frequently noticed under the microscope in the Făgăraş series rocks, but only in feldspathic paragneisses sequences this tendency occurs macroscopically, by the appearance of augen textures. The textures as such are actually secondary, being generated by post-M_2 lamination movements. The post-M_2 metamorphic evolution determines the changing of the physiographic type of porphyroblasts, but not the chemical composition of plagioclase as well. The initial physiographic characters are visible in unretrogressed rocks, where porphyroblasts are undeformed. They are xenomorphic, sometimes very sinuous and contain biotite, apatite, phengitic muscovite, quartz, reaction garnet and muscovite. Very rarely, antipertites were noticed.

Between plagioclase and biotite there are reactions resulting in garnet and muscovite. Idiomorphic and microblastic garnet is always included in plagioclase, sometimes in biotite contacting it directly. Reaction muscovite occurs on the biotite included in recrystallized plagioclase, either as a thin margin or as fine beaches.

Fibrolitic sillimanite revealing an important thermic effect develops intergranularly or included in the recrystallized quartz-feldspar phase, preferably within quartz. It occurs as isolated needles (0.03–0.25 mm) or grouped in nests, rows or bunches (0.58–2.20 mm). In many thin sections it grows directly over biotite, staurolite, garnet, and sometimes over kyanite. Fibrolitic sillimanite directly related to kyanite occurs also in the north-eastern Făgăraş Mts, in the Şinca Nouă micaschists.

The close succession of these two parageneses reflects an interesting evolution of the M_2 metamorphic event. The gliding movements on the S_2 planes induces the retrograde transformation of biotite into muscovite, a widely spread reaction in the Făgăraş series, as well as in other crystalline series. After the main movement moment, the temperature of the system grows and the metamorphic reactions become prograding. The prograding changes start by the thermic remobilization of the quartz-feldspar component, that locally reacts with biotite resulting garnet and reaction muscovite and ends by the intergranular blastesis of fibrolitic sillimanite. The direct transformation of kyanite into sillimanite shows that at the end of the M_2 metamorphism the temperature was higher than the one which governed the M_1 metamorphic transformations.

Şeclăman, Hârtopanu (1987) proves by thermodynamical calculations that in retrograding systems the temperature increases gradually, until it stops retrograding and initiates prograding reactions. The temperature increase is determined by the friction heat generated by lamination planes (S planes) and by the heat from exothermic retrograding reactions. The self-heating of the retrograding system is possible when the speed of producing heat is higher than the speed of losing it, under the circumstances of relatively quick lamination and retrograde reactions. In our opinion, these circumstances were achieved during the M_2 metamorphism. The fact that fibrolitic sillimanite-bearing paragenesis is discreet gives us reasons to suppose that, although important, the temperature increase was short-lived in the case of the Făgăraş series.

2.3 M_3 metamorphism

It is a regional retrograde metamorphism, sustained by the chlorite, quartz, sericite paragenesis, widespread in the Făgăraş series. Chlorite substitutes garnet and biotite not turned into muscovite. Quartz recrystallizes and partially substitutes garnet, biotite and, sometimes, muscovite. Sericite pseudomorphose aluminium silicates and incipiently substitutes plagioclase; beside chlorite it can also occur in the prejudice of garnet. The minerals of this paragenesis does not display obvious deformation traces, which makes us think that their blastesis developed mostly in static conditions.

In regionally retrogressed rocks, oligoclase porphyroblasts get augen forms, having pressure shadows in which quartz recrystallizes. Garnet inclusions are frequently broken, while the quartz ones, initially rounded, generate myrmekitic structures. During this stage, sometimes also sphene inclusions occur. Augen porphyroblasts are tightly adhered to micas and sometimes display rolling extinctions.

The only textural elements typical of the M_3 metamorphism are the kink-bands and the crenulation cleavages, which affect the muscovite beds. These cleavages, representing a possible S_3 foliation, were rarely remarked in outcrops. The forming of the augen textures in the S_2 foliation plane proves that a part of the movement

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quantity was consumed on the old S_2 planes. It is possible that the scarcity of the specific textural elements does not reflect the real situation of the M_3 moment; the reactivity of the S_2 foliation in Alpine movements had an almost complete destructive effect on textures belonging to the regional retromorphism.

2.4 The M_4 metamorphism

The last metamorphic event which affects the Făgăraş series is the Alpine dynamic retrogression. The Alpine paragenesis is made up of quartz and sericite. Quartz is the only mineral obviously recrystallized in M_4. It recrystallizes postkinematically segregating in bands and lenses, visible both microscopically and macroscopically, at the scale of the sample and the outcrops. The alternation of quartz bands and disaggregated micas give to the rocks affected by dynamic metamorphism a very characteristic microscopic banding. Sericite substitutes plagioclase, muscovite and garnet. Its belonging to the Alpine paragenesis is not very clear, because all the sericitization processes are set in during the regional retrogression. The greatest part of sericite develops against muscovite and plagioclase.

In dynamically metamorphosed rocks, plagioclase ocelli becomes fusiform, and micas that overlap them, as well as the pressure shadows are destroyed during this stage. There appear deformation twins and parallel fissures sets orientate at 45^0 as against the long axis of the ocelli. In a later stage of deformation, there are shearing fissures and all garnet inclusions are broken. From this moment on, ocullar porphyroblasts become porphyroclasts and they are invaded by sericite.

The M_4 dynamic metamorphism is characterized by a weak mineral neoformation, the deformation rate totally exceeding the blastesis rate. The lamination movements are extremely strong and act destructively upon the previous parageneses, especially upon those belonging to the M_1 and M_2 metamorphism. These movements occur along certain planes approximately concordant with the S_2 foliation, penetrative on a regional scale, but unpenetrative at lower scales. The sliding surfaces bend northwards, and are sometimes torsioned. They are marked in the field by the occurrence of certain metric or submetric mylonites or cataclastites.

The M_4 metamorphic events strongly develops in the northern half of the Transylvanian slope and in the area of the main crest, affecting both the Şerbota Formation and the Suru Formation.

Conclusions

The Făgăraş series is made up of two tectonically related formations, which are distinct from the point of view of the premetamorphic material: the Şerbota Formation, mostly bearing paragneisses and micaschists, and the Suru Formation, mostly made up of amphibolites and crystalline limestones. The unitary petrographical character of the Suru Formation is recognized by most geologists who work in the Făgăraş Mts.

We maintain that also the Şerbota Formation has an unitary character and, thus, it cannot be subdivided lithostratigraphically; all the lithostratigraphical subdivisions made within it up to now, are based upon petrographical differences generated by retrograde metamorphism. As the two formations have a common metamorphic evolution, it is unlikely that the tectonic plane between them should be an overthrust plane.

The detailed microscopic examination of the Făgăraş series rocks allowed us to establish a detailed image of the mineral paragenesis succession in this series. The succession of the post-M_1 transformations is that of a retrograde evolution, starting from the initial paragenesis, bearing staurolite and kyanite.

The four distinct metamorphic events occured in different thermodynamical conditions. The transformations that biotite underwent, specific of each moment of metamorphism, reflects perhaps the best this change of physical conditions. Between M_1 and M_2, over biotite minerals of the epidote group develop. During the first phase of the M_2 metamorphism, most of biotite transforms itself into phengitic muscovite. During the second phase, over biotite fibrolitic sillimanite develops; at the same time, it locally reacts with the plagioclase, resulting garnet or reaction muscovite. During the M_3 metamorphism, the biotite that did not turn into muscovite almost totally is chloritizised, while during the M_4 dynamic metamorphism the last biotite relics are mechanically destroyed.

During the M_2 metamorphism, we notice a continuous changing of the thermodynamical conditions. The self-heating of the retrograding system, during the second phase of the M_2 moment, starts reactions resulting in a discreet paragenesis, bearing fibrolitic sillimanite. At this moment there occur the thermic remobilization and the recrystallization of the quartz-feldspar component in the entire Făgăraş series, including in the Suru Formation amphibolites. In metapelites, the recrystallized plagioclase is a 15–20 % An oligoclase. The same oligoclase prevails also in amphibolites, where it coexists with an older phase (35 % An andesine) and a newer one (7–10 % An albite).

As for the dynamic regime, we think that the S_2 regional foliation functioned as a lamination plane all the time post-M_2, surely being reactivated during M_4, and probably during M_3. For M_4, this fact is proved by the

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Alpine tectonic planes, concordant with S_2 and penetrative only on a regional scale. The M_3 reactivation is suggested by the augen textures appearance, concordant with the S_2 foliation.

References

Balintoni I., Hann H., Gheuca I., Nedelcu L., Conovici M., Dumitraşcu G., Gridan T. (1986) Considérations on a Preliminary Structural Model of the South Carpathian Cristalline East of the Olt River. *D. S. Inst. Geol. Geofiz.*, 70-71/5, p. 23–44, Bucureşti.

Dimitrescu R. (1987) Lithostratigraphie des schistes cristallins du versant nord des monts de Făgăraş. *Anal. St. Univ. "Al.I. Cuza", Geol.-Geogr.*, XXXIII, p. 7–8, Iaşi.

Gheuca I. (1988) Versantul sudic al munţilor Făgăraş. Litostratigrafie şi tectonică. *D. S. Inst. Geol. Geofiz.*, 72–73/5 (1985, 1986), p. 93–117, Bucureşti.

Giuşcă D., Anastasiu N., Popescu Gh., Şeclăman M. (1977) Observaţii asupra şisturilor cristaline din zona centrală a Masivului Făgăraş (Cumpăna -v. Cirţişoara). *Anal. Univ. Buc. (Geol.)*, XXVI, p. 3–17, Bucureşti.

Gridan T., Dumitraşcu G. (1990) Notă asupra prezenţei formaţiunii de Valea Lupului în Făgăraşul central, versantul nordic. *D. S. Inst. Geol. Geofiz.*, 74/5, p. 29–40, Bucureşti.

Pană D., Ricman C. (1989) Post-nappe folding of metamorphics in NW Făgăraş Massif. *D. S. Inst. Geol. Geofiz.*, 74/1, p. 239–250, Bucureşti.

Pană D. (1990) Central and northern Făgăraş. Lithological sequences and Structure. *D. S. Inst. Geol. Geofiz.*, 74/5, p. 81–89, Bucureşti.

Şeclăman M., Hârtopanu I. (1987) Considerations on the Thermal Regim in Retromorphism. *Rev. roum. géol., géophys., géogr., Géologie*, 31, p. 39–46, Bucureşti.

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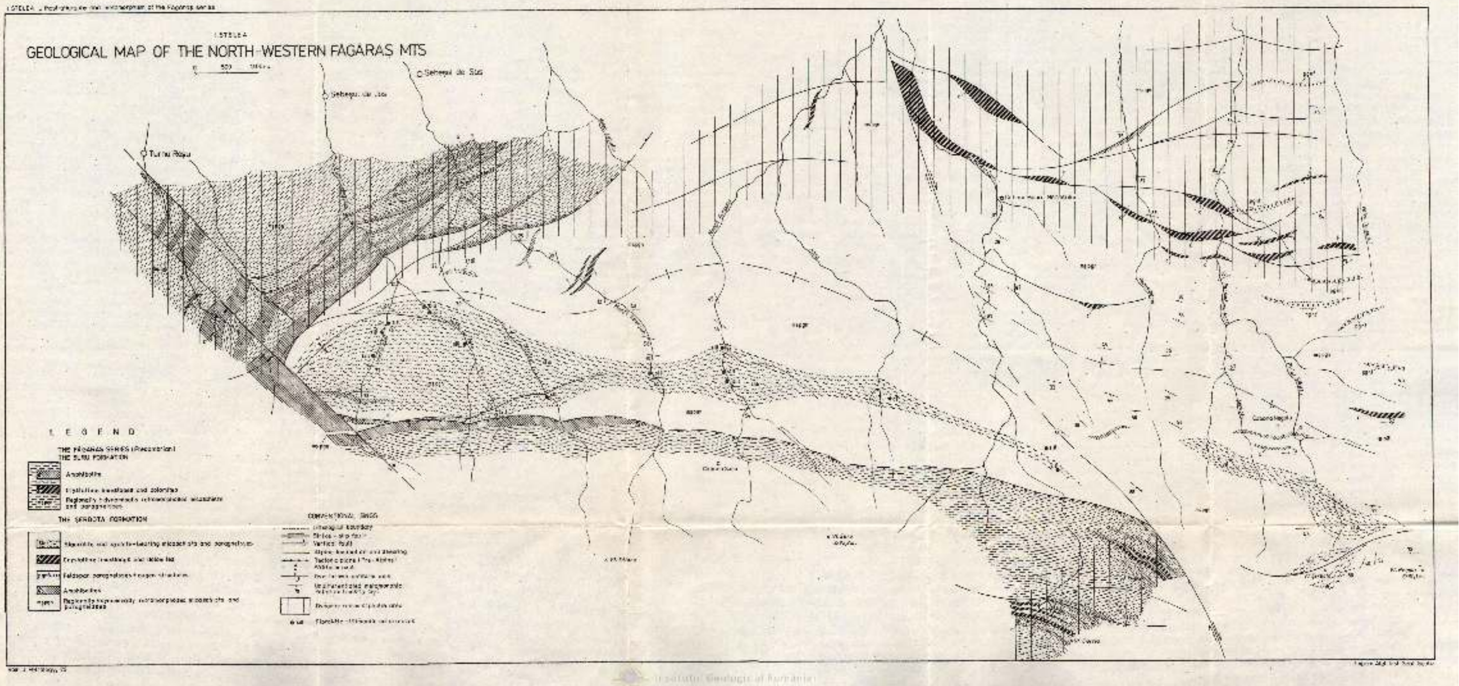

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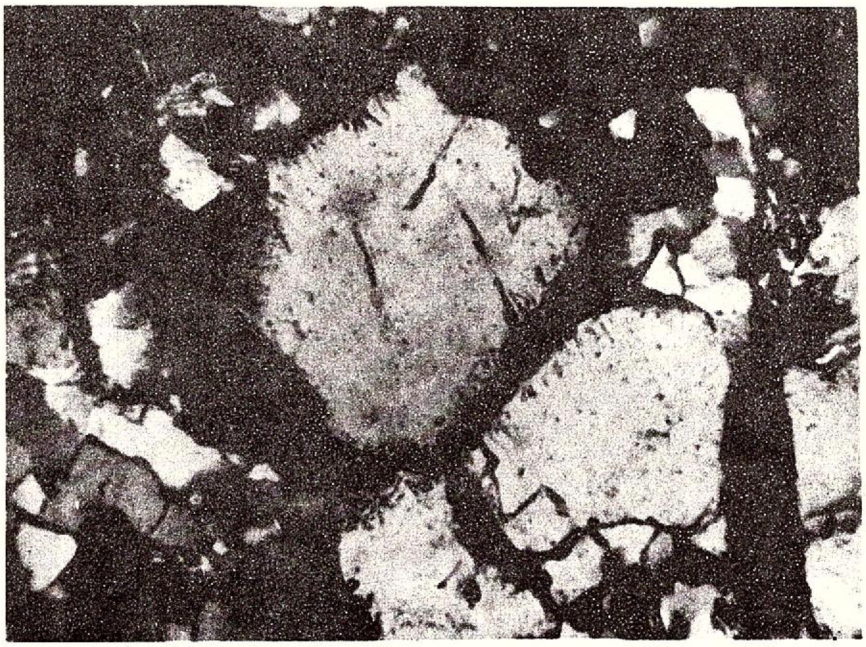

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Plate I

Fig. 1 – **Fibrolitic sillimanite faccicles in quartz. N II, 20 x. Sample 445.**

Fig. 2 – **Fibrolitic sillimanite needles in the marginal zone of the quartz grains. N +, 25 x. Sample 445.**

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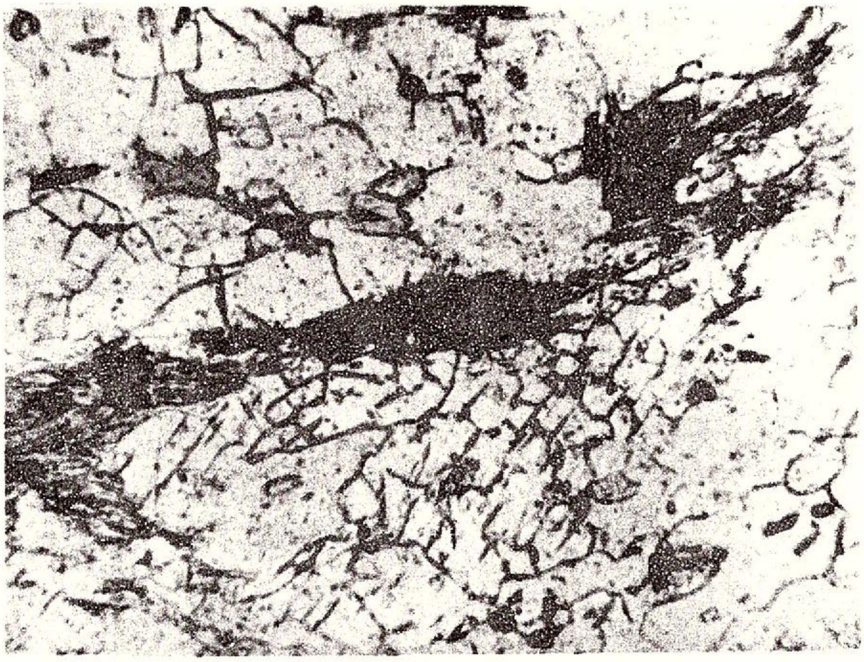

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Plate II

Fig. 1 – Fibrolitic sillimanite grown directly on disthene. N II, 20 x. Sample 517.

Fig. 2 – Reaction muscovite on the biotite included in oligoclase porphyroblast. N +, 25 +. Sample 517.

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Rom. J. Petrology, 1992, 75, p. 189–199

AMFIBOLITELE ŞI ULTRAMAFITELE ASOCIATE DIN ZONA PĂLTINIŞ – GURA RÎULUI (MUNŢII CIBIN)

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Key words: Amphibolites. Ultramafics. Mineral composition. Optical mineralogy. Major elements. Minor elements. South Carpathians – Crystalline Getic and Supragetic Realms – Cibin Mountains.

Abstract: *Amphibolites and Associated Ultramafic Rocks in the Păltiniş-Gura Rîului Area (Cibin Mountains).* This paper deals with aspects concerning the mineralogy, chemistry and petrogenesis of amphibolitic and associated ultramafic rocks occurring in the Păltiniş-Gura Rîului area (Cibin Mts). Within the amphibolitic rocks, orthoamphibolites and paraamphibolites have been separated and described. Orthoamphibolites have been formed by metamorphosis, under conditions of the almandine amphibolite facies, of basic magmatic rocks, with a composition similar to that of basalts, and paraamphibolites from pelitic sedimentary rocks, partially associated with magmatic material. The associated ultramafic rocks, represented by plagioclase-bearing wehrlites with transition towards melagabbros, are in different stages of transformation; finally, they are mostly constituted of phyllosilicates – especially chlorite from sheridanite to pennine, and antigorite and crysotile, iñ association with tremolite and opaque minerals; the last ones are represented by sulphides – pyrrhotite, chalcopyrite, cobalt-bearing pentlandite, and oxides-chromhercynite, ilmenite and magnetite.

În campania de vară a anului 1986, în partea nord-estică a munţilor Cibin, intr-un perimetru cuprins între Gura Rîului, valea Cibinului, valea Rîul Mare pînă la confluenţa cu pîrîul Dăneasa, aproximativ paralela Păltinişului şi valea Mărăjdiei (fig. 1) am făcut o probare sistematică a metabazitelor intîlnite şi a corpului de ultramafite de la obîrşia văii Mărăjdiei, in vederea studiului mineralogic, petrografic, chimic şi genetic al acestora.

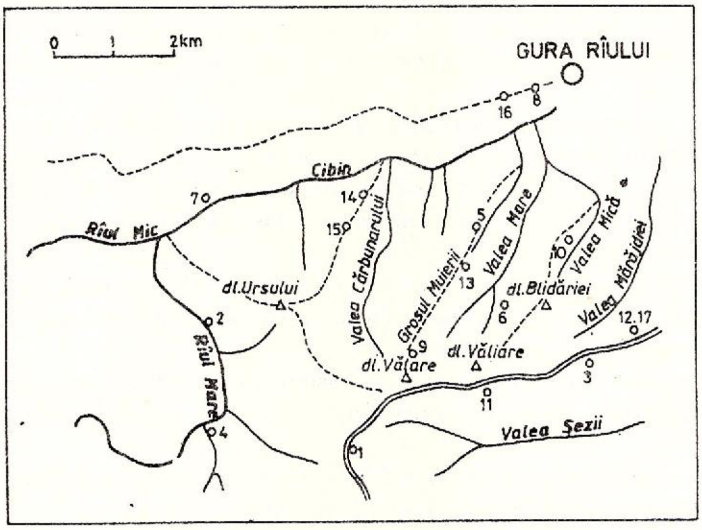

Fig. 1 – Schiţa topografică a perimetrului, cu localizarea probelor chimice.

Zona Păltiniş-Gura Rîului a fost cercetată de către Micu (1964, raport nepubl.), Şeclăman (1974, raport nepubl.), Apostoloiu et al. (1982, raport nepubl.), Hartopanu et al. (1987, raport nepubl.). Structural

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este situată în flancul nordic al sinclinoriului Lotru-Cibin-Sebeş, fiind constituită în întregime din metamorfitele grupului Sebeş-Lotru, metamorfozate în condiţiile faciesului amfibolitelor cu almandin, zonele cu staurolit, disten şi sillimanit. Fondul petrografic este format din plagiognaise muscovito-biotitice mediu pînă la grosier-granulare, în alternanţă cu plagiognaise biotitice fin granulare şi micaşisturi; în masa acestora se găsesc intercalaţii frecvente, sub formă de lentile de diferite dimensiuni, de roci amfibolice şi cuarţite. Întreaga suită este puternic migmatizată şi străbătută de numeroase corpuri pegmatitice.

I. Mineralogia şi petrografia rocilor amfibolice

Pe baza studiului microscopic şi a interpretării chimismului acestor roci (localizarea probelor analizate chimic este dată în figura 1 şi tabelul 1), am separat ortoamfibolite şi paraamfibolite.

1. Ortoamfibolite. In cadrul acestora pot fi descrise amfibolite, amfibolite zoizitice, amfibolite cu granat şi amfibolite eclogitice.

a)*Amfibolitele* au structură fanerocristalină, grosieră şi mai rar medie, aproximativ uniformă şi textură de la slab orientată la şistoasă, în funcţie de raportul cantitativ dintre mineralele principale, care în toate probele sînt amfibolul şi plagioclazul; în cantităţi reduse mai pot conţine epidot-clinozoizit, titanit, biotit, cuarţ, granat, apatit, ilmenit, magnetit şi calcopirită.

Amfibolul, de cele mai multe ori, este o hornblendă comună, cantitativ variind într-un interval foarte larg, frecvent putînd ajunge pînă la 75-80 % ; mai rar, hornblenda poate fi singurul mineral, rezultînd amfibolite monominerale (proba 5, de exemplu). Cînd se găseşte în cantitate ridicată, hornblenda formează poikiloblaste larg dezvoltate (pînă la 4-5 mm), scurt prismatice, conţinînd incluziuni de plagioclaz, cuarţ, granat, apatit şi epidot-clinozoizit, iar cînd se găseşte în proporţii aproximativ egale cu plagioclazul (totdeauna, în aceste cazuri, structura rocii este mai fină) formează nematoblaste mici (0,5-1,0 mm) , în general fără incluziuni. Pleocroismul este, de obicei, în nuanţe deschise, de la verde uneori cu nuanţă uşor brună, pînă la galben-verzui deschis, iar c : Ng = 20-23^0 ; poate apărea parţial cloritizată (clorit aluminos).

Plagioclazul este în majoritatea probelor net subordonat hornblendei, de obicei găsindu-se în proporţie de 15-25 % şi, mai rar, de 35-40 % . Formează atît granoblaste individuale mai larg dezvoltate (0,2-1,0 mm), xenomorfe în raport cu hornblenda, cît şi granule resorbite incluse în hornblendă; în unele probe este maclat după (010) şi subordonat după (001)/[010], uneori polisintetic însă cu puţini indivizi, iar în alte probe maclarea este rară. Cele mai multe granoblaste prezintă o uşoară zonare, manifestată prin extincţie neuniformă concentrică, compoziţia medie fiind cuprinsă în intervalul An_{25}-An_{30}; poate prezenta un început de transformare în epidot-clinozoizit.

b)*Amfibolitele zoizitice* (proba 14, de exemplu) au structură mediu-granulară şi textură slab orientată, fiind constituite din cantităţi aproximativ egale de amfibol şi clinozoizit, subordonat putînd conţine şi epidot.

Amfibolul este reprezentat prin hornblendă, cu caracteristici asemănătoare cu aceea din amfibolitele propriu-zise; în plus, poate fi parţial transformată în actinolit; dimensiunea celor mai multe nematoblaste este în jur de 1-1,5 mm.

Clinozoizitul apare idiomorf în raport cu hornblenda, formînd atît cristale mărunte, aproximativ echidimensionale, cît şi cristale prismatice alungite, larg dezvoltate (1,5-2 mm), frecvent maclate polisintetic.

Epidotul este în cantitate redusă, cu acelaşi mod de prezentare ca şi clinozoizitul.

Dacă epidotul pare să fie format pe seama hornblendei, clinozoizitul s-a format concomitent cu hornblenda, asociindu-se acesteia în locul plagioclazului.

c)*Amfibolitele cu granat* au, în general, structură grosieră, deosebindu-se de amfibolitele propriu-zise prin conţinutul ridicat în granat, uneori roca fiind constituită în cea mai mare parte din hornblendă şi granat (proba 16, de exemplu), plagioclazul fiind în cantitate redusă.

Granatul se dezvoltă poikiloblastic, cu dimensiuni ce pot atinge 4-5 mm; sub formă de incluziuni conţine hornblendă, cuarţ, titanit, dovedind o cristalizare simultană cu hornblenda; cantitativ poate ajunge la 15-20 % .

d)*Amfibolitele eclogitice* sînt formate dintr-o masă fin granulară, cu structură diablastică, în care se păstrează resturi de amfibol, granaţi şi minerale opace (proba 13, de exemplu).

Masa diablastică este constituită din actinolit, albit şi cuarţ, probabil şi puţin piroxen.

Amfibolul este reprezentat prin poikiloblaste de hornblendă, cu dimensiuni pînă la 3-4 mm, prezentînd pleocroism de la brun-verzui deschis la galben-verzui deschis; c:Ng este variabil, cele mai mari valori fiind însă în jur de 25^0; de obicei, hornblenda este opacizată pe clivaje, ca incluziuni primare conţinînd plagioclaz, granat şi minerale opace.

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Granatul formează, de asemenea, poikiloblaste aproximativ izometrice, rotunjite, conţinînd incluziuni de plagioclaz, actinolit şi, subordonat, cuarţ; cele mai multe granule au dimensiuni în jur de 0,3-0,5 mm.

Mineralele opace sînt reprezentate prin ilmenit şi calcopirită.

2. Paraamfibolite. Spre deosebire de ortoamfibolite, acestea au o constituţie mineralogică majoră mai puţin variată, fiind, în general, roci fin pînă la mediu-granulare, constituite în proporţii aproximativ egale de amfibol şi plagioclaz; în cantităţi mici pot conţine cuarţ, muscovit, biotit, granat, epidot-clinozoizit, calcit, titanit şi apatit.

Amfibolul, faţă de acela din ortoamfibolite, este, cel mai frecvent, reprezentat prin actinolit sau tremolit-actinolit, sub formă de nematoblaste lung-prismatice, cu lungimi diferite în probe diferite, în general de la 0,5 pînă la 2 mm şi mai rar pînă la 4 mm; prezintă pleocroism în nuanţe de verde deschis, c:Ng fiind de 15–17^{0}; mai rar, poate fi întîlnită şi hornblenda. Nematoblastele mai mari conţin incluziuni de plagioclaz, cuarţ, titanit; secundar, apare cloritul.

Plagioclazul formează granoblaste mărunte (0,3-1,0 mm), de obicei cu structură zonară concentrică; zona centrală este uşor mai bazică decît zona exterioară, compoziţia medie fiind în intervalul An_{35}-An_{40} ; poate fi nemaclat sau maclat în general după (010), dar şi (001)/[010], de obicei cu puţini indivizi.

In unele probe (proba 3, de exemplu) pe fondul constituit din amfibol şi plagioclaz se disting aglomerări lenticulare (pînă la 2,5-3 mm) constituite în principal din epidot-clinozoizit, subordonat găsindu-se clorit, cuarţ, plagioclaz alterat şi muscovit; acestea ar putea fi probabil elemente relicte, puternic transformate, din roca sedimentară iniţială.

II. Mineralogia şi petrologia ultramafitelor

Ultramafitele au fost întîlnite într-un singur loc (proba 17) unde sînt deschise artificial (în carieră) pe o grosime de aproape 10 m; spaţial, sînt strîns legate de ortoamfibolitele, străbătute discordant de numeroase corpuri pegmatitice, care aflorează către vest, aproximativ pe direcţie, pe o lungime de aproximativ 200 m (proba 12); reprezintă, probabil, o acumulare gravitaţională de fenocristale în primele faze ale consolidării bazaltelor care prin metamorfism au generat ortoamfibolitele menţionate. In cadrul corpului de ultramafite pot fi urmărite toate trecerile de la roca iniţială puţin transformată, conservată pe anumite porţiuni, pînă la roci complet transformate, constituite în prezent mai ales din filosilicaţi, cărora, ca un component major, li se poate adăuga amfibolul. In general, prezintă multe asemănări cu rocile de tip apowehrlitic şi cloritic descrise de Bercia şi Bercia (1962) în Banatul de Sud: de asemenea, Pavelescu (1955) citează un corp de serpentinite la sud de Răşinari - pe Valea Muntelui - dar acestea ar proveni din peridotite hornblenditice.

Roca iniţială puţin transformată, fanerocristalină, este constituită din cantităţi aproximativ egale de piroxen şi olivină magneziană, cărora, în unele probe li se adaugă plagioclazul; secundar apar filosilicaţi, epidot, clinozoizit, carbonaţi şi subordonat, tremolit; în plus, conţine minerale opace în cantitate ridicată.

Piroxenul formează cristale idiomorfe, scurt-prismatice, cu dimensiuni de pînă la 4-5 mm, înglobînd în masa sa granule de olivină; este aproape incolor, cu pleocroism slab în nuanţe de verde-gălbui deschis şi culori de birefringenţă galben-roşcate de ordinul I; c:Ng = 34^{0}, iar Ng - Np = 0,020, proprietăţi care indică un piroxen la limita dintre clinoenstatit şi clinohipersten. Contactul dintre piroxen şi plagioclaz, pe de o parte, sau olivina inclusă, pe de altă parte, este net, fără zone de reacţie.

Olivina formează cristale de diferite dimensiuni, atît izolate, mai mari, cît şi incluse în clinopiroxen; este parţial serpentinizată pe fisuri, pe lîngă filosilicaţi găsindu-se şi magnetit. Olivina în contact cu plagioclazul este totdeauna înconjurată de o zonă de reacţie formată din filosilicaţi.

Plagioclazul apare aproape complet transformat în calcit şi tremolit în apropierea contactului, mai ales cu olivina, dar şi cu clinopiroxenul, păstrîndu-se, totuşi, unele zone ceva mai puţin transformate în părţile centrale ale cristalelor, unde, pe baza caracterelor optice pare să corespundă labradorului.

Filosilicaţii formaţi pe seama olivinei, în acest stadiu, sînt reprezentaţi prin antigorit şi crisotil.

Cu aceste caractere, roca magmatică iniţială poate fi încadrată în grupa rocilor ultramafice (wehrlite) cu plagioclaz, cu trecere spre melagabbrouri.

Intr-un stadiu mai avansat de transformare a rezistat parţial doar olivina, piroxenul şi plagioclazul fiind înlocuite în cea mai mare parte prin bastit şi, respectiv, tremolit.

Bastitul pseudomorfozează aproape în întregime piroxenul, rezultînd şi o pulbere neagră care opacizează parţial fondul acestuia. Este incolor, cu culoare de birefringenţă cenuşie închisă, biax (-) şi alungire pozitivă (proprietăţi caracteristice antigoritului). Pseudomorfozele de bastit pot include olivina, anterior inclusă în piroxen; contactul dintre olivină şi bastit rămîne tot aproape brusc, pe cînd între olivină şi fostul plagioclaz se măreşte, existînd astfel tendinţa de transformare completă a olivinei în filosilicaţi.

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Tremolitul este aproape incolor, cu $c:Ng = 20^0$, dezvoltîndu-se în cea mai mare parte în zonele constituite iniţial din plagioclaz.

În anumite zone, pe fondul puternic transformat al rocii se observă tendinţa, uneori destul de clară, de formare, chiar sub formă de lamele larg dezvoltate, a cloritului aluminos.

În stadiul de transformare completă, roca apare constituită în cea mai mare parte (70-80%) din filosilicaţi formînd aglomerări lamelare (pînă la 4-5 mm), între aglomerări sau, uneori şi în masa acestora, găsindu-se mai ales tremolit.

Filosilicaţii sînt reprezentaţi prin diferiţi termeni ai cloritelor normale – sheridanit, clinoclor şi pennin – şi, numai subordonat, prin antigorit şi crisotil; astfel, din punct de vedere chimic, există probabil o serie continuă între cloritul aluminos (sheridanit) şi mineralele serpentinitice (antigorit sau crisotil).

Tremolitul este format anterior filosilicaţilor, păstrîndu-se sub formă de resturi în masa acestora sau ca fragmente izolate cu dimensiuni pînă la 1,5 – 2 mm; de obicei, în jurul resturilor de tremolit şi pe clivajul acestora se dezvoltă penninul.

Sub formă de produse secundare, rezultate în urma reacţiilor succesive de transformare, apar carbonaţii şi mineralele opace.

Mineralele opace din utramafite aparţin cel puţin la două generaţii.

– Generaţia mai veche, primară, reprezentată prin sulfuri (pirotină, calcopirită, pentlandit) şi oxizi (spinel, ilmenit şi, subordonat, magnetit); mai întîi au cristalizat sulfurile, după care au urmat oxizii, ultimul fiind ilmenitul.

Pirotina, calcopirita şi pentlanditul sînt în cantitate redusă, formînd concreşteri granulare mărunte, izolate sau incluse ori asociate spinelului şi magnetitului; dimensiunile nu depăşesc, în general, 0,1 mm. Imaginile de raze X au arătat, în toate cazurile o anumită cantitate de cobalt în compoziţia pentlanditului, fiind astfel vorba de un pentlandit cobaltifer (pl. I, fig. 1-6).

Spinelul se găseşte în cantitatea cea mai ridicată (2-3 %), sub formă de granule subidiomorfe de diferite dimensiuni (de obicei între 0,05 şi 1 mm, dar şi pînă la 2-3 mm), prezentînd frecvent dezamestecuri orientate sau zone marginale de ilmenit (pl. II, fig. 1-6); poate include, de asemenea, granule concrescute de pirotină şi calcopirită ± pentlandit. Imaginile de raze X obţinute la microsonda electronică au indicat prezenţa în cantităţi aproximativ egale a Fe^{2+}, Al şi Cr, raportul Al:Cr apărînd totuşi uşor supraunitar; de asemenea, este prezentă şi o cantitate mică de Mg. Cu această compoziţie – $Fe(Al,Cr_2)O_4$ – spinelul are compoziţia cromhercinitului, apropiat de compoziţia spinelilor cromiferi (în principal berezovskit) din Banatul de Sud (Kräutner, 1962).

Ilmenitul apare cel mai frecvent ca dezamestecuri în cromhercinit sau concrescut cu acesta, mai rar formînd şi granule individuale.

Cromhercinitul şi ilmenitul se asociază, de obicei, piroxenului, fiind incluse în acesta (ca şi olivina).

Magnetitul primar este în cantitate redusă, putîndu-se prezenta şi sub formă de concreşteri cu pirotina şi calcopirita.

Generaţia mai nouă, secundară, este reprezentată prin magnetit, sub formă de pulbere fină asociată serpentinei formată pe fisurile olivinei, sau bastitului format pe seama piroxenului; cantitatea de magnetit astfel formată este mare, probele deviind puternic acul magnetic; de asemenea, magnetitul poate pătrunde pe fisurile spinelului (pl. II, fig. 1, 2). Pe măsura transformării tot mai accentuate a rocii în filosilicaţi (clorite şi antigorit) are loc o mobilizare şi recristalizare a magnetitului secundar, sub formă de granule cu dimensiuni reduse, mai ales alungite, ducînd la o "curăţire" a probelor, în sensul că în locul pulberii negre, distribuite aproximativ uniform în toată roca, se formează granule mari şi, implicit, mult mai rare.

III. Chimismul şi petrogeneza

Consideraţiile chimice se bazează pe interpretarea unui număr de 17 analize complete de silicaţi şi spectrale (16 amfibolite şi un ultramafit puternic transformat).

Proiectînd valorile ACF (fig. 2), toate probele de amfibolite se plasează în cîmpul plagioclaz-hornblendă-almandin, în apropierea liniei plagioclaz-amfibol.

Pentru încercarea unei discriminări între orto- şi paraamfibolite am folosit o serie de diagrame de corelaţie dintre elementele majore. Astfel, figura 3 (simplificare după la Roche, 1972) arată că probele 1, 2, 3, 4, 6 şi 7 se proiectează în cîmpul sau în imediata vecinătate a rocilor pelitice, aproape de diorite, pe cînd toate celelalte probe urmează îndeaproape linia rocilor magmatice bazice; o delimitare şi mai clară apare pe diagramele din figurile 4 şi 5, în care am proiectat raportul $MnO:TiO_2$ şi, respectiv, $TiO_2:F$, $F = (FeO + Fe_2O_3):(FeO + Fe_2O_3 + MgO)$ (după Misra, 1971).

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TABELUL 1

Compoziţia chimică a rocilor amfibolitice şi a probei de serpentinit

Proba	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	CO_2	S	H_2O^+
1	52,36	16,8o	1,69	7,19	o,15	4,41	9,32	3,81	o,65	o,84	o,21	o,oo	o,12	1,95
2	51,46	18,94	1,63	5,52	o,15	5,59	9,15	4,17	o,27	o,5o	o,o5	o,oo	o,11	1,81
3	51,4o	18,34	1,4o	5,86	o,12	5,93	9,24	3,57	o,29	o,6o	o,11	o,oo	o,16	2,43
4	49,87	15,28	2,o3	9,45	o,15	5,o8	1o,31	3,66	1,o2	o,42	o,o5	o,oo	o,13	1,89
5	48,96	8,6o	3,o8	6,88	o,23	13,15	12,52	1,62	o,2o	1,66	o,o4	o,oo	o,23	2,o3
6	48,95	15,56	1,o5	6,39	o,14	8,5o	12,12	2,96	o,42	o,48	o,o4	o,oo	o,13	2,84
7	48,85	17,79	1,15	5,55	o,14	7,94	1o,53	2,92	1,61	o,92	o,o9	o,oo	o,o5	2,38
8	48,83	14,1o	2,27	9,33	o,27	6,73	1o,9o	2,15	o,17	2,22	o,o9	o,oo	o,18	2,17
9	48,43	14,91	o,97	6,36	o,14	9,98	1o,88	2,29	o,98	o,84	o,o4	o,oo	o,3o	2,98
1o	47,56	12,7o	1,9o	8,o6	o,16	9,39	1o,77	2,54	1,11	1,72	o,o6	2,14	o,o7	1,63
11	47,46	15,36	1,8o	8,49	o,17	6,94	1o,o8	3,1o	o,58	1,58	o,29	o,oo	o,12	3,36
12	46,98	13,44	2,82	8,81	o,17	7,78	12,21	2,o9	o,68	2,12	o,o4	o,oo	o,o8	2,58
13	46,96	13,59	2,7o	12,o2	o,16	6,89	1o,67	3,37	o,2o	1,9o	o,14	o,oo	o,22	o,66
14	46,29	14,59	7,92	3,98	o,2o	5,84	11,29	3,33	o,66	2,52	o,o1	o,oo	o,o6	2,88
15	46,28	14,8o	8,81	4,17	o,2o	6,72	11,91	3,o8	o,71	1,16	o,28	o,oo	o,16	1,27
16	44,17	13,65	3,34	11,68	o,23	4,88	13,87	o,82	o,33	3,9o	o,64	o,oo	o,19	2,57
17	39,4o	6,7o	3,39	7,oo	o,14	26,93	5,78	o,48	o,o3	o,28	o,o4	3,o5	o,2o	5,76

1-16 amfibolite: 1,Soseaua Păltiniş-Răşinari,vest de dealul Poplaca;2,Rîul Mare,aprox.17oo m amonte de confluenţa cu Rîul Mic;3,Soseaua Păltiniş-Răşinari,S 4o°E faţă de vîrful Blidăriei;4,pîrîul Dăneasa,aprox. 75 m de confluenţa cu Rîul Mare; 5,creasta Grosul Muierii,aprox. cota 1o7o m; 6,versantul drept al Văii Mari,vest de vîrful Blidăriei; 7,versantul stîng al Cibinului,aprox.1ooo m aval de confluenţa Rîului Mare cu Rîul Mic; 8,creasta din stînga Cibinului,vest de Gura Rîului (cota 7oo m); 9,aprox.45o m nord de dealul Vălare,pe creastă (cota 127o m); 1o,Valea Mică,aprox. cota 9oo m; 11,pe dreapta şoselei Păltiniş-Răşinari, sud de dealul Văliare; 12,pe stînga şoselei Păltiniş-Răşinari,Km 1o-carieră; 13,creasta Grosul Muierii aprox.cota 113o m; 14,interfluviul din stînga pîrîului Cărbunarului,aprox.cota 75o m; 15,idem 14,aprox. cota 85o m; 16,creasta din stînga Cibinului,aprox.cota 73om; 17,ultramafit cu plagioclaz complet transformat,idem 12.

TABELUL 2

Elemente minore (ppm) în probele de roci amfibolitice şi în proba de serpentinit

Proba	Pb	Cu	Zn	Sn	Ga	Ni	Co	Cr	V	Sc	Y	Yb	Zr	Nb	La	Be	Ba	Sr
1	3	46	58	<2	16	19	25	36	5oo	3o	18	2	85	<1o	37	<1	88	64o
2	2,5	18	46	<2	12	38	32	66	25o	27	12	<1	3o	<1o	<3o	<1	85	33o
3	<2	4,5	53	<2	14	5o	32	8o	25o	2o	17	<1	11	<1o	<3o	<1	13o	68o
4	<2	34	7o	<2	13	85	48	19o	55o	52	34	3,7	2oo	12	<3o	1,2	25o	36o
5	3	18	65	3,5	6	11o	66	65o	33o	85	13	<1	13	<1o	37	<1	32	11o
6	2,5	39	38	<2	9	95	41	37o	24o	4o	<1o	<1	14	<1o	<3o	<1	4o	23o
7	4,5	4	38	<2	8,5	24o	36	35o	26o	24	21	2,5	65	<1o	<3o	<1	26o	19o
8	3	34	12o	3	12	5o	29	3oo	6oo	54	28	3,5	85	<1o	<3o	2,6	44	78
9	13	2o	46	<2	1o	2oo	43	13oo	29o	33	2o	1,3	7o	<1o	37	1,4	6o	21o
1o	3,5	68	65	<2	11	24o	54	27o	27o	31	16	<1	42	<1o	<3o	<1	22o	4oo
11	2,5	46	53	<2	11	75	42	18o	54o	34	26	2,5	75	<1o	37	<1	21o	7oo
12	3,5	32	53	<2	11	8o	58	29o	115o	6o	16	2,5	24	<1o	<3o	1,7	65	22o
13	<2	53	145	5	13	65	55	12o	7oo	6o	53	6	145	<1o	<3o	<1	5o	85
14	4,5	13	<3o	<2	7,5	16o	36	13oo	26o	59	<1o	<1	<1o	<1o	<3o	<1	4o	2oo
15	3	12	95	<2	14	17o	46	33o	66o	46	5o	4,4	26o	18	<3o	1,5	13o	19o
16	5,5	52	95	5	14	25o	32	65o	42o	33	35	2	4oo	1oo	14o	1,8	5oo	4oo
17	3,5	24	38	<2	4	1ooo	46	25oo	63	12	<1o	<1	13	<1o	<3o	1,6	22	15

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TABELUL 3

Valorile Niggli pentru rocile amfibolitice

Proba	si	al	fm	c	alk	k	mg	ti	p	w	qz	c/fm
1	137,92	26,07	36,82	26,29	10,82	0,10	0,47	1,66	0,23	0,17	– 5,34	0,714
2	128,93	27,96	36,94	24,55	10,55	0,04	0,60	0,94	0,05	0,21	– 13,29	0,665
3	131,01	27,54	37,95	25,22	9,29	0,05	0,59	1,15	0,12	0,17	– 6,14	0,665
4	120,57	21,77	41,39	26,70	10,15	0,15	0,44	0,76	0,05	0,16	– 20,01	0,645
5	101,94	10,55	58,00	27,93	3,53	0,08	0,70	2,60	0,04	0,28	– 12,19	0,481
6	110,76	20,74	42,79	29,37	7,10	0,09	0,67	0,82	0,04	0,13	– 16,62	0,686
7	113,44	24,34	40,52	26,19	8,95	0,27	0,68	1,61	0,09	0,15	– 22,38	0,646
8	116,45	19,81	47,12	27,84	5,23	0,05	0,51	3,98	0,09	0,18	– 4,45	0,591
9	109,30	19,83	47,46	26,30	6,42	0,22	0,71	4,43	0,04	0,12	– 16,37	0,554
10	106,93	16,82	50,12	25,93	7,12	0,22	0,63	2,91	0,06	0,17	– 21,57	0,517
11	112,59	21,47	44,92	25,61	8,00	0,11	0,55	2,82	0,29	0,16	– 18,42	0,570
12	106,16	17,89	47,00	29,55	5,56	0,18	0,56	3,60	0,04	0,23	– 16,06	0,629
13	103,66	17,67	49,61	25,22	7,49	0,04	0,46	3,15	0,13	0,17	– 26,30	0,508
14	108,96	20,23	42,72	28,46	8,58	0,12	0,48	4,46	0,01	0,63	– 25,38	0,666
15	102,38	19,29	44,89	28,22	7,60	0,13	0,49	1,93	0,26	0,64	– 28,03	0,629
16	101,20	18,43	45,24	34,03	2,30	0,21	0,37	6,72	0,62	0,20	– 8,01	0,752

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TABELUL 4

Valorile A C F

Proba	A	C	F
1	22,3o	33,61	44,o9
2	24,54	31,58	43,87
3	24,59	31,16	44,25
4	17,33	34,14	48,53
5	1o,43	3o,75	58,81
6	17,15	34,49	48,36
7	2o,28	32,o5	47,67
8	19,o5	31,58	49,37
9	16,5o	2o,35	53,14
1o	14,61	24,78	6o,61
11	18,51	3o,27	51,22
12	17,o6	34,o7	48,87
13	15,o9	3o,11	54,81
14	24,62	37,51	37,88
15	24,83	35,71	39,47
16	21,o1	35,36	43,63
17	8,98	3,74	87,29

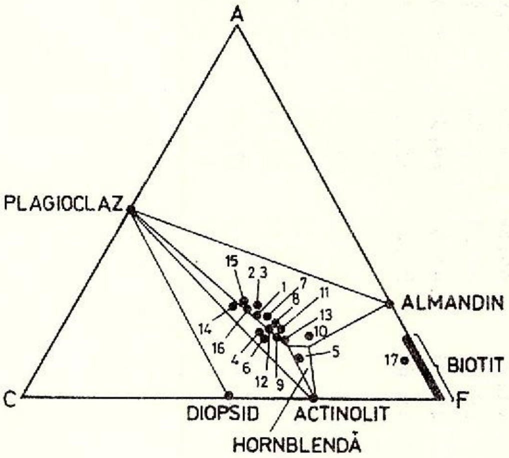

Fig. 2 – Diagrama ACF.

În ceea ce priveşte elementele minore, semnificativ poate fi raportul Ni:Co care este subunitar sau doar uşor supraunitar în probele care, şi pe baza corelaţiilor dintre elementele majore, tind către paraamfibolite, şi mult mai mare în majoritatea probelor care tind către ortoamfibolite.

Pe baza consideraţiilor de mai sus, pot fi apreciate ca fiind probabil ortoamfibolite, rezultate din roci magmatice cu compoziţia bazaltelor, probele 5, 8, 9, 10, 11, 12, 13, 14, 15 şi 16, iar paraamfibolite probele 1, 2, 3, 4, 6 şi 7, la acestea din urmă putîndu-se totuşi lua în seamă şi faptul că ar putea fi, măcar în parte, corespondentele metamorfozate ale unor amestecuri în proporţii variabile de material magmatic şi sedimentar (= tufite); astfel s-ar putea, eventual, justifica şi compoziţia mai acidă a rocii premetamorfice - compoziţia andezitului în toate cazurile (tab. 5).

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TABELUL 5

Norma Rittman pentru rocile amfibolitice

Probă / Mineral	1	2	3	4	5	6	7	8	9	1o	11	12	13	14	15	16
cuarț	2,18		o,52					2,75								3,43
sanidin				2,36			7,71		3,33	5,56		o,55				
andezin	68,9o	74,71	71,23	61,56	31,33	59,67				47,3o	62,76		53,55	54,11	31,38	
labrador							58,99	53,2o	54,16			51,o5				46,58
nefelin				o,57			1,83						1,25	3,8o	4,38	
augit	15,9o	11,o7	1o,45	25,55	44,99			23,64	23,68	17,81	2o,17	33,36				31,63
hypersten	9,58	6,84	15,13		17,82	o,18		16,32	7,25	14,28	4,34	5,95				11,39
diopsid						28,54	18,28						28,51	31,46	3o,o7	
olivină		4,94		6,5o	3,52	9,37	1o,64		8,72	6,34	8,76	5,7o	12,15	6,79	1o,26	
magnetit	1,91	1,58	1,45	2,72	1,o8	1,44	1,34	1,33	1,29	1,34	1,5o	1,35	2,2o	1,62	1,94	1,26
apatit	o,45	o,11	o,23	o,11	o,o9	o,o9	9,19	o,2o	o,o9	o,13	o,65	o,o9	o,32	o,o2	o,63	1,49
ilmenit	o,32	o,52	o,63	o,35	o,6o	o;42	o,9o	2,13	o,78	1,57	1,54	1,76	1,52	2,o5	o,97	3,77
calcit										5,52						
pirită	o,27	o,24	o,35	o,3o	o,55	o,3o	o,11	o,42	o,69	o,16	o,28	o,19	o,51	o,14	o,37	o,46
Denumirea rocii	andezit	andezit	andezit	andezit cu fiode	bazalt tholeitic	andezit cu olivină	latiandezit cu fiode	bazalt tholeitic	andezit cu olivină	latibezalt	hawaiit cu cuarț	bazalt tholeitic cu olivină	bazalt alcalin olivinic	bazalt alcalin cu olivină	bazalt alcalin olivinic	bazalt cuarțifer tholeitic

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In ceea ce priveşte proba 17, de ultramafit transformat, apreciem că ar putea reprezenta o acumulare gravitaţională de fenocristale de olivină, piroxen±plagioclaz în primele faze ale cristalizării magmatice care a generat rocile bazaltice prin metamorfozarea cărora au rezultat ortoamfibolitele.

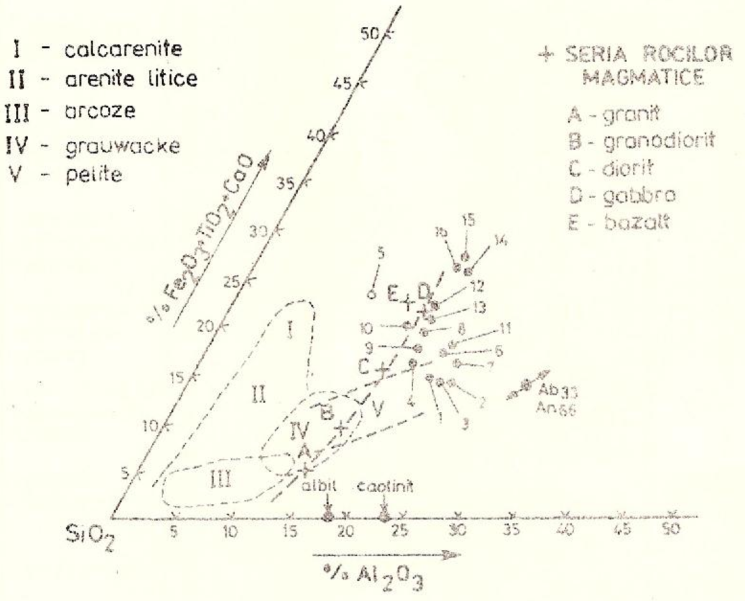

Fig. 3. — "Triunghiul rocilor totale" (după La Roche, 1972).

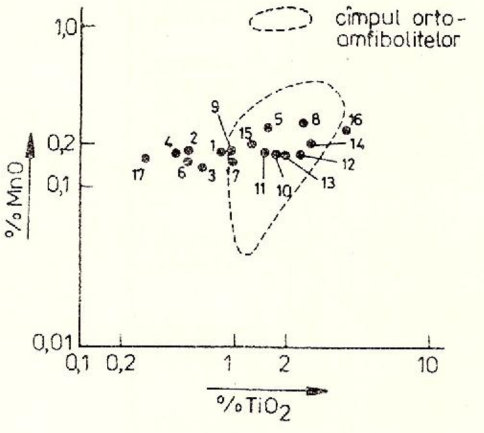

Fig. 4. — Diagrama TiO$_2$:MnO (după Misra, 1971).

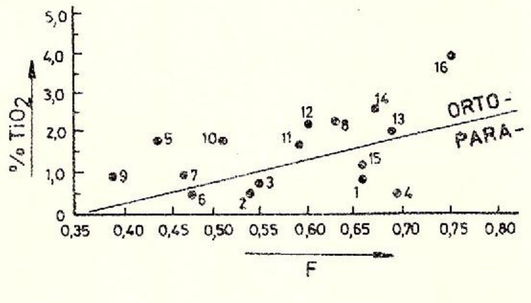

Fig. 5. — Diagrama TiO$_2$:F (după Misra, 1971).

IV. Concluzii

Rocile amfibolice, sub formă de intercalaţii lenticulare în masa plagiognaiselor grupului Sebeş-Lotru din partea nord-estică a munţilor Cibin, sînt corespondentele metamorfozate în faciesul amfibolitelor cu almandin ale unor roci magmatice bazice cu compoziţia bazaltelor = ortoamfibolite, sau ale unor roci sedimentare pelitice, eventual amestecate într-o anumită proporţie cu material magmatic = paraamfibolite.

In cadrul ortoamfibolitelor, pe baza constituţiei mineralogice majore şi, parţial structurii şi texturii, au fost separate amfibolite, amfibolite zoizitice, amfibolite cu granat şi amfibolite eclogitice.

Rocile ultramafice sînt asociate spaţial şi genetic ortoamfibolitelor, fiind reprezentate prin wehrlite cu plagioclaz, pe alocuri cu trecere spre melagabbrouri, în diferite stadii de transformare.

In stadiul final, roca este formată în proporţie de 75–80 % din filosilicaţi — în principal clorite (sheridanit, clinoclor şi pennin) şi numai subordonat prin antigorit şi crisotil, în cazul în care roca iniţială a conţinut

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plagioclaz în cantitate ridicată (melagabbro), sau mai ales antigorit şi crisotil, cînd roca iniţială a conţinut plagioclaz în cantitate redusă (wehrlit cu plagioclaz); restul este reprezentat prin tremolit şi minerale opace.

Prezenţa în cantitate mare a cloritelor în loc de minerale serpentinice şi, subordonat, a amfibolului calcic este datorată conţinutului de plagioclaz bazic în roca iniţială, prin descompunerea căruia au fost eliberate aluminiul şi calciul, incorporate, în final, în clorite şi, respectiv, tremolit; de asemenea, nu este exclusă şi posibilitatea existenţei în compoziţia rocii iniţiale, pe lîngă mineralele descrise, şi a hornblendei, deşi în probele noastre ea nu a fost intîlnită.

Mineralele opace din ultramafite sînt reprezentate prin sulfuri – pirotină, calcopirită şi pentlandit cobaltifer – şi oxizi – în principal spinel cu compoziţia cromhercinitului, magnetit şi ilmenit.

Bibliografie

Bercia I., Bercia E. (1962) Contribuţii la studiul serpentinitelor din Banatul de Sud. *An. Com. Geol.*, XXXII, p. 425–461, Bucureşti.

Kräutner H. (1962) Comportarea spinelilor cromiferi în procesul de serpentinizare (Banatul de Sud). *Stud. cerc. geol.*, VII, 3–4, Bucureşti.

Misra S. N. (1971) Chemical distinction of high-grade ortho- and parametabazites. *Norks. Geol. Tidsskr.*, 51, p. 311–316, Oslo.

Pavelescu L. (1955) Cercetări geologice şi petrografice în munţii Sebeş. *An. Com. Geol.*, XXVIII, p. 367–468, Bucureşti.

Roche de la H. (1972) Revue sommaire de quelques diagrammes chimico-minéralogiques pour l'étude des associations ignées ou sédimentaires et de leur dérivés métamorphiques. *Sci. de la Terre*, XVII, 1-2, p. 33–46.

Savu H., Berza T., Hârtopanu I. (1978) Precambrian in the Romanian Carpathians. A. South Carpathians (Guide to Excursions). Institutul de Geologie şi Geofizică, Bucureşti.

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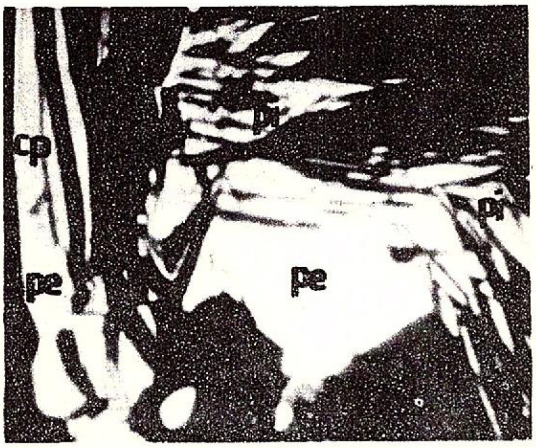

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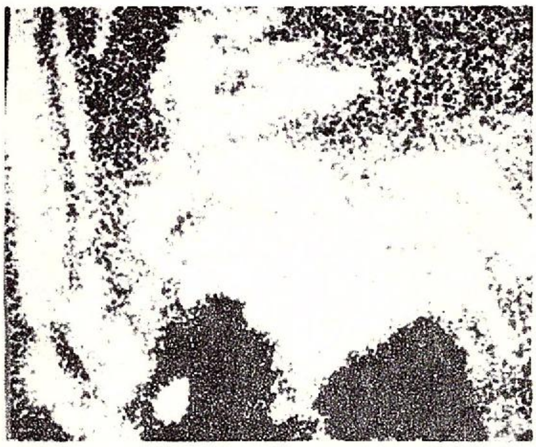

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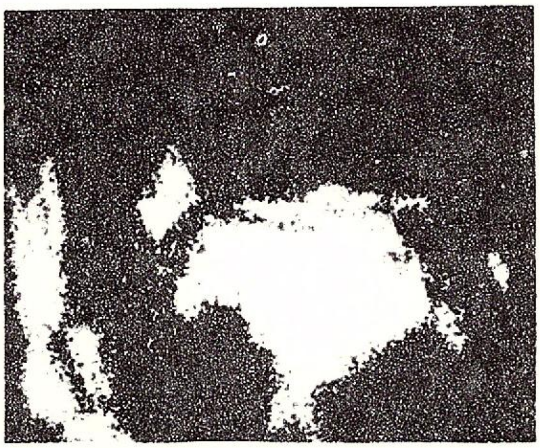

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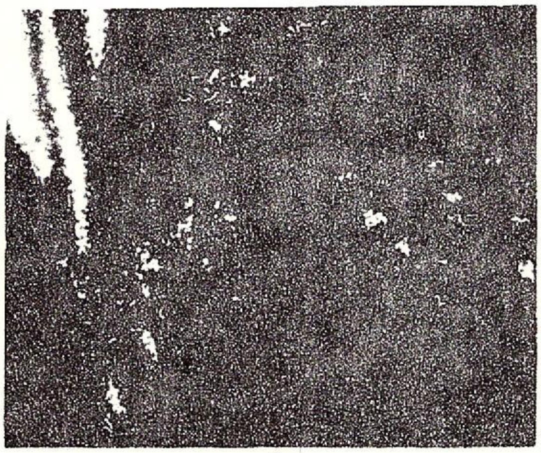

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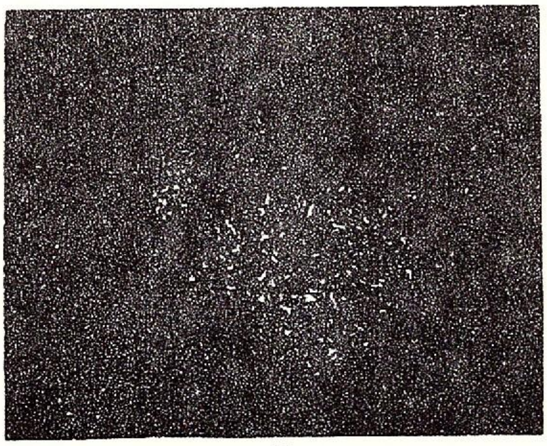

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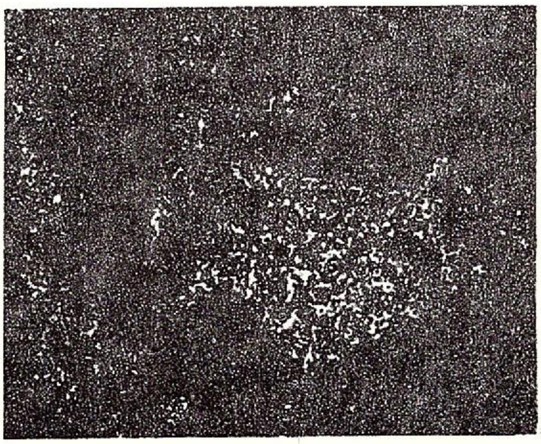

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Planşa I

Fig. 1 – Imagine de compoziţie, x 600; pe – pentlandit; pi – pirotină, cp – calcopirită.
Fig. 2 – Distribuţia Fe, x 600.
Fig. 3 – Distribuţia Ni, x 600.
Fig. 4 – Distribuţia Cu, x 600.
Fig. 5 – Distribuţia Co, x 600.
Fig. 6 – Distribuţia S, x 600.

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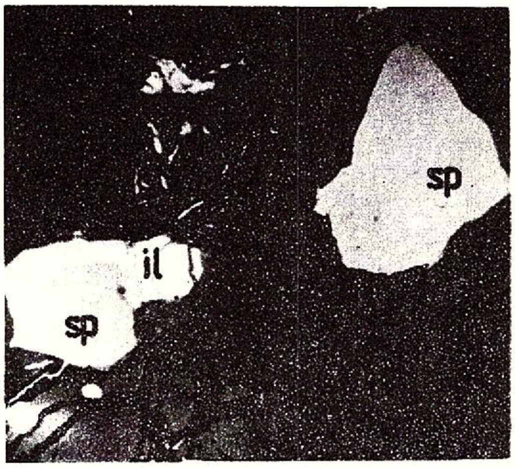

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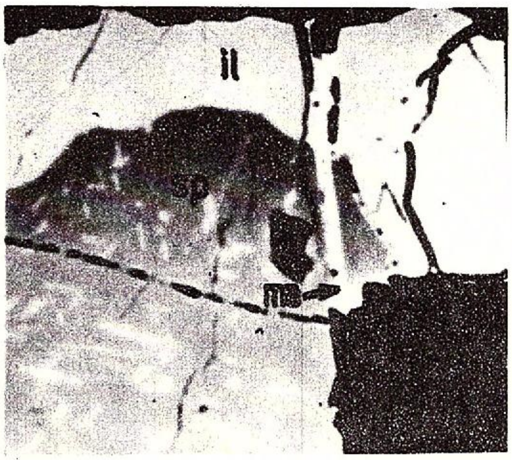

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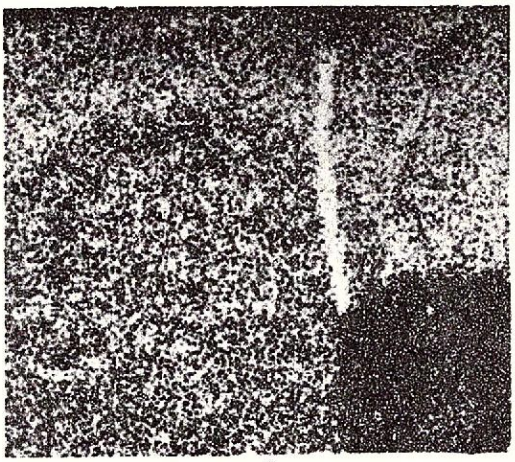

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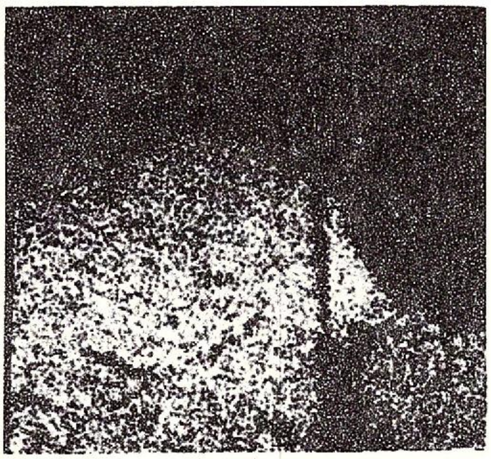

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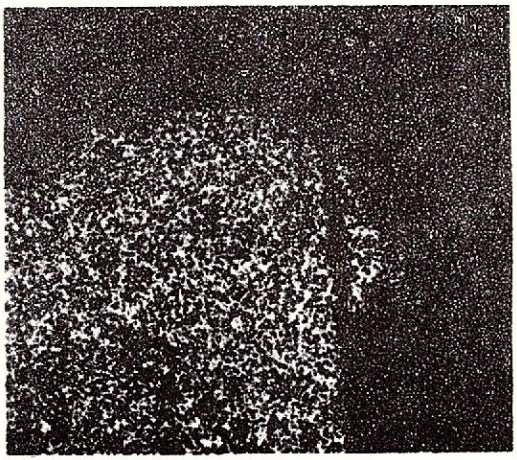

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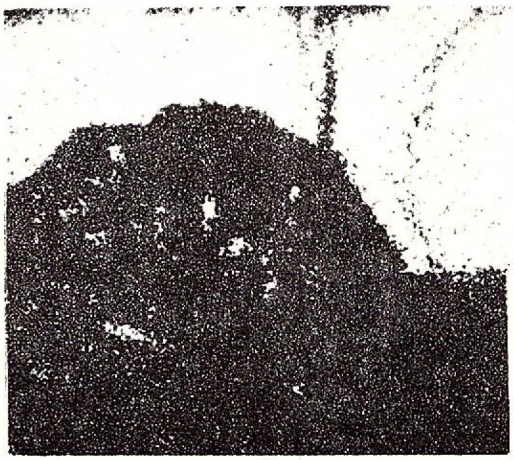

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Planşa II

Fig. 1 – Spinel alumo-cromifer (sp), ilmenit (il) şi magnetit (ma) în ultramafit puţin transformat, x 150.
Fig. 2 – Imagine de compoziţie, x 600; spinel (sp), ilmenit (il), magnetit (ma).
Fig. 3 – Distribuţia Fe, x 600.
Fig. 4 – Distribuţia Al, x 600.
Fig. 5 – Distribuţia Cr, x 600.
Fig. 6 – Distribuţia Ti, x 600.

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Rom. J. Petrology, 1992, 75, p. 201-207

NOTĂ ASUPRA CRISTALINULUI GETIC (GRUPUL DE SEBEŞ–LOTRU) DIN PETECUL DE VĂLARI (CARPAŢII MERIDIONALI)

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Key words: Metamorphic rocks. Polymetamorphism. Retrograde metamorphism. Deformation. Major elements. South Carpathians – Getic and Supragetic Crystalline Domains – Vîlcan Mountains.

Abstract: *The Getic Crystalline (the Sebeş–Lotru Group) of the Vălari Outlier (South Carpathians).* The Vălari outlier is made up of Getic type crystalline schists, which were atributed to two complexes: a lower, quartz-feldspar rocks complex, and an upper, micaceous rocks complex. Due to the intense laminations and the Getic overthrust, a large part of the crystalline turned into mylonites and even ultramylonites. The mineralogical, structural and textural observations prove a Barrovian type metamorphism (the almandine-bearing amphibolites facies), having a polymetamorphic character (there are five moments of metamorphic deformation±blastesis), overlapped by a retrograde metamorphism pointed out by intense laminations.

Scurt istoric

In regiune, primele menţiuni asupra unui "cristalin vechi", care aparţine "grupului I" cu caracter catamezozonal, sînt făcute de Mrazec (1895). Intr-o lucrare de sinteză asupra Carpaţilor Meridionali, prezentată la Congresul Internaţional de Geologie de la Stockholm, 1910, pe o schiţă tectonică, Murgoci indică cristalinul cuprins între Giulava şi Suseni (petecul de Vălari) ca fiind o "frontal region" a Pînzei Getice. Pe harta întocmită de Ionescu-Bujor (1911–1912) sînt figurate două apariţii izolate de cristalin getic: una în Valea Socilor şi alta pe valea Suseni. Ultima este descrisă ca fiind constituită din "roci granitice gnaisice". Streckeisen (1930), într-o lucrare asupra regiunii Vaideei, atribuie gnaisele şi micaşisturile de la Vălari–Giulava cristalinului getic. Manolescu (1937) întocmeşte o reprezentare cartografică şi dă o sumară descriere a şisturilor cristaline din regiunea Vălari, confirmînd opinia lui Murgoci (1905) că Pînza Getică avansează pînă la bordura meridională a munţilor Vîlcan.

Lucrări ulterioare, aparţinînd lui Codarcea (1941), Codarcea et al. (1967), Bercia et al. (1968) şi Pop (1973), fac scurte referiri la existenţa petecului de cristalin de la Vălari. O opinie cu totul contrară o au Savu et al. (1987), care, reluînd o idee mai veche (Popescu-Voiteşti, 1929) şi paralelizînd "filoanele de microdiorite cu amfibol", descrise de Murgoci (1923) în petecul de Vălari, cu cele din seria de Lainici-Păiuş descrise de Berza, Seghedi (1975), le consideră similare şi atribuie petecului de Vălari un caracter de "olistolit ce ar putea proveni din Autohtonul Danubian".

Date petrografice

Petecul de Vălari (fig. 1), cu o dezvoltare de aproximativ 7 km^2 şi o alungire SV–NE, se întinde între Valea Dobriţei la sud-vest şi Valea Şuşiţei la nord-est şi are o lungime maximă de 5,5 km şi o lăţime cuprinsă între 1 şi 2 km. Grosimea expusă a stivei de şisturi cristaline creşte din marginile petecului, unde are doar cîţiva metri, spre partea centrală, unde se estimează a fi cel puţin de ordinul sutelor de metri.

Fondul litologic este reprezentat, ca de altfel în tot grupul de Sebeş–Lotru, prin paragnaise micacee şi micaşisturi, cărora li se asociază, în proporţii variabile, paragnaise cu granaţi ± staurolit ± sillimanit, cuarţite micacee, paragnaise cuarţitice cu muscovit şi biotit, gnaise cuarţo-feldspatice cu biotit, amfibolite, gnaise tonalitice şi mici lentile de pegmatite. O caracteristică esenţială a acestui fond litologic este intensa laminare a rocilor, care, mai ales în părţile marginale ale petecului, conduce frecvent la milonite.

Litostratigrafic se pot distinge doi membri: în bază membrul rocilor cuarţo-feldspatice, iar la partea superioară membrul rocilor micacee, fără a putea pune limite cartografice nete între ele. Distincţia între cei

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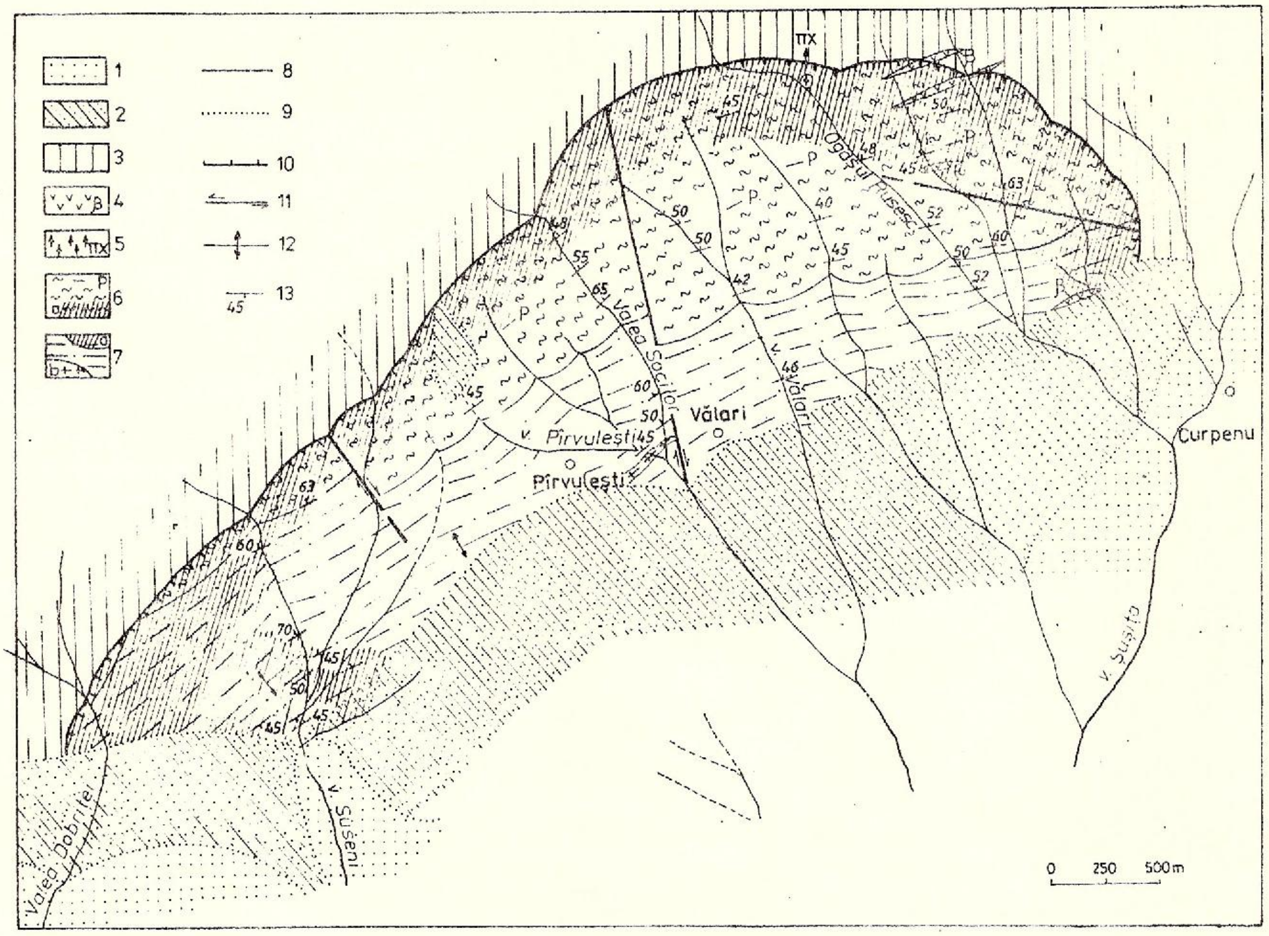

Fig. 1 – Metamorfitele getice de la Vălari.
1, Cuaternar: aluviuni, depozite de terasă; 2, Neogen: nisipuri, pietrişuri marne, calcare marnoase; 3, Cretacic superior: gresii, gresii argiloase, calcare, marno-calcare. Roci magmatice: 4, Mezozoic (Cretacic ?), dolerite; 5, Precambrian, roci ultrabazice (parţial serpentinizate). Cristalinul de Sebeş-Lotru: 6, Membrul rocilor micacee – micaşisturi cu granaţi, paragnaise micacee, paragnaise±staurolit±sillimanit, pegmatite (P), a) zonă cu milonite; 7, Membrul rocilor curţo-feldspatice – amfibolite (b), metatonalite, gnaise cu sillimanit, (a) zonă cu milonite; 8, limită geologică; 9, limită de discordanţă; 10, limita pînzei getice; 11, falie cu decroşare; 12, ax de anticlinal; 13, poziţia stratelor.

doi membri se face greu datorită şariajului getic, care, pe lîngă intensa laminare şi milonitizare, a prins sub unghiuri diferite secvenţele metamorfice în locul de origine, iar în timpul transportului către fruntea pînzei grosimea acestor secvenţe metamorfice a fost redusă tectonic.

Membrul rocilor cuarţo-feldspatice cuprinde gnaise cuarţo-feldspatice, tonalite gnaisice (metatonalite), gnaise cu sillimanit şi amfibolite. Acest complex se dezvoltă cu precădere în partea meridională a peticului de cristalin getic, între văile Suseni şi Socilor, fiind acoperit spre sud de depozite neogene.

Gnaisele cuarţo-feldspatice cu biotit sînt bine dezvoltate în bazinul mediu al văii Suseni (amonte de satul Suseni) şi în raza satului Pîrvuleşti. Sînt roci dure, de culoare cenuşie, uneori cenuşiu-verzuie, datorită cloritizării biotitului. Prezintă textură orientată (şistoasă) şi structură granoblastică. Sînt alcătuite, în principal, din cuarţ, plagioclaz (oligoclaz), microclin şi, în cantităţi reduse, biotit. Frecvent plagioclazul este opacitizat, iar biotitul este parţial cloritizat. După ambianţa litologică, după aspectele structurale şi texturale şi după proiecţia în tetraedrul Niggli a unei analize chimice (tab.), gnaisele cuarţo-feldspatice au caracter de pararoci, totuşi nu excludem posibilitatea prezenţei în stiva de sedimente iniţiale şi a unor nivele tufogene acide.

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Tabel

Compoziţia chimică, parametri Niggli şi valorile normative QAP ale şisturilor cristaline din petecul de Vălari

Proba	5	54	99	164	209
SiO_2	73,14	66,95	47,37	63,63	60,04
TiO_2	0,74	0,86	4,00	0,86	0,86
Al_2O_3	12,02	14,23	13,71	17,38	17,77
Fe_2O_3	0,57	0,68	3,00	1,51	1,43
FeO	3,18	4,34	12,45	3,61	3,61
MnO	0,05	0,07	0,30	0,06	0,07
MgO	1,10	2,40	4,70	1,40	2,50
CaO	2,52	1,40	7,00	2,52	1,82
K_2O	1,15	2,85	0,25	1,95	4,25
Na_2O	3,45	3,10	3,90	3,45	4,20
P_2O_5	0,18	0,18	0,40	0,22	0,21
H_2O^+	1,24	2,10	0,96	2,82	2,72
H_2O^-	0,38	0,32	0,34	0,45	0,45
CO_2					
S	0,10	0,03	0,16		0,05
Total	99,82	100,00	99,93	100,22	100,23
si	392,55	297,73	120,07	267,09	222,63
al	38,01	37,29	20,48	42,99	38,83
fm	25,60	34,58	50,51	26,41	28,78
c	14,49	6,67	19,01	11,33	7,23
alk	21,88	21,44	9,98	19,26	25,15
k	0,179	0,376	0,040	0,271	0,399
mg	0,343	0,459	0,351	0,331	0,479
ti	2,986	2,876	7,625	2,714	2,398
p	0,408	0,338	0,429	0,390	0,329
w	0,138	0,123	0,178	0,273	0,270
qz	205,00	111,93	-19,87	90,87	22,03
c/fm	0,566	0,192	0,376	0,429	0,251
Q	45,77	37,24	0,00	34.46	13,21
A	7,76	21,58	2,75	14,54	31,83
P	46,45	41,16	97,24	50,99	54,94

Tonalitele gnaisice (metatonalitele) apar cu precădere în bazinele afluenţilor de stînga ai văii Suseni, în amonte de satul Suseni şi în raza satului Pîrvuleşti. Au culoare alb-cenuşie şi prezintă textură vag orientată şi structură granoblastică. Sînt constituite din cuarţ, plagioclaz (albit-oligoclaz) şi ortoză, cărora li se asociază, într-o proporţie redusă, micele, îndeosebi biotitul. În secţiuni subţiri este evidentă cataclazarea mineralelor componente, mai ales a cuarţului. Aceste roci au fost descrise de cercetătorii anteriori drept "granite gnaisice". O analiză chimică (tab.) se proiectează în triunghiul QAP la limita dintre cîmpul tonalitelor şi cel al granodioritelor, motiv pentru care redefinim aceste roci drept tonalite gnaisice (metatonalite).

Gnaisele cu sillimanit apar sporadic pe valea Suseni. Prezintă textură orientată şi structură granoblastică. Sînt roci constituite, în principal, din cuarţ şi plagioclaz (oligoclaz), cărora li se asociază, în proporţii reduse, micele şi sillimanitul (var fibrolit). Efectele retromorfismului sînt evidente, ele fiind marcate de cloritizarea biotitului, sericitizarea muscovitului şi apariţia, în secţiuni subţiri, a fibrolitului insular într-o masă de sericit.

Amfibolitele se întîlnesc sub forma unor benzi înguste, de ordinul metrilor, dintre care cea mai bine dezvoltată este în raza satului Pîrvuleşti. Sînt roci verzui-negricioase, cu textură orientată şi structură granonematoblastică. Amfibolitele sînt constituite, în principal, din horblendă verde (cca 70 %), plagioclaz intermediar şi, într-o proporţie mai redusă, cuarţ. Ca minerale accesorii apar titanit, ilmenit, rutil, zircon şi apatit.

Membrul rocilor micacee cuprinde: paragnaise micacee, micaşisturi cu granaţi, paragnaise micacee cu granaţi ± staurolit ± sillimanit, paragnaise cuarţitice cu muscovit şi biotit, cuarţite micacee, amfibolite, pegmatite, milonite. În aria de răspîndire a acestui complex a fost întîlnit, în bazinul superior al Ogaşului Rusesc, un mic corp de peridotite, constituite preponderent din piroxen, parţial serpentinizat. De asemenea, în partea nord-estică a petecului de Vălari apar cîteva corpuri de dimensiuni reduse, ce par a fi dyke-uri, de roci bazice (dolerite). Raporturile acestor roci bazice cu cristalinul getic şi cu depozitele Cretacicului superior nu sînt clare.

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Paragnaisele micacee şi micaşisturile sînt roci cu granulaţie medie pînă la faneroblastică şi cu textură pronunţat şistoasă, dată de dispunerea orientată mai ales a micelor, dar şi a feldspatului şi cuarţului. La constituţia lor mai participă granatul (2-3 %), sub forma unor cristale de dimensiuni mici. Cristalele mari de granat sînt frecvent sparte şi întinse pe planele de laminare. Feldspatul, de obicei un oligoclaz cu 20 % An, apare în mod obişnuit argilizat. Biotitul se menţine proaspăt doar în zonele mai puţin afectate de laminare, în rest este adesea parţial cloritizat. În zonele de laminare mai intensă muscovitul trece în sericit, granatul, ca şi biotitul, se cloritizează, iar feldspatul suferă carbonatări şi argilizări.

Paragnaisele cuarţitice cu muscovit şi biotit şi cuarţitele micacee se deosebesc de paragnaisele micacee doar printr-o participare procentuală foarte redusă a feldspatului (cca 5 %). Structura lor este granolepidoblastică, iar textura orientată.

Paragnaisele micacee cu granaţi ± staurolit ± sillimanit se deosebesc de paragnaisele micacee prin prezenţa constantă a granaţilor şi prin apariţia staurolitului şi sillimanitului (var. fibrolit). Sillimanitul, identificat în petecul de Vălari pentru prima dată de Bercia, Hârtopanu (1980), apare sporadic sub forma unor mici "mănunchiuri" de fibre dispuse într-o masă de sericit. Staurolitul a fost identificat de noi şi se prezintă, de obicei, ca relicte (mici insule) într-o masă de sericit. Cristalele de staurolit proaspăt, cazuri foarte rare, au dimensiuni reduse. Uneori, în secţiunile subţiri cu relicte de staurolit în mase globuloase de sericit, apar şi aglomerări de sericit de formă dreptunghiulară sugerînd pseudomofoze după disten. Hârtopanu (informaţie verbală) consideră că unele agregate pinitice, după caracterele fiziografice, ar putea aparţine unui fost cordierit. Urmează ca cercetările de viitor să confirme sau să infirme această supoziţie, care are şi implicaţii petrogenetice vizînd îndeosebi condiţiile de presiune ale metamorfismului. Oricum, poziţia staurolitului este metastabilă în asociaţie cu sillimanitul.

Amfibolitele se întîlnesc foarte rar, ca intercalaţii de dimensiuni reduse (decimetrice) în bazinele văilor Socilor, Pîrvuleşti şi Răchita. Sînt roci masive constituite din hornblendă comună, plagioclaz (oligoclaz bazic-andezin) şi cuarţ. În cantităţi reduse apare şi granatul. Ca minerale accesorii se întîlnesc ilmenit, titanit, apatit, uneori zircon şi, foarte rar, rutil.

Pegmatitele apar ca mici budine de segregaţie metamorfică, mai ales în paragnaisele micacee şi în micaşisturi. În mod obişnuit sînt constituite din cuarţ, microclin, plagioclaz şi muscovit. O mică budină de pegmatite, interpretată drept filon de Manolescu (1937), ce apare în vecinătatea corpului de amfibolite de la confluenţa Văii Socilor cu valea Pîrvuleşti, oferă o asociaţie mineralogică cu structuri poikilitice între hornblendă şi plagioclaz.

Milonitele. În părţile marginale ale petecului de Vălari – şi cu siguranţă la baza cristalinului getic din acest petec – fenomenul de milonitizare s-a manifestat cu intensităţi variate, gradul de milonitizare, mai intensă sau mai slabă, depinzînd de fapt de tipul de rocă afectat. Milonitizarea deferitelor tipuri de paragnaise şi micaşisturi este progresivă şi poate fi urmărită pe teren de la un termen la altul, pînă la ultramilonite. Fenomenul, într-un stadiu incipient, se trădează la microscop prin apariţii de lamine micacee, extincţie ondulatorie a cuarţului, apariţia de striaţii şi fisuri, mai ales la cristaloblastele de cuarţ. Pe măsură ce creşte intensitatea laminării, efectele ei devin sesizabile şi pe teren. Într-un stadiu mai avansat milonitele devin compacte "afanitice" şi se mai pot observa rareori relicte extrem de mici de porfiroclaste de cuarţ şi plagioclaz, la paragnaise, şi de hornblendă la amfibolite. Micele sînt transformate în sericit şi clorit, hornblenda în clorit, iar dispunerea mineralelor secundare se face în lungul planelor de laminare. Cînd laminarea devine foarte pronunţată, nu se mai poate distinge macro- sau microscopic nici un mineral şi roca se defineşte drept ultramilonit. De menţionat efectele şariajului getic şi în depozitele cretacice peste care este şariat cristalin de Sebeş–Lotru. Ele se sesizează mai ales prin laminări în zona de contact.

Date chimice

Pentru obţinerea unei imagini asupra chimismului şisturilor cristaline din petecul de Vălari au fost executate cinci analize chimice (tab., analist Carmen Popescu), trei pentru complexul rocilor cuarţo-feldspatice (nr. 5 – tonalit gnaisic, valea Suseni; nr. 54 – gnais cuarţo-feldspatic cu biotit, valea Suseni; nr. 99 – amfibolit, valea Pîrvuleşti) şi două pentru complexul rocilor micacee (nr. 164 – paragnais micaceu, valea Răchita; nr. 209 – paragnais micaceu, Ogaşul Vălari). După datele analitice, parametrii Niggli şi valorile QAP (fig. 2a), rezultă că ceea ce a fost descris de Manolescu (1937) drept "granite gnaisice" sînt de fapt tonalite gnaisice (proba nr. 5 se proiectează în diagrama QAP la limita cîmpului tonalitelor cu cel al granitoidelor, suma alcaliilor fiind redusă), cu tendinţă trondjemitică (cantitatea foarte mică de femice şi cu o cantitate foarte mare de silice, 73,14 %). Gnaisele cuarţo-feldspatice se individualizează printr-o participare ceva mai ridicată a alcaliilor (K_2O – 2,85 %) şi a femicelor (FeO – 4,34 % ; MgO – 2,40 %), dar se înscriu în limitele normale ale acestui tip de

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rocă din cristalinul getic, ca şi amfibolitele (analiza 99). În tetraedrul Niggli (fig. 2b) primele se proiectează în cîmpul rocilor sedimentare, iar ultimele în cîmpul rocilor eruptive. La paragnaisele micacee, din complexul rocilor micacee, se constată o participare ceva mai ridicată a alcaliilor (îndeosebi analiza 209 cu 4,25 % K_2O şi 4,20 % Na_2O) faţă de normalul acestui tip de roci în cristalinul getic (Gridan, 1981), fapt ce se explică printr-o compoziţie iniţială mai alcalină a metapelitelor. Pentru cele două analize de paragnaise micacee, parametrii Niggli sugerează o origine sedimentară.

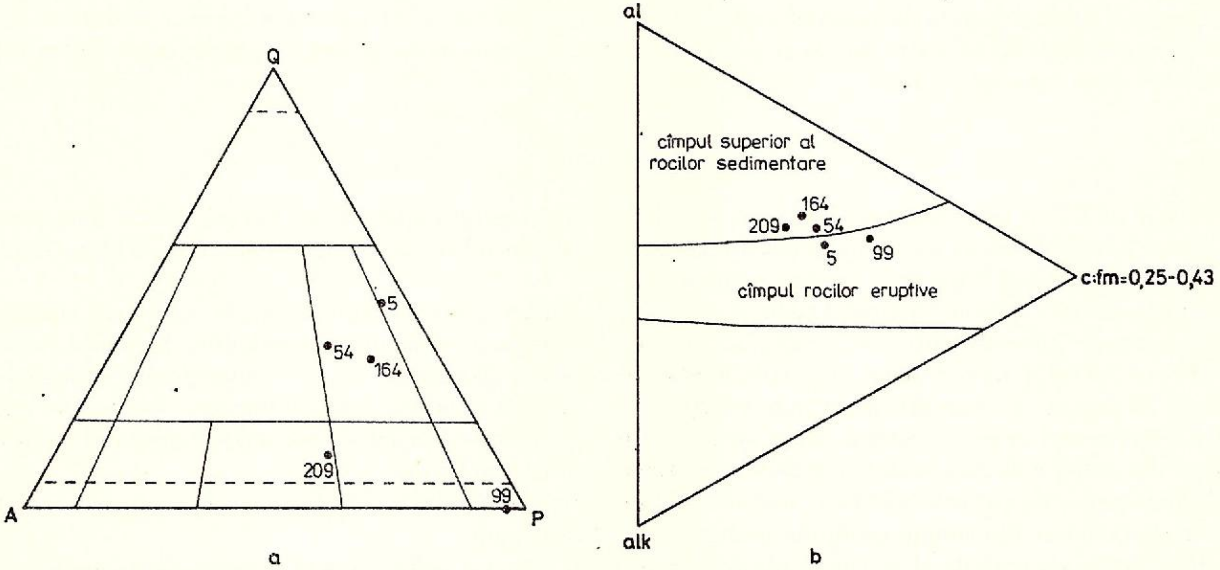

Fig. 2 - Diagrame ternare.
a, diagrama QAP; b, diagrama Niggli al-alk-c:fm;
5, tonalit gnaisic; 54, gnais cuarţo-feldspatic cu biotit; 99, amfibolit; 164, paragnais micaceu; 209, paragnais micaceu.

Date structurale şi petrogenetice

Observaţiile de teren şi cele microscopice, mai ales din paragnaisele micacee şi micaşisturi, oferă suficiente date structurale şi texturale, precum şi date legate de modul de prezentare al asociaţiilor mineralogice, pe baza cărora putem recunoaşte în cristalinul getic din petecul de Vălari un caracter polimetamorf cu următoarele momente metamorfice:

– M_1, cu deformare-blasteză, trădat în prezent doar de incluziunile de feldspat, cuarţ, rutil, ilmenit şi/sau magnetit din feldspat şi din granaţii foliaţiei S_2;

– M_2, cu deformare-blasteză, în condiţiile de metamorfism ale faciesului amfibolitelor cu almandin, ducînd la formarea staurolitului, distenului (?) şi sillimanitului. Dintre mineralele principale "reformate" acum, care dau foliaţia principală S_2, o menţiune aparte pentru un biotit (bi_1) vizibil în prezent în microlitoane transpuse în S_3;

– M_3, cu deformare şi blasteză de biotit (bi_2) în porfiroblaste cu dimensiuni pînă la 5 mm. Micele de neoformaţie, de obicei fengit, prezintă incluziuni de feldspat şi granaţi, ce apar ca granule înşirate pe anumite direcţii, care formează un unghi de 60–70° cu planele de clivaj ale fengitului.

Interesant este faptul că aceste direcţii de dispunere a incluziunilor sînt cvasiparalele cu foliaţia S_2. Legată de acest moment metamorfic este apariţia foliaţiei S_3, mai ales ca rezultat al transpunerii mineralelor din foliaţia S_2:

– M_4, cu caracter retromorf. De acest moment se leagă apariţia cloritului pe seama biotitului, granaţilor şi amfibolilor, precum şi apariţia carbonaţilor pe seama feldspatului;

– M_5, cu deformare-laminare şi realizarea microcutelor din păturile micacee ale foliaţiei principale S_2 şi cu apariţia seturilor de plane de laminare.

Aspectele litologice ale şisturilor cristaline din petecul de Vălari ne determină să considerăm că fondul petrografic al acestora provine dintr-o stivă sedimentară cu caracter pelitic, dar şi cu unele secvenţe psamitice.

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Conţinutul mineralogic al metasedimentelor este: cuarţ + plagioclaz (oligoclaz) + muscovit + biotit$_1$ ± granat ± staurolit ± disten(?) ± sillimanit ± biotit$_2$ + clorit ± calcit ± minerale argiloase.

Pentru intercalaţiile subţiri, decimetrice pînă la metrice, de amfibolite în masa paragnaiselor micacee, doar pe baza aspectelor litologice, mineralogice şi structurale (interstratificaţii, prezenţa cu caracter accesoriu a hornblendei în paragnaisele micacee vecine), este mai greu să acceptăm un caracter "para" al acestor tipuri de roci, mai ales că proiecţia lor în diagrama Niggli se face în cîmpul rocilor eruptive. Menţionăm că nu avem nici un indiciu asupra prezenţei unor nivele tufogene bazice în stiva de sedimente iniţiale, dar suprafaţa redusă a petecului de Vălari invită la circumspecţie.

In ceea ce priveşte tonalitele gnaisice (metatonalitele), aspectul masiv şi foliaţia metamorfică slab pronunţată, caracterele mineralogice structurale şi texturale, cît şi chimismul pledează pentru provenienţa acestor roci din vechi corpuri granitoidice.

Concluzii

Petecul de Vălari este constituit din şisturi cristaline de tip getic (grupul de Sebeş–Lotru), care se încadrează în faciesul amfibolitelor cu almandin şi pot fi repartizate la doi membri: membrul rocilor cuarţo-feldspatice, în bază, şi membrul rocilor micacee, la partea superioară.

Primul membru, întîlnit în arealul sudic al petecului de Vălari, este constituit din gnaise cuarţo-feldspatice, tonalite gnaisice (metatonalite, vezi analiza nr. 5, tab. 1), gnaise cu sillimanit şi amfibolite. Membrul rocilor micacee se dezvoltă cu precădere în părţile nordice şi cuprinde paragnaise micacee, micaşisturi, paragnaise micacee cu granaţi ± staurolit, paragnaise cuarţitice cu muscovit şi biotit, cuarţite micacee, amfibolite, pegmatite. In arealul acestui membru apare un mic corp de peridotite parţial serpentinizate, precum şi cîteva corpuri discordante de roci bazice (dolerite) ce secţionează cristalinul getic.

O mare parte din şisturile cristaline, mai ales cele ale complexului superior, au fost transformate de intensele mişcări de laminare din timpul şariajului în milonite şi ultramilonite.

Observaţiile structurale şi texturale pledează pentru un caracter polimetamorf al şisturilor cristaline. Se disting cinci momente metamorfice: primele patru (M_1, M_2, M_3 şi M_4) cu deformare şi blasteză, iar ultimul (M_5) cu deformare-laminare.

Bibliografie

Bercia I., Marinescu Fl., Mutihac V., Pavelescu M., Stancu J. (1968) Notă explicativă la harta geologică, scara 1:200.000, Foaia Tg.-Jiu (33), Bucureşti.

– , Hârtopanu I. (1980) Domaines à basse préssion dans la série de Sebeş–Lotru (Précambrien de la Nappe Gétique). *An. Inst. Geol. Geofiz.*, LVII, p. 297–303, Buureşti.

Berza T., Seghedi A. (1975) Asupra prezenţei distenului în complexul amfibolitic al seriei de Drăgşan din bazinul Motrului. *D. S. Inst. Geol. Geofiz.*, LXI/1, Bucureşti.

Codarcea Al. (1941) Contributions à la tectonique des Carpathes Méridionales. *C. R. Séances (1936–1937)*, XXV, p. 156–159, Bucureşti.

– , Bercia I., Boldur C., Constantinof D., Maier O., Marinescu Fl., Mercus D., Năstăseanu S. (1967) Geological Structure of the South-western Carpathians. Intern. Geol. Congr., XXIII Session, Prague 1968, Guide to excursion 49 AC, Bucureşti.

Gridan T. (1981) Petrologia Semenicului de nord-est. Ed. Acad. R.S.R.,291 p, Bucureşti.

Ionescu–Bujor D. (1911) Granitul de Şuşiţa. *An. Inst. Geol. Rom.*, V, p. 207–211, Bucureşti.

Manolescu G. (1937) Etude géologique et pétrographique dans les Monts Vîlcan. *An. Inst. Geol. Rom.*, XVIII, p. 79–172, Bucureşti.

Mrazec L. (1895) Contribuţiuni la studiul petrografic al rocilor din zona centrală a Carpaţilor de Sud, şi anume din judeţele Mehedinţi, Gorj şi Muscel. *An. Mus. Geol. Pal.*, 1894, I, Bucureşti.

– (1904) Sur les schistes cristallins des Carpathes Méridionales (versant roumain). C. R. IX-ième Congr. Intern. Géol. (1903), p. 635–648, Wien.

Murgoci G. M. (1923) Sinteza geologică a Carpaţilor de Sud. *D. S. Inst. Geol. Rom.*, I, p. 48–58, Bucureşti.

Pop Gr. (1973) Depozitele mezozoice din Munţii Vîlcan. Edit. Acad. R.S. România, p. 1–155, Bucureşti.

Popescu–Voiteşti I. (1929) Aperçu synthétique sur la structure des régions carpathiques. *Rev. Muz. Geol. Mineral.*, III, 1, Cluj.

Savu H., Udrescu C., Lemne M., Romanescu O., Stoian M., Neacşu V. (1987) Island Arc Volcanics Related to the Wildflysch on the Outer Margin of the Danubian Autochthon (Southern Carpathians) and Their Geotectonic Implications. *Rev. roum. géol., géophys., géogr., Géologie*, 31, p. 19–27, Bucureşti.

Institutul Geologic al României

Streckeisen A. (1930) Profilul de la Vai de Ei (Carpaţii Meridionali, jud. Gorj). *D. S. Inst. Geol.*, XVII, p. 93-101, București.

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– , Pavelescu L. (1954) Contribuţii la studiul cristalografic al mineralelor din zăcămîntul de la Muşca. Comm. Acad. Rom., IV, 11–12, p. 685–991, Bucureşti.

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