MINISTERUL MINELOR,
PETROLULUI ȘI GEOLOGIEI
INSTITUTUL DE GEOLOGIE ȘI GEOFIZICĂ

ANUARUL INSTITUTULUI de GEOLOGIE și GEOFIZICĂ

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Petrographic Investigation of Pegmatites Located between Teregova and Marga (Eastern Banat, South Carpathians)

Ș. VELICIU:

Geothermics of the Carpathian Area

VOL. 67 BUCUREȘTI – 1987

Couverture: C. Vasile

Les auteurs assument la responsabilité
des données publiées

MINISTERUL MINELOR, PETROLULUI ȘI GEOLOGIEI
INSTITUTUL DE GEOLOGIE ȘI GEOFIZICĂ

ANUARUL INSTITUTULUI DE GEOLOGIE ŞI GEOFIZICĂ

VOL. 67

BUCUREȘTI
1987

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Ilustrația : V. NIȚU
Traducător : RUXANDRA NEGREA

Dat la cules : iunie 1987. Bun de tipar : sept. 1987. Tiraj : 750 ex. Hîrtie scris IA 70x100/56 g. Coli de tipar 7 1/4. Comanda 231. Pentru biblioteci indicele de clasificare 55 (058)

Întreprinderea Poligrafică „Informația" str. Brezoianu nr. 23—25, București — România

PETROGRAPHIC INVESTIGATION OF PEGMATITES LOCATED BETWEEN TEREGOVA AND MARGA (EASTERN BANAT, SOUTH CARPATHIANS)[1]

BY

HORST PETER HANN[2]

Abstract

The occurrence of the pegmatite bodies is related to highly migmatized metamorphic terrains. Several genetic types are defined by the study of the shape of pegmatite bodies and of the relations to the adjoining rocks, the classification of the contact types and the interpretation of peripheral mafic accumulations. The study of zoning points to the succession of different stages of pegmatitic minerals genesis. The graphic textures are of metasomatic origin. The geochemistry of pegmatites shows the simultaneous development of the former and of adjoining paragneisses. The rare mineral content is low. The pegmatites originate in different petrologic processes, mostly of anatectic-metasomatic nature, within the metamorphic field.

Résumé

L'étude pétrologique des pegmatites situées entre Teregova et Marga (le Banat Oriental). La présence des corps pegmatitiques est reliée aux terrains métamorphiques intensément migmatisés. L'étude de la configuration des corps pegmatitiques et des relations avec les roches environnantes, la classification des types de contact et l'interprétation des concentrations mafiques périphériques indiquent la coexistence de plusieurs types génétiques. La zonation informe sur la succession des étapes de la genèse des minéraux pegmatitiques. Les structures graphiques sont d'origine métasomatique. La géochimie des pegmatites caractérise l'évolution simultanée avec celle des paragneiss environnantes. La teneur en minéraux rares est réduite. Les pegmatites se sont formées par des processus pétrologiques distincts, pour la plupart d'origine anatectique-métasomatique, dans le domaine métamorphique.

[1] Thesis of doctor's degree held on November 7, 1983 at the University of Bucharest; paper received on December 3, 1983 and accepted for publication on January, 1984.

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1. INTRODUCTION

The pegmatite investigation on Romanian territory is restricted to minor occurrences in the East Carpathians (Rodna Mts) and in the Apuseni Mts (Muntele Mare, Preluca Mts), and to a very large area in the South Carpathians wherein mesometamorphic rocks assigned to the Sebeș-Lotru Group contain numerous pegmatite mobilisates. Thus, the Teregova-Marga region containing many pegmatite bodies is of major importance.

The present study required, during the early stage, a revision of all geological investigations carried out by previous geologists in this area. Then, the different types of crystalline schists hosting pegmatite bodies and the problems regarding their structure and lithostratigraphy were taken into account. The relationships between the metamorphism grade, the migmatite textures and the numerous pegmatite bodies are also worth studying and will be treated further on.

The shapes of pegmatite bodies, their systematics, the classification of different types of contact zones between pegmatites and surrounding rocks, the description and discussion of the significance of mafic concentrations at the periphery of the pegmatite bodies are relevant for the interpretation of pegmatite genesis.

The investigation of the inner structure of pegmatite bodies accounts for the presence or absence of zoning and for its formation. The description of different types of zoning should be based on data concerning the grain size of minerals, the relative amounts of major and accessory pegmatite-bearing minerals and the occurrence of graphic texture. The study of the latter is a complex one — petrographic and geochemical — and is important with respect to their origin. Structural information reconstitute the sequence of pegmatite mineral formation and the thermobaric conditions which controlled their crystallization.

Concerning the pegmatite mineralogy the present author is interested in investigating both their major (feldspars, micas, quartz) and accessory minerals (tourmaline, beryl, garnet). The relationships between minerals, the study of the physico-chemical conditions are of peculiar interest. The geochemical characterization of minerals is also worth considering in order to complete the study; the mineralogical-geochemical data are also used as discriminating criteria of pegmatite genesis.

The geochemical study of pegmatites implies the interpretation of bulk analyses for defining the geochemical evolution and the relations between pegmatites and surrounding metamorphic rocks, which provide information on pegmatite genesis.

The pegmatite genesis is based on data concerning the morphostructural features of pegmatites and the interpretation of the mineralogic-geochemical investigation. Such analysis is an attempt to establish the origin of pegmatites from the investigated area, which represents in fact the purpose of the present study.

2. PREVIOUS STUDIES AND EVOLUTION OF IDEAS CONCERNING THE PEGMATITE BODIES

The oldest data on the pegmatite occurrences in the Armeniş-Sadova Veche region were provided at the end of the 19th century (Schafarzik, 1898). The author described pegmatite lenses and tabular bodies at Armeniş containing : potash feldspar, muscovite, quartz, garnet and light-green apatite prisms, 3 cm long. At Sadova he described three pegmatite veins which cut the crystalline limestones ; they contain perthite, oligoclase, quartz, black tourmaline nests, biotite ; feldspar and quartz form graphic intergrowths. The pegmatites exhibit zoned, symmetrical texture : tourmaline crystals surrounded by quartz in the inner zone, large size microcline, quartz, biotite crystals outwards, whereas both contact zones are marked by green tremolite stripes. These pegmatites deposited from former hot springs. This view on the pegmatite genesis was commonly accepted at that time. Schafarzik does not accept these pegmatites to be formed as a result of magmatism or lateral secretion.

Later on, in 1931, Dittler and Kirnbauer studied the pegmatite veins at Teregova and described some new mineral occurrences, that is beryl and columbite ; the pegmatites are considered of igneous origin.

The geological literature on pegmatites is rather sparse until 1951, when Gherasi presented a report on mica-bearing pegmatites from the Voislova-Măru region. The pegmatites are considered to represent lens-like bodies concordant with the surrounding crystalline schists. They contain orthose, microcline, oligoclase, quartz, muscovite, biotite, tourmaline, garnet, beryl.

Following this, several prospections were carried out in order to discover pegmatite bodies of economic interest. Thus, Avramescu (1954) described the pegmatite occurrences from the Luncaviţa-Teregova-Armeniş-Dalci zone. He described zoned textures and provided seven chemical analyses. The pegmatites bearing abundant plagioclase feldspar are considered of metamorphic origin, while those with prevailing potash feldspar and rare minerals are considered granitic pegmatites. Prospections were carried out by Roşca (1954), Micşa and Gall (1956), Marinescu and Ardeleanu (1956), Codarcea and Stoenescu (1957), Mînzatu and Mînzatu (1957, 1958), who described in detail the shape of pegmatite bodies and the relations between pegmatites and surrounding rocks. Zoned textures, mineralogical and geochemical features are discussed based on 13 chemical analyses.

Superceanu (1957) published a paper on the rare minerals contained by the pegmatites of the Banat region. Besides beryl and tourmaline, columbite, tantalite and montebrasite are described. Pegmatites are considered of granitic origin. All these data were later taken over by Schneiderhöhn (1961). Grosu and Angelescu (1960) described the zoning of some pegmatites in this area.

Savu and Micu (1964) described pegmatite veins and lenses in the Armeniş-Slatina-Timiş-Petroşniţa area and divided them into muscovite, biotite, muscovite and biotite pegmatites, feldspar pegmatites and tourmaline pegmatites. Suru (1966) presented a well documented report on pegmatites, mostly already exploited, including detailed mineralogic descriptions and problems concerning their zoning. These pegmatites are considered of magmatic origin. He stated that pegmatites with major potash feldspar contents and those characterized by graphic textures — potash feldspar and quartz — exhibit a low muscovite content.

Gherasi and Zimmermann (1968) mentioned concordant and discordant pegmatites occurring in the Getic Crystalline north of Muntele Mic, considering them of different ages (that is the discordant bodies are more recent).

Pomârleanu and Movileanu (1968) determined the temperature of 338—408°C for the muscovite formation by using the muscovite-paragonite diagram.

Deaş et al. (1968) presented a report of economic interest concerning the geological study of mica-bearing pegmatites, carried out during the 1961—1967 period of time.

According to Savu (1970) the sillimanite zone and the disthene zone contain the major bulk of the pegmatite bodies. Concordant pegmatites are mica-abundant and were formed *in situ*, while discordant bodies are feldspar-abundant, in association with rare minerals, and resulted from pegmatoids migrated from deep-seated sources.

Marinescu et al. (1973) described pegmatite occurrences from this region with emphasis on the chemistry of pegmatites.

The pegmatite occurrences from Teregova-Luncaviţa area are the object of a prospecting report (Hann, 1973) including the description of pegmatite shape, different types of relation between pegmatites and the host rock, their inner structure. The pegmatite genesis is interpreted in terms of secretion processes during metamorphic differentiation of rocks during almandine amphibolite facies.

Gherasi et al. (1974) divided the pegmatite occurrences north of Muntele Mic into muscovite-bearing calc-alkaline pegmatites and quartz-feldspar pegmatites. The largely crystallized tourmaline of pegmatites assigned to the muscovite-bearing group is considered as a result of pneumatolytic activity. The quartz-feldspar pegmatites are considered to result from lateral secretion since the plagioclase (An_{14}) of their marginal micropegmatite zone and the plagioclase of the host rock are similar.

The zoned textures, asymmetrical in places, of discordant pegmatites found in the metamorphic rocks of the Getic Nappe from the area situated north of Muntele Mic, and the different mineralogic features of pegmatites were described by Hann (1976).

Savu (1977) described several textural and petrographic characteristics of pegmatite occurrences from the Banat region. Taking into account the chemical analyses listed by Avramescu (1954) and Mânzatu, Mânzatu (1957, 1958) the author made some petrochemical and genetic comments. Thus, pegmatites are supposed to be formed by anatexis, on the level of sillimanite zone and disthene zone, at a temperature of 600—700°C.

Considerations on the evolution of pegmatitisation, the succession of different mineralogic types and the geochemical evolution were made by Hann (1977).

Pomârleanu and Movileanu (in Hann et al., 1977) gave the following geothermometrical data : 554—600°C for microcline, 560—585°C for biotite, 295—560°C for muscovite, 210—470°C for quartz and 237—470°C for beryl.

Gridan (1981) described the pegmatite occurrences from the north-eastern area of the Semenic Mts and their unhomogeneous mineralogic and structural features. According to this author the pegmatites are of quartz-feldspar and micaceous types; he found no evidence of their magmatic origin.

Diaconu (1979) reported exhaustive prospecting and exploration data in order to promote the discovery of other muscovite or feldspar-bearing pegmatite bodies.

3. LOCATION AND DESCRIPTION OF MAJOR PEGMATITE BODIES

The major pegmatite occurrences are found north of Muntele Mic, in the Tîlva, Măgura and Dalci-Var areas and in the Semenic Mts, that is Slatina-Timiș, Armeniș and Teregova areas (Fig. 1).

I. *Tîlva area.* The pegmatite bodies occur north of Măru and south of Voislova in the Curcan and Fața Lungă hills, in the basins of Slatina, Valea Mare, Plopi, Pîrîul Pascului valleys. The various pegmatite bodies have been entirely or partly mined out for muscovite and beryl, and partly for feldspar (e.g. the so-called "main body" — vein 1 Tîlva, situated north-west of Curcan hill). Other pegmatite bodies were exploited north of the Curcan hill or in the Fața Lungă hill — „Rîpi" pegmatite body. The pegmatites from this area are tourmaline-rich, too. Some pegmatite dykes are discordant (e.g. the body 1 Tîlva and another body situated at ca 200 m southwards) and most of the bodies are concordant and lens-like shaped.

II. *Măgura area.* West-southwest of Măgura pegmatite occurrences are frequent in the Pietroasa Valley basin, a left tributary of the Bistra Mărului Valley. An important body — „Cîrniș" (180 m long, 20 m wide) — cuts the main valley and is still mined. Discordant and beautifully zoned pegmatite veins occur in the right tributary of Pietroasa Valley-Pîrîul cu Mărul. Other bodies are located along the Ogașul Strîmb brook and Iedera Valley, right-side tributaries of the Pietroasa Valley. Several pegmatite bodies, usually concordant, occur in the Socetul Mare summit, south of Pietroasa Valley and north of Șasa Valley, left-side tributary of Bistra Mărului Valley. In this area, the pegmatite lenses may form elongated elevations, prominent due to their high erosion resistance. Pegmatite bodies lie in the Cermaz Valley, left-side tributary of the Șasa Valley, near its confluence with Bistra Mărului Valley. Concordant and discordant pegmatites, completely or incompletely zoned, or unzoned in places are also present. Pegmatite occurrences are reported at Dobrotin, Runcurelu Mare and Runcurelu Mic valleys, left-side tributaries of Bistra Mărului Valley, south of Pietroasa-Bistra Mărului confluence. The western margin of the Măgura area contains several lens-like, concordant pegmatite bodies along the upper course of the Scoarța Valley.

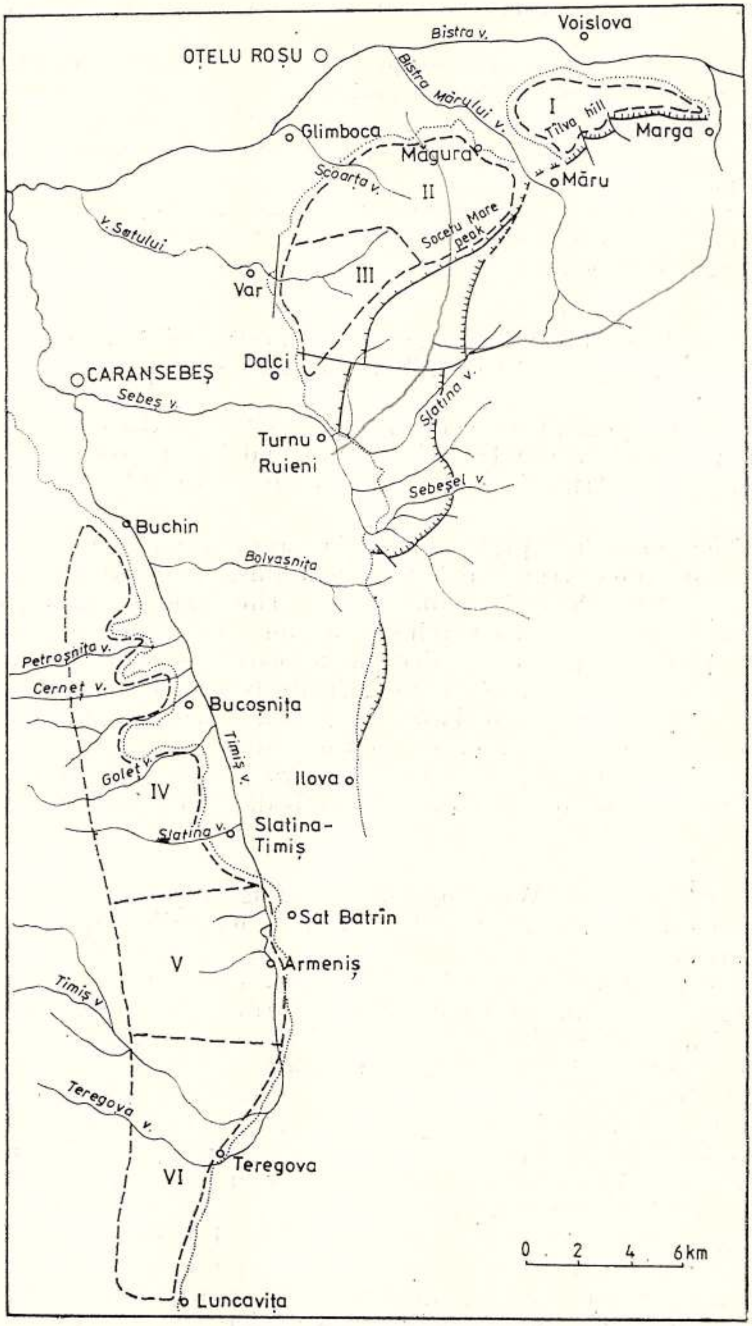


Fig. 1. Localisation of pegmatites in Teregova—Marga area, I, Tîlva area; II, Măgura area; III, Dalci-Var area; IV, Slatina Timiș area; V, Armeniș area; VI, Teregova area.

The muscovite contents of the pegmatite occurrences from the Măgura area are of economic interest. The bodies located along Runcurelu Mare and Runcurelu Mic valleys and the upper course of Scoarţa Valley are rich in potash feldspar.

III. *Dalci-Var area.* The bodies lying east and north-east of Var and Dalci localities occur in the basin of the Văruţ Valley, in Valea Satului and in its left-side tributary, Străului brook, in Dalci Valley and in Găina Mică brook, between Dalci and Var. The pegmatite bodies represent the south-western extension of Pietroasa Valley-ScoarţaValley portion from the Măgura area. These pegmatites are commonly concordant and have rather high muscovite and feldspar contents. Muscovite was exploited in Valea Satului and Străului brook, whereas feldspar in Dalci Valley.

IV. *Slatina-Timiş area.* It is located in the eastern part of the Semenic Mts. The pegmatite bodies occur south of Slatina-Timiş village in the Secaş brook, and then northwards, in the Slatina Valley, Goleţ Valley and Bucoşniţa brook. All these valleys are left-side tributaries of the Timiş Valley. An important body crops out in the Slatina Valley, ca 1.5 km off its confluence with the Timiş Valley. Other pegmatite occurrences are mentioned 600m upstream, in the left slope of the Slatina Valley, the main pegmatite bodies occur at 4.5—5 km off its confluence with the Timiş Valley. These pegmatites are characterized by their high feldspar or quartz contents.

V. *Armeniş area.* The pegmatite bodies occur both north and south of Armeniş along the Timiş Valley. To the north, important occurrences are mentioned up to Sat Bătrîn, in the right slope of the Timiş Valley, and to the south up to Valea Mare, near the tunnel, in the left slope of the Timiş Valley. Pegmatite occurrences are also reported in the Biban brook and Frîncu brook, left-side tributaries of the Timiş Valley. The most important body is situated near the Armeniş railway-station and crosses the Timiş Valley. It is the most important body of the region and one of the largest all over the country (550 m long, 40 m wide). It is mined for feldspar only in the right slope of the Timiş river where it crops out on 420 m; the open-pit working has developed along 340 m since 1954. The concordant body trends mainly NW—SE. It seems that, at least partly, it is delimited by disconformity from the adjoining gneisses.

The pegmatites do not exhibit large muscovite crystals, are mostly concordant and are mined or might be mined for feldspars.

VI. *Teregova area.* Pegmatite bodies are found north, south and west of Teregova locality. North-west of Teregova village, in the left slope of Teregova Valley, there are two large discordant and well zoned dykes mined intermittently for „glaze" potash feldspar since the beginning of the century till 1957. Big size beryl crystals have also been reported. The pegmatite veins are 220 m long and 10 m wide. They were mined at four horizons; no investigations were carried out beneath the lower horizon (IV). Commonly concordant pegmatite bodies occur between Teregova and Luncavița villages, in the Lazul hill situated south-west of Teregova, in the Cerbul ravine (right-side tributary of Teregovița Valley), in the Teregova and Timiș valleys, in the north of this area. All these pegmatite occurrences are important owing to their feldspar content and, subordinately, to the largely crystallized muscovite.

4. PETROGRAPHIC AND STRUCTURAL SETTING OF PEGMATITES — THE SEBEȘ-LOTRU GROUP

The mesometamorphic crystalline schists of the Sebeș-Lotru Group, which host the pegmatite bodies, constitute the Getic Nappe and occur both in the north of the Muntele Mic massif and east of the Semenic Mts (Pl. I; Pl. II). North of Muntele Mic, the Sebeș-Lotru crystalline schists thrust over the Măru amphibolites of the Danubian units. West of the thrust line, inside the Getic Nappe, a digitation builds up the Turnu Ruieni scale (Pl. I).

According to geological evidence, palynologic data (Visarion, in Savu et al., 1975) and the Rb/Sr dating of 837 m. y. (Bagdasarian, 1972), the Sebeș-Lotru Group is assigned to the Upper Precambrian A.

Since the second half of the last century, some studies have been designated to the general geologic characterization and description of the crystalline schists. It is to mention here Stur's data (1862—1863) cited by Hauer and Stache (1869), the studies of Böckh (1879, 1883), Inkey (1889) and Schafarzik (1898, 1899).

Mrazec (1897) assigned the South Carpathian crystalline schists to two groups with specific petrographic and metamorphic characteristics; following this, Murgoci (1905) put forward the alpine thrust of the Getic Nappe, that is the crystalline schists of the group I over the group II, which represents the Autochthon. In 1908, Murgoci figured on a map the Getic thrust line, which also crosses the present area of study.

Twenty-five years later, Streckeisen and Gherasi (1932) confirmed the tectonic relations between the Getic Nappe and the Danubian units in this region, and they drew up a tectonic sketch including some changes as compared to the former map by Murgoci (who had assigned the Măru amphibolites to the Getic Nappe based on Schafarzik's data). This tectonic sketch also includes the Socetul Mare area.

New data concerning the crystalline schists are given by Gherasi (1952), Roșca (1954), Avramescu (1954), Micșa and Gall (1956), Mânzatu (1957, 1958), Codarcea and Stoenescu (1957), Rădulescu and Rădulescu (1957, 1958), Hurduzeu (1962), Savu and Micu (1964), Popescu and Ștefan (1964), Savu (1965), Gherasi and Zimmermann (1968), Gherasi et al. (1969, 1970), Savu (1970), Hann (1973), Gherasi et al. (1974), Hann (1976), Hann et al. (1977), Gridan (1981), Hann (in Savu et al., 1981) Savu and Hann (1982).

4.1. Petrographic Characteristics

4.1.1. **Paragneisses.** They represent the main rock type. Either they are only biotitic or they contain both mica minerals; the increase of mica content is simultaneous with the decrease of feldspar content, and thus the initially granolepidoblastic rock becomes lepidoblastic and the rocks are to be named micaceous paragneisses. The common mineral association also includes plagioclase (15–20% An) and quartz which provide the ribbon structure of the rock. Microcline, almandine, disthene and sillimanite occur sparsely. Sillimanite (or fibrolite) is usually closely related to biotite and forms needles or sheaves. Epidote, zircon, apatite, rutile and opaque minerals occur in restricted amounts.

4.1.2. **Quartz-feldspar-bearing-leucogneisses.** They form an important level both north of Muntele Mic and to the east of the Semenic Mts (Savu, 1970). They are associated with amphibolites, rarely with micachists, and might be considered a leptino-amphibolite formation. The quartz-feldspar gneisses are interbedded in the paragneisses. Their texture is granoblastic to granolepidoblastic (determined by subparallel mica sheets). The main mineralogic components are microcline, plagioclase (15% An) and quartz. Mirmekite intergrowths are also present. Muscovite, biotite and isolated hornblende crystals occur subordinately.

4.1.3. **Micaschists.** Micaceous paragneisses often grade into micaschists which may also become phaneroblastic. Only thick micaschist interlayerings have been figured on the map. North of Muntele Mic, the micaschists and the micaceous paragneisses represent a main horizon and are usually migmatized. Most of the pegmatites from this region occur either in this horizon or in its vicinity. Besides muscovite and biotite, which are largely crystallized, they contain also quartz, subordinately plagioclase (10–20% An) and almandine porphyroblasts. Sillimanite and disthene do also occur, the former in association with biotite. Some thin sections show sillimanite needles and sheaves associated with biotite, but unrelated to short prismatic disthene crystals.

4.1.4. **Amphibolites, amphibole gneisses.** They occur as interlayerings of varying thickness and length and exhibit a mainly nematoblastic texture. They are either banded or massive. They consist of green hornblende, plagioclase (34–40% An, which may exceed 50% of the groundmass of the amphibole gneisses) and subordinate quartz. Some amphibolites also contain garnet porphyroblasts which lend to the rock the aspect of eclogitic amphibolite. The thin sections do not exhibit diablastic textures nor relict pyroxene which could account for the generation of amphibolite by eclogite retromorphism. However, the minor element content strongly suggests an obvious difference between this amphibolite. (Cr – 5 ppm, V – 60 ppm, Ni – 19 ppm, Zr – 630 ppm, analyst C. Udrescu) and the eclogitic amphibolites reported from the Sebeş-Lotru Group of the Căpăţîna Mts, and described by Hann (1983) as yielding a mean content of 335 ppm Cr, 315 ppm V, 113 ppm Ni, 113 ppm Zr.

Some amphibolites also contain biotite, epidote, chlorite. Ilmenite, titanite and uralite amounts point to the orthorock character. According to Savu, Vasiliu (1970) the amphibolites associated with the crystalline limestones and crystalline dolomites bearing magnetite lenses, in the Armeniș area, contain hastingsite resulting from intense migmatization.

4.1.5. **Crystalline limestones, dolomites.** They occur rarely and form lenses or bands and are white or grey in colour. Gherasi et al. (1974) mentioned from north of Pietroasa Valley some diverging grammatite prisms within granoblastic dolomite. Other crystalline limestone lenses, also occurring north of Muntele Mic, contain diopside, quartz, zoisite. In the Semenic Mts, the crystalline dolomites were described by Schafarzik (1898), Hurduzeu (1962), Savu and Micu (1964). At Armeniș, besides tremolite, quartz, diopside, plagioclase (45–50% An), phlogopite, titanite, zoisite and apatite, dispersed pyrrhotite and magnetite bands are also present.

4.1.6. **Quartz paragneisses.** They occur as thin, usually biotitic layers in the paragneiss pile. They are fine-grained rocks and exhibit typical granolepidoblastic texture. Quartz, biotite and minor plagioclase (15–25% An) are accompanied by subordinate muscovite, almandine, zircon and opaque minerals.

4.1.7. **Migmatites.** The migmatization is characteristic of the above mentioned crystalline schists. There is a correlation between the intensity of this process, namely the feldspathization of schists, and the frequency of pegmatite bodies.

Migmatization has often been conformable to the foliation resulting in the stromatitic structure. The ophthalmitic structure is often present too. All migmatic structures in terms of Mehnert's classification (1962) are encountered. The micaceous paragneisses and the micaschists were much more migmatized than the amphibolites and quartz paragneisses. The leucosome is from 1 mm to several cm thick. An increased thickness promoted an increased grain size of neosomatic microcline, plagioclase or quartz, resulting in pegmatoid structures.

4.1.8. **Serpentinized ultrabasites.** Most of such lenses are interlayered in the paragneisses in the Pietroasa Valley (left-side tributary of Bistra Mărului Valley), as well as in other areas. These rocks were described by Schafarzik (1899), Gherasi (1952), Mînzatu, Mînzatu (1957), Gherasi, Zimmermann (1968), Gridan (1979). If serpentinization is incomplete, the thin sections show cellular textures with olivine relics. Bastitized orthopyroxene relics are also present. The amphiboles are represented by anthophyllite and grammatite, both with needle-like habit and different extinctions. According to Mărunțiu (1978) the alteration of primary minerals of the ultramafic rocks is due to some retromorphic processes, and the orthopyroxene-rich rocks grade to anthophyllite and talc-rich rocks. Thus, the most common primary rocks correspond presumably to orthopyroxene peridotites (harzburgite).

4.1.9. **Granites.** South of Glimboca there is an elongated, conformable granitic body. The rocks are characterized by the coarse grains of white-pink feldspar, accompanied by biotite, usually chloritized, and quartz. Granite is highly tectonized due to the faults which either cut or delineate it; the most important fault delineates the Upper Cretaceous rocks of the Rusca Montană basin from the Sebeș-Lotru Group. In weathering areas granite is converted into typical gruss. The study of thin sections shows a hypidiomorphic structure; microcline, plagioclase and quartz are the main components. Chloritized biotite, epidote, calcite and sericite are also present.

These rocks also occur north-east of Glimboca in the Poiana Ruscă Mts and have been described by Maier et al. (1975) as the „Criva granitoid" of anatectic origin (Pl. II).

4.1.10. **Banatites.** Small discordant igneous bodies represented by hornblende and biotite granodiorites, biotite granodiorites and quartz-diorite porphyries occur in the Semenic Mts near Luncavița, Teregova and Armeniș. These rocks, which penetrated the crystalline schists along some fractures, are the result of Laramian (banatitic) calc-alkaline subduction magmatic activity in the South Carpathians (Rădulescu, Săndulescu, 1973) and occur along the eastern banatitic Teregova-Lăpușnicel alignment trending, according to Savu and Hann (1982), north-eastward to Turnu Ruieni.

The banatitic rocks have been described by Savu (1962), Paraschivescu, Serghie (1963), Hann (1973) and Gunnesch et al. (1978), according to whom the banatitic intrusions are related to a main north-south trending fault, and by Vlad et al. (1980) and Savu (1982) who consider that the banatite occurrences line a NNE–SSW trending fault system.

4.2. Metamorphism, Retromorphism

Two index minerals — kyanite and sillimanite — have been reported north of Muntele Mic. In some thin sections both minerals have been identified without mutual relations. In fact the direct alteration of kyanite into sillimanite is very rare. The transition of one index mineral to the other results from intermediate reactions (Hârtopanu, 1976). Therefore, kyanite is replaced by muscovite, while sillimanite is characterized by subsequent nucleation on muscovite or other mineral, usually biotite. A peculiar feature is the occurrence of both minerals on the isograde plane which thus preserves frozen reactions. This might be due to the fact that one mineral co-exists metastably with the other, namely kyanite coexists with sillimanite as far as it has not surpassed the kinetic limit of its decomposition. On the other hand, the spatial position of the isograde plane should be considered. Therefore, the different thermo-baric gradient potentials in different places of the area affected by metamorphism may result in an unconformity between the initial structure and the isograde plane. However it is possible that the two planes be conformable. In this case, the simultaneous occurrence of the two minerals accounts for the presence of the isograde plane approximately parallel to the present-day erosion level.

Pervasive retromorphism of schists is widespread between Şasa Valley and Slatina Valley. The retromorphism results in new minerals mostly of mimetic growth on pre-existing ones. Chlorite is the most characteristic mineral which replaces, more or less completely, garnet and biotite. Some zones also show textural redistribution following a young „*s*" plane, accompanied by the simultaneous formation of chlorite and albite. This is an incipient process which has not become penetrating nor almost homogeneous as reported in retromorphic areas elsewhere (e.g. the Chiril Series in the Rarău Mts — Kräutner et al., 1981 — and the Uria Formation in the Lotru Mts — Hann, Szász, 1983). This retromorphic zone is also characterized by some features which are due to mechanic deformations and laminations that affected the rocks. These deformations are partly related to the thrust line of the Getic Nappe or are present inside the nappe along some fractures or reverse faults. The discontinuous nature of the mylonitization area along the thrust line could be due to the fact that the mechanic effects of thrusting upon the rocks assigned to the two units depend on the position of the sliding plane with respect to the rock foliation. If the thrust plane parallels the metamorphic foliation plane, the frictional force is consumed along the latter and the effect is less spectacular. If the position of the two planes differs, the effet of mylonitisation is stronger. According to Higgins (1971) there are several types of cataclastic rocks, from breccias, microbreccias to ultramylonites.

The study of retromorphic processes and their different effects resulted in the actual image of the contact line between the Getic Nappe and the Danubian units; this line was estimated westward by early authors.

Savu (1970) recognized three metamorphic areas assigned, like the entire Semenic crystalline massif, to a metamorphic province of barrowian type. From east to west the sillimanite zone superposed on the sillimanite gneiss complex, the kyanite zone overlay the base of paragneiss and quartz-feldspar gneiss complex and the kyanite and staurolite zone superposed on several complexes, partly on the quartz-feldspar gneiss complex and on the micaschist complex from the investigated region.

4.3. **Lithostratigraphy**

According to Gherasi et al. (1974), the crystalline schists occurring north of Muntele Mic and belonging to the Sebeş-Lotru Group have been assigned to two horizons : the lower horizon including biotite paragneisses and amphibolites, metamorphosed at the sillimanite isograde level, and the upper micaschist horizon metamorphosed at the kyanite isograde level. In the Dalci-Măru area the lower horizon overlies the upper one. The crystalline schists are considered to form the reverse limb of a recumbant fold lying tectonically on the Danubian units.

The geologic map, scale 1 : 50,000, sheet Muntele Mic (Hann, in Savu et al., 1981) promoted a different image of the lithostratigraphic sequences characteristic of the Sebeș-Lotru Group from this region. The alterations are based on the following : 1) sillimanite occurs all over the area covered by the Sebeș-Lotru metamorphics, accompanies the kyanite amounts and is therefore inadequate for defining this horizon ; 2) the Turnu Ruieni scale was delimited within the Getic Nappe being characterized by a typical lithostratigraphic sequence showing a quartz-feldspar leucogneiss horizon ; 3) the lithostratigraphic sequence of the crystalline schists assigned to the Getic Nappe also includes a mainly micaschist and micaceous paragneiss horizon, usually highly migmatized. Most of the pegmatites are present in this horizon or in the vicinity. As a result, the following lithostratigraphic sequence is listed below : the base of the Turnu Ruieni scale shows alternating amphibolites, quartz paragneisses, quartz-feldspar gneisses and micaschists developed on two micas paragneisses. Then follows the horizon of quartz-feldspar leucogneisses with interbedded amphibolites and micaschists, and the top built up of alternating micaschists and amphibolites on biotite and muscovite paragneisses.

Kräutner (1980) named the quartz-feldspar leucogneisses the leptinoamphibolite formation and related it to similar sequences of the Sebeș-Lotru Group in the Semenic Mts (according to Savu, 1970), the Godeanu Mts (Bercia, 1975), the Poiana Ruscă Mts (according to Maier et al., 1975) and the Căpățîna Mts (according to Hann, Gheuca, in Lupu et al., 1978).

The lithostratigraphic sequence of the Sebeș-Lotru Group assigned to the Getic Nappe includes at its base several crystalline limestone levels and alternating micaschists and amphibolites on biotite and muscovite paragneisses. Between the base and the top lies the level of micaschists and micaceous paragneisses, while the top consists of alternating micaschists, amphibolites, crystalline limestones and quartz paragneisses on two mica paragneisses.

In the Semenic Mts, the stratigraphic sequence of the Sebeș-Lotru Group was described by Savu (1970) who distinguished five rock complexes ; three of them are found in the area of the present investigation. The sequence starts with the complex of sillimanite paragneisses and carbonate rocks (C_1) including biotite and sillimanite paragneisses and biotite, muscovite and almandine paragneisses with associated migmatites. The paragneisses contain interbedded amphibolites and crystalline limestones bearing lenses of silicates and crystalline dolomites. The next complex (C_2) consists of paragneisses and quartz-feldspar gneisses, muscovite and biotite-bearing paragneisses interbedded with kyanite micaschists or kyanite, staurolite and almandine micaschists, minor crystalline limestone lenses containing silicates or crystalline dolomites related in places to pyroxene gneisses. According to Gridan (1979) these two complexes constitute a unique unit named the lower complex of quartz-feldspar micaceous and carbonate rocks. Then the quartzite complex C follows, but it is not found in the area of the present investigation ; however the micaschist complex (C_4) is present and consists of kyanite, staurolite and almandine micaschists, muscovite and biotite micaschists, muscovite and almandine quartz schists interlayered with paragneisses, quartz-feldspar gneisses, amphibolites and quartzites.

4.4. Mineralizations of the Turnu-Ruieni Metallogenetic Field

Minor pyrite, pyrrhotite with related chalcopyrite and rare molybdenite occur as impregnations and thin veins (0.15–0.25 m thick) in the crystalline schists. Pyrite and pyrrhotite impregnations are reported by Rădulescu, Rădulescu (1957, 1958) in the gneiss in the Slatina basin, assigned at that time to the Danubian units. These ores have also been studied by Popescu, Stefan (1964). Later on, Gherasi et al. (1970, 1974) assigned the disseminated pyrite ores to the alpine cycle and mentioned that the minor molybdenum ore in the Slatina Valley might be related to a deep-seated banatite body.

The ores have been recently studied by Savu and Hann (1982); they are located along fractures and mylonitization areas of the crystalline schists of the Sebeș-Lotru Group. The lineation controlling the ores trends NNE–SSW parallel to the thrust plane. The emplacement was controlled by the Șasa Valley-Borlova reverse fault and a related fault. The ores are found on highly mylonitized rocks also affected by high temperature hydrothermal metamorphism. The chemical analyses show high Fe (22–24%) and S (12–14%) contents corresponding to pyrite and pyrrhotite rich ore. Minor Cu (0.044–0.49%), Ni (0.04–0.05%) and Mo (ca 0.05%) amounts have also been identified. Gold and silver contents are very restricted. The ores may be assigned to the catathermal type passing to the pneumatolitic one, as far as both the characteristic pyrite-pyrrhotite (± chalcopyrite; ± molybdenite) association and the results of hydrothermal metamorphism (garnet occurrence in hydrothermalized amphibolites) show a high temperature of formation (350–570°C). The ores located in a tectonic reactivation area associated with the constitution of the Getic Nappe. The pyrite crystals and the quartz gangue are not deformed tectonically and cement the mylonites and the breccias, thus proving their younger age. Therefore, one may consider that these ores are related to the Laramian (banatitic) magmatic activity. The ores from this metallogenetic field are located in the northward prolongation of the eastern banatitic alignment reported in the Semenic Mts between Mehadia and Teregova. The ores from the Turnu Ruieni-Borlova metallogenetic field are thus ascribed to the metallogenetic district related to the eastern Laramian lineament, which is to be named the Turnu Ruieni–Teregova–Lăpușnicel metallogenetic district. It belongs to the copper-bearing metallogenetic zone of the banatitic province (*sensu* Vlad, 1979).

4.5. Tectonic Data

The tectonic character of this region is given by alpine deformations, that is thrust of the Getic Nappe over the Danubian units following a WNW–ESE strike. In the Muntele Mic area, west of the thrust line, there are tectonic lines represented by faults, reverse faults and a digitation which borders the Turnu Ruieni scale. All these ruptures are approximately parallel to the thrust plane and they formed simultaneously with it. There are also transverse faults, some of them shifting the thrust plane. An important fault cuts the basins of Măloasa, Cornulețu and Slatina valleys shifting both the thrust plane and the other tectonic lines inside the nappe. The fault which delineates the Upper Cretaceous rocks of the Rusca Montană basin from the rocks of the Sebeș-Lotru Group, crossing the valleys of Scoarța, Măceș and Văruț is worth mentioning. In the Semenic Mts a generally N—S trending fault system is prevalent.

5. MORPHOSTRUCTURAL CHARACTERISTICS OF THE PEGMATITE BODIES

The characteristics regarding the shape of pegmatite bodies, their relations with the surrounding rocks and their inner structure are to be taken into account in relation with certain genetic features.

It is well known that the early meaning of pegmatite was confined to a specific texture, that is the graphic texture. This was used by Hauy before 1820. The notion of pegmatite synonymous with graphic granites was first used by Brogniart in 1813, who defined it as a rock built up of lamellar feldspar and quartz. Later, during the second half of the 19th century, according to new and more accurate petrographic data, pegmatites were considered coarse crystallized rocks, usually of granitic composition, often zoned due to the different grain size, the presence of graphic structure or the varying amounts of main or accessory component minerals.

5.1. The Pegmatite Shape and Their Relations with the Surrounding Rocks

The pegmatite bodies show rather various shapes, that is from lenses or veins to bodies with irregular and elaborate, even strange margins. According to Schneiderhöhn (1961) the pegmatite shapes within Precambrian crystalline schists, considered on the whole, are similar all over the world. However, there are several peculiar aspects rather misleading as regards their variety. This is also the case of the pegmatites occurring in the Teregova-Marga region. The common shapes are the following: lenses, concordant veins, dykes, nests and large irregular bodies.

5.1.1. **Lenses.** They are concordant pegmatite bodies which occur very frequently. The lenses exhibit slightly curved or very undulated margins following the folds of the adjacent rock. The high undulation of the contact line may show that its shape depends on the type of rock: a pegmatite occurring between quartz paragneisses and micaschists (Fig. 2) shows a curved shape as compared to the former and a sinuous one as compared to the latter. In other cases, the pegmatite lenses from the anticlinal zone of a fold show their maximum thickness in the area corresponding to the fold axis (Fig. 3) showing that the lens margin is directly influenced by the folding of schists and by the shape of that fold.

It may be thus inferred that pegmatites and folds formed simultaneously.

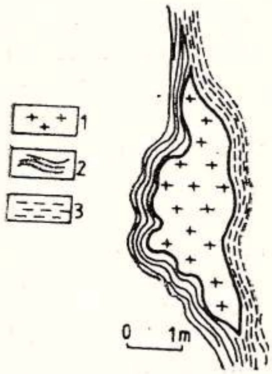


Fig. 2. Contour of a lens-like pegmatite body situated at the contact between two different rock types. 1, pegmatite; 2, micaschist; 3, quartz paragneiss.

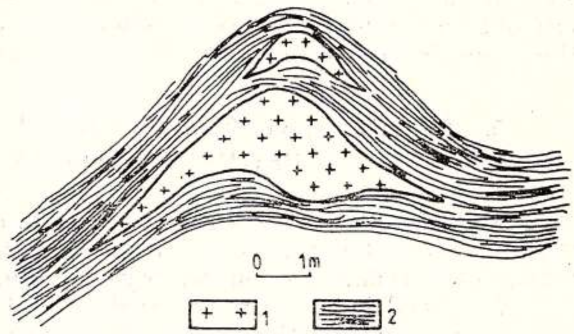


Fig. 3. Pegmatite lenses showing maximum thickness in the fold axis: the lens contour is influenced by the fold shape. 1, pegmatite; 2, paragneiss.

The pegmatite lenses are sometimes isolated in the paragneisses and commonly they form assemblages of distinct types:

a) clusters of lenses forming a swarm as a whole (Fig. 4);

b) rather large pegmatite lenses and smaller, elongated ones situated at a distance of a few cm, parallel to the contour of the big lens (Fig. 5);

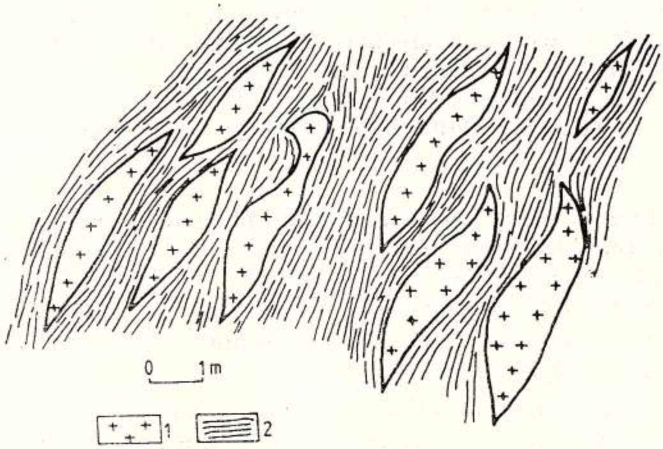


Fig. 4. Swarm of lens-like pegmatite bodies. 1, pegmatite; 2, paragneiss.

c) rows of lenses situated on the same level or on different levels within the paragneisses sequence (Fig. 6a, 6b, Pl. IV, Fig. 1).

It is to note that for type "*a*" the contact between pegmatites and the adjacent rock is usually diffuse [3]. For type "*b*" the pegmatite grades into the adjacent rock by several pegmatite stripes or bands parallel to the main body [4]. As for type "*c*" the contact is clear-cut and in certain

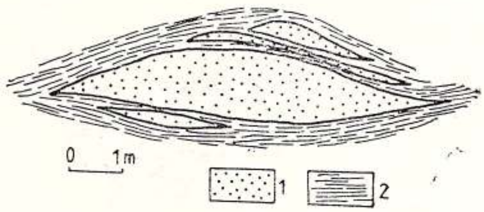
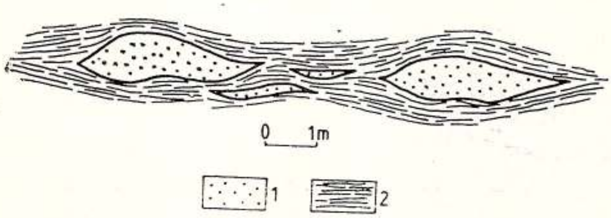


Fig. 5. Large pegmatite lens closely associated with thin, elongated, parallel lenses. 1, pegmatite; 2 paragneiss.



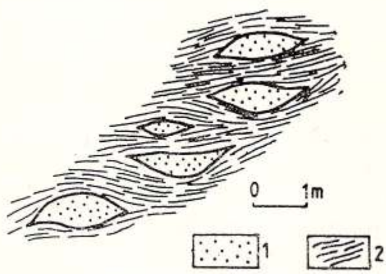


Fig. 6. a, Row of lenses resulting from the boudinage of some concordant vein-like pegmatite bodies. 1, pegmatite; 2, paragneiss.

Fig. 6. b, Pegmatite lenses situated at different levels and resulting from dyke fragmentation. 1, pegmatite; 2, paragneiss.

parts it may be delineated by slickensides. Thus these lenses have resulted from the boudinage of some concordant veins (Pl. IV, Fig. 1) or, according to Şeclăman (1972) from the breaking-down of a dyke due to unaffine lamination movements parallel to the gneiss foliation.

5.1.2. **Concordant veins.** The pegmatite bodies occur frequently as interlayerings within the migmatized paragneisses (Fig. 7) and show branchings out or thickening, then become thinner and thinner to their complete disappearance. In other cases they follow the schist undulated contours or, if secondary veins start from the main body, local unconformities are generated.

The contact with the adjacent rock is often sinuous, and it may be both diffuse and sharp. In other cases the contact is sharp and inside the pegmatite, near the contact with the schists, there is a narrow paragneiss stripe parallel to the contact line (Fig. 8).

As regards the first case, the pegmatite genesis may be related to the pegmatite evolution of quartz-feldspar mobilizates. The second case is characteristic of dilation pegmatites.

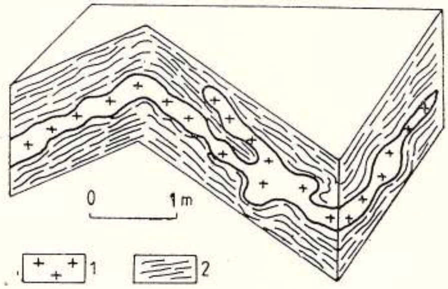


Fig. 7. Concordant pegmatite vein showing thinnings and thickenings, with secondary veins generating local unconformities. 1, pegmatite; 2, paragneiss.

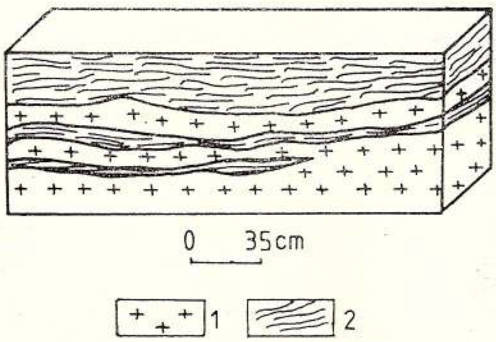


Fig. 8. Concordant pegmatite vein with parallel paragneiss stripes next to the contact line. 1, pegmatite; 2, paragneiss.

5.1.3. **Nests.** They occur less frequently, and form small bodies with irregular margins (Fig. 9a, b).

The contact with the adjacent rock may be sharp or diffuse. The pegmatite illustrated in Fig. 9 a shows both contact types within the same nest.

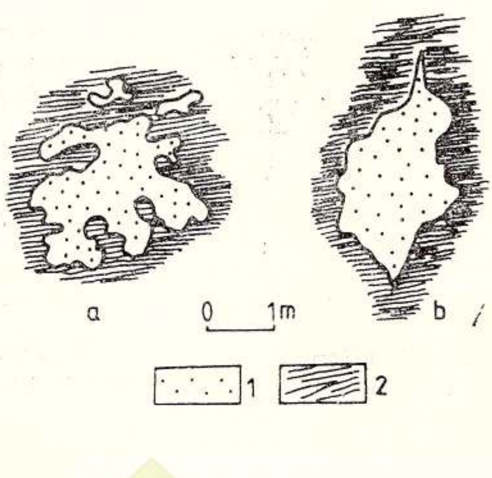


Fig. 9. Pegmatite bodies forming nests. a, generated by migmatization processes or assigned to concretionary pegmatites; b, generated by filling of rock fissures or cavities through secretion processes. 1, pegmatites; 2, paragneisses.

These pegmatite bodies may also form during the migmatization process, or may be ascribed to the concretionary pegmatites described by Ramberg (1952).

The pegmatite nests of type "*b*" result from filling rock fissures or cavities with quartz-feldspar mobilizates by secretion (Mehnert, 1971, Ramberg, 1952).

5.1.4. **Dykes.** They occur less frequently than the concordant lenses or veins, but may form several important bodies in places. The size of these bodies varies broadly. The unconformity with respect to the host crystalline schists varies and may be rapidly modified on short distances. As regards the shape of the contact line, there are dykes which cross the adjacent rocks along a predominantly straight line, resulting in a sharp contact (Fig. 10) and dykes characterized by a highly sinuous contact line, the pegmatitic body including parts of the adjacent rock (Fig. 11), the contact being both sharp and diffuse. Some dykes belonging to the former group show a contact line with fine, millimetric inlets a few cm long, which favour the advancement of pegmatites inside the paragneisses (Fig. 10 a, Pl. III, Fig. 1).

The contact is still sharp, but in this case the later intrusion of pegmatites along low resistance lines during some anatectic processes is obvious.

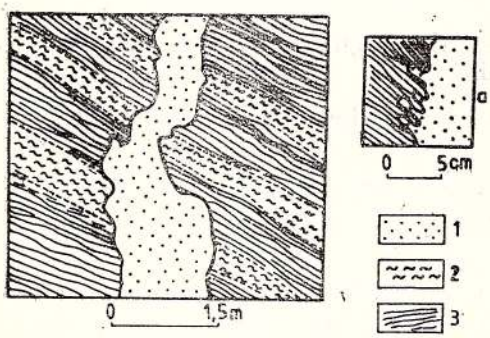


Fig. 10. Pegmatite dykes unconformably piercing the alternating paragneisses and micaschists. a, contact line showing fine pegmatite intrusions in the adjacent rocks. 1. pegmatites; 2. micaschists; 3, paragneisses.

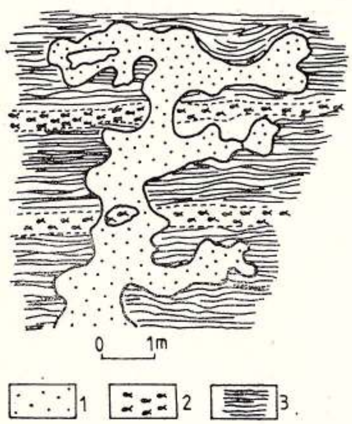


Fig. 11. Pegmatite dyke with a highly sinuous contact line and visible conformity features. 1, pegmatite; 2, amphibolite; 3, paragneiss.

The unconformable feature of dykes points to the late generation of the main metamorphism and migmatization stages of crystalline schists independent of their genesis (lateral secretion, intrusion along fractures of some anatectic mobilizates within open or closed systems, or metasomatic processes characteristic of "replacement" pegmatites, Ramberg, 1949).

5.1.5. **Large irregular bodies.** In the area of study there are also some pegmatite bodies, usually of large size and irregular shape (e.g. Armeniş, Dalci, Cîrniş—Valea Pietroasa, Curcan hill). Following the contact line, some parts of it are highly curved, then it becomes straight or slightly winding. The nature of the contact differs along the same body : it may be sharp (Armeniş body, in certain zones the contact is of tectonic nature and is due to slickenside) or diffuse. Their relation with the adjacent schists varies as well. Thus, at Armeniş the pegmatite body is conformable on the whole. However, following the contact line, on tens of meters, the pegmatite body is obviously unconformable. There is also the case (Pietroasa Valley, Curcan peak) in which the pegmatite body is mainly unconformable and, on limited areas, conformable. Thus, these pegmatite bodies show features which, when first examined, are not concordant. These apparent contradictions may be understood by taking into account that the different features of normal size dykes or lenses occur in the former case on a different scale. For example, if in the case of a small size body the unconformity is decimetric, in the other case it is of metric size and the unconformity between the body and the schists is obvious.

The complexity of shape and of the contact with the adjacent rocks might also be due to the inclusion of several pre-existing pegmatite bodies with peculiar genetic features which influenced differently their characteristics. Several genetic processes are supposed to have influenced successively or alternatively the constitution of these large size bodies and have thus generated obvious features.

5.2. Types of Contact Between the Pegmatite Bodies and the Surrounding Rocks

The above listed descriptions lead to the following classification of the different contact types :

A) Sharp contact : *a*, tectonic ; b, normal

B) Graded contact : *a*, diffuse ; *b*, alternating

C) Mixed contact

A) The sharp tectonic contact *(a)* (Fig. 12a) occurs along slickenside usually typical of young, alpine tectonic planes. This type of contact may not be associated with other mentioned types, as it is not involved in the genesis and evolution of pegmatite with respect to the adjacent rock. Therefore its assignment to this classification is formal. The budined or fragmented veins have previously been considered to show a tectonic contact, but at present this is no longer valid as the slickenside is commonly absent. The deformations seem to belong to a late metamorphic stage marked by similar metamorphic conditions. The recrystallization of minerals along the lamination planes between the pegmatite bodies and the adjacent rock took place to their complete closing. This results in a sharp normal contact *(b)* with an abrupt transition from the pegmatite to the adjacent rock (Fig. 12 b). On the other hand, this type of contact may be due to the petrographic features of the adjacent rock as inferred from the contact area between pegmatites and amphibolites (Fig. 12c).

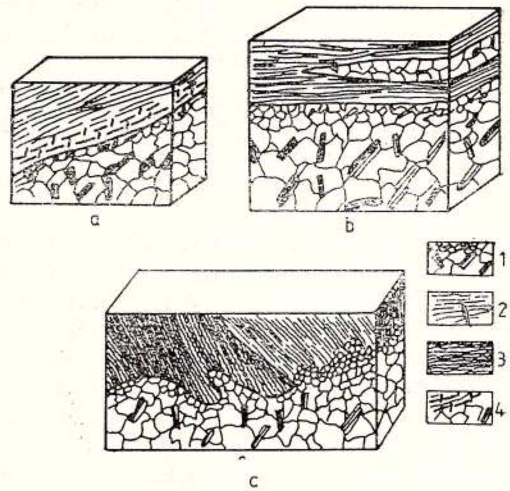


Fig. 12. Types of sharp contact: a, tectonic; b, normal; c, sharp contact resulting from the petrographic features of surrounding rock. 1, pegmatite; 2, paragneiss; 3, amphibolite; 4, mylonites on the friction mirror.

B) The graded diffuse contact *(a)* is of two types: 1, the transition results from the gradual decrease of the grain size of pegmatitic minerals and the simultaneous occurrence of minerals hosted by the adjacent rock (Fig. 13a); 2, the pegmatite groundmass exhibits parts built up of paragneisses and the paragneisses include pegmatitic zones, the transition zone being represented by the area between the pure pegmatite and the paragneiss unaffected by pegmatitization (Fig. 13b, Pl. III, Fig. 2).

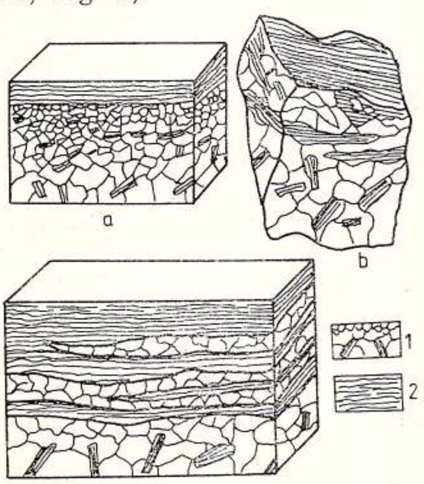


Fig. 13. Types of graded contact: a and b, diffuse. a, by gradual decrease of grain size of pegmatitic minerals; b, paragneiss stripes in pegmatite groundmass and pegmatite zones in paragneiss; c, alternating graded contact. 1, pegmatite; 2, paragneiss.

The alternating graded contact *(b)* is represented by some rock stripes within the pegmatite, alternating with pegmatite stripes to final settlement of one or other feature (Fig. 13a), depending on the direction in which it is looked at.

C) Mixed contact ; in some cases different contact types occur along the boundary of a pegmatite body. The most common example is represented by a tectonic contact partly accompanied by another contact type. There are also some other associations of different contact types within the same pegmatite body, such as the sharp normal contact and the diffuse contact.

Considering the data presented above, one may infer that the contact type, associated with the shape of the body, may provide genetical information ; however, it is not advisable to draw hastly general conclusions based on these relations.

5.2.1. **Mafic accumulations at the periphery of some pegmatite bodies.** The last characteristic implied by the study of the boundary between the pegmatite bodies and the adjacent rocks concerns the accumulation of mafic minerals (biotite, tourmaline or hornblende) in the host paragneiss, in the contact area with some pegmatite bodies. These accumulations are up to 1 cm thick and the minerals are always parallel to the contact plane. This feature is noted in connection with both the mafic cover of conformable pegmatite bodies and the one of unconformable bodies, pointing to a sharp contact. The occurrence of a mafic border is irrelevant at first sight as its presence is related to a certain shape or a certain genetic type of pegmatite bodies.

As it is known, these accumulations of mafic minerals may be considered restites resulting from anatectic differentiation processes (Scheumann, 1937 ; Mehnert, 1951, 1962). However, they may also result from metasomatic processes due to the reaction between the host rock and the mobilised pegmatite, constituting according to Reynolds (1946, 1949) the "basic front". These stripes of mafic mineral accumulations are generated by lateral secretion, as well, in which case the mineral substance is carried away by diffusion, marked by the migration of active minerals — quartz, feldspar — and the simultaneous accumulation of inert minerals — biotite, tourmaline, hornblende (Ramberg, Şeclăman, 1972).

Taking into account the fact that the stripe of mafic minerals resulting from anatectic differentiation may be then destroyed by recrystallizations or local revival of anatexis, one may account for its absence from some pegmatite bodies of this genetic type, therefore for its apparently unstable feature. The mafic mineral accumulations resulting from metasomatosis processes may be similarly approached : the absence of basic phenomena might be due to incomplete reactions or to their revival, or even the later annihilation of their effects because of, for example, some mechanic deformations. As regards the mafic cover of the pegmatite bodies generated by lateral secretion, it may be destroyed by subsequent blastesis of leucocrate minerals.

It is to infer that the mafic accumulation as stripes at the periphery of pegmatite bodies takes place during some distinct petrological processes. In other words, different phenomena determine the same features.

5.2.2. **Conclusions.** The study of the different characteristics of pegmatites from the Teregova-Marga region shows primarily the co-existence of several genetic types. Pegmatite bodies with different shapes formed during the same genetic process and the other way round, similar shapes were generated by different genetic processes. The assignment of these pegmatites to the different genetic types is difficult because of the concentration of phenomena.

It is possible to determine the genetic type of the pegmatite bodies by using all data available and by studying the different aspects in detail.

5.3. Inner Structure

The study of the inner structure of the pegmatite bodies allows to reconstruct the successive stages of pegmatitic minerals formation and the thermo-baric conditions of their crystallization.

All data relevant of the inner structure regard the grain size, the relative amounts of major and accessory minerals and the presence of the graphic structure. Considering these data, the zoned structure may be recognized, as well as the occurrence inside pegmatites of some tabular bodies which represent in fact fractures or fissures filled with younger pegmatitic minerals, or the presence of substitution bodies *sensu* Cameron et al. (1949).

5.3.1. **Zoning.** Most of the pegmatite bodies exhibit this structural feature, even if restricted cases consist of an aplite contact zone contrasting to a largely crystallized, homogeneous pegmatitic mass. When several zones are delineated, they are proved to be generally parallel to the outer contour of the pegmatite body. Both zoning and other structural features show a wide range of aspects often difficult to recognize. These zonings may be complete, incomplete, symmetrical, asymmetrical, simple or composed. There are also numerous bodies marked either by incipient zoning or by destroyed zoning. The most representative example is provided by the large size Armeniș body. It also exhibits wonderful graphic structures, big clusters of largely crystallized microcline-perthite or large areas consisting of plagioclase (oligoclase) associated with quartz, subordinate microcline, muscovite, garnet, biotite. Each area, apart, shows features characteristic of a zone, but the lack of their continuity makes us speak of "pieces" of zones, whereas the pegmatite body is unzoned on the whole. A similar instance is represented by the big Cîrniș body in the Pietroasa Valley, Măgura sector. These situations may result from the occurrence of substitution bodies which pierce and obliterate the initial zonings, as well as from intense cataclasis which changed the primary inner structure of pegmatite bodies.

Zonings have been mentioned by different authors, concerning both the pegmatite bodies in the Semenic Mts and those occurring north of Muntele Mic. Thus, Superceanu (1957) recognizes, within the two large pegmatite dykes at Teregova, an outer aplite zone also containing tourmaline, biotite and hornblende, an intermediate zone with large microcline blocks, quartz and abunding rare minerals such as beryl, niobate, tantalite, monazite and zircon, and an inner zone mainly built of quartz. The contact area between the inner and the intermediate zone includes column-like tantalite crystals weighing 0.5 kg, zircon, niobate, lepidolite crystals and large beryl crystals weighing up to 11 kg. This author also describes the complex zonings of the pegmatite bodies in the Tîlva hill.

Mînzatu and Mînzatu (1957, 1958) also mention zoned structures of pegmatites from Slatina-Timiș area and from Socetul Mare peak lying north of Muntele Mic. They contain an outer aplite zone, then a „pegmatite'' zone of varying grain size, which also contains quartz and both micas besides feldspars, then a central (inner) zone of coarse grain size including quartz cores with miarolitic cavities.

Suru (1966) treats the zoning of some relevant pegmatite bodies in the Tîlva hill (body 1 Tîlva) and in the Pietroasa Valley (Pîrîul cu Mărul body). Thus he describes a micropegmatitic outer zone, of centimetric size, a medium grained marginal zone with important muscovite accumulations, an intermediate zone with „block''-like structure in which the crystals show the largest dimensions, and an inner zone represented by the quartz core. As regards the body 1 Tîlva, the marginal zone consists of plagioclase, quartz, muscovite, as well as garnet and tourmaline, the intermediate zone with block-like structure contains microcline, plagioclase, scarce quartz and muscovite, and the central zone consists of quartz with large beryl crystals. The Pîrîul cu Mărul body shows a quartz core in marginal position next to the dyke hanging-wall with a poorly developed and discontinuous marginal zone. Nevertheless, the muscovite accumulations of economic interest are present in this zone.

Zoned structures are also mentioned by Savu (1977) and have been described by Hann (1973, 1977).

The review of the different shapes of the pegmatite bodies points to the following remarks on their zoning.

*Lenses.*Most of the small size lenses (2–3 m long, as much as 1 m wide) are unzoned. The grain size increases almost always from the margin to the core of the lens wherein feldspars abund. Concomitantly the mica content increases slightly next to the contact with the surrounding rocks. After detailed investigation,most of these lenses exhibit randomly spread, small size cores abunding in one or two minerals, such as quartz and micas, or isolated zones with coarse-grained minerals. Thus, the inner structure of these lenses is however heterogeneous, unzoned and unhomogeneous. These data are relevant as far as these individualized spots account for an incipient zoning and allow to assign the pegmatite body to the sequence of processes resulting in pegmatites, in our case an early evolution stage.

Graphic textures are scarce, mainly consisting of plagioclase and quartz. Plagioclase is however the main component of these lenses. Quartz and mica amounts are relatively high. Microcline ones are reduced. Under the microscope potash feldspar grows by corroding plagioclase. The decrease of the grain size from inside to the periphery might be due to deformations resulting in the boudin structure, namely the breaking up of some pegmatite veins or dykes. Șeclăman (1972) has shown that decrease of grain size points in fact to the increased intensity of mineral cataclasis.

The same explanation may be given for the limited occurrence or the absence of graphic textures, which, as shown by Drescher-Kaden (1969) and Şeclăman (1971, 1972) are easily destroyed during deformation.

The larger lenses frequently show the central zone which consists almost exclusively of quartz and may be discontinuous. There is also an intermediate zone of plagioclase, more and more substituted by microcline, as well as of quartz and micas. It grades into the outer zone by decrease of grain size, quartz enrichment and concomitant increase of mica amount characterized by biotite prevalence and gradual decrease of microcline content.

Concordant veins. The small size ones are never zoned. The size increase is accompanied by grain size increase and the appearance of zoning, which in certain cases is complete. The small size veins consist of plagioclase, microcline, quartz, biotite, muscovite. Under the microscope, the potash feldspar is noticed to replace the plagioclase. Biotite and muscovite amounts are equal or muscovite ones are prevailing. Muscovite formed at the expense of biotite is scarce.

Most of the specialists called the large conformable veins lenses, although there is an obvious difference between the typical lens-shape and the shape of these bodies. The confusion is probably due to both their layered aspects and their gradual thinning out to pointed ends. Most of these veins are clearly zoned. Each body stands out owing to its inner structure. Some veins exhibit between the quartz core and the contact (outer) zone an intermediate zone built of largely crystallized microcline, frequently graphically intergrown with quartz, and a marginal zone mainly consisting of plagioclase, muscovite and quartz; other veins show only the contact zone, an intermediate zone and the central zone. In most cases, the intermediate zone may not be clearly delimited from the marginal one. All deviations from an ideal zoning pattern are due to the succession of the different stages of pegmatite formation and to the conditions under which they took place. For example, in the vein located below Faţa Lungă hill, to the left of the road to Măru village, the quartz core is placed laterally and almost the entire pegmatite groundmass consists of largely crystallized microclineperthite, frequently graphically intergrown with quartz. This microcline mass contains only plagioclase, quartz and muscovite nests representing the marginal zone relics. Thus, it is to infer that the potassic metasomatosis phase was prevailing and took place in its anhydrous form, within an open system, accompanied by the decrease of the muscovite amount. The body located to the east, below the Curcan peak (Tilva area), shows an intermediate zone consisting of largely crystallized perthites, which grades outwards into a marginal zone containing plagioclase, abundant quartz and large muscovite plates. The muscovite amounts result from biotite substitution, the latter being however present as relict mineral. This is proved by plates in which the two minerals are entwinned, the muscovite area showing "brown" patches of biotite or small iron oxide inclusions resulting from biotite deferrization. In other instances biotite disappears completely and only the opaque inclusions are present. In these pegmatites the potassic metasomatosis took place initially in the conditions of water preservation which favoured the crystallization of muscovite, and then anhydrous metasomatosis and large crystallization of potash feldspar occurred as a result of system opening.

On the ridge to the left of the Pietroasa Valley there is a vein showing a zone built of oligoclase and quartz graphic intergrowths (Pl. VI, Fig. 2). Similar intergrowths have been reported from other veins too, preserved as relict structures of limited extent within perthites. The potash feldspar occurs subordinately inside the body, replacing the plagioclase, while the central quartz zone is discontinuous. Therefore, the largely developed plagioclase, characterized by a high temperature of formation, represents a first stage of pegmatite generation. At the end of the plagioclase phase, due to quartz aggressiveness and relatively low water pressure, graphic intergrowths of the two minerals do form. A zone of plagioclase graphic textures stands thus out, and it has been preserved in the absence of any other stage of pegmatite evolution.

Nests. These bodies are irrelevant with respect to zoning. Generally, the grain size is directly proportional to the size of the body. As far as the nests are always of reduced dimension, the grain size is never exceptional. No proper zoning is noted; micas occur in somewhat higher amounts at the periphery of the body. Microcline amounts are at least equal to plagioclase ones, or in other instances are subordinate, but the former always corrodes the latter.

Dykes. These pegmatite bodies exhibit in places complete zonings often of complex type owing to the diversity and importance of their mineralogic composition. However, there are also unzoned dykes or simple zoned dykes, which include those resulting from metasomatic processes of "replacement" type which frequently preserve inclusions of the surrounding rock, as shown in Figure 10. As a result of more intense recrystallization innerwards there is a difference between the periphery and the inner zone. No mineralogic zoning is obvious, the whole body consisting of plagioclase, microcline, fine grained muscovite and scarce biotite. Another similar dyke (lying below the Ogaș hill, right slope of Bistra Mărului Valley, Tîlva area) exhibits a lateral, asymmetrical quartz zone and the rest of the body consists of plagioclase, microcline partly graphically intergrown with quartz, muscovite and biotite, no zoning being noticed so far. However, the quartz zone and the microcline substitution by graphic quartz show that at the end of a mainly alkaline metasomatic stage the quartz was redistributed and became active again. The evolution of these processes may result in the obvious zoning of these smaller dykes.

The large dykes, unfortunately already completely mined, show quite exceptional size in this region (at Teregova they are 10 m thick and over 200 m long). It is worth mentioning that these bodies underwent all or almost all the evolution stages characteristic of pegmatites and they exhibit different zoning patterns, while the typical rare minerals (beryl, monazite, tantalite) contribute to their complex features. Thus, these bodies show features similar to pegmatites of magmatic-granitic origin, with the difference that they had been generated by anatectic metamorphic processes, as shown further on.

These similarities, which are in fact a consequence of converging phenomena, allowed the assignment of these pegmatite bodies to a magmatic granitic origin (Dittler, Kirnbauer, 1931 ; Avramescu, 1954 ; Superceanu, 1957 ; Suru, 1966).

The first, eastward located, dyke at Teregova has a still unmined area (only 4.5 m wide) which shows asymmetrical zoning (Fig. 14). It has an outer contact zone, finely crystallized, of aplite nature, 3–4 cm wide. Under the microscope, the rock shows equigranular texture and leucotonalite composition, consisting of albite and quartz (Pl. VI, Fig. 1).

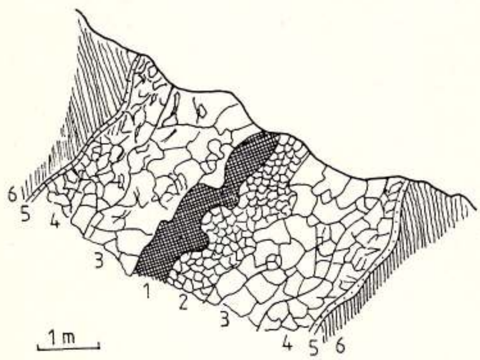


Fig. 14. Pegmatite dyke with asymmetric zoning (Teregova, eastern body). 1, central zone built of quartz ; 2, intermediate micropegmatite (aplite) zone ; 3, largely crystallized intermediate zone built of microcline, subordinate muscovite, quartz, plagioclase ; 4, finer grained outer zone built of abundant muscovite, quartz and subordinate microcline ; 5, contact aplite zone with tourmaline ; 6, migmatized paragneisses.

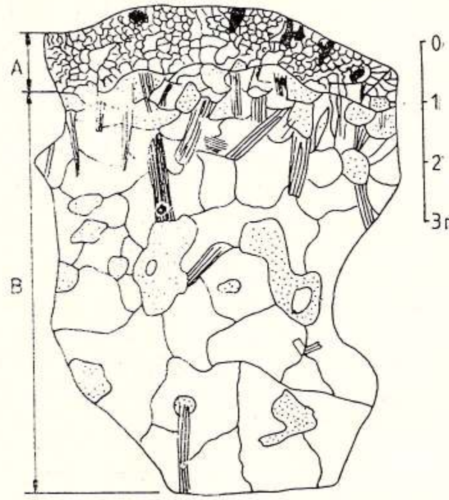


Fig. 15. Contact area (A) and part of the external area (B). Sketch drawn according to polished hand specimen — Teregova eastern body. It is to note the sharp boundary between the two zones and the constant width of the contact area within the aplitic, quartz-feldspathic groundmass also including tourmaline crystals (black). The outer area consists of muscovite (hatched), quartz (dots) and feldspars (white).

The symmetric extinctions of albite crystals are of at least 14° pointing to pure sodic feldspar composition. Considering that the surrounding rock never contains albite and only oligoclase (An_{14}), it is excluded that the contact zone had formed by secretion process. On the other hand, Schneiderhöhn (1961) shows that the contact areas built of albite and quartz are characteristic of pegmatites yielding lithium minerals, as confirmed by Superceanu (1957) who notes the scarce occurrence of lepidolite and spodumene in the Teregova pegmatites. The contact area is not even. A few meters away the aplitic features are still present, but they are uneven as the finely grained groundmass also includes larger quartz, oligoclase and tourmaline crystals (Fig. 15).

The contact area is continuous, of constant width, different from the other areas of varying width characteristic of numerous pegmatite occurrences described by different authors, such as Cameron et al. (1941) and Jahns (1955). Taking into account the texture and mineralogic features of the contact area presented above, it is considered to result from the undercooling of anatectic-palingenetic fluids, which, according to Tuttle (1952), occurred at the same time with volatile loss.

An outstanding characteristic is the presence of a micropegmatite zone next to one side of the quartz core; although it is not as fine grained as the contact area, it contrasts obviously with the surrounding intermediate area, lending an asymmetric character to the zoning of these pegmatites. The microscopic mineralogic study points to oligoclase, quartz, scarce microcline and to a slightly inequigranular texture. These rocks might occur inside the pegmatites due to the local loss of volatiles and the concomitant increase of crystallization nuclei (Jahns, 1955).

The proper intermediate area consists almost exclusively of largely crystallized microclineperthite with subordinate quartz, plagioclase and muscovite. The outer area contains acid plagioclase, quartz, muscovite, subordinate microcline and less crystallized minerals as compared to the intermediate area.

Considering the inner texture of the pegmatites under discussion one may get the graphic representation (Fig. 16) of the relations among zoning, grain size and mineralogic features.

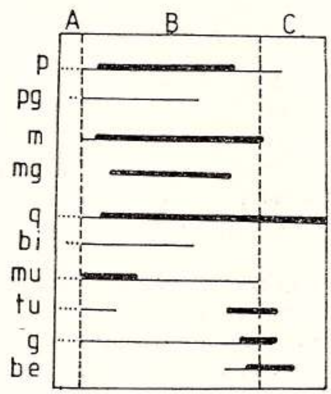


Fig. 16. Relationship among zoning, grain size and mineralogic composition characteristic of pegmatites located between Teregova and Marga localities.

The appearance and growth of different minerals may be studied by taking into account their texture, thus defining the successive stages of pegmatite evolution.

5.3.2. **Graphic texture.** The graphic textures are of special importance to the investigation of pegmatites. Micrographic intergrowths of quartz and feldspar are also present in granites, but only pegmatites show exceptionally large ones, due to the specific growth conditions of pegmatite minerals with peculiar crystal size. Therefore, the pegmatite fluids should contain an increased amount of volatiles and a reduced number of crystallization nuclei. Regardless of their origin, the graphic intergrowths of pegmatites are largely developed depending on these conditions. The graphic textures of pegmatites are also the result of quartz aggressiveness. Quartz may be often redistributed within the different stages of pegmatite evolution.

The study of the graphic fabrics, of their genesis and changes also implies the investigation of pegmatite genesis. Therefore, many researchers speak of "graphic granites" or "graphic pegmatites" or "runites".

The origin of graphic intergrowths is rather controversial. According to different interpretations, there are three different means of understanding the constitution of graphic textures.

1) Graphic textures are the result of simultaneous crystallization of feldspar and quartz. This is the classical view supported, among others, by Brögger, 1881 (in Schneiderhöhn, 1961) and mainly by Fersman (1915, 1929, 1952) whose contribution is outstanding. According to this author, there is a definite reciprocal orientation within quartz and feldspar intergrowths owing to simultaneous crystallization (trapezohedron rule). Vogt (1929) mentions the crystallization of quartz and feldspar in eutectic conditions, based on a 26% quartz within graphic intergrowths. Later studies pointed to frequent important variations of quartz/feldspar ratio, in the absence of eutectic conditions (Barth, 1962), as also proved by some other relatively recent studies (Winkler, 1967).

2) The quartz of graphic textures has formed by partial corrosion and replacement of feldspar, as considered by several researchers such as Sederholm, 1916, (in Schneiderhöhn, 1961), Landes (1925), Schaller (1927), Schädel (1961), Augustithis (1962, 1974), Drescher-Kaden (1969), Şeclăman (1971), Şeclăman, Constantinescu (1972).

3) The quartz and feldspar intergrowths are the result of both simultaneous crystallization and replacement of one of the intergrown minerals by the other. This is Wahlstrom's (1939 a, b) opinion, who, on the one hand denies Fersman's law and on the other supports the idea of the constitution of graphic textures as a result of simultaneous crystallization as well as partial replacement of feldspar by quartz. Erdmansdörfer (1941) accepts both possibilities and according to Eskola the simultaneous crystallization of quartz and feldspar should not respond to the eutectic parameters, while graphic textures are the result of late substitution of feldspar by quartz. Jahns (1955) and Schneiderhöhn (1961) have shown that the graphic textures formed in different ways.

In the area under discussion there are graphic intergrowths of quartz and plagioclase (Pl. VI, Fig. 2 a, b), perthite and microcline (Pl. VI, Fig. 3), characteristic of intermediate or outer zones of pegmatite bodies. The study of three-dimensional graphic quartz shows that it is mostly skeletal and does not occur as isolated inclusions. Some faces of the graphic quartz exhibit a lamellar character (Fig. 17 a). The cut perpendicular to this face points to an entirely different image (Fig. 17 b) of proper graphic quartz (Pl. V, Fig. 1 a, b, c). The quartz lamellae intersect at depth other lamellae almost perpendicular to the former, unseen on the face bearing lamellar quartz as they are cut almost parallel to the plane on which they occur, or, owing to their thinness they have been detached or represent ununiform patches in case they had been cut aslant.

These features are typical of graphic quartz formed by selective replacement of feldspar along its low resistant zones represented by fissures or irregular dislocations, and mainly by distortion stripes, described by different authors, such as Seifert (1965) and Drescher-Kaden (1969). These distortion stripes are reticular labile zones as most of the

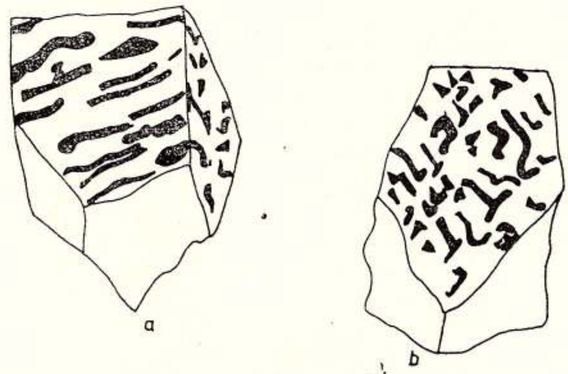


Fig. 17. a, Graphic quartz with lamellar aspect; b, face cut perpendicular to the former (a), with graphic quartz resulting from the intersection at depth of different quartz lamellae.

components of their network are not in equilibrium. The distortion stripes are easily noticed in relatively large crystals and concentrate in one or several systems, out of which the most frequent ones are parallel to the cleavage, to the face (100) and form an angle of 45° with the cleavage (Şeclăman, 1971). It is important to note that the different positions of the distortion stripes may be encountered in the case of the quartz lamellae. This conformity is accounted by the substitution of feldspar by quartz along the low resistant zones represented by the distortion stripes. The graphic textures recognized at Armeniş and Voineasa (Şeclăman, 1971) show quartz lamellae obviously parallel to *B* crystal axis of feldspar. In other instances they are parallel to *A* and *C* crystal axes. Therefore, these lamellae correspond to the intersection of distortion stripes parallel to faces (100), (010) and (001), accounting for the graphic shape which results from the selective substitution of feldspar by quartz along the intersection areas between the distortion stripes. This phenomenon has been minutely described by Drescher-Kaden (1969) and then convincingly substantiated by Şeclăman (1971) and Şeclăman, Constantinescu (1972).

The typical graphic feature is the result of selective substitution of feldspar by quartz along the intersection between the distortion stripes. The quartz also shows remarkable optical continuity on large areas. Thus, one may consider that quartz was intruded irrespective of the position of the distortion stripes. In other cases, however, this is not achieved as the optical orientations of quartz are different. This may point to a difference in time of substitutions, and to subsequent recrystallization of quartz in small networks resulting in independent fragments. Recrystallizations resulted from quartz fragmentation, its network being destroyed by relatively slight efforts, while intergrown feldspar is more plastic and preserves its network and optical continuity intact.

The hypothesis of graphic structure of pegmatite formed by metasomatic corrosion and replacement is also supported by the presence of radial-symmetric textures (Fig. 18).

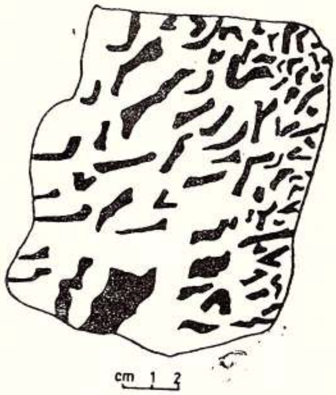


Fig. 18. Radial symmetric graphic texture (Armeniș). Graphic quartz resulting from metasomatic corrosion and becoming coarser and coarser next to the source.

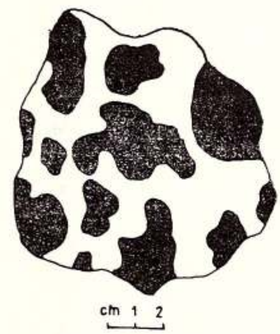


Fig. 19. Graphic quartz with ocellar aspect (Tîlva--Curcan hill area).

One may easily see the outer rods and fine lamellae of graphic quartz which become coarser and coarser innwards resulting in a common quartz grain which contrasts, in size and shape, with the characteristic outer graphic quartz. In fact, the feldspar represents a single large relict grain differently substituted by quartz. Similar examples are given by Augustithis (1974), too. In the neighbouring areas of the silica source the substitution was complete, and the quartz developed freely. Because of the environment which favoured the development of quartz, there are some special features, such as the "ocellar (augen) quartz" (Fig. 19) from the widely crystallized microcline area next to the quartz core of a pegmatitic body located in the Curcan hill (Tîlva area). Considering the radial texture it is to note that far from this largely crystallized "ungraphic" quartz substitution was selective, resulting in the graphic aspect which, according to Șeclăman (1971) depends on the ratio between the crystallizing capacity of the neosoma and the displacement strength of the paleosoma.

A graphic intergrowth of quartz and plagioclase (10% An) substituted together with muscovite by microcline (Fig. 20) was noted. It is to mention that quartz stands substitution better than plagioclase. Therefore, owing to these pseudomorphic graphic textures subsequent to plagioclase ones, in the case in which the latter mineral had been wholly replaced, one encounters graphic quartz as isolated inclusions in microcline. These features should be systematically studied in order to avoid inaccurate interpretations. As regards the quartz-plagioclase graphic intergrowths prior to microcline, it is worth considering that there are no proofs of their constitution by metasomatic processes or simultaneous crystallization.

Graphic quartz is easily affected by deformation, may be easily broken, and the different fragments resulted have their own optical

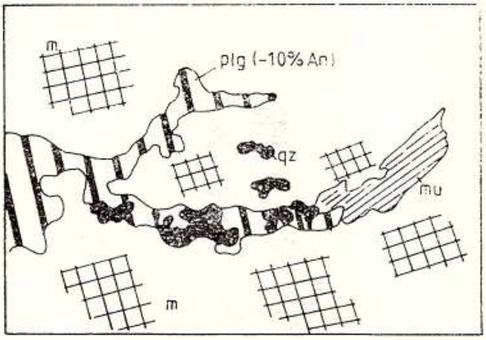


Fig. 20. Graphic intergrowth of quartz and plagioclase partly substituted by microcline; quartz resists substitution better. plg, plagioclase; qz, quartz; m, microcline; mu, muscovite.

orientations. At the same time, deformed quartz rods underwent recrystallization. Graphic textures are completely destroyed by intense cataclasis. These characteristic features are encountered in the large Armeniş body and in several other places.

The ten chemical analyses presented in Table 1 inform on the composition of graphic granites. Thus, the normative composition (Table 2) shows the amount of normative minerals such as quartz, plagioclase and potash feldspar (estimates acc. to R. Constantinescu). The variation of muscovite, chlorite, magnetite and apatite amounts is also considered.

It is obvious that the amount of normative muscovite decreases concomitantly with the increase of the amount of normative potash feldspar and the other way round (sample 2133 represents a graphic intergrowth of plagioclase and quartz and yields the highest muscovite amount), accounting for their constitution within the intermediate zone, in a closed or open system, in the presence or absence of water respectively. Normative quartz varies between 18.80 and 33.22%, potash feldspar represents 34.27–60.21% and plagioclase 11.90–47.92%, most of the graphic granites contain ca 26% quartz, 47% potash feldspar, 26% albite. These ratios are close to those presented by Simpson (1962) – 4–60% quartz, 5–85% plagioclase and 4–70% potash feldspar, most of the graphic granites yielding ca 25% quartz, 55% potash feldspar and 20% albite – and point out the variable character of quartz and other mineral amounts (Fig. 21); therefore, it is not a granitic eutectic as, according to Vogt (1929) and Fersman (1929) it should show a more constant composition. Winkler's (1967) data show that granitic eutectics themselves have a composition which varies owing to certain factors. However the ratio variations are correlated and quartz amounts reach 35%, which is not valid for the graphic textures under discussion.

The major and minor elements content (Tables 1, 3) of graphic granites also informs on the graphic texture zones within the geochemical evolution of pegmatites.

TABLE 1

Chemical composition of some graphic granites from Teregova-Marga area

Sample	2206	2133	2518	2279/2	1136 A	1105	35	41	54_1	55
Location	Cermez Valley Măgura area	Pietroasa Valley Măgura area	Valea Rea Dalci-Var area	Armeniş Valley quarry	Rîpi hill Tilva area	Scoarţa Valley Măgura area	Cioaca cu Tei Măgura area	Mărul brook Măgura area	Armeniş quarry	Armeniş quarry
No.	1	2	3	4	5	6	7	8	9	10
SiO_2	72.58	72.72	75.80	72.10	73.09	70.72	70.95	70.61	71.85	72.51
Al_2O_3	14.24	17.45	13.12	14.93	12.95	16.00	16.70	15.45	17.00	17.15
Fe_2O_3	1.78	0.19	—	—	2.05	0.10	0.46	0.18	0.03	0.13
FeO	0.08	0.16	—	—	0.14	0.36	0.13	0.06	0.11	0.16
MnO	—	0.03	—	traces	—	—	0.025	—	—	—
MgO	0.34	0.54	0.07	0.06	0.55	0.35	—	—	0.10	0.15
CaO	0.45	1.36	0.14	0.12	0.60	0.70	0.70	0.59	0.40	0.52
Na_2O	1.76	5.20	1.47	2.67	1.60	1.40	1.40	2.50	2.60	3.20
K_2O	9.30	1.45	8.96	9.18	9.00	10.20	9.50	9.40	9.00	5.80
TiO_2	—	0.02	0.05	0.08	—	—	—	—	—	—
P_2O_5	0.08	0.07	0.27	0.22	0.12	0.11	0.07	0.095	0.070	0.070
CO_2	—	—	—	—	—	—	—	—	—	—
S	0.04	0.03	0.03	0.02	—	—	—	—	—	—
H_2O^+	—	0.35	0.53	0.24	0.10	0.02	0.24	0.05	0.30	0.34
Fe(S)	—	—	0.03	0.02	—	—	—	—	—	—
H_2O	—	—	—	—	0.10	0.24	0.04	0.12	0.08	0.10
Total	100.65	99.59	100.47	99.64	100.30	100.20	100.21	100.05	101.54	100.13

Analysts: 1—2 V. Neacşu; 3—4 C. Vlad; 5—10 A. Movileanu

TABLE 2

Normative composition (%) of graphic granites

Sample	2206	2133	2518	2279_2	1136_A	1105	35	41	54_1	55
Quartz	25.46	32.22	32.54	21.09	27.90	21.61	24.88	18.80	21.28	33.39
Albite	14.88	41.17	11.60	22.07	13.88	11.83	9.68	22.20	21.99	27.07
Anorthite	2.21	5.75	0.31	0.37	2.33	3.41	2.22	2.55	1.97	2.55
Orthoclase	54.90	—	51.70	53.52	54.21	60.21	48.27	55.36	53.12	34.27
Chlorite	1.21	1.94	0.27	0.21	1.40	1.82	0.06	0.03	0.53	0.70
Muscovite	—	17.87	3.32	2.93	—	—	14.66	1.06	—	—
Magnetite	—	0.32	—	—	0.18	0.05	0.39	0.21	0.01	0.16
Apatite	0.19	0.17	0.63	0.51	0.28	0.26	0.16	0.21	0.16	—

The plots of graphic granites on Na_2O-K_2O diagram (Fig. 22) form a well-defined field with respect to the pegmatite plots (chemical analyses presented in Table 10). One of the two plots outside the field

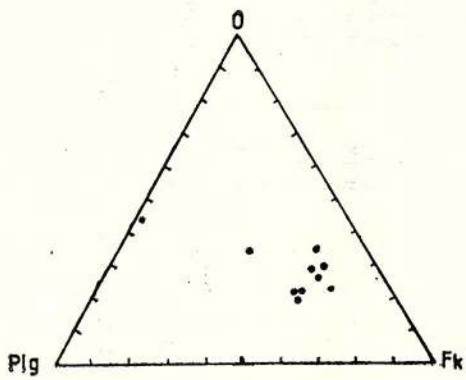


Fig. 21. Variation of mineral ratios within graphic intergrowths in Teregova-Marga pegmatites.

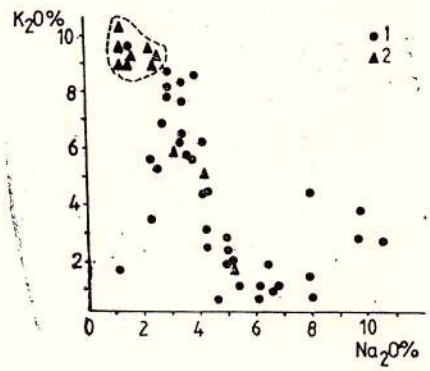


Fig. 22. Na_2O : K_2O diagram of graphic granites and pegmagites from Teregova-Marga region: 1. pegmatites; 2. graphic granites.

represents a plagioclase graphic texture, while the other is of intermediate character, possibly a plagioclase graphic intergrowth mostly replaced by microcline.

TABLE 3

Minor element content of graphic granites from Teregova—Marga area

No.	Sample	Location	Pb	Cu	Sn	Ga	Mo	Ni	Co	Cr	V	Zr	Be	Ba	Sr	Li
1	2206	Cermez Valley, Măgura area	85	3	2	4	2	2	2	1	2	12	1	4000	500	3
2	2133	Pietroasa Valley, Măgura area	39	3	2,5	12	2	2	2	2	2	10	7	60	70	5,5
3	1105	Scoarţa Valley, Măgura area	75	4	3	7	2	2	2	2	3	10	1	2800	440	3
4	35	Cioaca cu Tei, Măgura area	155	5.5	2	6	2	2.5	2	1	2	10	1	5000	500	3
5	41	Mărul brook, Măgura area	90	4.5	2	6	50	3	2	1	2	10	1	860	75	3
6	54_1	Armeniş, quarry	70	4	2	8	2	2	2	1.5	2	10	1	400	50	3
7	55	Armeniş, quarry	90	4	2	9.5	2	2	2	1	2	10	1.2	500	65	3
8	1245	Rîpi hill, Tîlva area	115	6.5	2	4.5	2	3	2	1	2	10	1	3600	480	3
9	1126_2	Văruţ Valley, Dalci-Var area	16	3.5	6.5	16	2	3	2	2.5	2	10	5.8	165	29	9
10	1108/A	Armeniş	39	3	2	15	2	4	2	4	2	10	2.1	90	67	3
11	1108/B	Armeniş	90	4	2	9	2	3	2	2.5	2	10	1.4	790	67	3
12	2274	Armeniş	100	3	3	8	2	2	2	2.5	2	10	1	1400	100	3
13	T_3	Teregova	115	5	2	6.5	2	2	2	1	2	17	1	950	300	3

Analysts: 1—2, 4—8 — Constanţa Udrescu; 3.9—13 — Ana Şerbănescu

The diagram $SiO_2-Al_2O_3$ (Fig. 23) also shows that the plots of graphic granites enter a unitary field quite distinct from the field of pegmatite plots. The field of graphic granites is located inside a larger and relatively uniform pegmatite zone, to its base.

Graphic granites show a positive correlation of Ba and K (Fig. 24), while pegmatites have an irrelevant behaviour. This relationship informs

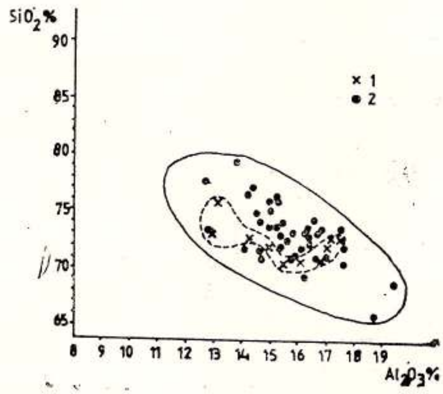


Fig. 23. $SiO_2 : Al_2O_3$ diagram of graphic granites and pegmatites from Teregova-Marga region: 1. graphic granites; 2, pegmatites.

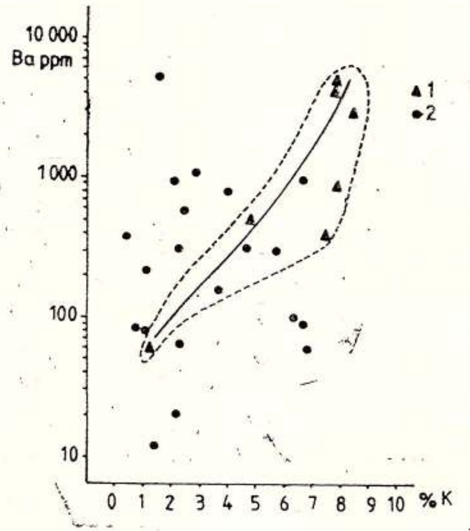


Fig. 24. Ba ppm : %K diagram of graphic granites and pegmatites from Teregova-Marga region: 1. graphic granites; 2, pegmatites.

on both the geochemical evolution inside the graphic texture zones and the geochemical differentiation of graphic granites compared to the geochemical evolution of pegmatites.

The Ba : K, Ba : Sr and Sr : Ca relations on the one hand, and Na and K amounts on the other, are also important for the geochemical character of graphic granites (Fig. 25). All three cases are marked by two correlation fields, the left side one of positive nature. At the same time, graphic textures are noted within pegmatites. The right side field, including most of the plots, represents the graphic intergrowths of quartz and microcline (or microclineperthite), and the left side field stands for the intergrowths of plagioclase and quartz which pass (by metasomatic resorption) to quartz-microcline ones. Taking into account the geochemical data the graphic granites are supposed to have resulted from pegmatites by metasomatic processes.

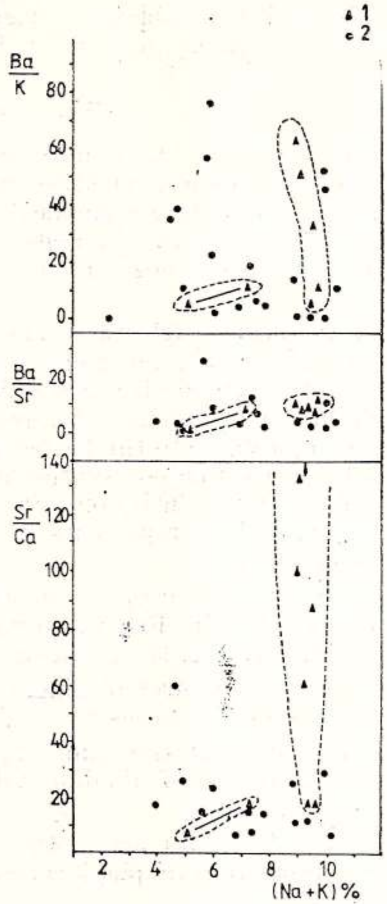


Fig. 25. Ba : K, Ba : Sr, Sr : Ca ratios and (Na + K)% of graphic granites and pegmatites in Teregova-Marga region : 1, graphic granites ; 2, pegmatites.

6. MINERALOGY OF PEGMATITES

Most of the pegmatites under discussion show simple mineralogic features, characteristic of metamorphic pegmatites. The main mineral phase (which in the case of anatectic pegmatites constitutes the stable components of the pegmatitic fluid) consists of biotite-acid plagioclase-microcline, microclineperthite-muscovite-quartz. These minerals may occur in all pegmatites irrespective of their origin, and they represent stable minerals of the pegmatite stage sensu Vogt (1929), Fersman (1952), generate the characteristic zoning (Lacroix, 1922, Andersen, 1938, Vlasov, 1952), account for the geochemical evolution of pegmatites (Ghinsburg, 1960), and by replacing each other point to the evolution stage of pegmatites.

Pegmatites with rare minerals or other characteristic minerals are subordinate, while those containing numerous accessory minerals as a result of pneumatolysis, are sparse. The different major mineralogic stages include: garnet, tourmaline, beryl, apatite, columbite, monazite, orthite, kyanite, sillimanite, hornblende, magnetite, hematite, pyrite, calcite and chlorite (as pseudomorphoses after biotite).

6.1. Feldspars

Feldspars are the most important minerals of pegmatites both in number and influence in pegmatite genesis. The feldspar accounts for the evolution stage and the geochemical phase of pegmatite. Feldspars also exhibit the deformations, implicitly the change of tectonic conditions, which had affected the pegmatites.

6.1.1. **Plagioclase feldspars.** The acid plagioclase (albite-oligoclase, oligoclase) of grey colour represents the main component of several pegmatites (especially small size, concordant bodies). There are pegmatites which contain equal amounts of microcline and plagioclase feldspar, or the latter is subordinate to the former or is absent. Under the microscope, the plagioclase with fine polysynthetic twins is substituted by microcline wherein it forms relict inclusions (Pl. VII, Fig. 1) resistant to metasomatosis. Thus, plagioclase represents the first evolution stage (I) resulting in initial pegmatites.

Albite occurs in the aplite contact zone of a discordant pegmatite from Teregova (Pl. VI, Fig. 1), forming in places the cover of altered plagioclase (Pl. IX, Fig. 1) and being present in different perthite types.

The chemical features of plagioclase are shown by analytical data (Table 4, analysts V. Neacșu, C. Vlad).

It is to note that the sum total of alkaline oxides ($Na_2O + K_2O$) is inferior to that one of alkaline feldspars (Table 5), varying between 9.30 and 9.72%.

The analytical data led to the crystallochemical formulas and the normative composition of plagioclases:

Sample:

· 11062 $Ca_{0.51} K_{0.04} Na_{3.17} Al_{4.23} (Al_{0.19} Si_{11.81} O_{32})$
$Or_{1.07} Ab_{85.21} An_{13.71}$

· 2492 $K_{0.11} Ca_{0.35} Na_{3.11} Al_{4.05} (Al_{0.08} Si_{11.92} O_{32})$
$Or_{3.08} Ab_{87.11} An_{9.80}$

With respect to the geochemical distribution of minor elements, sample 11062 (analyst A. Șerbănescu) shows higher Ba amounts than Sr ones (790 ppm and 78 ppm respectively), both elements pointing to smaller contents as compared to alkali feldspars (Table 6). The following amounts are presented: Pb 50 ppm, Cu 30 ppm, Zn 30 ppm, B 30 ppm, Nb 10 ppm, Be 5 ppm, Co less than 2 ppm (similar to potash feldspars).

6.1.2. **Alkali feldspars.** The investigated pegmatites contain alkali feldspars with triclinic system, represented by microcline or microcline-perthite not all the crystals being cross-hatched. The microscopic study points to their metasomatic development at the expense of plagioclase. Plagioclase is corroded by the potash feldspar and antiperthites, substi-

TABLE 4

Chemical composition of some plagioclases from pegmatites in Tilva area

Sample	OXIDES %													
	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	CO_2	H_2O^+	S	TOTAL
1106_2	67.72	20.84	0.02	0.03	0.0	0.15	2.70	9.10	0.20	0.00	0.07	0.00	0.45	0.00	101.38
2492	68.30	20.08	0.00	0.00	0.02	0.11	1.91	9.22	0.50	0.10	0.11	0.00	0.15	0.03	100.53

TABLE 5

Chemical composition of some microclineperthites yielded by Teregova-Marga pegmatites

No.	1	2	3	4	5	6	7	8
Sample	Fk 1136	1245 M	1101/5	3673	1101	Fk 1126	64_2	64_1
Location	Faţa Lungă hill, Tilva area	Rîpi hill Tilva area	Scoarţa Valley Măgura area	Dobrotin Valley Măgura area	Scoarţa Valley Măgura area	Văruţ Valley Dalci-Var area	Curcan hill Tilva area	Curcan hill Tilva area
SiO_2 %	62.74	71.09	63.40	64.27	72.51	64.32	65.04	66.16
Al_2O_3 %	18.90	15.15	19.30	18.44	14.80	15.75	20.17	18.00
Fe_2O_3 %	—	0.03	0.05	0.02	0.10	3.50	0.03	0.06
FeO %	0.18	0.02	0.07	0.04	0.36	1.58	0.15	0.13
MnO %	0.009	—	—	0.11	0.010	sld	0.00	sld
MgO %	0.94	0.32	0.32	0.81	0.40	0.45	0.00	0.25
CaO %	1.04	0.43	0.59	0.45	1.05	0.77	0.30	0.20
Na_2O %	2.90	2.15	2.40	3.00	2.70	3.0	3.05	3.00
K_2O %	12.50	9.65	12.65	11.85	7.60	10.2	11.70	12.00
TiO_2 %	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
P_2O_5 %	0.08	0.15	0.10	0.17	0.090	0.120	0.41	0.205
CO_2 %	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
S %	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
H_2O^+ %	0.59	0.45	0.61	0.88	0.27	0.13	0.00	0.07
H_2O^- %	—	—	—	—	0.26	0.066	0.00	0.08
Total	99.87	99.44	99.42	100.04	100.15	99.88	100.88	100.15

Analysts: 1—4, 7 — Neacşu Vasilica; 5, 6, 8 — Aurelia Movileanu;

TABLE 6

Minor element content of some microclineperthites from Teregova-Marga area

No.	Sample	Location	Pb	Cu	Sn	Ga	Ni	Cr	V	Be	B	Ba	Sr	Li
1	1136	Fața Lungă hill, Tilva area	165	2.5	<2	6.5	3	<1	1	<1	<30	4000	350	<3
2	1245 M	Ripi hill, Tilva area	170	4.5	<2	5	4	1.5	<1	<1	<30	2100	260	<3
3	367_3	Dobrotin valley, Măgura area	115	6.5	<2	6	5.5	1	1.5	<1	<30	3000	380	<3
4	64_2	Curcan hill, Tilva area	125	7	<2	5	<2	5.5	<2	2.6	<30	180	32	<3
5	1101	Scoarța valley, Măgura area	73	3	<2	10	5	2	<2	<1	60	64	150	20
6	1101/5	Scoarța valley, Măgura area	100	3	3	7	<2	7	3	<1	<30	>3000	500	<3
7	1101/5a	Scoarța valley, Măgura area	105	4.5	<2	6	2.5	<1	<2	<1	<30	5200	460	<3
8	8M	Ogaș hill, Măgura area	130	3.5	<2	6.5	3.5	1	<2	2.1	<30	200	42	<3

Analysts: 1–4, 7–8 Constanța Udrescu, 5–6 Ana Șerbănescu tution perthites or relict plagioclase inclusions in microcline are often present. It is thus inferred that potash feldspar represents a second evolution stage (II) of pegmatite genesis. The absence of water from the potash feldspar network accounts for its genesis by "dry" potassic metasomatosis.

Microclineperthite is the prevailing alkali feldspar. The perthite textures defined may be of mixed-metasomatic and exsolution nature, as far as the microclineperthite composition consists of a potassic stage, an exsolved sodic stage and a relict plagioclase stage. Irregular shapes (Pl. VII, Fig. 2 a, b) of low temperature perthites resulted from potash feldspar substitution by plagioclase are prevailing (Wahlstrom, 1939; Drescher-Kaden, 1969). Perthites with fine albite bands uniformly arranged and originating in high temperature exsolutions are sparse (Vogt, 1908; Andersen, 1938; Mehnert, 1971).

A microclineperthite sample of irregular shape (collected from the Armeniş quarry) was studied by JEOL 100 microsonde at ICEM laboratory for physical metallurgy (Pl. VIII). The topographic surface shows a relatively homogeneous, fine grained groundmass abunding in light elements (Al, K, Si). Considering the element distribution, Al abundances are uniformly spread, followed by Si. K, Na and Ca show complementary distribution, with irregular and uneven boundaries. K contains Na in places, while the reduced Ca amount is related to Na one and determines the unhomogeneous character of the mineral groundmass, visible to the right of the image.

The analytical data from Table 5 point to the chemical composition of alkali feldspars. The silica excess is due to quartz inclusions. A single sample yields an anomalous Fe content (3.5% Fe_2O_3) probably due to a hematite or magnetite inclusion. The potassium content varies from 12.50 to 7.60% K_2O, with a mean content of 11.18%. The sodium content varies from 2.15 to 3.05% Na_2O, with a mean content of 2.79%. The calcium content (1.05–0.20% CaO) shows the normal values characteristic of these minerals.

The crystallochemical formulas and the normative composition of alkali fedspars were also calculated:

Sample:

· Fk 1136	$K_{2.99}Na_{1.05}Ca_{0.21}Al_{3.92}(Al_{0.24}Si_{11.76}O_{32})$ $Or_{70.35}Ab_{24.70}An_{4.94}$
· 1101/5	$K_{3.00}Na_{0.86}Ca_{0.12}Al_{4.03}(Al_{0.20}Si_{11.80}O_{32})$ $Or_{75.37}Ab_{21.60}An_{3.01}$
· 367	$K_{2.80}Na_{1.07}Ca_{0.08}Al_{3.98}(Al_{0.05}Si_{11.95}O_{32})$ $Or_{70.88}Ab_{27.08}An_{2.02}$
· 1245 M	$Or_{73.66}Ab_{23.53}An_{2.79}$
· 1101	$Or_{61.50}Ab_{31.36}An_{7.12}$
· 1126	$Or_{67.90}Ab_{28.60}An_{3.49}$
· 64_2	$Or_{71.78}Ab_{26.76}An_{1.45}$

Taking into account the minor element distribution within microclineperthites (Table 6) the most important substitution types result from K replacement by Ba and Pb. Ba amount varies within wide limits

(180—5200 ppm) and is higher than Sr one. Ba : K ratio (Fig. 26) shows that K increase entails Ba increase. Side values account for albite occurrence in the perthite texture.

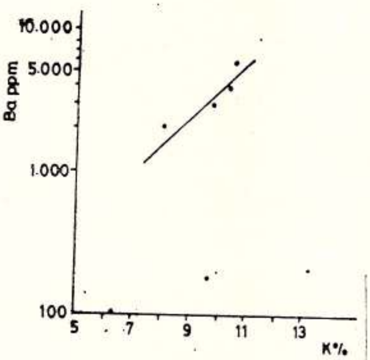


Fig. 26. Ba : K diagram of microclineperthites from Teregova-Marga area

The mean Ba content is of 2.218 ppm, close to the mean value of 2.486 ppm reported for pegmatites from Rodna Mts (Murariu, 1979). According to the same author, a comparison with the alkali feldspars of granites (mean Ba content of 5.093 ppm, Liahovici, in Murariu, 1979). shows that the Ba content of pegmatites is highly decreasing. Taking into account the relation between Ba content of feldspars and crystallization temperature, the alkali feldspars of the investigated pegmatites were generated at lower temperatures than those of granites or than the alkali feldspars of some igneous, granitic pegmatites.

Sr content varies from 32 to 500 ppm and is lower than Ba content.

The similar size of ionic radiuses of K (1.33 kX) and Pb (1.32 kX) results in identical position of Pb and K ions within the potash feldspar. The investigated pegmatites show Pb accumulations of alkali feldspars varying from 73 to 170 ppm.

6.2. Micas

Considering their textural relations with the other minerals micas characterize different stages of pegmatite genesis. According to the alterations affecting them (biotite especially) and to the response to mechanic deformations (noticeable in the case of muscovite) micas point to both the alteration of chemistry typical of different evolution stages and the tectonic conditions of pegmatite genesis.

6.2.1. **Biotite**. Although biotite occurs subordinately, its presence is of petrogenetic importance. It is frequently fresh in small, mainly plagioclase pegmatite bodies. The large zoned bodies contain biotite, largely crystallized in places, associated with plagioclase or microcline, which is to be wholly altered to muscovite, seldom chlorite, and represents a relict mineral. Biotite is characteristic of stage I of pegmatite genesis and crystallizes together with plagioclase, quartz, subordinate microcline. Then, it becomes unstable, is deferrized and the increased pressure of water vapours results in its alteration to muscovite. This phenomenon is also represented by brown biotite patches on some muscovite grains and small hematite or magnetite grain accumulations. The microscopic study of muscovites shows hematite grains along the cleavage planes.

6.2.2. **Muscovite.** Large muscovite plates are often present in the so-called "muscovite pegmatites". Muscovite may show greenish shades due to different hematite (especially resulted from biotite alteration), tourmaline, apatite or garnet inclusions.

Muscovite is subordinate in "microcline pegmatites". There are also pegmatites containing both minerals assigned to different zones. The microscopic study shows plagioclase replaced by muscovite, accompanied by quartz, the two minerals being closely related. Muscovite was subsequent to plagioclase and was assigned to another stage (II A) of pegmatite genesis, characterized by potassic metasomatosis abunding in water vapours. During this stage, the crystallization of muscovite plates associated with quartz becomes prevailing and plagioclase occcurs as relict mineral. In the absence of water vapours, muscovite crystallization may cease and microcline may form, both minerals occurring in high amounts within the same body. It is also possible that potash feldspar hydrolysis may generate postmicrocline muscovite according to the reaction

$$3\ KAlSi_3O_8 + 2H_2O \rightarrow KAl_2(OH)_2Si_3AlO_{10} + 2KOH + 6\ SiO_2$$

Muscovite, commonly not too largely crystallized, occurs along some fissures or fracture areas, filling them together with quartz. This muscovite is subsequent to the main evolution stages of pegmatite genesis, due to their partial revival as a result of the gaps inside the pegmatite bodies generated by tectonisation. Intense and prevailing mechanic deformations and the conditions characteristic of muscovite genesis favoured the appearance of "acicular muscovite" with bent lamellae, associated with fine-grained quartz, which accounts for the pressure conditions that had influenced the crystal growth.

The analytical data of Table 7 point to constant silica content and high Al_2O_3 amount varying from 29.95 to 41.06%, also in agreement with other authors' data (Ianovici, 1939; Murariu, 1979). These analyses also point to Fe^{3+}, Fe^{2+}, Mg^{2+}, Ti^{4} ions contained by muscovite which may substitute octahedral aluminium. Ferrous oxide contents are almost always higher than ferric oxide ones, agreeing with literature data mentioned above.

The diagram $(FeO + Fe_2O_3)/TiO_2$ (Fig. 27) shows two fields; the right side one with the plots of high TiO_2 contents characteristic of muscovites originating in biotite alteration and the left side field with plots of muscovite resulted from plagioclase or microcline hydrolysis.

The K content of muscovites varies from 8.10 to 11.12% K_2O with a mean value of 9.92% K_2O. The Na content is of 0.50–1.00% Na_2O.

The $K_2O : Na_2O$ ratio (Fig. 28) shows plots grouped within a horizontally elongated field, which accounts for independent K_2O and Na_2O contents.

TABLE 7

Chemical composition of some muscovites from pegmatites in Teregova – Marga area

Sample	377	369	Mu 1136	Mu 1106	35	41	49	57	64	70	1020	2206	2060_1	1009_5	B_1	N_1	N_2	T_1
Location	Ogaș Valley Măgu-ra area	Dobro-tin brook Măgu-ra area	Rîpi hill Tîlva area	Curcan hill Tîlva area	Cioaca cu Tei Măgu-ra area	Mărul brook Măgu-ra area	Cioaca Pie-troasă Măgu-ra area	Arme-niș	Tîlva hill Tîlva area	Scoar-ța Va-lley Măgu-ra area	Cioaca cu Mă-rul Măgu-ra area	Cer-mez Valley Măgu-ra area	Scoar-ța Va-lley Măgu-ra area	Arme-niș	Bău-țari Tîlva area	Tere-gova	Tîlva hill	Tîlva hill
SiO_2 %	46.00	44.58	44.42	45.08														
Al_2O_3 %	35.20	36.58	36.42	36.72	41.06	34.23	34.64	33.32	36.02	32.55	33.47	30.36	37.80	34.49	34.49	36.49	36.49	29.95
Fe_2O_3	1.10	0.89	1.09	0.87	1.20	1.30	0.82	1.40	1.17	0.97	0.81	1.00	0.70	1.54	1.45	0.58	0.58	1.07
FeO	0.67	0.95	0.81	0.59	0.97	0.82	0.67	0.82	0.82	0.82	0.90	0.75	0.82	0.60	0.42	0.97	0.97	0.67
MnO	0.015	0	0.015	0.018	0	0	0	0	0	0	0	0	0	0	0	0.02	0.02	0.03
MgO	1.48	0.29	1.08	1.30														
CaO	0.82	0.58	0.98	1.04														
Na_2O	0.80	1.00	0.90	0.90	0.87	0.87	0.50	0.62	0.67	0.75	0.80	0.50	1.00	0.50	0.75	0.75	0.92	0.75
K_2O	9.70	9.90	9.90	8.70	11.12	10.12	10.25	10.00	9.75	9.75	10.37	10.17	9.62	10.12	9.75	10.17	9.62	9.62
TiO_2	0.58	0.44	0.20	0.20	0.21	0.26	0.12	0.05	0.34	0.46	0.22	0.52	0.22	0.05	0.08	0.05	0.06	0.18
P_2O_5	0	0.07	0	0														
CO_2																		
H_2O	4.45	4.92	3.73	4.77														
TOTAL	100.81	100.20	99.54	100.18														

Analyst : Vasilica Neacșu

TABLE 8

Minor element content (ppm) of muscovites in Teregova-Marga area

No.	Sample	Location	Pb	Cu	Sn	Ga	Ni	Co	Cr	V	Sc	Be	Nb	Ba	Sr	Si
1	369	Dobrotin valley	6.5	2	15	67	6.5	12	1	19	48	1.8	35	1000	36	76
2	Mn 1106	Tîlva hill	10	3	30	77	13	6.5	5	55	10	7.5	60	1200	26	22
3	Mn 1136	Rîpi hill	8.5	<2	20	70	5	3.5	<1	2	20	<1	22	430	12	50
4	8 Mn	Ogaș valley	9	75	200	65	7	2	2	26	5.5	16	60	500	14	110

Taking into account the minor element content (Table 8), it is to mention the low Sn content (15–65 ppm) and the varying Cu one (2–75 ppm). Co and Ni amounts are sparse, 2–12 ppm and 5–13 ppm respectively.

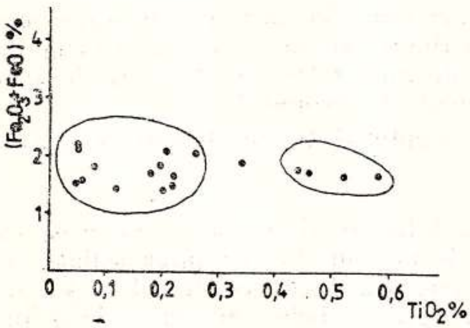


Fig. 27. ($FeO + Fe_2O_3$)/TiO_2 diagram of muscovites from Teregova-Marga area.

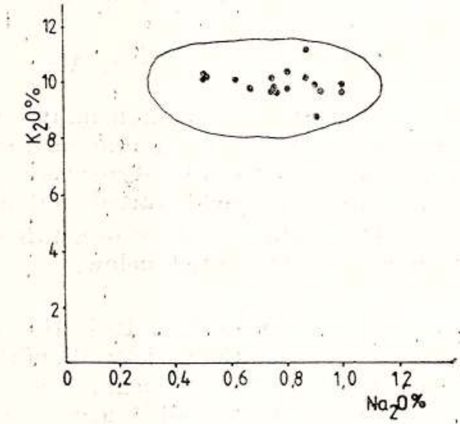


Fig. 28. Na_2O/K_2O diagram of muscovites from Teregova-Marga area.

The data mentioned above show that the muscovites yielded by these pegmatites are characterized by low minor element concentration coefficients, which, according to Stern (1966), is typical of metamorphic pegmatites.

6.3. Quartz

Quartz occurs in all pegmatite types, is associated with all pegmatite minerals and characterizes the different stages of their evolution. Quartz appears at the end of each crystallizing stage and contributes to the development of metasomatic textures. It is partly redistributed in each evolution stage, which makes difficult the recognition of different rock generations. The quartz from the central zone (III) usually marks the end of pegmatite evolution. It is translucent or of dark-grey colour due (acc. to Holden, 1920) to some free Si atoms resulted from irradiation. Far from the core, the quartz becomes of light grey or whitish colour and is of high temperature type.

During stage I quartz occurs together with plagioclase and biotite or forms graphic textures with plagioclase. During stage II it occurs together with microcline and microclineperthite or represents graphic textures within the latter. The transition from one stage to another is marked by myrmekite quartz prior to microcline generation. Stage II A is characterized by peculiar affinity between quartz and muscovite.

The minor element contents of a quartz sample collected from the central area of a pegmatite exposed on Mărul brook (Măgura area, analyst

Ana Şerbănescu), and of two samples collected from the quartz core of some pegmatites lying in the Ogaş hill (Măgura area, analyst Constanţa Udrescu) are the following : Cu 3–9 ppm, Ni 2–3.5 ppm, Zr 3–9 ppm, Ba 17–38 ppm, and only the first sample yielded B 30 ppm as a result of abundant tourmaline crystals near the central area.

6.4. Accessory Minerals

They show varying, usually low, contents in all pegmatite zones and associate with the main minerals. On the whole, most pegmatites contain very reduced amounts of accessory minerals or lack in them, while very few pegmatites yield varied and abundant amounts.

The main accessory minerals encountered in the area under discussion will be presented below.

6.4.1. **Tourmaline.** It is the most frequently encountered accessory mineral. The dark variety (Schörl) is present in the intermediate (to central) zone of some pegmatites (Teregova, Pîrîul cu Mărul – Măgura area, Curcan hill – Tîlva area), some grains being 20–30 cm long and associated with both microcline and quartz. Tourmaline also forms small prisms in the marginal aplite zone of several pegmatites, or radial prisms ("Sonnenturmalin"), parallel to the contact line with the surrounding rock, pointing to a volatile-abunding environment. Tourmaline may form at the expense of biotite (especially outside the pegmatite bodies) under the influence of B-rich solutions (Kunitz, 1929). In this case biotite is not altered to muscovite, but to schörlite. Owing to its low alkali content, tourmaline occurs in all stages of pegmatite genesis and associates with final quartz.

The large tourmaline grains were frequently corroded by quartz, resulting in „graphic tourmaline" (Pl. IX, Fig. 2 a, 2 b, Fig. 3). Besides quartz, fine-grained muscovite occurs on fissures inside the grains.

TA

Minor element content of some tourmaline samples

Sample	Location	Pb	Cu	Zn	Sn	Ga	Mo	Ni	Co	Cr
64_1	Curcan hill	5	6.5	850	43	48	<2	70	28	95
1106/80	Curcan hill	6.5	9	1150	16	41	3.5	15	<2	18
1126_3/80	Văruţ Valley	13	3.5	60	6.5	17	2	3	<2	<2

The minor element content of some largely crystallized tourmaline grains from the intermediate zone (Table 9) points to Zn 60–1150 ppm. The isomorphism between zinc and ferrous iron and magnesium accounts for the zinc substitution of these elements contained by tourmaline.

Relatively high Sn (16—43 ppm), Ga (17—48 ppm), Zr (10—130 ppm), Mn (10—1000 ppm) and Li (3—56 ppm) contents are worth mentioning; these values are close to those reported by Pomârleanu and Murariu (1970) from Teregova and Voislova tourmaline occurrences.

6.4.2. **Beryl.** Some large, discordant pegmatite bodies (Teregova, Pîrîul cu Mărul, Tîlva hill) contain important beryl amounts as greenish, hexagonal prisms, of decimetric size in places. It occurs in the intermediate zone within quartz or microcline. The pegmatites abunding in tourmaline also contain beryl accounting for similar genesis of the two minerals. Dittler and Kirnbauer (1931) are the first to report beryl occurrences at Teregova and to analyse them chemically; it is also cited by Avramescu (1954) and Savu (1977), who also calculated its crystallochemical formula:

$$(Be_{2,71}\,Na_{0,25}\,K_{0,06}\,Ca_{0,04})\ 3.06\ (Al_{1,74}\,Fe^{3+}_{0,06}\,Mg_{0,14})\,1.94$$

$$(Si_{5,78}\,Al_{0,22})\ 6\ (O_{17,27}\,OH_{0,73})\ 18$$

6.4.3. **Garnet.** The investigated pegmatites contain brown-dark red garnet, of probably almandine character, in intermediate zones (e.g. Văruț Valley, Dalci-Var area), represented by large idiomorphic grains (ca 1 cm in diamater) within plagioclase, quartz and few muscovite groundmass (Pl. IV, Fig. 2).

Cherry red coloured garnet associates with abundant translucid final quartz which may be slightly corroded, as well as with microcline or microclineperthite, and a few muscovite (e.g. Pîrîul cu Mărul body). Its colour is probably due to a high Mn content which resulted in spessartite enrichment, these garnets being characteristic of the last stages of pegmatite genesis. According to Vlasov (1961) garnets alter their composition during pegmatite processes, the increase of Mn content taking place from outside innwards.

BLE 9

rom pegmatites in Teregova — Marga area

V	Sc	Y	Yb	Zr	Be	B	Nb	Ba	Sr	Mn	Ti	Li
120	13	14	<1	130	8.5	> 3000	<10	80	90			40
8	<2	16	2.5	60	5.5	> 3000	14	34	<10	> 1000	370	56
2.5	<2	<10	<1	10	13	1000	18	10	12	80	72	<3

These large garnet grains point to the high pressure which influenced the pegmatite genesis.

Fine grained, rosy garnet also occurs on the fissures which cross the less chloritized pegmatite body (e.g. Armeniș).

TA

Chemical composition of some pegmatite samples

No	Sample	Location	OXI				
			SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO
1	40	Mărul brook — Măgura area	73.74	15.00	sld	0.46	0.13
2	41/4	Mărul brook — Măgura area	71.26	17.00	,,	0.21	0.13
3	50	Cioaca cu Mărul-Măgura area	77.67	12.65	,,	0.31	0.13
4	51	Mărul brook — Măgura area	74.19	15.45	,,	0.03	0.11
5	50/1	Cioaca cu Mărul-Măgura area	65.68	18.75	,,	0.08	0.16
6	52	Cirniș-(Mărul brook)-Măgura area	73.23	16.70	,,	0.19	0.10
7	53	Armeniș	73.23	15.80	,,	0.09	0.10
8	54	Armeniș	70.97	16.65	,,	0.09	0.10
9	56	Armeniș	74.36	16.65	,,	0.86	0.13
10	57	Armeniș	72.33	16.40	,,	0.23	0.06
11	58	Teregova	69.40	16.25	,,	0.24	0.10
12	58/1	Teregova	70.77	15.65	,,	0.34	0.10
13	59	Teregova	71.17	15.90	,,	0.14	0.10
14	60	Armeniș	73.48	17.50	,,	0.36	0.22
15	62	Sat Bătrîn-Armeniș area	73.42	16.75	,,	0.26	0.13
16	63	Slatina Timiș	72.97	16.40	,,	0.43	0.16
17	64	Tîlva	87.58	8.40	,,	0.33	0.16
18	67	Var	76.46	15.25	,,	0.18	0.16
19	70	Scoarța Valley — Măgura area	69.08	19.40	0.05	0.46	0.08
20	74	Scoarța Valley — Măgura area	70.57	17.66	0.04	0.03	0.16
21	1754	Găina Mare Valley — Măgura area	73.67	15.23	0.075	0.39	0.55
22	1767	Strau Valley — Dalci-Var area	71.80	16.15	sld	0.36	0.31
23	2273	Armeniș	73.69	16.40	sld	0.57	0.39
24	2492	Curcan hill — Tîlva area	71.96	15.37	sld	0.55	0.23
25	2494 A	Curcan peak — Tîlva area	72.08	14.10	sld	0.16	0.31
26	2494 B	Curcan peak — Tîlva area	73.33	12.73	sld	0.36	0.31
27	2495	Curcan peak — Tîlva area	71.95	17.02	0.075	0.64	0.32
28	2518	Rea Valley (Valea Satului)-Dalci-Var area	71.30	14.70	sld	0.15	0.23
29	2275/2	Armeniș	75.20	15.03	traces	0.10	0.11
30	2273/2	Armeniș	76.00	14.90	0.07	0.05	0.14
31	Rb 4B	Ogaș hill — Tîlva area	79.40	13.73	0.08	0.10	0.11
32	1108/80	Armeniș	73.00	15.33	0.03	—	—
33	1106/80	Curcan hill — Tîlva area	73.30	16.30	0.10	0.86	0.56
34	1108/60/1	Armeniș	74.30	14.63	0.06	0	0.07
35	Rn A 1	Ogaș hill — Tîlva area	77.20	14.35	0.03	0.06	0.11
36	1107	Slatina-Timiș	72.50	15.60	—	0.22	0.36
37	1126	Var	74.93	14.50	—	0.20	0.28
38	1136	Rîpi hill — Tîlva area	76.22	13.00	0.025	0.21	0.72
39	1238	Găina Valley-Dalci-Var area	75.77	15.25	—	0.10	0.14
40	2103/2	Sasa Valley — Măgura area	72.05	17.62	—	0.92	0.38
41	352	Dobrotin brook — Măgura area	73.69	15.22	—	0.11	0.12
42	367	Dobrotin brook — Măgura area	71.78	17.62	—	0.17	0.03

Analyst: samples 1—28 Movileanu Aurelia, 29—35

BLE 10
in Teregova-Marga area (Semenic-Țarcu mountains)

DES %

MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O^-	H_2O^+	S	Total	Fe_2O_3 Total
0.013	0.05	0.87	4.30	4.40	0.085	0.18	0.79		100.01	
sld	0.10	0.56	3.00	8.00	0.095	0.16	0.47		100.98	
0.005	0.40	0.59	5.00	2.70	0.050	0.26	0.24		100.00	
0.005	sld	0.77	3.40	6.10	0.105	0.08	0.50		100.73	
sld	0.27	1.68	9.80	3.60	0.105	0.28	0.22		100.62	
0.010	sld	0.80	3.70	5.60	0.105	0.14	0.02		100.59	
sld	sld	1.12	8.00	1.30	0.105	0.28	0.07		100.09	
sld	sld	0.56	3.50	8.20	0.075	0.18	0.12		100.44	
0.013	sld	0.40	4.30	3.00	0.025	0.16	0.97		100.86	
sld	0.10	0.35	3.50	7.60	0.70	0.14	0.07		100.85	
0.020	sld	0.45	10.60	2.50	0.41	0.14	0.20		100.31	
0.029	sld	0.45	9.80	2.60	0.54	0.14	0.19		100.60	
0.005	0.05	0.35	8.00	4.20	0.35	0.12	0.33		100.71	
sld	0.15	0.87	6.80	1.00	0.040	0.08	0.53		101.03	
0.010	0.15	0.87	8.00	0.50	0.025	0.16	0.37		100.64	
0.005	0.07	0.50	4.30	5.00	0.050	0.06	0.57		100.51	
sld	0.27	0.20	1.20	1.60	0.05	0.06	0.34		100.19	
0.005	0.07	0.50	5.10	2.30	0.095	0.20	0.53		100.84	
sld	0.40	0.50	4.20	6.10	0.10	0.14	0.18		100.69	
0.16	0.02	0.52	3.50	6.40	0.025	0.18	0.28		99.54	
0.030	0.55	0.70	4.40	4.90	0.110	sld	sld		100.60	
0.012	sld	3.22	6.50	1.70	0.550	,,	,,		100.61	
0.020	0.30	1.61	6.70	0.80	0.090	,,	,,		100.57	
sld	0.45	0.98	3.00	7.80	0.320	,,	,,		100.66	
0.020	0.28	0.70	4.00	8.50	0.100	0.10	0.57		100.42	
0.024	0.57	0.70	3.00	8.70	0.580	0.04	0.24		100.58	
0.020	sld	2.10	4.40	2.40	0.250	0.12	0.04		100.52	
sld	0.79	1.05	1.70	9.50	0.850	1.23	0.20		100.51	
0.03	0.09	0.71	2.37	5.55	0.12		0.42	0.03	99.79	
traces	0.10	1.08	0.23	0.94	0.03		0.64	0.03	100.24	0.26
0.02	0.19	0.83	4.70	0.49	0.05		0.45	0.03	100.21	0.25
—	0.09	0.32	3.02	8.06	0.15		0.22	0.03	100.28	0.26
0.28	0.25	1.25	3.50	0.94	0.12		0.54	0.03	100.02	—
—	0.09	0.63	4.33	4.83	0.17		0.44	0.03	99.61	1.48
0.02	0.13	1.44	6.09	0.52	0.16		0.42	0.03	100.59	0.11
0.040	0.40	1.26	3.90	5.50	0.090	0.16	0.42		100.45	0.22
0.010	0.50	0.70	4.50	4.50	0.135	0.34	0.27		100.86	
0.012	0.50	1.05	2.60	5.20	0.275	0.20	0.58		100.58	
—	0.50	1.05	5.00	1.80	0.125	0.16	0.56		100.46	
0.02	0.87	0.70	2.46	3.40	0.13	0.13	1.11	0.05	99.77	
0.018	0.64	1.84	5.30	1.90	0.07	—	0.95		99.85	
—	0.85	1.70	2.80	6.80	0.06	—	0.88		99.68	

Vlad Catrinel, 36—42 Neacșu Vasilica.

TABLE 10 A

Chemical composition of some pegmatites in Teregova — Marga area

No	Location	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P.C.	Total
1	Slatina Valley, Semenic Mts	69.21	14.50	0.20	—	5.25	traces	6.14	3.72	traces	0.02	0.15	99.49
2	Goleţu Valley, Semenic Mts	69.07	18.04	0.81	—	0.10	0.21	6.04	5.10	absent	traces	0.53	99.92
3	Făurar Valley, Semenic Mts	69.58	17.86	0.81	—	0.23	0.78	5.12	5.34	traces	traces	0.80	100.52
4	Cerneţ Valley, Semenic Mts	69.85	17.40	0.22	—	0.80	0.47	6.03	3.95	traces	0.04	0.68	99.44
5	Goleţu Valley, Semenic Mts	70.00	16.46	0.71	—	0.10	0.36	3.80	8.82	absent	traces	0.23	100.48
6	Slatina Valley, Semenic Mts	71.56	17.90	0.15	—	0.72	0.55	5.64	2.78	traces	0.01	0.43	99.74
7	Lung hill-Luncaviţa, Semenic Mts	71.30	19.00	0.31	0.22	0.25	2.70	4.70	2.06	—	—	—	100.54
8	Poienii hill-Teregova, Semenic Mts	67.00	20.50	0.17	0.12	0.07	1.00	7.50	3.84	—	—	—	100.20
9	Bibanul brook-Armeniş, Semenic Mts	71.30	17.80	—	1.20	0.40	1.00	5.00	1.90	—	—	—	98.60
10	Bibanul brook-Armeniş, Semenic Mts	65.40	18.40	—	0.50	0.40	0.50	9.10	2.40	—	—	—	96.70
11	Cerbul brook-Teregova, Semenic Mts	67.00	21.60	—	1.00	0.20	1.10	5.10	1.90	—	—	—	97.90
12	Dalci, Muntele Mic	69.10	18.85	0.16	—	0.06	0.90	4.02	6.65	—	0.04	—	99.78
13	Socetu Mare, Muntele Mic	71.52	17.03	0.27	—	0.10	0.63	1.88	7.00	—	0.02	—	98.45
14	Socetu Mare, Muntele Mic	71.55	17.48	0.24	—	0.13	1.39	4.55	4.12	—	0.03	—	99.49
15	Socetu Mare, Muntele Mic	73.06	19.02	0.47	—	0.18	0.48	3.36	2.68	—	0.06	—	99.31
16	Pietroasa Valley, Socetu Mare-Muntele Mic	71.34	16.36	0.31	—	0.07	0.32	3.09	8.31	—	0.04	0.53	100.37
17	Pietroasa Valley, Socetu Mare-Muntele Mic	71.06	17.42	0.39	—	0.10	0.60	3.28	6.25	traces	0.02	0.61	99.93
18	Pietroasa Valley, Socetu Mare-Muntele Mic	74.09	15.50	0.43	—	0.09	0.71	2.81	5.96	—	0.05	0.74	100.38
19	Găina brook, Dalci-Muntele Mic	69.80	17.70	0.27	0.19	0.49	1.00	6.80	4.20	—	—	—	100.45
20	Dîlma brook, Dalci-Muntele Mic	67.90	21.80	0.21	0.15	0.32	1.40	4.40	3.80	—	—	—	99.98

Analysts : samples 1—6 Mînzatu et al. (1958), 12—18 Mînzatu, Mînzatu (1957) 7—11 and 19—20 Avramescu (1954).

Different pegmatite types correspond to the different structural-mineralogic stages of pegmatites genesis. Pegmatites with intermediate features are also added. The relations among pegmatite minerals point to the succession of evolution stages, which, by partial superposition, generate zoned pegmatites. Each zone of a pegmatite body represents a petrogenetic type which may constitute an independent pegmatite.

Plagioclase and biotite are typical of stage I, followed by quartz which under certain conditions generates graphic textures. Stage II — microcline or stage II A — muscovite depend on the pressure of water vapours. In an open system and low H_2O pressure microcline crystallizes, while in closed system and high H_2O pressure muscovite is generated, the entire process being influenced by the prevailing tectonic conditions. During this stage the "microcline block" zone defined by Vlasov (1952), namely a zone with large muscovite plates, is formed. Biotite is deferrized, but it may occur as relict mineral. Quartz is present again, substituting microcline and resulting in graphic textures or (during stage II A) acting as catalyser which facilitates the development of muscovite grains. This quartz type (stage III) marks the end of pegmatite evolution by concentrating in a quartz core. In certain cases, this was followed by slight albitisation which generated the substitution perthites, but not a proper albite zone illustrative of an independent stage as mentioned by Ghinsburg (1960), Vlasov (1952). However, the pegmatite evolution is concluded by quartz, graphically developed in replacement microclineperthites and generating the quartz core.

The different features implied by the minor element content of pegmatite minerals account for the relatively reduced intensity of substitution, which together with the reduced contents of accessory minerals account for the metamorphic origin of pegmatites under discussion.

7. GEOCHEMISTRY OF PEGMATITES

The chemical sampling of pegmatites is rather difficult, as an accurate method based on all their characteristics is needed ; the large grain size of minerals, zoning, fissures or fractures filled with later formed minerals, accidental distribution of different accessory minerals, etc. are to be considered. Even if all these features are taken into account, the result will be subjective, as the selection imposed by practical conditions is objective.

Taking into account the chemical analyses (Tables 10, 10 A) several features related to the evolution and characteristics of major and minor elements of pegmatites may be defined.

The sequence of different evolution stages of pegmatite genesis reflected by the mineralogic characteristics may also be transferred to the chemical properties : NaCa (I), K (II) or $K + H_2O$ (II A) and SiO_2 (III), each stage being marked by the presence of quartz, and the appearance of final quartz is preceded by albitisation (Na) (replacement microclineperthites) which does not represent a proper stage within these pegmatites, as no independent crystals are present.

The characteristics of pegmatites geochemical evolution are shown by different diagrams. Thus, on SiO_2/Al_2O_3 diagram (Fig. 29) were plotted the results of the chemical analysis of some paragneisses collected from the rock pile which hosts the pegmatites (Table 11). It is to note a main

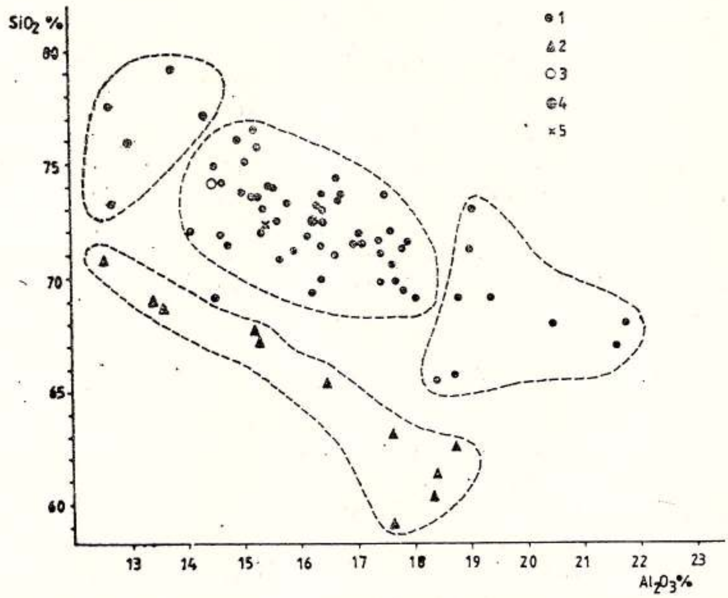


Fig. 29. SiO_2 / Al_2O_3 diagram of pegmatites from Teregova-Marga area. 1, pegmatites; 2, paragneisses; 3, Tröger mean; 4, mean value of Teregova-Marga area; 5, mean of graphic textures.

TABLE 11

Chemical composition of some paragneisses from Teregova-Marga area

Sample	359	360	366	367	368	369	370	371	372	373	374	375
SiO_2	67.64	67.25	61.03	62.98	62.32	70.85	68.74	57.48	58.81	59.98	65.46	69.01
TiO_2	0.90	0.88	0.87	0.88	0.93	0.57	0.73	1.11	1.23	0.93	0.77	0.61
Al_2O_3	15.21	15.24	18.41	17.61	18.77	12.51	13.59	19.44	17.66	18.31	16.46	13.37
Fe_2O_3	0.57	0.72	0.73	1.05	0.65	0.73	0.80	1.51	1.41	1.23	0.68	1.01
FeO	4.25	4.35	4.99	4.89	4.88	4.71	4.76	5.76	6.36	5.42	4.36	4.69
MnO	0.08	0.07	0.06	0.06	0.05	0.04	0.04	0.10	0.12	0.12	0.06	0.03
MgO	2.03	1.97	3.25	2.55	2.52	2.23	1.82	3.35	3.85	2.79	2.44	1.97
CaO	1.53	1.52	2.01	1.70	1.89	2.09	2.40	1.92	1.60	1.77	1.86	2.75
K_2O	2.70	2.85	2.64	2.54	2.50	1.60	1.52	3.67	3.83	3.46	2.22	1.50
Na_2O	2.31	2.19	3.20	2.90	3.06	2.60	3.15	3.20	2.54	3.23	3.19	3.05
P_2O_5	0.24	0.24	0.22	0.19	0.17	0.16	0.27	0.34	0.22	0.17	0.31	0.18
H_2O^+	1.66	1.66	2.11	2.14	1.82	1.36	1.58	1.87	1.92	2.01	1.85	1.30
CO_2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
S	0.26	0.34	0.09	0.07	0.09	0.14	0.13	0.03	0.08	0.08	0.07	0.18
Fe(S)	0.22	0.29	0.08	0.06	0.08	0.12	0.11	0.02	0.06	0.06	0.05	0.15
Total	99.60	99.57	99.69	99.62	99.73	99.71	99.64	99.80	99.69	99.56	99.78	99.80
Fe_2O_3tot.	5.70	5.96	6.38	6.56	6.18	6.13	6.24	7.94	8.55	7.33	5.59	6.47

Analyst: Elena Colios

field of accumulation of plots between 75% SiO_2 and 18% Al_2O_3. The Tröger mean (acc. to Schneiderhöhn, 1961) occurs at the top of this field, unlike the mean of investigated pegmatites. There is a left field and a right field with fewer plots. The left field represents the mainly muscovite pegmatites (quartz rich and attached to the preceding plagioclase member), and the right field contains the plots of mainly microcline pegmatites. At the base of this diagram the plots of paragneisses form an elongated field parallel to the three pegmatite fields pointing to the simultaneous evolution and interdependence of the two rock types during the metamorphic processes.

The diagram Na_2O/K_2O (Fig. 30) points to an almost uniform and widespread occurrence of pegmatite plots, and the concentration of paragneiss plots at the base of the diagram. The Tröger mean appears to the right of the diagram and does not correspond to the pegmatite mean.

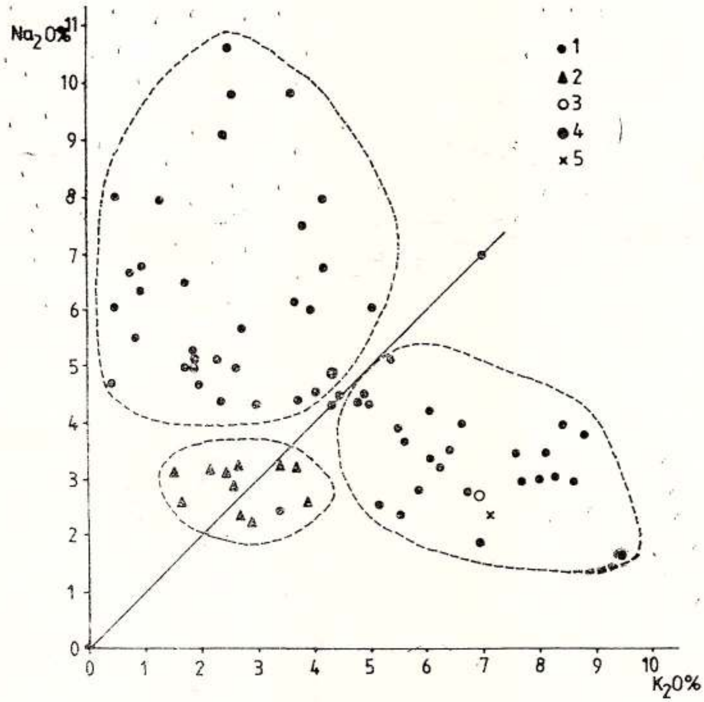


Fig. 30. Na_2O/K_2O diagram of pegmatites from Teregova-Marga area. 1, pegmatites; 2, paragneisses; 3, Tröger mean; 4, mean value of Teregova-Marga area; 5, mean of graphic textures.

The plots of the mean of graphic textures appear to the right of the diagram too, as most of them are of microcline character.

The diagonal line delimits the plots in two fields : above and below it. According to Mehnert (1971) the plots above the diagonal line correspond to the plots of graywacke gneisses. It may be inferred that some pegmatites were generated *in situ* from anatectic solutions of pegmatoid character mobilized from surrounding micaschists and paragneisses, while other pegmatites originate in anatectic pegmatite fluids generated by gneisses, which may migrate from greater depth and may form complex pegmatites in certain conditions.

Diagrams SiO_2 : Al_2O_3/ K_2O : Na_2O and $Na_2O + K_2O$: K_2O + + Na_2O (Figs. 31, 32) have been drawn in view of a varied characterisation of pegmatites based on their major element contents.

According to the diagram of Figure 31, the different pegmatite petrographic types may be divided horizontally as follows : a field with maximum concentration of plots within a field with fewer plots, both

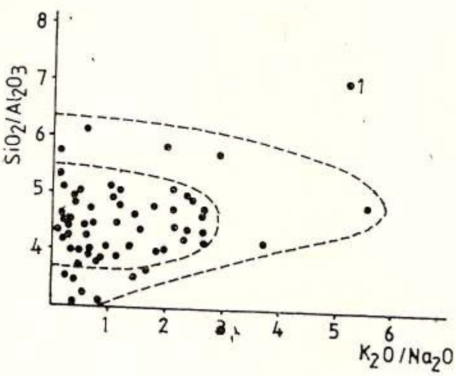


Fig. 31. SiO_2 : Al_2O_3 /K_2O : Na_2O diagram of pegmatites from Teregova-Marga area. 1. pegmatites

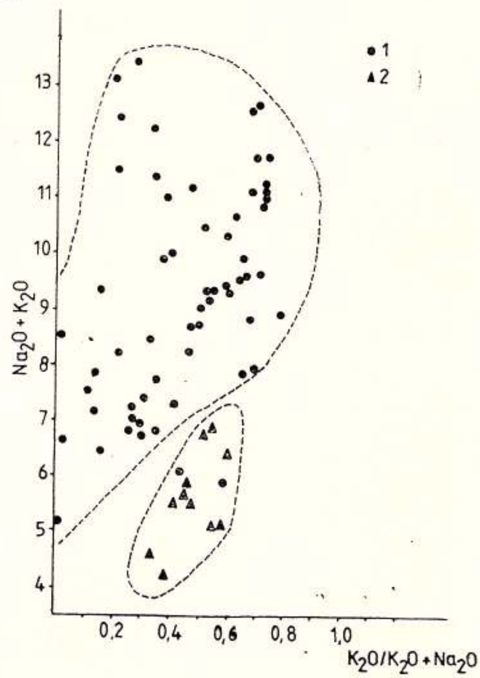


Fig. 32. $Na_2O + K_2O$: $K_2O/K_2O + Na_2O$ diagram of pegmatites from Teregova-Marga area. 1, pegmatites ; 2, paragneisses.

TABLE 12

Minor element content (ppm) of some pegmatites samples in Teregova — Marga area

No	Sample	Location	Pb	Cu	Sn	Ga	Ni	Co	Cr	V	So	Y	Yb	Zr	Be	B	Ba	Sr	Li
1	40	Mărul brook-Măgura area	48		5	15	2.5	2	1	2	2	15	1	60	2	30	165	42	7
2	41/4	Mărul brook-Măgura area	80	4.5	2	15	2	2	1	2	2	10	1	10	1.7	30	950	100	8.5
3	50	Cioaca cu Mărul-Măgura area	17	4.5	7	9	3	2	1.5	2	2	10	1	10	7.6	30	65	10	5
4	50_1	Măgura area	42	5	2	11	3	2	1	2	2	10	1	10	5.8	30	310	90	3
5	52	Pietroasa (Cirniș) Valley Măgura area	60	6.5	2	13	2	2	1	2	2	10	1	12	1	30	315	45	7
6	54	Armeniș	90	4.5	2	10	2.5	2	1	2	2	10	1	10	1.4	30	60	50	3
7	56	Armeniș	55	8	5	16	2.5	2	1	2	2	10	1	24	5	36	570	65	3
8	57	Armeniș	90	4.5	2	7.5	2	2	1	2	2	11	1	10	1.3	30	100	30	3
9	58	Teregova	26	6	3	15	2	2	1	2	2	10	1	22	130	30	950	95	3
10	58_1	Armeniș	24	8.5	2	15	2	2	1	2	2	10	1	26	140	30	20	10	3
11	64_2	Tilva	18	7.5	2	3.5	2	2	2	2	2	10	1	10	4.8	30	19	10	3
12	35_2	Dobrotin brook-Măgura area	30	3	2	12	2.5	2	1	2	2	10	1	10	1	30	5200	200	3.5
13	367	Dobrotin brook	70	5	2.5	10	2	2	1	3	2.5	10	1	21	1.9	30	300	170	4.5
14	$1108/80_B$	Armeniș	39	3	2	15	4	2	4	2	2	10	1	10	2.1	30	90	67	3
15	$1108/80_A$	Armeniș	90	4	2	9	3	2	2.5	2	2	10	1	10	1.4	30	790	67	3
16	$1126_1/80$	Var Dalci-Dalci Var	36	4	6.5	15	4	2	2	2	2	14	1	16	3.1	35	38	12	3
17	1106	Curcan—Tilva hill-Tilva area	48	3	2	10	4	2	1	2	2	11	1.4	10	5	30	85	230	3
18	1126_a	Var, Dalci-Var area	20	3.5	4	15	2	2	1.5	2	2	10	1	17	11	30	43	30	4.5
19	1126_b	Var, Dalci-Var area	12	3	4	16	2	2	1.5	2	2	10	1	17	11	30	42	35	7
20	Rn 4.B	Ogaș hill-Tilva area	22	10	2	12	9.5	2	8	7	3	10	1	10	1.9	30	380	100	3
21	Rn $A.A_1$	Ogaș hill-Tilva area	42	9	2	23	7	2	5	6	2	10	1	10	6	30	110	130	3
22	Rn $A.A_2$	Ogaș hill-Tilva area	34	8	2	18	8	2	4.5	2.5	3	10	1.1	38	5.6	30	130	380	3
23	53	Armeniș	55	4	2	13	2	2	1.5	2	2	26	1	12	2.3	30	80	55	3
24	2103_2	Strîmba Valley-Dalci Var area	28	4.5	2	13	8	3	5	16	4	14	1	49	1.7	30	1100	300	10
25	1027	Teregova	22	16	5.5	38	2	2	1.5	3.5	2	10	1	10	280	70	50	16	6
26	1011_3	Armeniș	75	5.5	3	18	2	2	2	3	2	17	1	10	2.6	30	200	72	3

27	1015_b	Armeniș	42	4	2	13	2	2	1	2	2	10	1	12	2.3	30	75	84	3
28	40_2	Mărul brook-Măgura area	10	5	8.5	28	2.5	2	2	3	4	10	1	10	1	30	85	10	3
29	65	Var-Satului Valley-Dalci-Var area	20	5	2	9	2.5	2	2	3	2	10	1	35	6.2	30	85	230	3
30	1016_2	Armeniș-highroad	30	10	3.5	19	2	3	1	9.5	4	18	1	37	9	37	550	300	3
31	1021_1	Cioaca Pietroasă-Măgura	30	3.5	4.5	13	2	2	1	2	2	10	1	33	6.2	600	30	15	5
32	1012_5	Armeniș	05	3	2	6	2	2	1	2	2	10	1	10	1	30	2800	110	3
33	1027	Teregova	82	4	2	9.5	3	2	1	2	2	10	1	10	44	30	300	16	3
34	1008_1	Armeniș	54	3.5	2	11	2	2	1	2	2	11	1	22	1	30	240	40	3
35	1017_1	Armeniș-highroad	25	7.5	2	20	2	4.5	1	8	2	10	1	15	6.7	40	200	910	3
36	1007_1	Armeniș	24	3	5	24	2	2	2.5	2	3.5	10	1	50	1.8	30	15	52	3.5
37	1009_3	Armeniș	110	4	4	18	2	2	1	2	2	10	1	24	1.7	30	340	55	3
38	1191	Dalci	28	6	2	10	6.5	2	4	3	2	10	1	10	2.5	30	5600	550	3
39	1022	Mărul brook-Măgura area	37	3	2	10	2	2	1	2	2	10	1	36	2.7	20	375	50	3
40	1125	Var	20	4	3	10	3	2	1	2	2	10	1	11	5.5	30	160	950	3
41	1119	Var	18	5	4	10	7	2	2	2	2	10	1	10	5.6	30	120	850	9
42	1023	Pietroasa (Cirniș) Valley Măgura area	36	3.5	2	16	2	2	1	2	2.5	17	1	16	1.3	30	100	36	9

Analyst : samples 1–10, 12–13, 17, 20–22 and 30–38 Constanța Udrescu, samples 11, 23–29 Irina Bratosin, samples 14–16, 18–19 and 39–43 Ana Șerbănescu pointing to the pegmatite evolution owing to the alkali content and thus accounting for its importance to the geochemical characterisation of pegmatites.

The diagram of Figure 32 shows distinct fields with the plots of paragneisses and pegmatites respectively. However, it is to note the similarity of the two fields all over the area of paragneiss occurrence. At the top of the paragneiss field, their plots are similar to pegmatite plots, partly parallel to them. Then, the pegmatite field covers the top of the diagram. This accounts for the pegmatite origin in the gradual division and migration of alkaline mobilisates (together with quartz) from the rock piles during anatectic metamorphic processes.

In view of a thorough geochemical characterisation the pegmatite samples were also analysed spectrally (Table 12). The minor elements (B, Li, Be, Zr, Y, Yb, Sn, etc.) characteristic of granitic pegmatites do not show high contents. The results of these analyses have been compared to the minor element content of some paragneisses (Table 13) : the vana-

TABLE 13

Minor element content of some paragneisses from Teregova-Marga area

No.	Sample	Pb	Cu	Ga	Sn	Ni	Co	Cr	V	Sc	Zr	Nb	Y	Yb	La	Ba	Sr
1	359	3	7	16	2	60	16	95	100	19	320	17	45	4.8	50	600	170
2	360	3	7	16.5	2	50	13	85	85	16	240	14	32	3.2	56	600	160
3	366	10	16	23	3	48	10	80	95	17	180	10	19	2.0	30	400	320
4	367	8	19	23	3.5	48	12	90	105	17	210	12	24	2.3	38	420	300
5	368	8	8.5	22	2.5	48	11	67	80	15	160	14	20	1.8	32	420	500
6	369	3	5	16.5	3.5	50	13.5	90	90	16	260	12	30	2.8	30	270	90
7	370	3	8	18	4	48	15	85	95	19	260	14	27	3.2	30	300	110
8	371	11	6.5	20	2.5	66	22	125	130	21	170	12	33	3.4	44	500	240
9	372	8	9	19	3	66	25	110	130	21	190	13	27	2.2	46	500	170
10	373	13	38	23	3	65	23	120	140	22	210	10	30	2.0	42	600	270
11	374	13	10	19	3	62	14	95	110	20	270	12	29	3.0	44	380	400
12	375	3.5	6	24	4	55	19	95	120	20	280	13	29	3.6	30	280	100

Analyst : Constanța Udrescu

dium of paragneisses did not migrate to pegmatites as the latter yielded very reduced contents (26 ppm at the most, compared to 130 ppm reported from paragneisses) ; zinc contents of pegmatites are of 10—60 ppm, and of rocks are of 160—320 ppm, while chromium and nickel contents of paragneisses are of 67—125 ppm and 48—66 ppm respectively, higher than those of pegmatites (1—4.5 ppm 2—8 ppm respectively). Beryl content of pegmatites is relatively high (1.3—280 ppm). Lithium amounts are usually reduced. The biotite-rich pegmatites also have high Li contents, as it may substitute Mg and Al of the mineral network. Proper minerals are exceptionally formed (Superceanu (1957) mentions spodumene and lepidolite at Teregova). It may also lack, mainly in microcline pegmatites, as for example those from southern Norway (Björlikke, 1937).

The rather monotonous and reduced minor element content of pegmatites does not favour the drawing up of diagrams relevant of the petrogenetic characteristics. However, the diagram of Figure 33 is an attempt at rendering the variation of some minor elements compared to the sum total of alkalis. Besides Pb and partly Ba, marked by positive correlation, the other elements (Sr, Cu, Ga) occur in reduced amounts, which correspond to Hornung's data (1962). Stan(1977) studied lead and barium substitution during feldspathisation. The differentiation of alkaline fluids of the paragneiss pile results in pegmatite quartz-feldspar mobilisates.

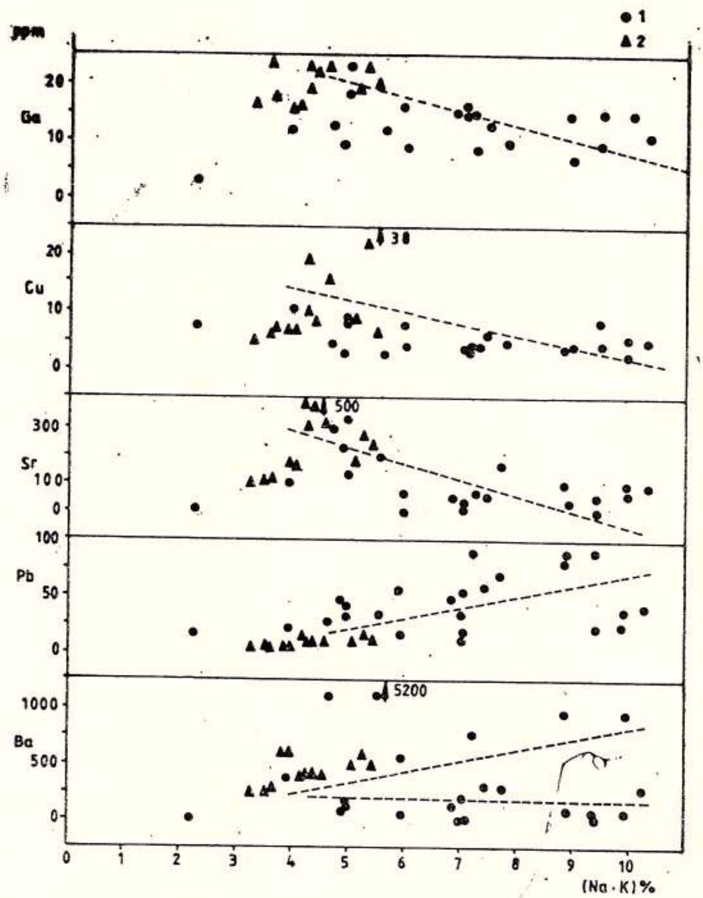


Fig. 33. Variation of some elements compared to alkali sum total of Teregova-Marga pegmatites. 1, pegmatites; 2, paragneisses.

8. PEGMATITE GENESIS

All structural, mineralogic and geochemical data account for the metamorphic origin of the investigated pegmatites, their genesis by magmatic processes being excluded. There is no evidence of any connection with an igneous mass, no feeding channels are present and there is no trace of shifting or disappearance during the deformations which affected the rock pile. No systematic zonal variations of pegmatite characteristics are noted and no areal distribution against surrounding granitoid massifs relevant of the more or less rhythmical pulsatory character of pegmatite emplacement is present. In fact, Savu (1977) demonstrated the absence of petrogenetic relations between the pegmatites in the Semenic Mts and the Poniasca granitoid pluton.

The metamorphic genesis of pegmatites implies the concomitant or successive development of independent genetic processes within the same rock pile affected by metamorphism. This may be proved by the different morphostructural features, also coinciding by convergence of phenomena.

An example is given by the lens shape. In folded areas, the maximum thickness of lenses is shown by the fold axis, where local pressures are lower and the pegmatites formed concomitantly with the folds. Some lenses show features accounting for migmatisation processes (diffuse graded contact, Fig. 13 a, b ; "swarm"-like arrangement, Fig. 4), others show the characteristics of metamorphic pegmatites of ,,dilation" type (alternating graded contact), while a third category of lenses originates in the boudinage of concordant veins or the fragmentation of some pegmatite dykes due to unaffine lamination, parallel to paragneiss foliation. The concordant pegmatite veins may also be the result of some "dilation" metamorphic processes, as described by Goodspeed (1940) and recognized by Gherasi et al. (1974) in the Măru-Voislova region, or of the pegmatite evolution of quartz-feldspar mobilisates.

The morpho-structural features also account for other pegmatite types of metamorphic origin. Thus, there are discordant pegmatites resulted from metasomatosis and marked by "nondilation" pegmatitization (Ramberg, 1952), while the structural features of surrounding rock have been partly preserved as relics within the pegmatite groundmass (Fig. 11). The boundary of these pegmatites is very sinuous and the contact with the surrounding rock is of mixed nature – either clear or diffuse. Pegmatites of this type have been reported by Maier et al. (1975) from Băuțari, in the Poiana Ruscă Mts. Pegmatite dykes are also generated by anatectic metamorphic processes. These pegmatite bodies show clear contact (usually of large size), and along the contact line there are fine-grained pegmatite intrusions in the surrounding rock (Fig. 10 a) accounting for the subsequent intrusion of anatectic mobilisates, probably along some fractures.

,,Concretion" pegmatites (Ramberg, 1952, Barth, 1962) resulted from concretion growth, of porphyroblastic character outside, starting from some cores at the expense of a solid host rock (Fig. 9 a).

The study of mafic concentrations at the periphery of some pegmatite bodies also points to the different petrological processes which influenced the constitution of metamorphic pegmatites. The mafic border may result from anatectic differentiation (restites – Scheumann, 1937 ; Mehnert, 1951, 1962), or from metasomatic processes between the host rock and the pegmatitic mobilisate ("mafic front" – Reynolds, 1946, 1949), or from lateral secretion due to diffusion (Ramberg, 1952).

Some of the pegmatite bodies result from lateral secretion (Fig. 9 b). The metasomatism is proved by the spatial distribution of stages during these processes, marked by matter supply and release. The generation of pegmatites by lateral secretion was described by many authors, as for example Ramberg (1952), Barth (1962), Gresens (1967), Şeclăman (1972). It is to mention that the rock cavities or fissures due to high pressure and temperature are incompatible with the equilibrium state. Therefore, the basic gradients (free energy gradients implicitly) appear and cause the migration of rock fluids and mineral components from high pressure areas to lower pressure cavities or fissures. As the mobilisation speeds of different minerals are not equal, the cavities are filled with the easiest mobilized minerals (feldspar and quartz), namely the minerals consisting of alkalis and silica easily soluble in aqueous fluids. The initiation and development of lateral secretion processes are due to the constitution of rock cavities during the folding accompanied by the generation of shearing fissures and strain fissures. In ductile metamorphic environments deprived of any proof of vein shearing, the fissures are the result of hydraulic force (Yardley, 1975) : as far as the crystalline schists represent anisotropic environments, the hydraulic fissuring with a certain trending may also occur in the absence of any stress deviation, along some low resistance planes. According to Şeclăman (1972) the fissures result from unaffine laminations of rock piles. The mineral substance is transported by diffusion, represented in the solid stage (intracrystalline) by the crystal network of minerals or by their discontinuities, imperfections respectively, and by the intergranular fluid solutions or intergranular veins through stationary pore fluid contributing to substance transfer (Yardley, 1975). The aqueous fluid of rocks contains volatile elements which influence the partial water pressure affecting the formation of pegmatitic minerals. For example, potash feldspar is subsequent to plagioclase and it appears by decrease of water pressure below the lower limit of mica stability. The diffusion of a certain substance is caused by its chemical potential. Diffusion takes place from the high chemical potential area to the low chemical potential and low pressure area, resulting in the thermo-dynamic condition of diffusion (Şeclăman, 1972).

Most of the pegmatites were generated by metamorphic processes called "anatectic-metasomatic" by Barth (1962). They may be generally characterized as "lateral secretion" processes according to Holmquist (1924) who mentioned the "lateral secretion venites" originating in the surrounding rocks. The mobilisation of initially solid rock fragments and the presence of liquid components as molecular dispersion within the solid/ liquid boundary area result in the accumulation of mobile components in veins and irregular masses — metatects, obviously different from the solid components — restites (Scheumann, 1937, Mehnert, 1951, 1968). The characteristic features are effaced in the palyngenetic field and lateral secretion processes are no longer noted.

The crystalline schists of the Sebeş-Lotru Group were generated by high-grade metamorphism which also implied some anatectic processes, as proved by the different types of migmatitic textures. Winkler (1966, 1967) and von Platen (1965) present the thermo-baric conditions (pressures of 10 kb and temperatures of 600—700°C) which influenced these processes and caused the segregation of leucocrate mobilisates from micaschists and gneisses. In appropriate conditions (the prevalence of a volatile stage, favourable tectonic conditions) a pegmatite evolution takes place. According to the evolution stages crossed simple or compound pegmatite bodies appear.

The pegmatites commonly occur in areas with highly migmatized rocks and lack from those areas deprived of these occurrences, thus accounting for their anatectic genesis. The sillimanite (Armeniș) genetically related to muscovite (muscovite $\pm$ quartz $=$ sillimanite $+$ potash feldspar) also demonstrates the anatectic genesis of pegmatites. The subsequent pegmatite evolution is marked by gradual cooling of pegmatites, as demonstrated by the temperature at which some pegmatite minerals were formed (Pomârleanu, in Hann et al., 1977): microcline between 554° and 600°C, biotite between 560°C and 585°C, muscovite between 285°C and 470°C and beryl between 237°C and 460°C. The relation between the Ba content of feldspars and the crystallisation temperature accounts for decreased temperature of alkali feldspar generation in the Semenic-Țarcu pegmatites.

With respect to the structural-mineralogic features of the pegmatite evolution, it is to note their generation by a main mineral stage (the hardly mobile anatectic fluids) consisting of plagioclase, biotite-microcline, microclineperthite-muscovite-quartz, which replace each other and account for the pegmatite evolution stage. Plagioclase and biotite are characteristic of stage I, followed by quartz, which also generates graphic textures. Microcline and muscovite mark the second stage. Microcline is metasomatically developed on plagioclase, and muscovite results from biotite deferrization or plagioclase replacement, at the expense of a preexisting sericite. The mineralogic features depend on the pressure of water vapours. An open system and decreased H_2O pressure result in microcline (II) crystallization, while an equally open and increased H_2O pressure generate muscovite (II A). Thus, the tectonic framework influences the entire process. Quartz substitutes microcline and generates the graphic textures, or (stage II A) favours the growth of muscovite grains. The metasomatic character of this mineral is also evinced by the study of graphic textures. The data mentioned above account for the selective substitution of feldspar by quartz. Quartz (III) marks the end of pegmatite evolution and concentrates in the quartz core. Albitisation is also present, resulting in perthitic textures. The microscopic study points to its metasomatic character. A proper albite zone representing an independent stage is absent. However, the pegmatite evolution ends with quartz graphically developed in the substitution microclineperthites, subsequently forming the quartz core.

Therefore, one may state that different pegmatite types correspond to the different structural-mineralogic stages. The relations between pegmatitic minerals point to the sequence of evolution stages, which commonly replace each other and by partial superposition result in zoned pegmatites. Each zone of a pegmatite body represents a petrogenetic type which, isolated, constitutes an independent pegmatite.

The geochemical study evinces that the sequence of pegmatite evolution stages is also revealed by the chemical characteristics: NaCa

(I) — K (II) or K + H_2O (II A) and SiO_2 (III), that is a sequence of alkaline stages (with quartz intermediate stages) ending with final quartz. The geochemical data usually account for the interdependence between pegmatites and surrounding rocks during the metamorphic processes, as well as for the segregation and gradual migration of alkaline mobilisates and quartz contained by the rock pile during the anatectic processes. The reduced minor element content of pegmatites informs on their metamorphic origin, while the diagrams show that the differentiation of alkaline fluids results in quartz-feldspar pegmatite mobilisates.

The accessory minerals occur in reduced amounts and accompany the different major mineralogic stages. Metamorphic pegmatites are also characterized by reduced minor element contents of different pegmatite minerals, as proved by the Sebeș-Lotru pegmatites too.

The large pegmatite dykes (e.g. Teregova, Pîrîul cu Mărul, Tîlva hill), characterized by complete evolution and varied mineralogic features, resulted from ascending migration of pegmatite fluids along several important fractures and imply the occurrence of some anatectic-palyngenetic cores, thus engendering the characteristics of typical magmatic pegmatites. The convergence of phenomena is significant again, as far as different petrogenetic processes, even opposite ones, determine similar mineralogic and structural features. These pegmatites are the first products of the palyngenetic processes and show the same features as the last products of magmatic differentiation.

9. CONCLUSIONS

There are different opinions on the pegmatite genesis. Most of the authors, however, assign the large dykes to the magmatic origin; as regards the other pegmatite bodies, some are considered magmatic, others metamorphic.

The main petrographic type from the area under discussion is represented by paragneisses, but pegmatite bodies are frequently reported from micaschists or mica paragneisses and pierce the amphibolites in places. Due to the high grade of metamorphism (sillimanite isograde) frequent migmatisations are noted, and a positive correlation is mentioned between the intesity of this process and the pegmatite occurrence. The pegmatite bodies are lithostratigraphically related, at least north of Muntele Mic, to highly migmatized micaschists and mica paragneisses.

Among the different pegmatite shapes, the most frequent are: lenses, concordant veins, dykes, nests and large irregular bodies. The contact between the pegmatite bodies and surrounding rocks is also described: sharp contact (tectonic or normal), graded contact (diffuse or alternating) and mixed contact. The contact type and the shape of the pegmatite body may inform on their genesis by different metamorphic processes. The mafic accumulations at the periphery of some pegmatite bodies may result from distinct petrologic processes within a metamorphic field; it is worth mentioning that different phenomena generate identical features.

The investigated pegmatite bodies are zoned (complete, incomplete, symmetrical, asymmetrical, simple or complex zoning), unzoned or show incipient zoning; other bodies show destroyed zoning. Most of the small lenses are unzoned and their inner structure is heterogeneous, unzoned and unhomogeneous. The prevailing minerals of these lenses are plagioclase and quartz. The larger lenses also exhibit zoning, and contain besides plagioclase, quartz and micas, microcline which substitutes the plagioclase. The large concordant veins are completely zoned : a contact aplite zone, a marginal zone consisting of plagioclase, quartz and muscovite $\pm$ biotite, an intermediate zone, mainly containing microcline and a central zone built up of quartz, which is the last to crystallize. Graphic textures characterize both the marginal and the intermediate zones. In other cases, it is to note only a contact zone, an intermediate zone and the quartz core. If microcline prevails in the intermediate zone, plagioclase and muscovite are relics of a former marginal zone. It is to infer that during this stage of pegmatite genesis, the K-metasomatosis phase was prevailing in anhydrous state within a closed system. If muscovite prevails in the intermediate zone and associates with quartz and plagioclase, K-metasomatosis took place in the conditions of water preservation which favoured the crystallization of muscovite. The small dykes are unzoned or show simple zoning and contain enclaves of the surrounding rock accounting for their generation by metasomatic processes. The large dykes, less numerous, often show complete zoning, which means that all evolution stages characteristic of pegmatite genesis had been crossed, while some rare minerals (beryl, monazite, tantalite) determine their complex nature.

The graphic textures result from quartz intergrowths with both plagioclase and microcline or microclineperthite. They formed by metasomatic corrosion, that is due to selective substitution of feldspar by quartz along deformation stripes, which constitute the reticular labile zones associated in one or several systems intersecting each other. Their metasomatic origin is demonstrated by the fact that the different positions of deformation stripes correspond to the positions of the quartz lamellae, by the radial symmetrical graphic textures pointing to ever increased feldspar substitution by quartz next to the source. The geochemical study of graphic textures points to the graphic texture areas within the geochemical evolution of pegmatite bodies. The varying amounts of quartz and other minerals, as well as the correlation fields plotted on Ba : K, Ba : Sr, Sr : Ca and $(Na + K)\%$ diagrams account for graphic textures generated by metasomatic processes, not by simultaneous crystallization.

The main mineral phase of pegmatites consists of biotite, acid plagioclase, microcline, muscovite, quartz. These minerals determine the characteristic zoning, the geochemical evolution of pegmatites and, by replacing each other, point to the evolution stage of pegmatites. Accessory minerals — garnet, tourmaline, beryl, apatite, columbite, orthite, kyanite, sillimanite — accompany the different stages of major mineralogic composition. Most of the pegmatites under discussion show simple mineralogic features, characteristic of metamorphic pegmatites.

Feldspars are the most important pegmatite minerals as regards both their amount and the fact that the feldspar type points to the evolution stage attained by a pegmatite. Feldspars also account for the deformations which had affected the pegmatites, implicitly the changes of tectonic conditions. The acid plagioclase (albite-oligoclase, oligoclase) represents the first evolution stage (I) generating initial pegmatites. Alkali feldspars represented by microcline or microclineperthite occur in the triclinic modification, being metasomatically developed at the expense of plagioclase and constituting the second evolution stage (II). Most of the perthitic textures are metasomatic due to abundant irregular shapes and uneven disappearance as shown by the microprobe analysis.

Mica occurrences also characterize different stages of pegmatite genesis. Owing to the changes underwent (especially by biotite) and to their sensitiveness to mechanic deformations (noticed in the case of muscovite), micas evince both changes of chemistry typical of different evolution stages and the control of tectonic conditions during pegmatite evolution. Biotite is characteristic of stage I, crystallizing together with plagioclase. Later on, it becomes unstable, is deferrized and altered to muscovite by increase of pressure of water vapours. Muscovite amounts are considerable and constitute large plates associated with quartz and plagioclase; however, muscovite is subordinate in microcline pegmatites. Muscovite replaces plagioclase, at the expense of pre-existing sericite, being accompanied by quartz. It may also result from biotite alteration. It represents stage II A of pegmatite evolution marked by K-metasomatosis abunding in water vapours.

Quartz occurs in all pegmatite types, associates with all pegmatite minerals and characterizes the different stages of their evolution. It is redistributed in the successive evolution stages, while central quartz (III) ends the pegmatite evolution.

Accessory minerals are few or even absent from most of the pegmatites and only in a few instances they are varied and abundant. The most frequent accessory mineral is tourmaline, preserved in all stages owing to its low alkaline character. The pegmatites abunding in tourmaline also contain beryl amounts in the intermediate zone. Garnet, characteristic of the high pressure which influenced the pegmatite genesis, occurs in the intermediate zone and next to the quartz core, altering its composition from outside innerwards by increasing its Mn content.

Finally, the different structural-mineralogic stages of pegmatite genesis are represented by different pegmatite types. The relations among pegmatite minerals point to the sequence of evolution stages, commonly replacing each other, and generating zoned pegmatites by partial superposition. Each zone of a pegmatite body represents a petrogenetic type, which isolated, may form an independent pegmatite.

The different features implied by the minor elements yielded by pegmatitic minerals account for the relatively low intensity of substitutions, which, together with the reduced amount of accessory minerals, accounts for the metamorphic origin of these pegmatites.

The sequence of evolution stages, shown by the chemical features, is the following: NaCa (I), K (II) or $K + H_2O$ (II A) and SiO_2 (III); quartz precedes each stage and final quartz is preceded by albitisation

(Na) — substitution microclineperthites which do not represent an independent stage as no independent crystals are formed.

The different diagrams, based on bulk analyses of pegmatitic rocks, point to the simultaneous evolution of pegmatites and paragneisses as well as to the interdependence of the two rock types during metamorphic processes. Thus, the geochemical data prove the generation of pegmatites by segregation and gradual migration of alkaline mobilisates and of quartz from the rock pile during the anatectic metamorphic processes.

The minor element content of the investigated pegmatites is reduced and rather monotonous. The diagrams regarding the differentiation of alkaline fluids from the paragneiss pile point to the generation of quartz-feldspar pegmatitic mobilisates.

The origin of studied pegmatites should be assigned to the metamorphic field, as proved by all mineralogic and geochemical structural data. No relations with a granitic magmatic mass are defined. The metamorphic genesis of these pegmatites implies the simultaneous or successive development of some independent petrologic processes within the same rock pile affected by metamorphism. Thus, some pegmatites have resulted from metasomatosis, marked by „nondilation" pegmatitisation, while others are of „dilation" or „concretion" type. Some pegmatite bodies are the product of lateral secretion : the fluids and mineral components of rocks migrate toward the rock cavities or fissures and because of varying mobilisation speeds, the latter are filled with the easiest mobilized minerals — feldspars and quartz. The mineral substance is transported by diffusion. The aqueous fluid of rocks consists of volatile components which influence the genesis of pegmatitic minerals.

However, most of the pegmatites under discussion were generated by „anatectic-metasomatic" metamorphic processes which are in fact (prior to palyngenesis) „lateral secretion" as they consist of mobilisates supplied by surrounding rocks. Under favourable conditions (a volatile phase, adequate tectonic conditions) some of these mobilisates undergo pegmatitic evolution. According to its duration, the evolution results in simple or complex pegmatite bodies. Commonly, these pegmatites accompany the highly migmatized rocks, while the sillimanite amounts of pegmatites account for anatectic genesis.

The large pegmatite dykes, characterized by complete evolution and varied mineralogic features, due to convergent phenomena, resembling to magmatic-granitic pegmatites. They resulted from ascending migration of pegmatitic fluids along some fractures and account for the occurrence of some anatectic-palyngenetic cores.

The investigation of pegmatites from Teregova-Marga region also informs on the main trends of prospect studies of pegmatites. Further investigation of pegmatites is based on both their special petrogenetic features and the considerable amounts of muscovite, rare minerals and feldspars of real economic interest.

New data may be obtained from field investigation of different pegmatite types and of the contact relations with surrounding rocks, as far as new features and new interpretations are to be expected.

The structural-mineralogic study of pegmatites, mainly the relations among pegmatite minerals, also informs on the pegmatite genesis.

The geochemical study, based on bulk analyses of pegmatites, brings no new data because of the practical difficulty of obtaining accurate mean values. However, the most important means of future investigations of pegmatites are offered by the study of pegmatitic minerals, of the deformations which had affected them and of their geochemical characteristics. New progress is expected by using up-to-date methods of investigation of crystallochemical characteristics of pegmatitic minerals. All detailed data informing on pegmatite genesis improve the conclusions inferred.

Acknowledgements

I would like to offer my grateful thanks to professor dr. doc. Dan Rădulescu for having guided my geologic thinking and scientific activity since student years.

I am grateful to dr. Nicolae Gherasi for having suggested this topic of study to me and for his first useful advice in the field work.

I thank my colleague Radu Constantinescu for numerous useful discussions on this subject and for helping me with the processing of chemical data.

I am grateful to dr. Mircea Săndulescu for having facilitated my investigations in the area under discussion.

I also thank my colleagues Ana Șerbănescu, Antoneta Seghedi, Ana Baralia, Ovidiu Șerbănescu and Dumitru Danci for helpful contribution to elaborating the graphic material.

[3] These characteristics account for the relation with migmatisation processes being characteristic of a certain stage of leucosoma pegmatite evolution.

[4] Features considered typical of dilational type metamorphic pegmatites (Goodspeed, 1940, Gherasi et al,, 1974).

REFERENCES

Andersen O. (1938) The genesis of some types of feldspar from granite pegmatites. *Norsk. Geol Tidskrift.*, 10, p. 114–205, Oslo.

Augustithis S. S. (1962) Researches of blastic process in granitic rocks and later graphic quartz in pegmatites (pegmatoids) from Ethiopia. *Nova Acta Leopoldina*, N. F. 25, 156, p. 5–17.

— (1974) Atlas of the textural pattern of granites, gneisses and associated rock types. Elsevier Scientific Publishing Company. Amsterdam, London, New York.

Avramescu C. (1954) Report, archives of the Institute of Geology and Geophysics, București.

Bagdasarian G. P. (1972) Despre vîrsta absolută a unor roci eruptive și metamorfice din masivul Ditrău și munții Banatului din România. *Acad. R.S.R. Stud. cerc. geol. geofiz. geogr., Ser. geol.*, 17, 1, p. 13–21, București.

Barth T. W. F. (1962) Zur Genese der Pegmatite im Urgebirge. *N. Jahrb. Mineral., Beih. Bd.*, 58 A, p. 385–432.

Bercia I (1975) Metamorfitele din partea centrală și de sud a masivului Godeanu. *Stud. tehn. econ. Inst. Geol. Geofiz.*, I, 12, p. 3–159, București.

Björlikke H. (1937) The granite pegmatites of Southern Norway. *Amer. Mineral.*, 22, p. 241–268.

Böckh J. (1879) Auf den südlichen Teil des Komit. Szöreny bezügliche geologische Notizen. *Földt. Közl.*, IX, p. 20–35, Budapest.

— (1883) Geologische Notizen von den Aufnahmen des Jahres 1883 im Komitate Krasso-Szöreny. *Földt. Közl.*, XIII, p. 50–81, Budapest.

Cameron E. N., Jahns R. H., McNair A. H., Page L. R. (1949) Internal structure of granitic pegmatites. *Econ. Geol. Mon.*, 2, 115 p, Urbana (Ilinois).

Codarcea F., Stoenescu V. (1957) Report, archives of the Institute of Geology and Geophysics, București.

Constantinescu R. (1974) Report, archives of the Institute of Geology and Geophysics, București.

Deaș A., Isăilă N. (1957–1958) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— Rădulescu D., Ștefănescu D. (1968) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Demeter I. (1971) Report, archives of the Enterprise for Geological Prospectings and Explorations, Caransebeș.

Diaconu Fl. (1979) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Drescher-Kaden F. K. (1969) Granitprobleme. Akademie-Verlag, 586 p., Berlin.

Erdmannsdörffer O. H. (1941) Myrmekit und Albitkornbindung in magmatischen und metamorphen Gesteinen. *Zentralbl. Min.*, A, p. 41–55, Stuttgart.

Eskola P. (1946) Kristalle und Gesteine. 397 p., Wien.

Fersman A. E. (1915) Pismennaia structura pegmatitov i pricinî eë vozniknovenia. Izbrannîe trudî, I, p. 1–30, Moscova.

— (1929) Die Schriftstruktur der Granit-Pegmatite und ihre Entstehung, *Z. Krist.* 69, p. 77–104, Stuttgart.

— (1952) Les pegmatites. I. Les pegmatites granitiques. (French translation of the first edition – 1931). Louvain et Bruxelles, 1952, 671 p.

Fischer W. (1929) Zu Tom Barth's Bemerkungen über die Natur des Schriftgranites. *Zentralbl. f. Min.*, A, p. 391–396.

Gherasi N. (1951) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— Zimmermann P. (1968) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— , Zimmermann P., Matsch E., Hann H. (1969, 1970) Reports, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— , Zimmermann P., Zimmermann V. (1974) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Ghinsburg A. J. (1960) Specific geochemical features of the pegmatitic process. Report of the 21st session Norden, 17, Kopenhagen.

Goodspeed G. E. (1940) Dilation and replacement dikes. *J. Geol.*, 48, p. 175–195.

Gresens R. L. (1967) Tectonic-hydrothermal pegmatites. *Centr. Mineral. Petrol.*, 15, p. 345–355, Washington.

Gridan T. (1981) Petrologia Semenicului de nord-est. Edit. Acad. R.S.R., 194 p., București.

Grosu A., Angelescu M. (1960) Metodica de cercetare a pegmatitelor din țara noastră. *Rev. Min.*, anul XI, p. 1–12, București.

Gunnesch K., Gunnesch M., Vlad C. (1978) Considerații petrografice și petrochimice asupra banatitelor din zona Teregova-Lăpușnicel (munții Semenic). *St. cerc. geol. geofiz., geogr., ser. Geol.*, 23, 2, p. 239–248, București.

Hann H. P. (1973, 1976) Reports, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— Pomârleanu V., Movileanu A. (1977) Report, archives of the Institute of Geology and Geophysics, București.

— Szász L. (1983) Structura geologică a văii Oltului între Ciineni și Brezoi (Carpații Meridionali). *D. S. Inst. Geol.*, LXVIII/1, p. 23–27, București.

— (1983) Zur Deutung der Eklogitvorkommen im Căpățina-massiv (Süd-Karpaten). *Rev. roum. géol. géophys., géogr., sér. Géol.*, p. 15–21, București.

Hauer Fr., Stache G. (1885) Geologie Siebenbürgens. Wien.

Higgins M. D. (1971) Cataclastic rocks. *Geological Survey Professional Papers*, 687, Washington.

Hârtopanu I. (1976) Cristalinul getic : metamorfism polifazic sau polimetamorfism?. *St. cerc. geol. geofiz. geogr., ser. Geol.*, 23, 2, p. 185–193, București.

Heritsch H., Sonnleitner P. (1950) Über die Verteilung von Rechts und Linksquarzen in Schriftsgraniten. *Öster. Akad. Wiss., math,-nat. Kl. Anzeiger*, 87, p. 338–339, Wien.

Holden E. F. (1920) The cause of color in smokey quartz and amethyst. *Amer. Min.*, 10, p. 203–252.

Holmquist P. J. (1924) Typen und Arten der Adergesteine. *Geol. Fören Stockholm Förh.*, 43, p. 612–631.

Hornung G. (1962) Wall rock composition as a guide to pegmatite mineralization. *Econ. Geol.*, 57, p. 1127–1130.

Hurduzeu L. (1962) Cercetări geologice și petrografice în partea centrală a munților Semenic. *D. S. Com. Geol.*, XLV (1957–1958), p. 215–227, București.

Ianovici V. (1939) Etude minéralogique sur les micas de la région de Voineasa dans les Monts Lotru. *Ann. Sci. Univ. Iași*, XXV/2, p. 463–473.

Inkey B. V. (1891) Die Transilvanischen Alpen vom Rothenturmpass bis zum Eisernen Tor. *Math. Naturwiss. Ber. aus Ungarn.*, IX, I. Hälfte, Budapest.

Jahns R. H. (1955) The study of pegmatites. *Econ. Geol. Annivers.* 1955, p. 1025–1130.

Kirnbauer F., Dittler E. (1931) Über das neue Beryll-Vorkommen von Teregova in Rumänien. *Zeitschr. f. prakt. Geol.*, 39, 4, p. 59–64.

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

— , Kräutner Fl., Hann H., Iliescu V., Udrescu C., Colios E. (1981) Report, archives of the Institute of Geology and Geophysics, București.

Kunitz W. (1929) Die Mischungsreihen in der Turmalingruppe und genetische Beziechungen zwischen Turmalinen und Glimmer. *Chemie der Erde*, 4, Jena.

Lacroix A. (1922) Minéralogie de Madagascar, Paris.

Landes K. K. (1925) The paragenesis of the granite pegmatites of Central Maine. *Amer* Min., 10, p. 335–411.

Lupu M., Popescu B., Szász L., Hann H., Gheuca I., Dumitrică P., Popescu Gh. (1978) Harta geologică a R. S. România, sc. 1 : 50.000, foaia Vînturarița (Olănești). Inst. Geol. Geofiz., București.

Maier O., Solomon I., Zimmermann P., Zimmermann V. (1975) Studiul geologic și petrografic al cristalinului din partea sudică a munților Poiana Ruscă. *An. Inst. Geol.*, XLIII, p. 67–189,București.

Malloe S. (1974) The cristallization of simple pegmatites in Moss area, southern Norway. *Norsk Geologisk Tidskrift*, 54, 2, p. 149–167, Oslo.

Marinescu E., Ardeleanu F. (1956) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

— Marinescu I., Biriescu I. (1973) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Mînzatu S., Mînzatu E. (1957, 1958), Reports, archives of the Institute of Geology and Geophysics, București.

Mărunțiu M. (1978) Considerații preliminare asupra caracterelor morfostructurale și a evoluției geologice a rocilor ultrabazice asociate seriilor cristaline precambriene din Carpații Meridionali, *St. cerc. geofiz. geogr. geol.*, 23, 2, p. 215—227, București.

Mehnert K. R. (1951) Zur Frage des Stoffhaushalts anatektischer Gesteine. *Neues Jahrb. Mineral., Abhandl.*, 82, p. 155—198, Stuttgart.

— (1962) Zur Systematik der Migmatite. *Krystallinikum*, p. 95—110, Praga.

— (1971) Migmatites and the origin of granitic rocks. 2nd impression, Elsevier publishing company, 406 p., Amsterdam, London, New York.

Micșa L., Gall V. (1956) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Mrazec L. (1897) Essai d'une classification des roches cristallines de la zone des Carpates roumaines. *Arch. sc. phys. et nat., Génève*, III, p. 1—4, Génève.

Murariu T. (1979) Studiul mineralogic, geochimic și structural al pegmatitelor din munții Rodnei. *Inst. Geol. Geofiz. Stud. tehn. econ.*, I 15, 264 p., București.

Murgoci G. M. (1905) Sur l'existence d'une grande nappe de recouvrement dans les Carpates Méridionales. *C.R. Ac. Paris*, 1905, 31 juillet *(Bull. Soc. Științe)*, XVI, 1907, p. 50—52 București.

— (1908) Terțiarul din Oltenia. *An. Inst. Geol. Rom.*, p. 1—128, București.

Paraschivescu C., Serghie R. (1963) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Platen H. von (1965) Experimental anatexis and genesis of migmatites. Controls of metamorphism. Edit. Pitcher and Flinn, Oliver and Boyd, p. 203—218, Edinburgh, London.

Pomârleanu V., Murariu T. (1970) Beitrag zum Studium von Turmalinen aus Pegmatiten der S. R. Rumänien. *Ges. geol. Wiss. B. Miner.* Lagerstätten, 15, 2, Berlin.

— , Movileanu A. (1968) Temperatura de cristalizare a muscovitului din diverse pegmatite. *Rev. Min.*, 10, București.

Popescu A., Ștefan R. (1964) Report, archives of the Institute of Geology and Geophysics București.

Rădulescu D., Săndulescu M. (1973) The Plate-Tectonics Concept and the Geological Structure of the Carpathians. *Tectonophysics*, 16, p. 155—166, Amsterdam.

Rădulescu I., Rădulescu L. (1957, 1958) Reports, archives of the Institute of Geology and Geophysics, București.

Ramberg H. (1949) The facies classification of rocks : a clue to the origin of quartzo-feldspathic massifs and veins. *Journ. Geol.*, 57, Jan.-Nov., 1949, p.18—54, Univ. of Chicago Press.

— (1952) The origin of metamorphic and metasomatic rocks. Univ. of Chicago Press, 317p.

Reynolds D. L. (1946) The sequence of geochemical changes leading to granitization. *Quart. J. Geol. Soc. London*, 102, p. 389—446.

— (1949) Observations concerning granite. *Geol. Mijnbouw.* 11, p. 241—263.

Roșca L. (1954) Comunicare preliminară asupra cercetărilor geologice și petrografice din regiunea munților Semenic de nord. *D. S. Com. Geol.*, XXXVIII (1950—1951), București.

— (1954) Report, archives of the Enterprise for Geological and Geophysical Prospectings, București.

Savu H. (1962) Report, archives of the Institute of Geology and Geophysics, București.
— Micu C. (1964) Contribuții la cunoașterea geologiei și petrografiei părții centrale a munților Semenic. *D. S. Com. Geol.*, XLIX, p. 39—50, București.
— (1965) Structura în virgație a cristalinului munților Semenic. *D. S. Com. Geol.*, LI/1 (1963—1964), București.
— , Vasiliu C. (1970) Asupra unui amfibol hastingsitic din zona cu sillimanit a munților Semenic. *St. cerc. geol. geofiz. geogr. ser. Geol.*, 15, 2, p. 547—552, București.
— (1970) Stratigrafia și izogradele de metamorfism din provincia metamorfică prebaicaliană din munții Semenic. *An. Inst. Geol.*, XXXVIII, p. 223—311, București.
— , Maier O., Bercia E., Schuster A. C., Berza T., Hârtopanu I. (1975) Report, archives of the Institute of Geology and Geophysics, București.
— (1977) Genesis of pegmatites from Banat (Romania). *D. S. Inst. Geol. Geofiz.*, LXIII (1976), p. 99—111, București.
— Hann H. P., Năstăseanu S., Marinescu Fl., Morariu A., Rogge-Țăranu E. (1981) Harta geologică a R. S. România, sc. 1 : 50. 000, foaia Muntele Mic. Inst. Geol. Geofiz., București.
— , Hann H. (1982) Mineralizațiile de sulfuri de la Turnu Ruieni-Borlova (Banat). *D. S. Inst. Geol. Geofiz.*, LXVI (1979), p. 127—138, București.
Schädel J. (1961) Untersuchungen zur Bildungsfolge der Mineralien in den Drusen der Granite von Striegan (Schlesien). *Nova Acta Leopoldina*, N. F., 135, p. 1—32.
Schaller W. T. (1927) Mineral replacement in pegmatites. *Am. Mineral.*, 12, p. 59—63.
Scheumann K. H. (1937) Metatexis und Metablastesis. *Tschermaks Mineral Petrogr. Mitt.*, 48, p. 402—412.
Schneiderhöhn H. (1961) Die Erzlagerstätten der Erde. Band. II. Die Pegmatite, 720 p., Gustav Fischer Verlag, Stuttgart.
Seifert K. E. (1965) Deformation bands in albite. *Amer. Miner.*, 50, p. 1469—1472.
Șeclăman M. (1971) Contribuții la cunoașterea structurilor grafice. *St. cerc. geol. geofiz. geogr. ser. Geol.*, 1, 16, p. 133—147, edit. Acad. R.S.R., București.
— (1972) Studiul rocilor cuarțo-feldspatice din cursul superior al văii Streiului. Teză de doctorat, Universitatea București.
— , Constantinescu E. (1972) Metasomatic origin of some micrographic intergrowths. *Amer. Mineral.*, 57, p. 932—940.
Simpson D. R. (1962) Graphic granite from the Ramona pegmatite district, California. *Amer. Miner.*, 47, 1, p. 123—138.
Stan N. (1977) Feldspathization processes in the crystalline Lainici-Păiuș Series (Vilcan Mountains-Romania). *An. Inst. Geol. Geofiz.* VII, p. 6—98, București.
Stern W. B. (1966) Zur Mineralchemie von Glimmer aus Tessiner Pegmatiten. *Schweiz. Miner. Petr. Mitt.*, 46, 1, p. 137—189.
Streckeisen A., Gherasi N. (1932) Recherches géologiques dans la partie orientale des Carpates Méridionales. *C. R. Inst. Géol. Roum.*, XX, Bucarest.
Superceanu C. (1957) Mineralele rare din pegmatitele granitice din Banat. *Rev. min.*, 3, p. 140—155, București.
Suru V. (1966) Report, archives of the Enterprise for Geological Prospectings and Explorations, Caransebeș.
Tuttle O. F. (1952) Origin of the contrasting mineralogy of the extrusive and plutonic salic rocks. *J. Geol.*, 60, p. 107—124.
Uspensky N. M. (1943) On the genesis of granite pegmatites. *Amer. Min.*, 28, p. 437—447.
Vlad Ș. (1979) A survey of banatite (Laramian) metallogeny in the Banat region. *Rev. roum. géol. géophys. géogr., ser. Géol.*, 23/1, p. 39—44, Bucarest.

— Gheorghiță J., Gheorghițescu D., Hann H., Iosof V., Vlad C., Bratosin I., Vanghelie L., Anastase Ș. (1980) Report, archives of the Institute of Geology and Geophysics, București.

Vlasov K. A. (1952) Texturno parageneticeskaia klassifikatiia granitnih pegmatitov. *Izv. A. N., S S S R., ser. gheol.*, 2, Moskva.

— (1961) Prinţipi klassifikaţii granitnih pegmatitov i ih texturno-parageneticeskie tipi. *Izv. A.N.S.S.S.R., ser. gheol.*, 1, Moskva.

Vogt J. H. L. (1908) Über anchi-monomineralische und anchi-eutektische Eruptivgesteine. *Vidensk Selsk. Skr. Kristiana, M—N. Kl.*, 10, Oslo.

— (1929) The physical chemistry of the magmatic differentiation of igneous rocks. *Skrifter Utgitt av Det Norske Videnskaps-Akademi i Oslo, I Mat. Naturw. Klasse*, Oslo.

Wahlström E. E. (1939 a) Graphic granite. *Amer. Miner.*, 24 (11), p. 681—694.

— (1939 b) Graphic granite. *Amer. Miner.*, 42, p. 859—888.

Winkler H. G. F. (1966) Der Prozess der Anatexis : Seine Bedeutung für die Genese der Migmatite. *Tscherm. Min. Petrogr. Mitt.*, XI, 3—4, p. 266—287, Springer Verlag, Wien-New York.

— (1967) Die Genese der metamorphen Gesteine. Springer Verlag, p. 1—237, Berlin—Heidelberg—New York.

Yardley B. W. D. (1975) On some quartz-plagioclase veins in the Connemara schists, Ireland. *Geol. Mag.*, 112 (2), p. 184—189.

STUDIUL PETROGRAFIC AL PEGMATITELOR DINTRE TEREGOVA ȘI MARGA (BANATUL DE EST, CARPAȚII MERIDIONALI)

(Rezumat)

În cadrul teritoriului ocupat de rocile mezometamorfice ale grupului Sebeș-Lotru din Carpații Meridionali, regiunea cercetată, situată atît în estul munților Semenic cît și în nordul Masivului Muntele Mic, prezintă o importanță deosebită pentru studiul pegmatitelor datorită apariției în această zonă a unui număr însemnat de diferite corpuri de pegmatite. Localizarea și prezentarea principalelor corpuri de pegmatite se face ținînd cont de faptul că acestea au în anumite sectoare o răspîndire mai mare. Astfel deosebim în regiunea situată la nord de Muntele Mic, sectoarele Tîlva, Măgura, și Dalci-Var, iar în cadrul munților Semenic, sectoarele Slatina-Timiș, Armeniș și Teregova. Șisturile grupului Sebeș-Lotru reprezintă unitatea Pînzei Getice, care, la nord de Muntele Mic încalecă, în lungul planului de șariaj getic peste amfibolitele seriei de Măru care aparțin domeniului Danubian. La vest de linia de șariaj apare în interiorul pînzei getice o digitație care delimitează solzul Turnu Ruieni.

Fondul petrografic al regiunii este constituit din paragnaise $\pm$ sillimanit $\pm$ disten în care se întîlnesc intercalații și nivele constituite din gnaise albe cuarțo-feldspatice, micașisturi, amfibolite, calcare cristaline etc. Stiva de roci prezintă frecvent intense migmatizări. Succesiunea litostratigrafică a grupului Sebeș-Lotru din pînza getică de la nord de

Muntele Mic este caracterizată de prezența orizontului micașisturilor și al paragnaiselor micacee intens migmatizate, în interiorul sau în vecinătatea căruia apar majoritatea corpurilor de pegmatite.

Caracterele morfo-structurale ale corpurilor de pegmatite

Forma și relațiile cu roca înconjurătoare. Cel mai frecvent se întîlnesc l e n t i l e l e care reprezintă corpuri concordante și apar izolate sau formează asociații de tipuri distincte: a) lentile de dimensiuni asemănătoare dispuse în roi; b) lentile mai mari alături de care apar lentile mici paralele; c) șiruri de lentile situate la același nivel sau la nivele diferite. La tipul "*a*" contactul cu roca înconjurătoare este de obicei difuz ceea ce indică legătura cu fenomenele de migmatizare, la tipul "*b*" contactul este gradat, pegmatitele fiind de tip dilational (Goodspeed 1940), iar la tipul "*c*" contactul este net, lentilele formîndu-se prin budinarea unor filoane de pegmatite concordante sau prin fragmentarea unor dykeuri (Șeclăman 1972). F i l o a n e l e c o n c o r d a n t e reprezintă corpuri mari cu gîtuiri și îngroșări, limita față de roca înconjurătoare poate avea atît un caracter difuz cît și net. C u i b u r i l e de pegmatite se întîlnesc mai rar, au mărimi reduse și contururi neregulate, contactul fiind net sau difuz. Ele s-au format în timpul proceselor de migmatizare, pot fi pegmatite de concrețiune (Ramberg 1952) sau formate în urma unor procese de secreție laterală. D y k e-u r i l e (filoane discordante) se întîlnesc mai rar, dar formează cîteva corpuri importante (Teregova) și s-au format în cadrul unor procese de natură anatectică prin pătrundere în lungul unor zone de minimă rezistență, sau sînt de tip „replacement" (Ramberg 1949) și s-au format prin intermediul unor procese metasomatice. C o r p u r i l e m a r i c u f o r m e n e r e g u l a t e prezintă în toate privințele caractere heterogene și au luat naștere prin înglobarea, în timpul formării lor, a mai multor corpuri de pegmatite preexistente cu caractere deosebite sau, la formarea lor au participat mai multe procese genetice a căror influență a predominat succesiv sau alternant, generînd aspecte distincte.

C l a s i f i c a r e a t i p u r i l o r d e c o n t a c t (net și gradat) și caracterul formei corpurilor ne ajută la stabilirea unor indici genetici.

C o n c e n t r a ț i i l e m a f i c e p e r i f e r i c e unor corpuri sînt constituite din biotit, turmalină sau hornblendă și pot fi interpretate ca „restite" (diferențiere anatectică — Scheumann 1937, Mehnert 1951), se pot forma prin procese metasomatice (Reynolds 1946), sau datorită proceselor de secreție laterală, transportul substanței minerale producîndu-se prin difuzie (Ramberg 1952, Șeclăman 1972). Remarcăm deci, că procese petrologice distincte generează aspecte asemănătoare. În concluzie, rezultă că în regiune coexistă mai multe tipuri genetice de pegmatite, iar la stabilirea acestora trebuie să ținem cont de dificultățile create de convergența fenomenelor.

2. **Structura internă.** Z o n a r e a se rezumă la unele corpuri la prezența unei zone de contact aplitice în contrast cu o masă pegmatitică larg cristalizată și omogenă. Întîlnim însă și corpuri cu zonări complete,

incomplete, simetrice, asimetrice, simple, complexe, corpuri cu o zonare incipientă sau a căror zonare a fost distrusă. Lentilele mici nu sînt zonate sau prezintă o zonare incipientă, ceea ce denotă că aparțin unui stadiu evolutiv timpuriu. Predomină plagioclazul, microclinul apare corodîndu-l pe acesta. Zonări complete întîlnim în cazul filoanelor concordante mari. Cînd zona intermediară este constituită din microclin deducem prezența unei faze de metasomatoză potasică într-un sistem deschis, în forma ei anhidră. Cînd apare muscovitul larg cristalizat în locul microclinului, metasomatoza s-a produs în condițiile păstrării apei în sistem. Dyke-urile mari (Teregova) prezintă zonări complete și complexe, datorită varietății și semnificației conținutului lor mineralogic (beril, columbit, monazit).

Structura grafică este dată de concreșteri ale cuarțului cu plagioclaz, cu pertit și cu microclin și este caracteristică zonelor intermediare. Apar toate aspectele caracteristice formării cuarțului grafic prin înlocuirea selectivă a feldspatului, deci prin coroziune metasomatică. Compoziția normativă a granitelor grafice indică caracterul variabil al proporțiilor de cuarț, plagioclaz și feldspat potasic, fapt caracteristic formării metasomatice și nu unui eutectic granitic.

Mineralogia pegmatitelor. Majoritatea pegmatitelor din regiunea cercetată au o mineralogie simplă, fapt care constituie o trăsătură caracteristică pegmatitelor de origine metamorfică. Principala fază de minerale este constituită din biotit — plagioclaz acid — microclin, microclin pertit — este constituită din biotit — plagioclaz acid — microclin, microclin pertit — muscovit — cuarț. Minerale rare sau accesorii, care ar putea reprezenta produsul pneumatolizei, apar doar izolat. Asociate diverselor stadii ale mineralogiei majore au fost întîlnite următoarele minerale: granat, turmalină, beril, apatit, columbit, monazit, ortit, disten, sillimanit, hornblendă, magnetit, hematit, pirită, calcit și clorit.

Feldspații. Tipul de feldspat indică treapta evolutivă la care a ajuns un pegmatit. Reflectă și deformările suferite de pegmatite, deci modificările regimului tectonic. Feldspații plagioclazi (albit, albit-oligoclaz, oligoclaz) reprezintă componentul mineralogic principal al corpurilor mici, uneori este substituit de către microclin și reprezintă un prim stadiu (I) de formare al pegmatitelor. Feldspații alcalini, reprezentați prin microclin sau microclinpertit, corodează plagioclazul și reprezintă un al doilea (II) stadiu evolutiv. Structurile pertitice întîlnite sînt de origine metasomatică. *Micele* caracterizează prin prezența lor anumite etape de formare a pegmatitelor, iar prin transformările suferite (biotitul) reflectă modificări ale chimismului caracteristic diferitelor stadii evolutive. Biotitul cristalizează alături de plagioclaz și aparține stadiului (I) de formare. Ulterior este deferizat sau se păstrează ca mineral relict. Muscovitul se formează prin deferizarea biotitului sau înlocuirea plagioclazului și este întotdeauna însoțit de cuarț, reprezentînd stadiul de formare II A, caracterizat de o metasomatoză potasică, bogată în vapori de apă. Prin hidroliza feldspaților potasici este însă posibilă și formarea unui muscovit postmicroclinic. *Cuarțul* apare în toate tipurile de pegmatite, intervine la sfîrșitul fiecărei faze de cristalizare, contribuind la dezvoltarea structurilor metasomatice. Cuarțul zonei centrale (III) încheie evoluția proceselor de formare a pegmatitelor.

Geochimia pegmatitelor. Pe baza a 62 analize chimice complete de silicați și 42 analize spectrale s-au evidențiat unele caracteristici geochimice ale pegmatitelor. Evoluția elementelor majore prezintă următoarea succesiune: NaCa (I), apoi K (II) sau K + H_2O (II) și SiO_2 (III) cu precizarea că între fiecare stadiu se interpune cuarțul, iar înainte de cuarțul final apare uneori o albitizare (Na — microclinpertitul de substituție), dar care nu formează în acest caz un stadiu independent.

Diferitele diagrame construite indică evoluția cvasiconcomitentă și interdependența dintre paragnaise și pegmatite în timpul proceselor metamorfice care le-au generat. Conținutul în elemente minore ale pegmatitelor este redus și are un caracter monoton. Din diagrame rezultă sensul proceselor de diferențiere a fluidelor alcaline din masa paragnaiselor înspre mobilizatele pegmatitice.

Geneza pegmatitelor. Datele structurale, mineralogice și geochimice converg spre ideea originii metamorfice a pegmatitelor. Dealtfel Savu (1977), demonstrează absența oricărei legături petrogenetice a pegmatitelor din munții Semenic cu plutonul granitoid de Poniasca. Prin geneza metamorfică a pegmatitelor înțelegem desfășurarea în paralel sau succesiv a unor procese petrologice distincte în cadrul aceleiași stive de roci supuse metamorfismului. Dovezi se găsesc începînd cu studiul aspectelor morfostructurale, care atrag totodată atenția și asupra coincidențelor posibile ca urmare a convergenței fenomenelor. Astfel, pegmatitele s-au format în urma proceselor de migmatizare și pot fi de tip „dilational", „concrețional", și au luat naștere în urma unor procese metasomatice, pegmatitizarea producîndu-se nedilational, sau reprezintă produsul unor procese de secreție laterală. Majoritatea pegmatitelor acestei regiuni au fost însă generate de procese „anatectic metasomatice" (Barth, 1962). Prezența proceselor anatectice este demonstrată de diferitele tipuri de structuri migmatice întîlnite. Unele dintre mobilizatele leucocrate urmează o evoluție pegmatitică și care, în funcție de desfășurarea ei în timp, generează pegmatite simple sau complexe, după cum au fost parcurse stadiile evolutive. Plagioclazul și biotitul reprezintă stadiul I de formare, după care intervine cuarțul, ce poate genera în anumite condiții structuri grafice. Caracterul mineralogic al stadiului II este în funcție de presiunea vaporilor de apă: cînd sistemul se deschide și presiunea H_2O scade, cristalizează microclinul (II), cînd rămîne închis și presiunea H_2O crește, cristalizează muscovitul (II. A). Deci, desfășurarea întregului proces este controlată de regimul tectonic dominant în perioada respectivă. Cuarțul intervine din nou, formînd concreșteri grafice cu microclinul sau însoțește muscovitul. Evoluția se încheie prin cuarțul (III) din nucleu. În concluzie se poate afirma, că diferitelor stadii structural mineralogice ale formării pegmatitelor le corespund diverse tipuri petrografice de pegmatite. Relațiile dintre mineralele pegmatitelor arată ordinea succesiunii stadiilor de evoluție, care în general, se înlocuiesc unele pe altele, rezultînd pegmatite zonate. Fiecare zonă a unui corp pegmatitic reprezintă un tip petrogenetic, care, atunci cînd apare singur, constituie un pegmatit independent.

Dyke-urile de dimensiuni mari, cu o evoluţie completă şi o mineralogie variată, prezintă datorită convergenţei fenomenelor aspecte asemănătoare cu pegmatitele de origine magmatică — granitică. Ele s-au format însă în urma migrării ascensionale a fluidelor pegmatitice în lungul unor fracturi şi presupun existenţa unor sîmburi anatectici palingenetici.

Semnificaţia petrogenetică deosebită şi importanţa economică a pegmatitelor impun cercetarea lor şi în viitor, progrese fiind posibile atît prin studiul relaţiilor dintre minerale cît şi prin acela al caracteristicilor geochimice şi cristalochimice ale acestora.

EXPLANATION OF PLATES

Plate III

Fig. 1. Some dykes belonging to the former group show a contact line with fine, millimetric inlets a few cm long, which favour the advancement of pegmatites inside the paragneisses.

Fig. 2. The pegmatite groundmass exhibits parts built up of paragneisses and the paragneisses include pegmatite zones, the transition zone being represented by the area between the pure pegmatite and the paragneiss unaffected by pegmatitization.

Plate IV

Fig. 1. Pegmatite lenses resulted from the boudinage of some concordant veins.

Fig. 2. Garnet (almandine) represented by large idiomorphic grains (ca 1 cm in diameter) within plagioclase (plg), quartz (qz) and scarce muscovite (mu) groundmass.

Plate V

Fig. 1a, 1b, 1c. Graphic quartz. Some faces exhibit a lamellar character. The cut perpendicular to this face points to an entirely different image of proper graphic quartz.

Plate VI

Fig. 1. Contact zone with equigranular aplite texture and leucotonalite composition, consisting of albite (ab) and quartz (qz).

Fig. 2a, 2b. Graphic intergrowth of quartz (qz) and plagioclase (plg).

Fig. 3. Graphic intergrowth of quartz (qz) and microcline (mi).

Plate VII

Fig. 1. Plagioclase (plg) with fine polysynthetic twins is substituted by microcline (mi) wherein it forms relict inclusions.

Fig. 2a, 2b. Irregular shapes of low temperature perthites resulted from potash feldspar substitution by plagioclase are prevailing.

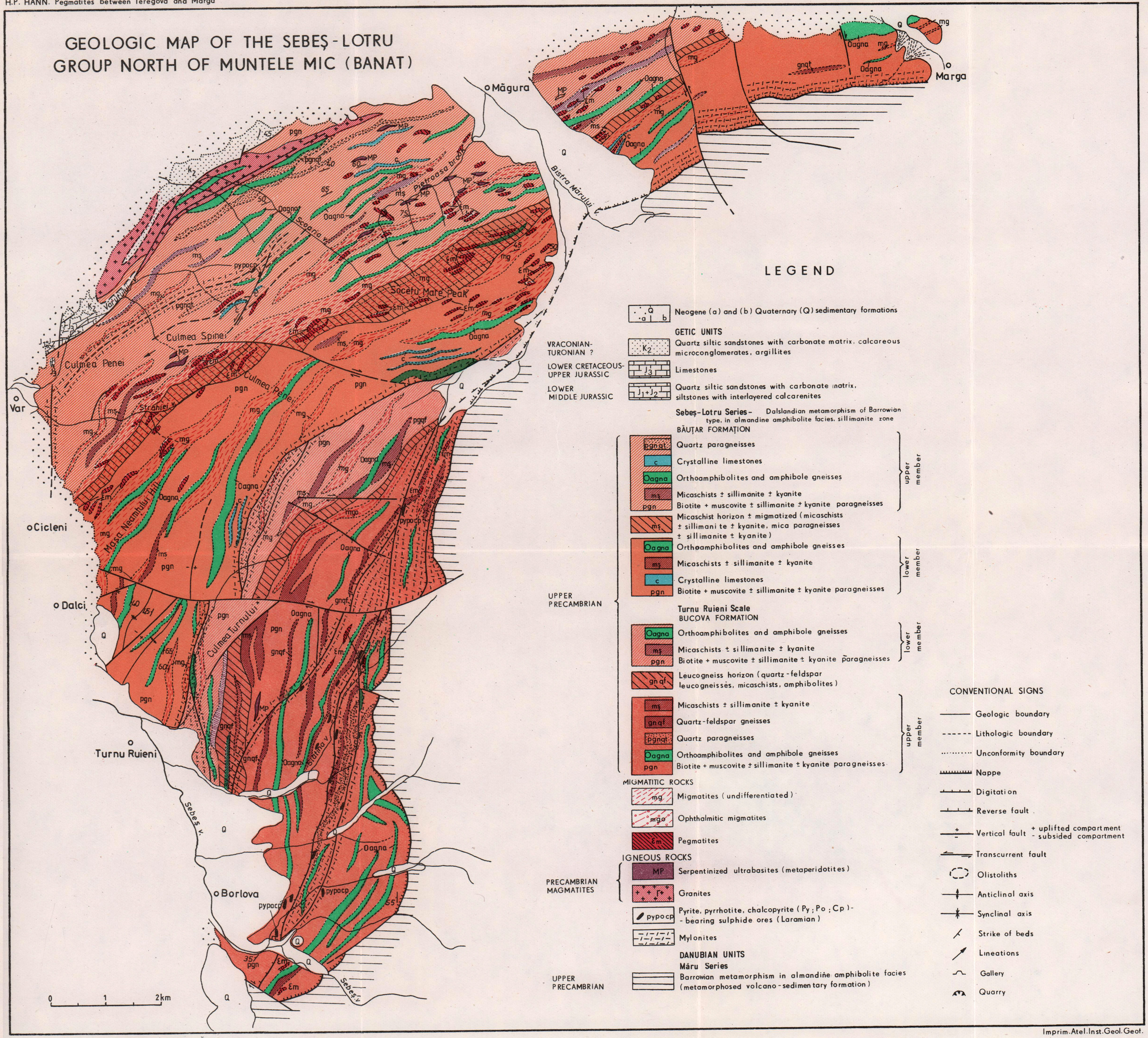
Fig. 3. Irregular shapes of perthites (plg) cut by a quartz (qz) vein.

Plate VIII

A microclineperthite sample of irregular shape studied by microprobe analysis (x 1200).
Fig. 1. K-distribution.
Fig. 2. Na-distribution.
Fig. 3. Ca-distribution.
Fig. 4. Al-distribution.
Fig. 5. Si-distribution.
Fig. 6. The topographic surface (composition pattern).

Plate IX

Fig. 1. Albite (ab) forming in places the cover of altered plagioclase (plg).
Fig. 2a, 2b. Large tourmaline (tu) grains corroded by quartz (qz) resulting in "graphic tourmaline".
Fig. 3. Tourmaline (tu) is substituted by quartz (qz) wherein it forms relict inclusions.



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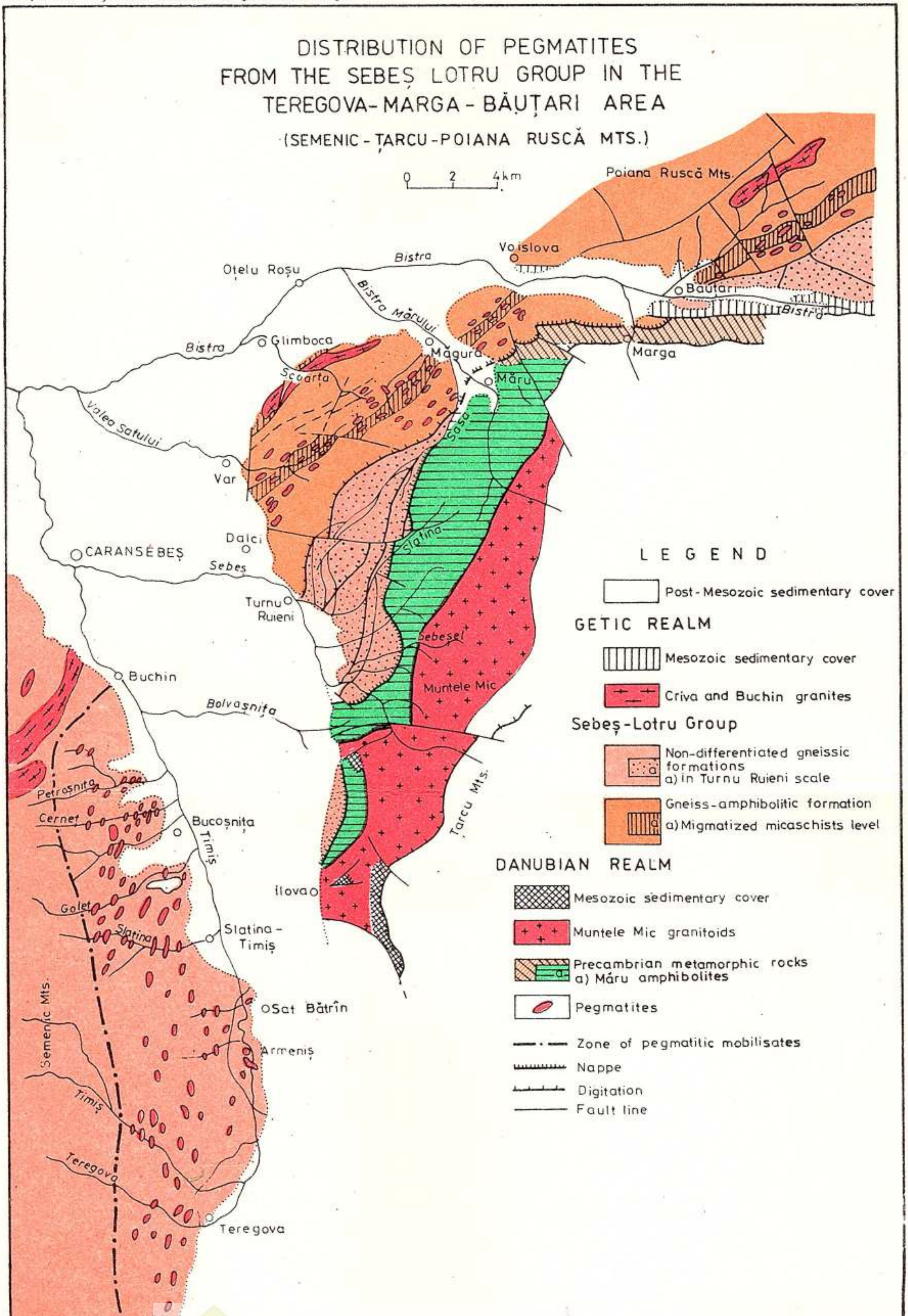

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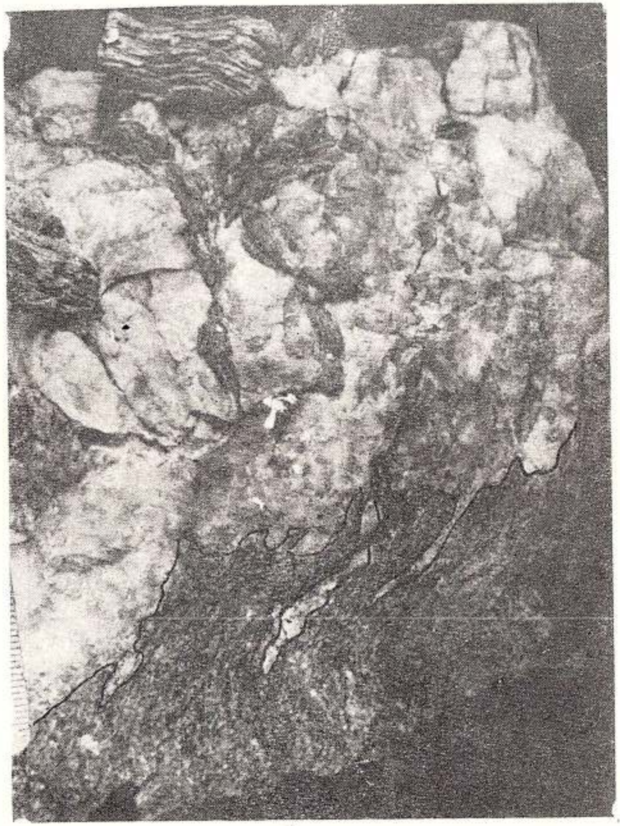

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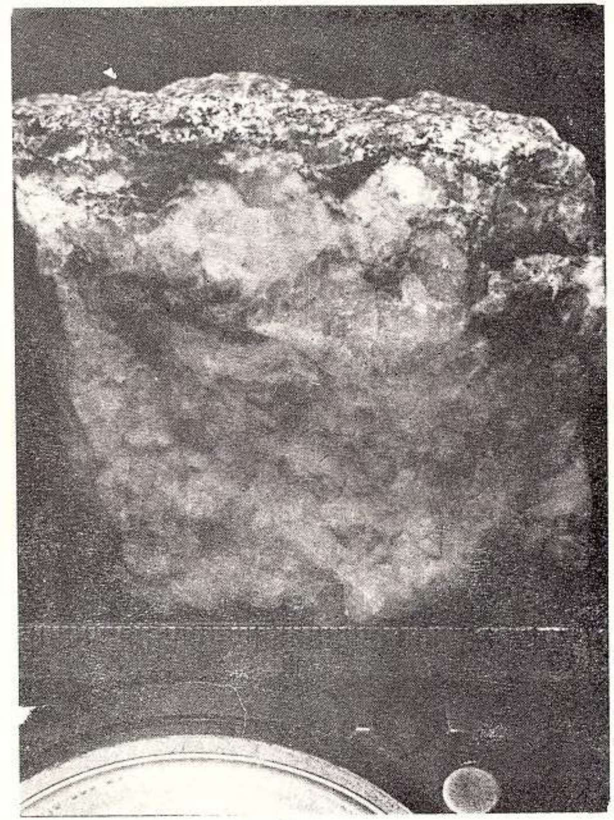

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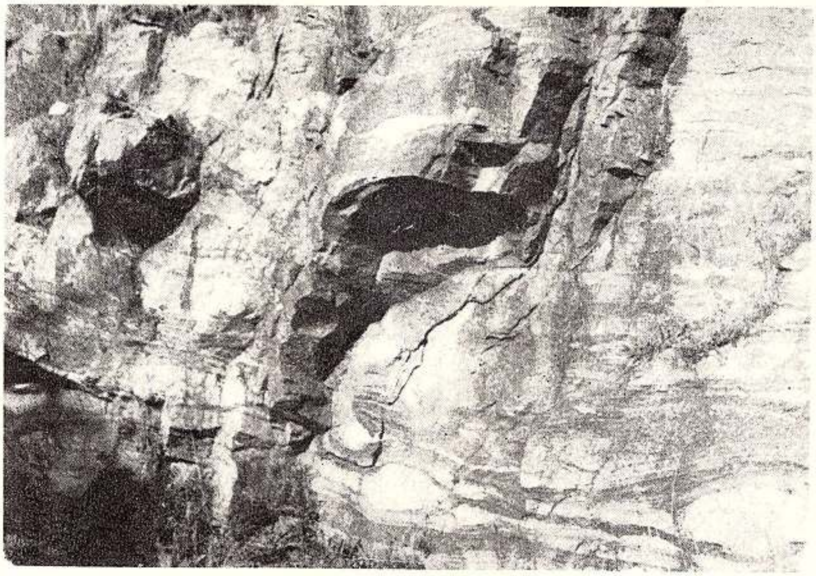

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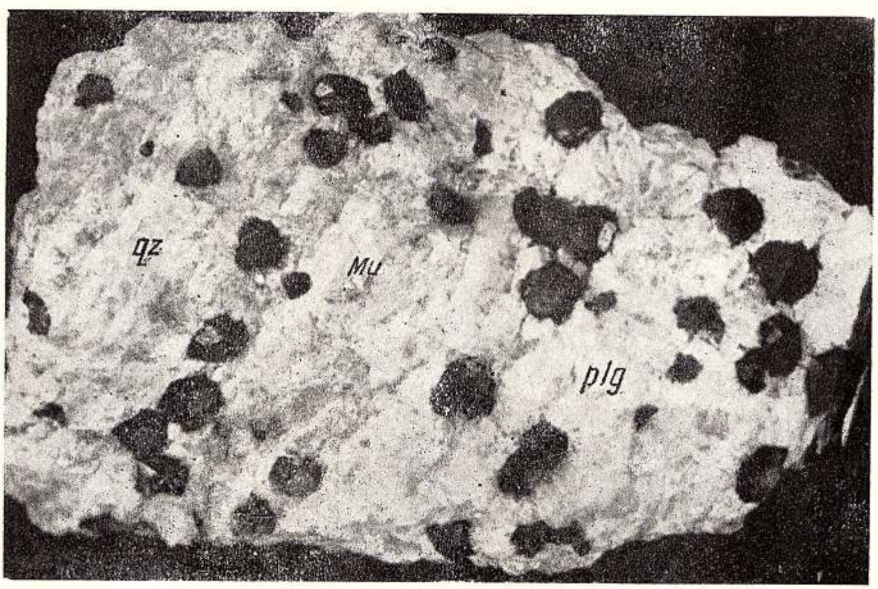

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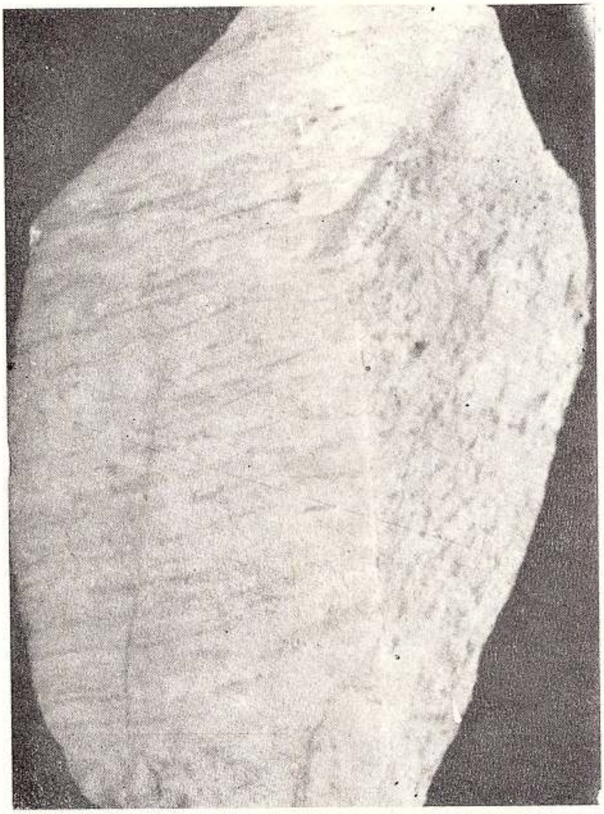

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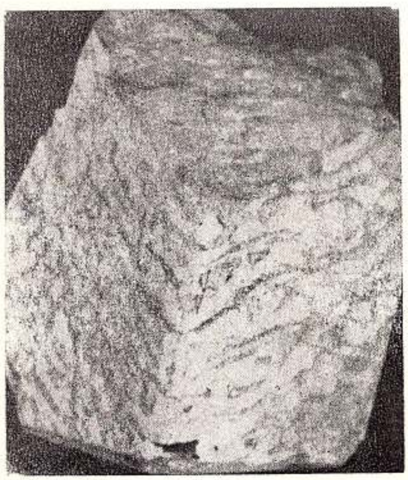

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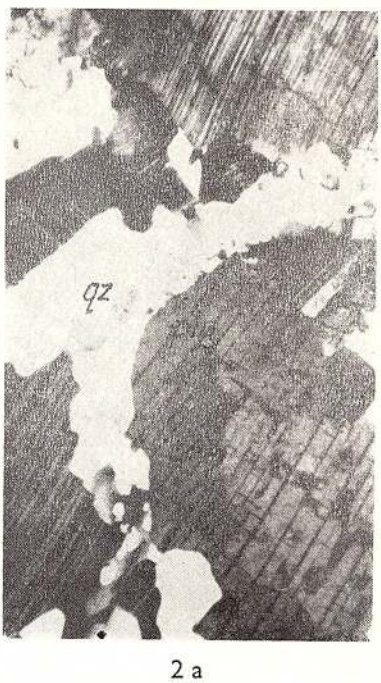

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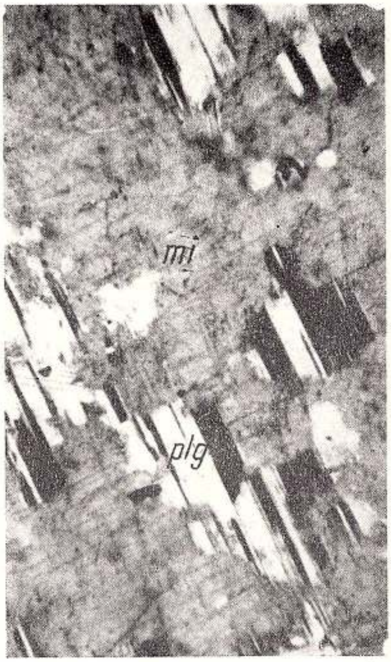

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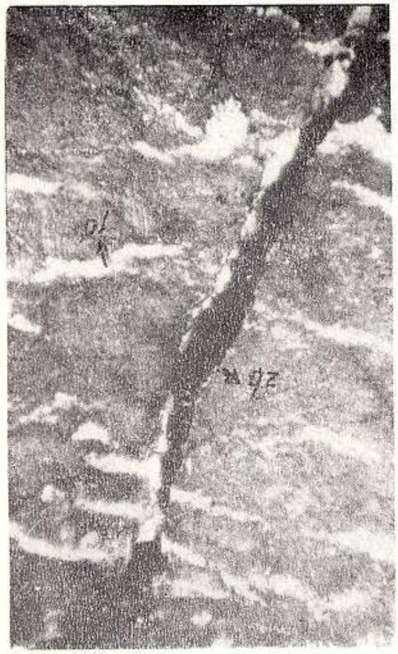

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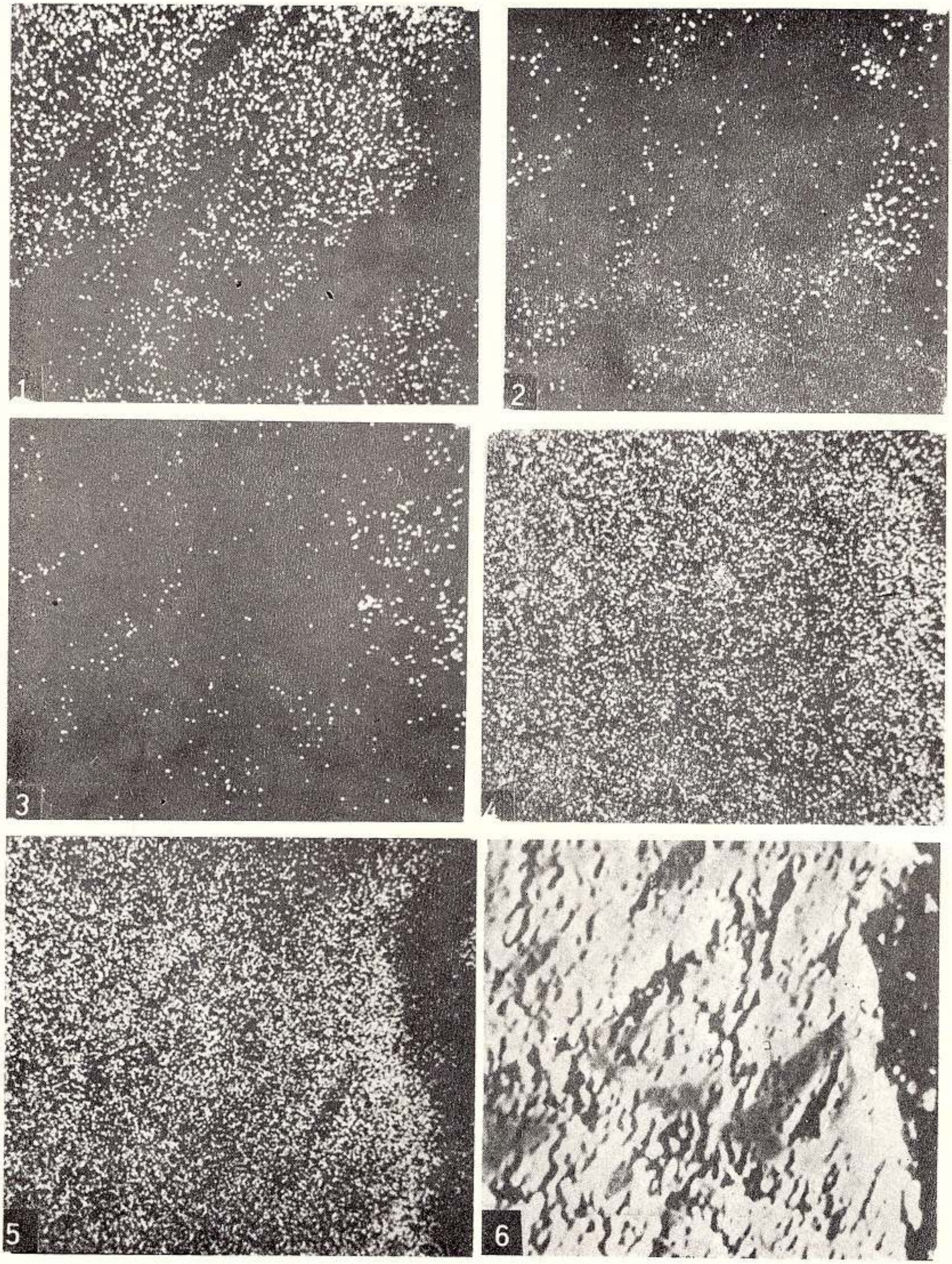
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GEOTHERMICS OF THE CARPATHIAN AREA[1]

BY

ŞERBAN VELICIU[2]

Abstract

The study presents the geothermal regime of the Romanian Carpathians area on the basis of the measured values of heat flow, heat conductivity and radiogenic heat generation of the rocks. Heat flow values in 54 deep wells in various geological units were obtained. Correlations among heat flow within the Carpathian folded chain (40–126 mWm^{-2}, including post-tectonic depressions), the Moldavian Platform (39–55 mWm^{-2}) as well as the Moesian Platform (39–78 mWm^{-2}) and structural features of these major units have been established. The distribution of temperature *vs.* depth and its interrelation with geological and geophysical observations was examined with a particular regard to the East Carpathians, Pannonian Basin and Transylvanian Basin. The territory under discussion represents a relatively young area, of less than 25 m. y. since the last thermal-tectonic event thus some dynamic geothermal models have been adopted. In order to explain the high heat flow (73–126 mWm^{-2}) observed at the inner part of the East Carpathians, a model which takes into account the radiogenic heat generation and the descent of a lithospheric plate has been elaborated. Calculated geotherms indicate the presence of a high temperature anomaly located in the upper mantle. The geothermal data suggest that both the high heat flow and the building up of the Neogene volcanic chain are consequences of the lithospheric Alpine subduction process in the Carpathian area. The Transylvanian Basin exhibits abnormally low heat flow which makes it distinct as compared with other typical ensialic intra-arc basins. Lithospheric stretching as described for extensional basins (i.e. Pannonian Basin) is not very likely to be present here.

Résumé

Le régime géothermique de l'aire carpathique. L'étude présente le régime géothermique des Carpathes roumaines en considérant les valeurs mesurées du flux géothermique, la conductibilité thermique et la génération thermique radiogénique des roches. On a obtenu des valeurs du flux géothermique pour 54 forages profonds dans différentes unités géologiques. On a établi aussi des corrélations entre le flux géothermique à l'intérieur de la chaîne plissée des Carpathes (40–120 mWm^{-2}, dépressions post-tectoniques y incluses) et de la Plate-forme Moesienne (39–78 mWm^{-2}) et les traits structuraux de ces unités majeures. La distribution de la température par rapport à la profondeur et les corrélations avec les données géologiques et géophy-

[1] Paper presented on 5. IV. 1985, accepted on 16. IV. 1985

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siques sont examinées surtout en tenant compte des Carpathes Orientales, le Bassin Pannonien et le Bassin Transylvain. L'aire investiguée est relativement jeune, à moins de 25 m.a. dès le dernier événement thérmo-tectonique et par conséquent on a adopté des modèles géothermiques en régime dynamique. Le flux géothermique élevé (73–126 mWm^{-2}) à l'intérieur des Carpathes Orientales est expliqué par un modèle basé sur la génération thermique radiogénique et la plaque lithosphérique baissée. Les géothermes calculées indiquent une anomalie de température élevée dans le manteau supérieur. Les données géothermiques suggèrent que tant le flux géothermique autant que l'édification de la chaîne volcanique néogène sont le résultat du processus de subduction lithosphérique alpine dans la région des Carpathes. Le Bassin Transylvain se caractérise par un flux géothermique particulièrement bas qui le différencie d'autres bassins interarc ensialique typiques. L'extension lithosphérique décrite pour les bassins post-tectoniques (ex. le Bassin Pannonien) n'est pas en concordance avec les données ci-présentées.

Introduction

The geological structure of the earth's crust is now viewed as a collection of effects of the plate tectonics (Mc Kenzie, Parker 1968 ; Morgan 1968 ; Le Pichon 1968 ; Dewey, Bird 1970). On the other hand, it is possible to construct various geothermal models that provide some insights into the tectonic history due to the long relaxation time of the heat conduction in rocks. Von Herzen (1967) demonstrated that the time interval required by the surface heat flow is below 150 m.y., in order to be in equilibrium with the inner heat sources distributed within a lithosphere 100 km thick. If it is taken into account that the most important contribution of the radiogenic heat sources is concentrated within the first 30–40 km of the earth's crust, less than 50 m.y. is necessary to reach thermal equilibrium.

Such geothermal models should be reconciled with the knowledge of the structure of the crust, the petrophysical properties of the rocks and the thermal regime of the earth as inferred from the observations at its surface. The geothermal models for oceanic basins are mainly conditioned by the time elapsed since the formation of the oceanic crust in the spreading zones. Simple conductive geothermal models have been derived from the paleomagnetic data, cooling and contraction of the initial magma as well as the distance from the spreading centres. A review of the geothermal models has been performed by Palmason (1973) for the Mid-Atlantic Ridge and by Lubimova, Nikitina (1978) for the Western Pacific.

Near old trenches and extensional basins the situation is quite different as compared to the ridges and oceanic basins ; a broader region is subjected to thermal disturbances by a variety of processes, the details of which are not as well understood yet. Models that take into consideration both the heat sources associated with descent of the lithosphere or with extension of the ensialic post-tectonic basins, provided reasonable explanations for the gross features of the observed heat flow density at the earth's surface (Oxburg, Turcotte 1971 ; Lachenbruch, Sass 1978 ; Mc Kenzie 1978).

It is the aim of this study to present the geothermal regime of the Romanian Carpathians area based on the measured values of heat flow density, heat conductivity and radiogenic heat generation of the rocks. So, the distribution of geotherms (temperature vs. depth) and its interrelation with geological and geophysical observations is examined with a particular regard to the East Carpathians, Pannonian Basin and Transylvanian Basin.

The territory under discussion represents a relatively young area, of less than 25 m.y. since the last thermal-tectonic event, thus some dynamic geothermal models should be adopted.

A general outlook on the geology shows that the Carpathian area, belonging to the northern branch of the peri-Mediterranean Alpine folded chains has been divided into structural units (Dumitrescu, Săndulescu 1968 ; Săndulescu 1975) which generally group together nappes of similar type and of synchronous age of tectogenesis. From the last point of view the Carpathians show two main periods of compressions : Cretaceous and Miocene. The Cretaceous Carpathians group the Dacides, the Transylvanides and the Danubian (Fig. 1). The Miocene Carpathians (Moldavides) cover the outer zone of the chain. Two big Neogene molasse basins (Transylvanian and Pannonian) are superposed on the folded units and a Sarmato-Pliocene molasse foredeep borders the folded chain outwards.

Pre-Alpine crystalline formations crop out inside the Dacidian areas. The Transylvanides are the main suture of the Carpathians con-

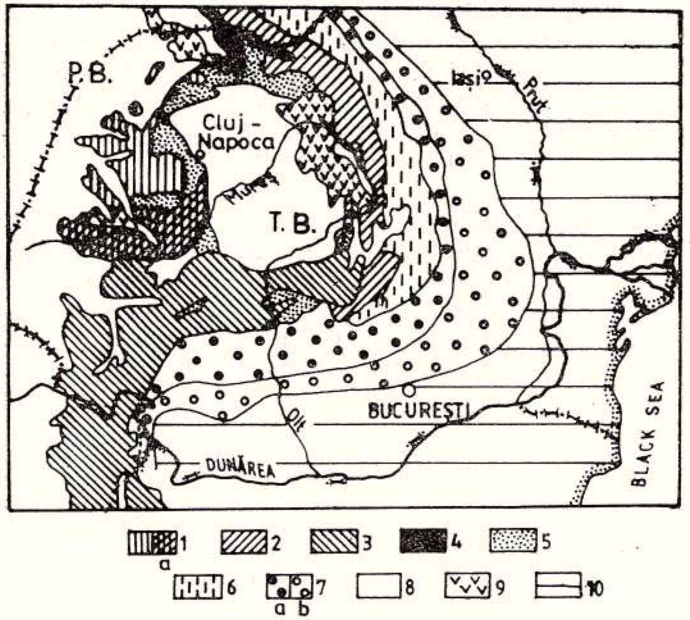


Fig. 1. Schematic tectonic map of the Carpathian area (after Săndulescu, 1975). 1-Western Dacides, a-zone of the Metaliferi Mts. (Southern Apuseni) ; 2-Eastern Dacides ; 3-Southern Dacides ; 4-Pienides ; 5-post-tectonic cover ; 6-Moldavides ; 7-Foredeep, a-inner, b-outer ; 8-Neogene depressions : T. B. -Transylvanian basin ; P. B. -Pannonian basin ; 9-Neogene volcanics ; 10-Foreland.

taining units with ophiolitic complexes. The outer Dacides are the second suture, showing flysch deposits and ophiolitic rocks. Large development of flysch is known inside the Moldavides.

Except the ophiolitic assemblages, the Alpine magmatic activity exhibits three igneous periods : an ensialic predominantly alkaline moment (Jurassic) known in the South Carpathians and the East Carpathians, and two subduction calc-alkaline moments, the first in the Upper Cretaceous and Paleocene time, the second during the Neogene.

The Carpathians foreland groups platform areas of different ages. The oldest one is the Moldavian Platform (belonging to the Epi-Algomian East-European craton) situated in front of the East Carpathians. An Epi-Hercynian platform (Moesian Platform) runs south and westward to the former. The outer part of the Danubian realm has been considered a direct prolongation of the Moesian Platform.

Radiogenic heat generation in the Carpathians

It is generally accepted that the distribution of radioactive heat sources is the major motive for the temperature field in the continental crust. All natural radioactive isotopes generate heat to a certain extent but only the contribution of the decay series U^{238}, U^{235}, Th^{232} and of the isotope K^{40} are significant for the heat flow from the earths' interior.

Two independent methods were developed for the determination of radiogenic heat generation constants (Birch, 1954). The first summed the kinetic energy of the particles emitted and of the recoil nucleus and the electromagnetic radiation involved in the radioactive decay. The second method used the mass difference between the parent and daughter nuclides.

The heat generation constants based upon the work of Birch (1954) which gave the heat generated per unit time and gram U, Th, K, were extensively used in the literature. However, decay schemes, half lives and mass differences have been subsequently revised. This fact called for the redetermination of the heat generation constants (Hamza, Beck 1972 ; Rybach 1976). Table 1 presents a compilation of the heat genera-

TABLE 1

Heat generation constants after Birch, Hamza, Beck and Rybach

Element	Heat generation, in $\mu W\ kg^{-1}$			
	Birch (1954)	Hamza, Beck (1972)		Rybach (1976)
		with neutrino energy	without neutrino energy	
U	97 (0.73)	98 (0.74)	106 (0.80)	92 (0.718)
Th	27 (0.20)	27 (0.20)	28 (0.21)	26 (0.193)
K	0.0036 (27×10^{-6})	0.0034 (26×10^{-6})	0.0068 (51×10^{-6})	0.0035 (26.2×10^{-6})

The more familiar figures in cal $yr^{-1}g^{-1}$ are given in parantheses tion constants determined for the main natural radioactive elements. The differences found between Birch's values and those of other authors are minor.

In the present study on the Carpathian area the radiogenic heat generation (A) for a certain rock has been calculated using the constants revised by Rybach (1976) :

$$A/\mu\mathrm{Wm}^{-3}/ = 0.133\rho(0.718\ c_U + 0.193\ c_{Th} + 0.262\ c_K)$$

The computation involved knowledge on the U, Th and K contents taking also the density ρ (g cm^{-3}) into account. Practical concentrations units are weight ppm for U and Th, and weight per cent for K.

A gamma-ray spectrometric technique using a NaI (T1) crystal was utilised because it enabled simultaneous determinations of the significant heat-producing radio-elements. The basic principles and applications of gamma-ray spectrometric technique have been thoroughly discussed by Lemne (1968, 1970).

The selection of the rock samples for the heat generation determinations has been performed based on two criteria : (1) the samples should be representative for the uppermost part of the crust in the Carpathian area and (2) they should offer the possibility to investigate a variety of petrographic types ranging from the acid rocks to the basic ones. From the first point of view, deserved consideration the crystalline and magmatic formations of Archean, Proterozoic and Paleozoic age belonging to the Danubian Autochthon (South Carpathians) and the Gilău Autochthon (Apuseni Mts).

Some magmatic rocks (Paleozoic granites, Paleocene granodiorites, Neogene andesites and Jurassic ophiolites) from the Apuseni Mts have also been analysed (Table 2).

The number of analysed samples corresponds roughly to the relative surface abundance (in %) of the respective petrographic type in the Romanian Carpathians.

Table 3 lists the average heat generation figures for the Romanian Carpathians grouped according to the petrographic facies (magmatic and metamorphic, respectively). For comparison the values determined for characteristic surface-rock types from the Carpathians are presented together with values from Swiss Alps (Rybach 1976). In terms of the surface radiogenic heat generation of the rocks, the differences found between these two Alpine orogenic regions are minor.

Data from Table 3 clearly imply a marked upward concentration of radiogenic heat-producing isotopes in the crust. So, from a theoretical point of view, the crust in the Carpathians area constituted only of granodiorites ($A = 1.87\ \mu\mathrm{Wm}^{-3}$) could generate enough heat in order to sustain the average surface heat flow (69 ± 26 mWm^{-2}) measured in the Carpathians (including the post-tectonic Neogene depressions), without any caloric contribution from the mantle. However, on petrological and geophysical arguments it has been assumed (Smithson, Deker 1974) that the lower crust has to exhibit a granulitic facies which is less endowed with radiogenic heat-producing elements. This assumption seems to be valid also for the Carpathian area.

TABLE 2

Radiogenic heat generation of characteristic rocks from the Romanian Carpathians

Tectonic unit	Rock type	Age (m.y.)	Number of samples	Heat generation (μ Wm^{-3})	
				Width of variation	Mean value
Southern Carpathians (Southern Dacides):					
Danubian Autochthon	Epizone series (sericitic-chloritic schists)	Late Proterozoic (850)	24	0.42–1.06	0.74
	Granitoids	Early Paleozoic (550)	300	1.83–4.13	2.98
Getic Nappe	Mesozone series (microschists; paragneisses)	Late Archean (1600)	391	1.81–2.22	2.02
	Granites	Late Proterozoic (800)	50	1.94–3.10	2.52
Apuseni Mts (Western Dacides)	Granites	Early Paleozoic (530)	49	1.86–2.48	2.17
	Neogene andesites	Miocene (15–20)	61	0.52–1.18	0.85
	Banatites (granodiorites)	Paleocene (60)	41	1.71–1.99	1.85
	Ophiolites	Lower Jurassic (200)	53	0.14–0.56	0.35
Gilău Autochthon	Epizone series (Arada Series)	Early Paleozoic (500)	22	0.70–1.43	1.09
	Mesozone series (Someş Series)	Archean (900–1600)	56	1.74–3.11	2.43
	Cretaceous Flysch	Maastrichtian (70)	91	0.86–1.31	1.09

The data from the Carpathians indicate a variation of radiogenic heat generation over two order of magnitude reflecting the geochemical conditions during the formation of the respective rocks (magmatic differentiation, metamorphism or sedimentation). The mechanism by which

TABLE 3

A comparison of heat generation values for the Carpathians and Swiss Alps

Rock type	Romanian Carpathians (this work)			Swiss Alps (Rybach 1976)		
	Number of samples	Width of variation (μWm^{-3})	Mean value	Number of samples	Width of variation (Wm^{-3})	Mean value
Magmatic rocks: Granite	50	1.94–3.10	2.52	8	1.88–6.06	2.5
Granodiorite	41	1.71–1.99	1.87			1.5
Andesite	61	0.52–1.18	0.85			1.1
Basalt	53	0.14–0.57	0.35	8	0.08–1.05	0.3
Metamorphic rocks: „Green schists" facies (epizone)	22	0.70–1.49	1.09	18	0.25–2.42	1.5
Amphibolites facies (mesozone)	391	1.74–3.11	2.43	55	0.86–5.02	2.42
Sedimentary rocks Cretaceous flysch (sandstones)	91	0.86–1.31	1.09			
Carbonate rocks				12	0.03–0.92	0.33
Continental crust			0.72			0.80

the segregation of the heat-producing isotopes occurred is not well understood yet (Rybach 1976) but clearly these are closely associated with the above mentioned geological processes.

Roy et al. (1978) first recognized that the regional variation in surface heat flow within a "heat flow province" is linearly related to the heat generation of the surface rocks. The relationship is expressed as: $q_0 = q_r + b\,A_0$ (where q_0 is the surface heat flow, A_0 is the heat generation of the surface rocks, q_r is the "reduced heat flow" and b characterizes the vertical distribution of the heat sources within the crust). This observation has been confirmed in 17 different "heat flow provinces" and the best fitting line through the data points after Vitorello, Pollack (1980) is shown in Fig. 2. On the same graph the data from the Carpathian area have been plotted (Fig. 3). The graph suggests that the proportion $q_r/q_0 = 0.4/0.6$ empirically established by Pollack, Chapman (1977) seems to be valid also for data from the Carpathians.

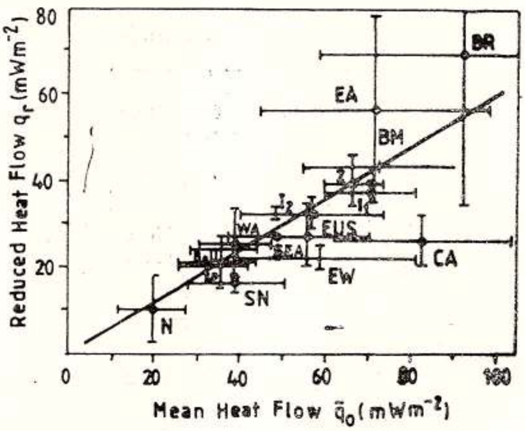


Fig. 2. Reduced heat flow vs. mean surface heat flow for different 17 "heat flow provinces" of the Globe, after Vitorello, Pollack (1980).

In the Carpathian area, affected by young tectogenetic phases (Cretaceous and Miocene) and returning to thermal equilibrium by passive cooling, it should be a decrease in the crustal radiogenic contribution to

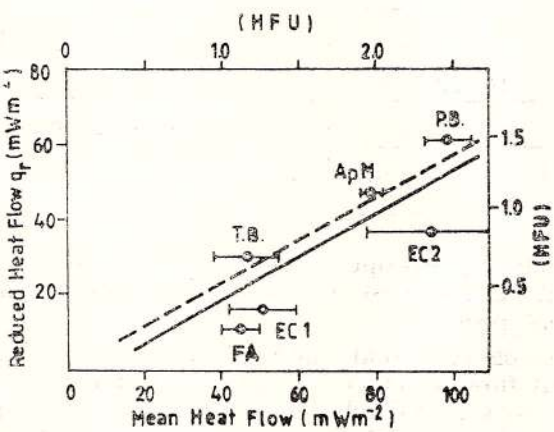


Fig. 3. Reduced heat flow vs. mean surface heat flow for six tectonic units from the Carpathians (East Carpathians, EC_1-Crystalline-Mesozoic zone and flysch zone, EC_2-Neogene volcanic zone; F-Foredeep; Ap. M.-Apuseni Mts.; T. B.-Transylvanian Basin; P. B.-Pannonian Basin). The bars represent the standard deviation of the mean heat flow. Correlation factor of the regression line is 0.87. The dotted line represents the best fitting line through the data points from Fig. 2.

maintain the ratio of 0.4/0.6. This is very likely accomplished by the removal of radiogenic heat-producing elements through erosional processes.

Heat flow density distribution

The first references in the scientific literature to measurements of temperatures in boreholes on the Romanian territory (Bungeţeanu 1910, Tănăsescu 1912) appear at the beginning of the 20-th century (fide Airinei 1981). However, it was not until the 4-th decade that the research in geothermics was intensified directly connected with the programmes of oil industry boreholes logging. Accumulated in a huge amount (over 4000 investigated boreholes) the temperature data have been collected, analysed and interpreted in order to disclose the thermal regime of the oil and gas fields (Cristian et al. 1969, 1971, Negoiţă 1970, Neguţ 1972, Paraschiv, Cristian 1973, 1976). Some attempts were made to use these data for heat flow calculation. However, the reported values (Negoiţă 1970, Paraschiv, Cristian 1973, 1976) were mere estimations based on thermal conductivities quoted from literature, for the rock types from different regions with distinct geological characteristics.

In the last years several heat flow researchers have carried out measurements, appropriate to the international standards for this kind of studies, on tectonic units of the Carpathians and their foreland. Veliciu et al. (1977) reported 21 values of heat flow covering the whole territory of Romania ; 19 values for the Transylvanian Basin and the Neogene volcanic chain of the East Carpathians were determined by Demetrescu (1979, 1981) ; 11 values for the eastern part of the Moesian Platform were determined by Neguţ (1982) ; Veliciu, Visarion (1984) published 4 values of heat flow from Oaş-Gutîi and Călimani Mts. Consequently the set of heat flow reliable data for the Romanian territory contains now 54 values (see Appendix I).

For this study the mentioned set of data has been supplemented by the heat flow values determined in U.S.S.R. (Kutas, Gordienko 1970 ; Kutas 1978) and Hungary (Horváth, et al. 1981). This enabled the compilation of the heat flow map from Plate I.

A non-uniform geographic distribution of the heat flow observation points is a first remark about the heat flow map of Romania. Moreover, despite a higher density of determinations as compared with neighbouring territories, this density is still relatively low as referred to the geologic-structural complexity of the Carpathian area. Most observations were concentrated in the central and eastern part of the Transylvanian Basin, on the Neogene volcanics, the Outer Flysch and Foredeep zones of the East Carpathians as well as in the central and eastern part of the Moesian Platform.

Heat flow values indicated in Plate I suggest some possible correlations between the thermal regime and the characteristics of the major structural elements of the Carpathians.

Such correlations have been previously attempted for different regions of the Globe, taking into consideration an empirical relationship regarding the heat flow and the age of the earth's crust. In a detailed and systematic analysis of the data from Europe, Polyak, Smirnov (1968) first demonstrated the decrease of heat flow versus tectonic age. Recently, similar analyses performed on more than 1500 observations from the continents (Chapman, Furlong 1977 ; Sclater at al. 1980) confirmed this decrease (Tιble 4).

TABLE 4

Variation of heat flow with age of the most recent tectonic-thermal event (compiled from Chapman, Furlong 1977; Sclater et al. 1980)

Age group	Mean heat flow (mWm^{-2})	Standard deviation	N	Age group (m.y.)	Mean heat flow (mWm^{-2})	Standard deviation	N
Cenozoic	71	37	587				
Mesozoic	73	29	85	0—250	76	53	398
Late Paleozoic	61	18	514				
Early Paleozoic	52	17	88	250—800	63	21	500
Late Proterozoic	54	20	265				
Early Proterozoic	51	21	78	800—1700	50	10	138
Archean	41	11	136	>1700	46	16	375
		Total	1753			Total	1411

„N" represents number of analysed observations.

For all geothermic studies the definition of the "tectonic age" constituted a key point. Within the continents this was usually considered as the age of the last tectonic-thermal event which mobilized the region in which the heat flow values were determined. Measurements performed in the undeformed platform sediments carry the age of stabilization of the platform basement; measurements from the magmatic areas were referred to the radiometric age of the magmatic extrusion or intrusion; a fold belt carries the youngest age of deformation and a metamorphic terrain carries the age of the latest thermal metamorphic episode. Obviously, the assignment of age is somewhat subjective, particularly in partially or weakly remobilized terranes and this subjectivity contributes to some of the scatter about the age group-heat flow means. Moreover, there are certain geologic and tectonic settings where the regional heat flow may depart significantly from the appropriate age-group mean from Table 4.

One of the uses for the heat flow versus age relationship is the estimation of heat flow in unsurveyed areas on the basis of the tectonic-thermal age of the terrain, supplementing in this way the existing heat flow observations.

The decay of continental heat flow with age extends over a longer period of time and it is a complex phenomenon comprising the following three components (Vitorello, Pollack 1980) as is shown in Fig. 4 : component I, crustal radiogenic heat with a time-dependent decrease introduced through erosion; component II, heat from a transient thermal perturbation associated with tectogenesis and component III, background heat flow from deeper sources within the earth.

On a similar graph (Fig. 5) data for the Romanian territory have been plotted (Table 5), using the tectonic age from Rădulescu, Dimitrescu

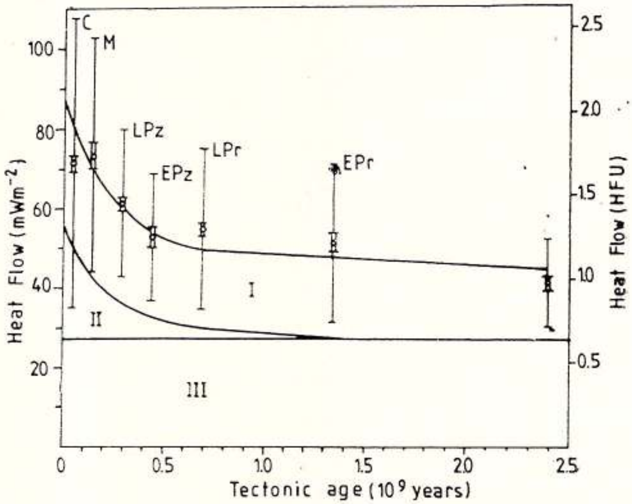


Fig. 4. Decrease of continental heat flow with age and its three principal components after Vitorello, Pollack (1980): component I — radiogenic heat from the crust; component II — heat from a transient thermal perturbation associated with tectogeneses; componet III — background heat flow from deeper sources. EPr-Early Proterozoic; LPr-Late Proterozoic; EPz-Early Paleozoic; LPz-Late Paleozoic; M-Mesozoic; C-Cenozoic.

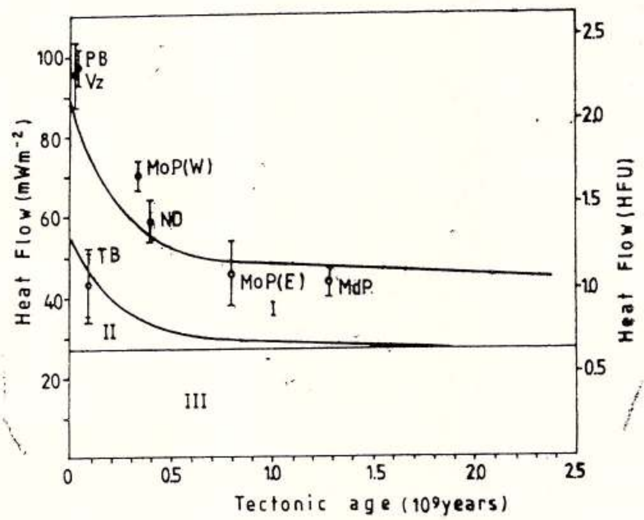


Fig. 5. Decrease of heat flow with age in the Carpathian area. N. V. -neo-volcanic zone of the East Carpathians; T. B. -Transylvanian Basin; P. B. -Pannonian Basin; N. D. -North Dobrogea; Mo. P. (w)-Moesian Platform (western part); Mo. P. (e)-Moesian Platform (eastern part); Md. P. -Moldavian Platform. Heat flow data are from Veliciu, Demetrescu (1979). The three principal components of heat flow from Fig. 5. Age data from Rădulescu, Dimitrescu (1982).

TABLE 5

Variation of heat flow with age of the most recent tectonic-thermal event in the Carpathian area. Age data from Rădulescu, Dimitrescu (1982)

Tectonic unit	Last tectonic-thermal event and its age (m.y.)	Number of observations	Mean heat flow (mWm^{-2})	Standard deviation	References
Neogene volcanic zone of East Carpathians	Mio-Pliocene- Quaternary calcalkaline volcanics (3–12)	18	96	16	1,2, 3,4
Pannonian Basin (eastern limit)	post-Miocene extension (10–25)	4	96	8	1,5
Transylvanian Basin	pre-Senonian folded basement (80–100)	14	44	15	1,6
North Dobrogea	Permian and Triassic volcanics (250–390)	5	60	8	2,7
Moesian Platform: western part	magmatic intrusions during Breton tectogenetic phase (400)	4	70	7	1
eastern part	metamorphosed and folded epi-Hercynian basement (700)	6	47	9	1,8
Moldavian Platform	metamorphosed and folded Proterozoic basement (1000–1600)	7	44	6	1,7

References: (1) Veliciu et al. (1977); (2) Veliciu, Visarion (1984); (3) Demetrescu (1979); (4) Kutas, Gordienko (1970); (5) Horváth et al. (1981); (6) Demetrescu et al. (1981); (7) Kutas (1978); (8) Neguţ (1982).

(1982). The observational data from the Carpathian area fit the mean values calculated by Chapman, Furlong (1977) and Sclater et al. (1980), except for the Transylvanian Basin which is characterized by surprising low heat flow (mean 44 mWm^{-2}).

In areas affected by Tertiary tectogeneses, the three components of the regional heat flow contribute with 36 mWm^{-2}, 27 mWm^{-2} and, respectively, 27 mWm^{-2} to establish the mean value of 90 mWm^{-2} usually encountered in terrains younger than 50 m.y.

The component II, due to the cooling of the crust, diminishes effectively to zero after 300–400 m.y. since the thermal transient perturbation associated with tectogenesis. Consequently, in the Hercynian terrains, only the radiogenic component I and "background" of heat flow are present. Moreover, the radiogenic component I is reduced to approximately 20 mWm^{-2} by the erosion of the radiogenic isotopes from the uppermost crust. The two components I + II give only 45 mWm^{-2} in the Precambrian terrains.

The Moldavian Platform has low heat flow values (39–55 mWm^{-2}) which is typical for a platform with a Proterozoic basement.The lowest heat flow values are located on the contact with the Alpine orogenic zone of the Carpathians. This fact is explained by the "blanketing" thermal effect due to the thickening of the sediments on the epi-platformic side of the Foredeep. So, the mean of 44 mWm^{-2} calculated for the Moldavian Platform (Table 5) is slightly under the heat flow mean established by Sclater et al. (1980) for the terrains affected by the Proterozoic tectogeneses (50 mWm^{-2}, Table 4).

The values in the Moesian Platform can be grouped into two distinct areas : a western part with relatively high heat flow (59–78 mWm^{-2}) and an eastern one with lower heat flow (39–56 mWm^{-2}).

Remarks regarding this particularity have been previously made by Neguț (1972) and Paraschiv, Cristian (1976) on the basis of geothermal gradients from oil industry boreholes as well as from geographic distribution of heat flow values (Veliciu et al. 1977 ; Veliciu, Demetrescu 1979 ; Neguț 1982). The high heat flow in the western part of the Moesian Platform could be explained by the existence of the radiogenic heat sources represented by acid magmatic intrusions into its basement. Indeed, the deep drilling data (Paraschiv 1979) show that the Breton tectogenetic phase resulted in the intrusion of gabbros, diorites, granitic porphyries and granitic aplites, which pierced the Devonian formations. Such intrusions were not encountered in the eastern part of the Moesian Platform.

It is interesting to notice that the mean heat flow, calculated for the entire territory of the Moesian Platform (58 mWm^{-2}), is in good agreement with the mean (61 mWm^{-2}) established by Chapman, Furlong (1977) for the terrains affected by Hercynian tectogeneses (Table 4).

The Predobrogean Depression exhibits relatively high values of heat flow (75 mWm^{-2} in the Danube Delta). The high values are related to the Alpine structures south of the Scythian Platform and these can be correlated with the structures from the North Crimea (Săndulescu 1980), where a geothermal activity has been observed.

The Carpathian chain (including depressions) is characterized by heat flow ranging from 26 to 126 mWm^{-2}. This broad interval is a consequence of various geological and tectonic features : the Alpine type structure of the Carpathians, the large development of Cretaceous and Paleogene flysch deposits, the presence of ophiolites and important masses of Neogene volcanics. These features are an expression of peculiar structure of the crust in a region where the Alpine orogenic belt comes in direct local contact with the old East-European craton.

The average heat flow, calculated for the whole Carpathian area (including depressions) is 69 mWm^{-2} (standard deviation : $\pm$ 16 mWm^{-2}), that is closely to the mean heat flow from Table 4, established for the Cenozoic folded regions of Eurasia (71 mWm^{-2}). This observation could be surprisingly taking into account the very low heat flow recorded in the

Transylvanian Basin (40—50 mWm^{-2}); however, this low heat flow is probably compensated by the high heat flow from the Neogene volcanic zone (73—126 mWm^{-2}) and from the Pannonian Basin (85—108 mWm^{-2}).

Temperatures within the crust on the Romanian territory

Terrestrial heat flow values observed at the earth's surface, radiogenic heat generation as well as heat conductivity data have been associated to the geological-structural cross-section from Fig. 6—B (modified after Rădulescu et al. 1976).

As for the adopted geothermal model (Fig. 6—F), the gamma ray spectrometric measurements on the rock samples offered direct information regarding the heat generation in the first few kilometrs of the crust. Radiogenic heat generation in the sedimentary covers ranges from 0.33 μWm^{-3} (carbonatic rocks) to 1.5 μWm^{-3} (molasse and flysch). For the metamorphic schists an average value of 2.1 μWm^{-3} has been considered.

The deeper zone of the crust has a radiogenic heat generation value of 0.25 μWm^{-3} representative for intermediate-composition granulite-facies rocks. Beneath the base of the crust, a depleted ultrabasic zone with a characteristic heat production of 0.01 μWm^{-3} has been assumed. Both values are in good agreement with the heat generation —seismic waves velocity relationship, in the range 5.0—9.0 km s^{-1} (Rybach 1976).

Heat conductivity values of 2.3 $Wm^{-1}\,°K^{-1}$ for the upper crust and 3.3 $Wm^{-1}\,°K^{-1}$ for the lower crust have been inserted into the calculations.

Temperature vs. depth distribution has been obtained by automatic computation using the solution of the heat conduction equation under the following assumptions (Haenel 1979): (1) the analytical solution of the equation of heat conduction can be greatly simplified by assuming stationary conditions and an one dimensional temperature distribution, i.e. temperature depending only on depth; (2) the model used takes into account the planeparallel stratification (within a certain portion of the geological cross-section from Fig. 6—B), where the source strength and heat conduction are uniform within each layer; (3) no dependence of thermal properties was considered in respect of temperature and pressure.

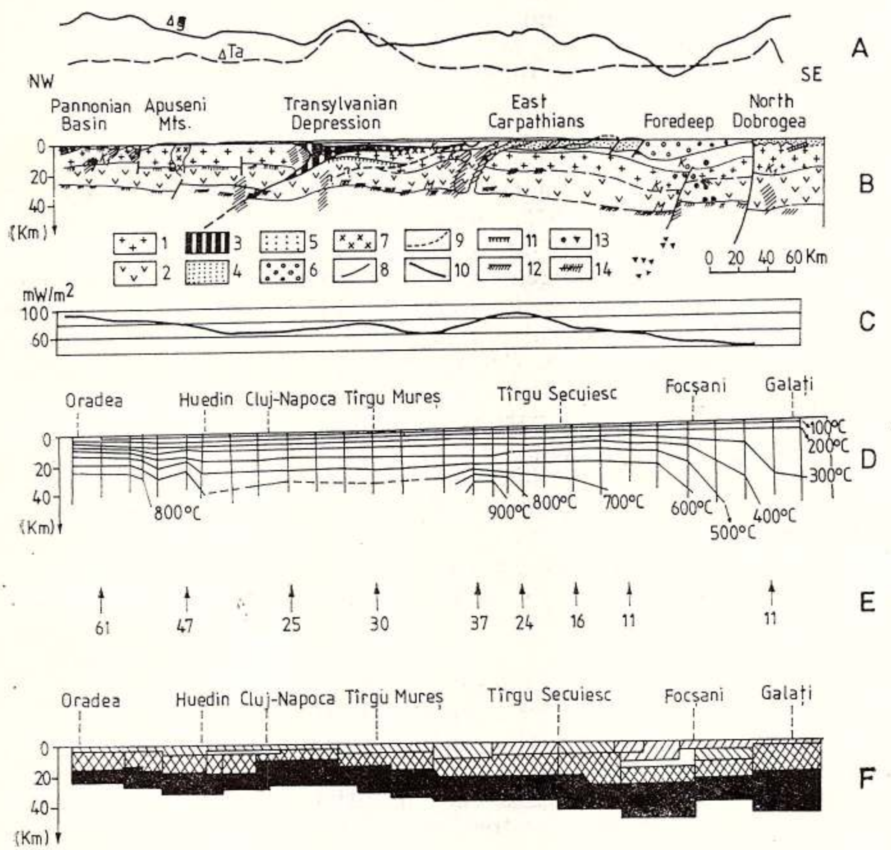
The calculated isotherms are presented in the lower part of Fig. 6—D. It is worth noticing that similar temperature-depth distributions have been obtained by Kutas (1973) for the Soviet Carpathians (in the same tectonic units of the East Carpathians) and by Bodri (1976) for the Pannonian Basin.

The geoisotherms from Fig. 6—D show a possible temperature difference exceeding 500 °C at the base of the crust under the East Carpathians Bend (Vrancea region) and a horizontal gradient of 200—300 °C/100 km is not to be excluded. This fact supplies new evidence that the region is still tectonically active, the accumulated tensions in the lower crust being probably released in shallow earthquakes.

The mantle heat flow has been calculated according to:

$$q_M = q_0 - \int_0^{z_M} A(z)\, dz$$

where q_0 is the heat flow observed at the surface; q_M represents the heat generated at the base of the crust (Moho-discontinuity) or transferred from the deeper zones (approximately equal to the so-called "reduced heat flow"); z_M is the depth at the base of the crust and $A(z)$ describes the productivity of radiogenic heat sources within the crust.



G

A(μWm^{-2})	0,33–1,0	1,1–1,5	2,1	2,6–3,2	0,25
K(Wm^{-1}°K^{-1})	1,9–2,4	1,7–1,9	1,7–2,0	2,3	3,3
SYMBOL					

Fig. 6. Characteristics of the geotraverse in the Carpathian area: A-regional gravity (Δg) and magnetic (ΔT_a) anomalies; B-geological-structural cross-section after Rădulescu et al. 1976). (1-upper crust; 2-lower crust; 3-oceanic crust; 4-flysch; Foredeep, 5-inner part; 6-outer part; 7-volcanic rocks; 8-thrust and overthrust; 9-boundaries; 10-faults; data revealed by DSS and seismology, 11-Conrad discontinuity K_1; 12-Moho-discontinuity M; 13-hypocentres of normal and intermediate earthquakes; 14-faulting zone); C-surface heat flow; D-calculated isotherms; E-calculated mantle heat flow (mWm^{-2}); F-adopted geothermal model; G-identifying symbols for geothermal model.

The study performed so far indicated that the heat flow from the upper mantle increases eastwards. Below the foreland tectonic units of the Carpathians a mantle heat flow may exist being 2–4 times lower as compared to the Alpine orogenic area (11–16 mWm^{-2}, respectively 24–61 mWm^{-2}). On the eastern border of the Pannonian Basin the contribution of the mantle to the surface heat flow exceeds probably 50%, whereas for the Moldavian Platform and the easthern part of the Moesian Platform this contribution may be less than 25%.

It is also interesting to notice that on Romania's territory the Moho discontinuity is clearly an unisothermal surface. Consequently, this possibly marks a petrographic (chemical) change but no phase transition.

Geothermal regime of the East Carpathians

A "classical" division of the East Carpathians (Dumitrescu, Săndulescu 1968; Săndulescu 1975, 1980) pointed out the following zones (from west toward east); Transcarpathian Zone, Crystalline-Mesozoic Zone, Flysch Zone and Neogene Zone. Otherwise, taking into account the age and the type of different groups of nappes, the East Carpathians exhibit the following structural units: Inner Dacidian Nappes (Pienides, Transylvanian Nappes System, Central East Carpathians Nappes System), Outer Dacidian Nappes, Moldavides and Foredeep. Post-nappe covers are also known. The eastern part of the Transylvanian Basin as well as the Neogene-Quaternary molasse depressions are superposed on parts of the deformed East Carpathians or on their post-nappe covers. The inner part of the East Carpathians is characterized by Neogene-Quaternary calc-alkaline magmatic activity.

The geological structure of the East Carpathians is now viewed as a result of effects of the plate tectonics (Bleahu et al. 1973; Rădulescu, Săndulescu 1973; Hertz, Savu 1974; Săndulescu 1980). The present geothermal data are generally consistent with the geodynamic model suggested by Rădulescu, Săndulescu (1973). This model envisages that the Carpathian area comprises two suture zones: the Transylvanides (inner position) and the Outer Dacides (outer position) each of them corresponding to an Alpine paleoplane of „consumption" of the oceanic and/or thinned lithosphere. The "consumption" process was connected with the main compressive tectogenetic "Carpathian phases".

The first important compressive period occurred between the Barremian and the Late Albian, affecting successively the Transylvanian ophiolitic realm and the East Carpathian Dacidic Zone. Consequently both the obduction in the Transylvanian area and the basement shearing nappes in the East Carpathians were generated.

The Outer Dacidic megatrough was subjected to less important compressions; nevertheless it was involved in the crustal shortening process since the end of the Upper Cretaceous.

During the Early Miocene-Middle Sarmatian, the subduction was still active in the East Carpathians area having as a result the overthrust of the Moldavian cover nappes and simultaneously inducing the extra-heat that produced the partial melting above the subducting plate. The hot upwelling mantle material expanded laterally at the base of the continental crust and locally pierced it where a large mass of andesites extruded.

Heat flow values for the East Carpathians have been reported by Veliciu et al. (1977), Demetrescu (1978), Veliciu, Visarion (1984). The geographic distribution of the surface heat flow shows that the Neogene volcanic chain is characterized by high values ranging from 73 to 126 mWm^{-2}, exceeding by a factor of 1.5 or 2.0 the value accepted for the "normal" heat flow (approximately 60 mWm^{-2}, Pollack 1982). The high heat flow anomaly overlaps also the outer Alpine paleoplane of lithospheric "consumption".

It is an interesting feature that the youngest andesitic rocks are located in the southern part of the chain (Harghita Mts) where values of 83–118 mWm^{-2} were measured; however, similar high heat flow values (85–126 mWm^{-2}) have been recorded on the older andesitic formations from the northern part (Oaş Mts). So, despite a southward migration of the volcanic activity versus time (Rădulescu 1972), the high heat flow anomaly is almost constant along the entire neo-volcanic chain.

A future review of the radiometric age data from the northern part of the neo-volcanic chain could offer new elements for the correlation of the geological and geothermal features mentioned above.

In order to study the inter-relations among the geodynamics of the East Carpathians, the building up of the Neogene volcanic chain, the inner heat sources and the lithospheric heat transfer, it has been necessary to elaborate a dynamic geothermal model (Rădulescu et al. 1981; Veliciu, Visarion 1984). This model takes into consideration not only the radiogenic heat sources and the steady-state heat conduction but also the subducting process during the Miocene time. The dynamic model is imposed by the relatively young age (less than 25 m.y.) since the last thermal-tectonic event in this region.

As for the construction of the geotherm families, it has been based on the observed surface heat flow associated with data of heat conductivity and heat generation, estimated from petrologic and geophysical arguments. The adopted model for the East Carpathians comprises an upper crustal region enriched in radioactive sources (heat generation 2.1–2.4 μWm^{-3}), an intermediate-composition granulite facies lower crust (0.25 μWm^{-3}) and a depleted ultrabasic zone (0.01 μWm^{-3}) overlying the pyrolite mantle (details in Veliciu 1977, and a more complete discussion in Pollack, Chapman 1977). All the chosen values fit the graph reported by Rybach (1976), according to the heat generation – seismic wave velocity relationship in the range 4.5–8.7 km s^{-1}.

Deep seismic soundings revealed depth of 16–20 km for the Conrad-discontinuity and 37–43 km for the Moho (Rădulescu 1979). The lithospheric thickness of 150 km was inferred from magnetotelluric measurements (Stănică, Stănică 1981).

A resulting family of geotherms is presented in Fig. 7; the various heat flow values correspond to the different tectonic units. The diagrams indicate that the highest temperatures may be reached below the inner part of the East Carpathians: 500–600°C at the boundary between upper and lower crust and over 900°C at the base of the crust. This fact constitutes new arguments for some previous volcanologic hypotheses regarding the Neogene-Quaternary magmatic activity in the region (Rădulescu, Borcoș 1969). These hypotheses took into consideration the

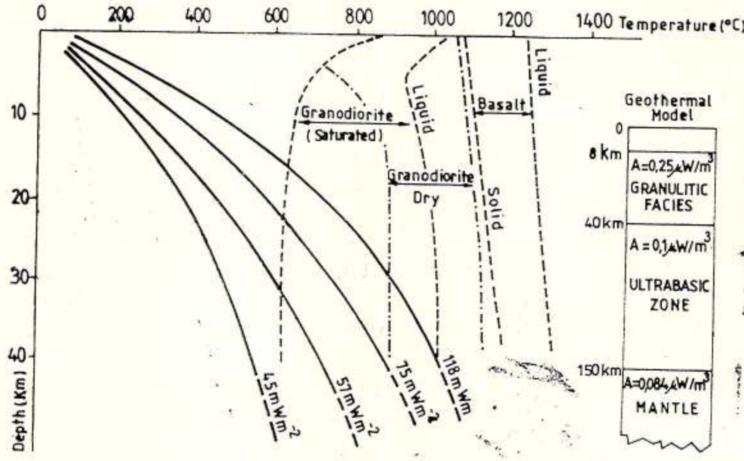


Fig. 7. Geotherms family calculated for the East Carpathians (steady-state conduction). Family parameter is surface heat flow (mWm^{-2}) corresponding to the following tectonic units: 75–118, neo-volcanic zone; 57, Central East Carpathians nappes system and Outer Dacides; 45, Moldavides and Moldavian Platform. Phase transition for intermediate and basaltic crustal rocks after Wyllie (1971).

possible existence of some phase transitions ("solidus") in the crust, at the level of the Conrad-discontinuity, where calc-alkaline magmas stationed during their ascent.

A very similar idea has been illustrated on the crustal cross sections in a work published by Socolescu et al. (1964).

The calculation performed for the East Carpathians shows a value of 25–30 mWm^{-2} penetrating the base of the crust. Accordingly, the major part of the heat is generated by radiogenic elements within the crust, but it may be a significant contribution from the young subcrustal processes.

The vertical temperature distribution (Fig. 6–D) suggests the presence of a positive temperature anomaly (over 1000°C) located in the upper mantle. The origin of this anomaly is possibly due to the Alpine subduction in the East Carpathians area. Some theoretical models for collision of two lithospheric plates generating subduction exhibit similar temperature anomalies for time constants ranging between 150 and 25 m. y. (Hasebe et al. 1970; Toksöz et al. 1971; Lubimova, Nikitina 1978). The models of the subduction mechanism at the top of a descending plate show the stretching in the plate as it bends downwards into the underlying mantle; this implies subsequent compression as the plate experiences resistance to its motion. The compression results in an increase in temperature of the down-going slab that can be expressed by the general equation :

$$\rho c\left(\frac{\delta T}{\delta t} + v \nabla T\right) = K \nabla^2 T + A$$

where ρ is density, c is specific heat capacity, T is temperature, t is time, v is subduction rate, K is the thermal conductivity and A represents the radioactive heat generated in the crust.

In Fig. 8 is presented a simplified two-dimensional geothermal model from Lubimova, Nikitina (1978), adapted to the Alpine subducting mechanism in the East Carpathians. Three distinct zones have been separated within the section from Fig. 8 : zone (I) with continental crust (lithosphere) ; zone (II) with oceanic crust (lithosphere) ; an intermediate zone (III) between zone (I) and (II) where the "consumption" process of the lithosphere (Benioff Zone) takes place.

In the x-y coordinates system, the continental lithosphere is represented as a semi-infinite plate $(-\infty \leqslant x \leqslant 0 ;\ 0 \leqslant z \leqslant l_1)$ with the following constant geothermal parameters : $v_I = 0$; K_I ; T_I ; T_0 ; l_I (l_I is the lithospheric thickness). For the steady-state continental plate ($\delta T/\delta t = 0$) the heat conduction equation has the form of Poisson's equation, which can be solved without any mathematical difficulties :

$$K_1 \nabla^2 T(x, z) + A_I(z) = 0$$

Owing to the possible physical effects of heating related to the displacement and evolution of a continental margin, it is necessary to consider the friction, phase transitions and adiabatic compression. Such effects are described by a temperature function $T = f(z)$ on the vertical boundary of the semi-infinite plate (I). The boundary conditions are :

$$T_{z=l_I} = T_0 ;\ T_{z=0} = T_I ;\ T_{x=0} = f_I(z) ;\ \delta T/\delta x|_{x\to-\infty} = 0.$$

In this simplest formulation the continental plate (I) is represented by a "static thermal intrusion model". The general solution to such a

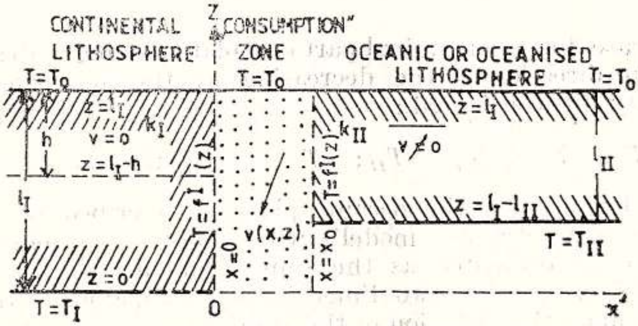


Fig. 8. Simplified two-dimensional model adapted from Lubimova, Nikitina (1978) for the thermal effect produced by Alpine subduction process in the East Carpathians (see explanations in text).

physical-mathematical problem is expressed by the sum of two terms: the "normal" temperature $T_0^I(z)$ with uni-dimensional variance and "anomalous" temperature $T_{add}^I(x, z)$ with a bi-dimensional variance:

$$T = T_0^I(z) + T_{add}^I(x, z)\,;\ (-\infty \leqslant x \leqslant 0\,;\ 0 \leqslant z \leqslant l_I)$$

The function $f_I(z)$ can be defined on the basis of observational data $(f_{obs}^I(z))$ from the actual island-arc structures of the Pacific Ocean (Fig. 9).

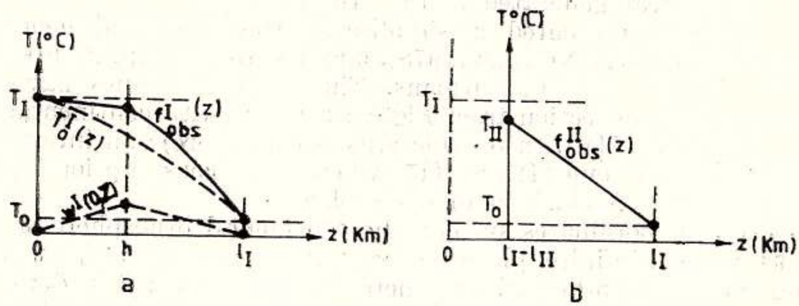


Fig. 9. Schematic diagrams of temperature distribution; *(a)* under the island arc and *(b)* under the border of the moving oceanic plate (after Lubimova, Nikitina 1978).

A similar model has been elaborated for the oceanic (or oceanised) lithosphere of the semi-infinite plate (II), considered in a solid within limits: $x_0 \leqslant x \leqslant +\infty\,;\ l_I - l_{II} \leqslant z \geqslant l_I$ and parameters $v_{II}(v_x, 0, 0)$; K_{II}, l_{II}, T_0, T_{II} (Fig. 8). For this plate, omitting the terms $\delta T/\delta t$ and A_{II}, which do not limit the generality, and taking into account the oceanic plate motion towards the continental one, the heat conduction equation with a convective term is obtained:

$$K_{II}\nabla^2 T(x, z) + \rho c v_{II}\,\frac{\delta T(x,z)}{\delta x} = 0$$

In this case, for the marginal part of a plate moving with a horizontal rate v_{II} in the direction of the decreasing x, the following boundary conditions may be considered:

$$T_{z'=l_I} = T_0\,;\quad T_{z=l_I=l_{II}} = T_{II}\,;\quad T_{x=x_0} = f_{II}(z)\,;\quad \delta T/\delta x|_{x\to+\infty} = 0$$

In this formulation the oceanic plate is described in terms of a "dynamic thermal intrusion model" with $v \neq 0$. The general solution of the problem is presented as the sum of one-dimensional "normal" temperature $T_0^{II}(z)$ and a two-dimensional "anomalous" temperature $T_{add}^{II}(x, z)$. The analytical solution of the problem for v_{II} = constant and K_{II} = constant, was given by Lubimova, Nikitina (1978). There is also the possibility of defining the function $f_{obs}^{II}(z)$ on the basis of the available experimental data concerning heat flow over the oceanic plate border and on the ocean floor in the north-western Pacific Ocean (Fig. 9).

For the adopted model, the intermediate "consumption" zone (i.e. the subduction zone of an oceanic or oceanised plate which descends to the marginal part of the continental plate) is located in the area : $0 \leqslant x \leqslant x_0$; $0 \leqslant z \leqslant l_1$. All the values $z \leqslant 0$ in Fig. 8 are pertinent to the mantle's lower layer (asthenosphere).

The temperature field $T(x, z)$ in the intermediate "consumption" zone has been studied by the general heat conduction equation which takes into account both the convective heat-mass transfer $v_{III}\ (v_x, 0, v_z)$ and the generation of heat $A \neq 0$:

$$K_{III}\nabla^2 T(x, z) - \rho c v_{III} \nabla T(x, z) + A^{III}(x, z) = 0$$

where A^{III} is the superposition of all types of heat sources in the "consumption" zone. Two functions of temperature distribution are the boundary conditions.

For a rigorous mathematical solution of the formulated two-dimensional problem, not only geothermal parameters $K_{III}(x, z)$, ρ, c, should be known, but also the velocity-vector $v_{III}(x, z)$ within the "consumption" zone. The assumption was that the material in the "consumption" zone acts, on the geological time scale, as a fluid. The vector v_{III} can be obtained from the solution of a problem based on Navier-Stokes equation.

The problem was investigated by a computer-aided solution and the distribution rate obtained was introduced into the heat conduction equation. In this way the equation has a definite mathematical formulation and, consequently, gives a unique solution for the temperature field T_{III}.

The result of the computation performed for the East Carpathian area is presented in Table 6. The geothermal parameters inserted into the calculation have been presented above. It is to point out that the Alpine subduction rate in the Carpathians was considered similar to the rate observed for the actual island-arc structures from the Pacific Ocean. A time scale referred to the main compressive tectogenetic "Carpathian phases" has been used.

A problem that deserves consideration is whether any igneous-derived thermal anomaly may exist in connection with the Pannonian and/ or Pliocene volcanic and/ or subvolcanic formations from the East Carpathians. Mathematically the thermal calculations contain two major assumptions (Smith, Shaw 1975) : the heat transfer in the rocks surrounding the magma is by solid-state conduction and effects of magmatic pre-heating and increases of magma after the time of the last eruption are ignored. Such estimations involve knowledge on the age and volume of magmatic manifestations.

In Fig. 10 the age-volume data for seventeen volcanic systems from the East Carpathians are plotted to show the approximate present position of each system in relation to its probable cooling state. A pair of lines is drawn to represent cooling models after Carslow, Jaeger (1959), Shaw (1974), that identify igneous systems that are now approaching the ambient temperature.

The following hypotheses have been taken into consideration. The inner part of the East Carpathians is characterized by a calc-alkaline magmatic activity manifested effusively during the Mio-Pliocene

TABLE 6

Generation of extra heat due to the Alpine subducting process in the East Carpathians area

Depth (km)	Subduction rate (cm yr^{-1})	Energy released (J m^{-2} yr^{-1})	Temperature (°C)
20	0.20	110	600
30	0.49	300	675
40	0.60	400	850
50	0.72	500	950
60	0.85	600	1000
70	1.00	675	1030
80	1.25	800	1050
90	1.60	1000	1080

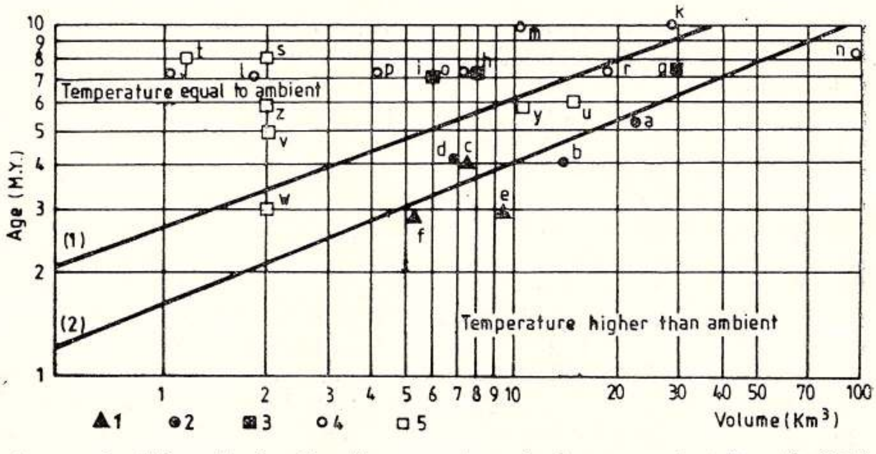


Fig. 1(. raph of theoretical cooling time *vs.* volume for igneous systems from the East Carpathians. Pair of lines represents cooling models after Shaw (1974). Symbols represent youngest age and estimated volume. 1-Harghita Mts. (c-Vîrghiş, e-Luci, f-Cucu); 2-Gurghiu Mts. (a-Fincel-Lăpuşna Caldera, b-Şumuleu-Fierăstrae, d-Ciumani); 3-Călimani Mts. (g-Călimani Caldera, h-Călimani Spring, i-Zebrac-Mermezeu); 4-Tibleş-Rodna zone (j-Botiza, k-Hudin, l-Stegioara group, m-Grohot-Tomnatec, n-Ţibleş-Măgura Neagră, o-Cormaia, p-South Rodna group, r-Vinului Valley-group); 5-Oaş-Gutîi Mts. (s-Comlăuşa Caldera, t-Batarci Caldera, u-Săpinţa Caldera, v-Pietroasa Stratovolcano, z-Herja, y-Igniş Stratovolcano, w-Gutin).

and Pleistocene volcanism in the Gurghiu-Harghita Mts representing the final stage of the subsequent Alpine magmatic activity (Rădulescu 1973). From a petrographic point of view, the Neogene magmatic rocks cover the whole range of calc-alkaine magmas but they exhibit a quite clear predominance of andesites in respect of rhyolites, dacites and basalts.

Taking into account the petrographic composition, with the prevalence of andesites, the following parameters have been used into the calculations : initial temperature of magma 1000°C, mean density 2.9 g cm^{-3}, latent heat of crystallization 65 cal g^{-1}, heat capacity 0.3 cal g^{-1} $°C^{-1}$.

As for the age of the magmatic systems, the volcanism in the Oaș-Gutîi Mts started during the Lower Badenian on a basement constituted by sedimentary formations of Senonian-Oligocene age.

Numerous hypabyssal structures outcrop in the Țibleș, **H**udău, Bîrgău and Rodna Mts. This region is geographically located within the volcanic andesitic chain of the East Carpathians, but no definite volcanic structure has been yet identified (Rădulescu, Dimitrescu 1982). Magmatic rocks usually come in direct contact with Paleogene formations which are metamorphosed ; so, the information on the age in this situation is of post-Paleogene time. On the graph in Fig. 10 the youngest possible age has been plotted (Săndulescu, Udubașa, personal communication).

The beginning of magmatic activity in the Călimani-Gurghiu-Harghita Mts is considered in the Late Pannonian (Rădulescu 1973 ; Mihăilă, Peltz 1977). Here, Quaternary sediments with interbeddings of volcanic horizons have been identified. Consequently, the continuity of magmatic activity was supposed at least till the Middle Pleistocene time.

The radiometric age determinations by K-Ar method (Rădulescu 1973) offered two extreme elements for the Călimani-Gurghiu-Harghita Mts : 7.37 ± 0.66 m.y. (andesite with augite and olivine from the South Călimani) and 3.92 ± 0.2 m.y. (andesite with hypersthene from the Vlăhița pit). In respect of these age determinations, the volcanics of the upper compartment are situated between 7 and 3 m.y.

Data regarding the magnitude of the magmatic systems are plotted in Fig. 10 on the basis of cartographic surface occupied by the systems, tectonic-magmatic characteristics, superstructure volume and geophysical mapping (gravity and airborn magnetic anomalies ; Suceavă 1974, Cristescu 1977).

The points plotted on the graph from Fig. 10 indicate that almost all igneous-systems from the East Carpathians have reached the ambient temperature or are very near to this. This fact is due to the relatively great age of the last eruption in the region, ranging between 7 and 1.5 m. y.

Considering the geological evolution of the East Carpathians area and the suggested geothermal models, it may be concluded that both the high heat flow and building up of the neo-volcanic chain are consequences of the lithospheric Miocene subducting process. The source of the high surface heat flow observed at the inner part of the East Carpathians seems to have the following partition : approximately 60—70 % of heat is released by the radiogenic generation, 25—30 % is produced by the heat connected with the Alpine subduction and preserved in the upper mantle and lower crust, and a few percent is due to the heat content of the upper levels in the crust where the andesitic magma accumulated during its ascent and the thermal equilibrium has not been reached yet.

Some geothermal characteristics of the Pannonian Basin

Both the geothermal gradients and heat flow data demonstrated that the Pannonian Basin is characterized by a high heat flow anomaly (mean 96 mWm^{-2} after Horváth et al. 1981). Average temperatures at a depth of 1 km are of 70–80 °C.

A remarkable feature of this geothermal anomaly is its uniform distribution, with high heat flow values over the whole Pannonian area. This fact has been demonstrated by the heat flow determinations performed on the eastern border of the basin, where values of 94 mWm^{-2} (Arad) and respectively 85 mWm^{-2} (Sîniob) have been recorded (Veliciu et al. 1977).

The downward continuation of the geothermal data suggested the existence of the temperatures of 800°C at the base of the crust (Fig. 6-D) and of 1000–1200°C at a depth of 60 km (Fig 13). The calculated mantle heat flow has a value of 60 mWm^{-2} beneath the eastern border of the Pannonian Basin (that means approximately 50% of the surface heat flow).

Deep seismic soundings (DSS) indicated a thin crust (23–30 km), despite of a normally developed upper crust (17–19 km, after Posgay et al. 1981). The low velocity zone (LVZ) is situated in an elevated position. The compressional seismic wave velocity increases with depth to 9.1 km s^{-1} as far as 50 km depth, then it diminishes to 7.7–7.8 km s^{-1} at a depth of approximately 60 km.

Magneto-telluric measurements (Adam 1965) revealed a rise of the high conductivity layer (HCL).

The Bouguer anomaly within the basin has an average from +10 to +15 mgali, that is surprisingly low taking into account the thinning of the crust and the rise of the upper mantle. Computed gravity models (Horváth, Stegena 1977) argumented for a density of the upper mantle lower than normal.

Geothermal data correlated with other geophysical information (LVZ and HCL in an elevated position, low density) provide consistent arguments toward the partial melting hypothesis of the upper mantle beneath the Pannonian Basin.

The Pannonian Basin was formed during and after the Miocene tectogeneses which affected the Carpathian area, by a Badenian subsidence process continued by a rapid subsidence in the Pannonian. The Pannonian and post-Pannonian sediments have a maximum thickness of 5 km depending on location.

The mechanism of the subsidence as well as the high heat flow anomaly can be explained using an extensional lithospheric model. Mc Kenzie (1978) noticed that in the sedimentary post-tectonic basins, where geological and geophysical evidence for an extending lithosphere exist (thinner crust and lithosphere, a certain pattern of the tensional faults, recent basic magmatism within the bäsin), the surface heat flow is abnormally high (over 85 mWm^{-2}). In order to explain this phenomenon, Lachenbruch, Sass (1978) studied two different extensional models : (1) a model with extension viewed as stretching in the homogeneous plastic sense produces a thinner lithosphere and (2) a model that represents homogeneous stretching of a lithosphere whose thickness is maintained constant by accretion of crystalline material at its base and with a vertical convective transport by fluid basalt in dyke-like intrusions.

A horizontal extension at a constant rate "s" accounts for the relatively uniform thickness of the lithosphere and the high heat flow on the entire basin. On the other hand, the lithospheric thinning is partly compensated by the supply (accretion) of melted basaltic material from the asthenosphere which crystallizes at the base of the lithosphere (updoming of the asthenosphere or "mantle diapir").

The equation that describes this process is :

$$K \frac{d^2T}{dt^2} + v\rho c \frac{dT}{dz} = A_0 e^{-\frac{Z}{D}}$$

where the terms are heat conduction, heat convection and respectively, radiogenic heat generation ; v represents the vertical migration velocity of the melted basalts.

Using the error function, Carslaw, Jaeger (1959) gave an analytical solution :

$$T = e^{h^2/\beta^2}\left[q + \frac{Lh}{c\beta^2}\right]\frac{\beta}{K}\left(\frac{\pi}{2}\right)^{1/2}\operatorname{erf}\left(\frac{z^2}{2\beta^2}\right)^{1/2} + A_0 D^2 \frac{1}{K}(1 - e)^{Z/D} + \ldots)$$

where :$\beta^2 = K/s\rho c$; $s = v/z$.

In Fig. 11 the calculated family of geotherms is presented for an extensional velocity s expressed as % per m.y., with family parameter being the surface heat flow. The temperatures within the lithosphere are remarkably increased as compared with the simple conduction in steady-state thermal regime. The adopted extensional model suggests that the high regional heat flow anomaly from the Pannonian Basin may be referred to the abnormal heat flow from the asthenosphere (approximately 60 mWm^{-2}) plus the thermal effect of the extensional process at a rate of 1—3% per m.y.

The generalized development of an extensional basin may be divided into two stages (Mc Kenzie 1978). During and immediately after extension there is a rapid subsidence. This occurs in isostatic response to net density changes resulting from the lithospheric thinning and from the heating and thermal expansion. The second stage of subsidence is a relatively long-term process caused by cooling and contraction of the lithosphere following the extensional phase. The overall subsidence is generally amplified by the effects of sediment loading. If original crustal thickness, elevation and temperatures are known, a detailed analysis of subsidence history can be used to determine the magnitude of extension.

The extension (more correctly "the distension") of the Pannonian Basin was described as a result of large scale strike-slip faulting along a system of conjugate shears striking NE-SW and NW-SE respectively; NE trending faults exhibit sinistral offsets whereas NW trending faults are dextral. This conjugate set implies E-W extension.

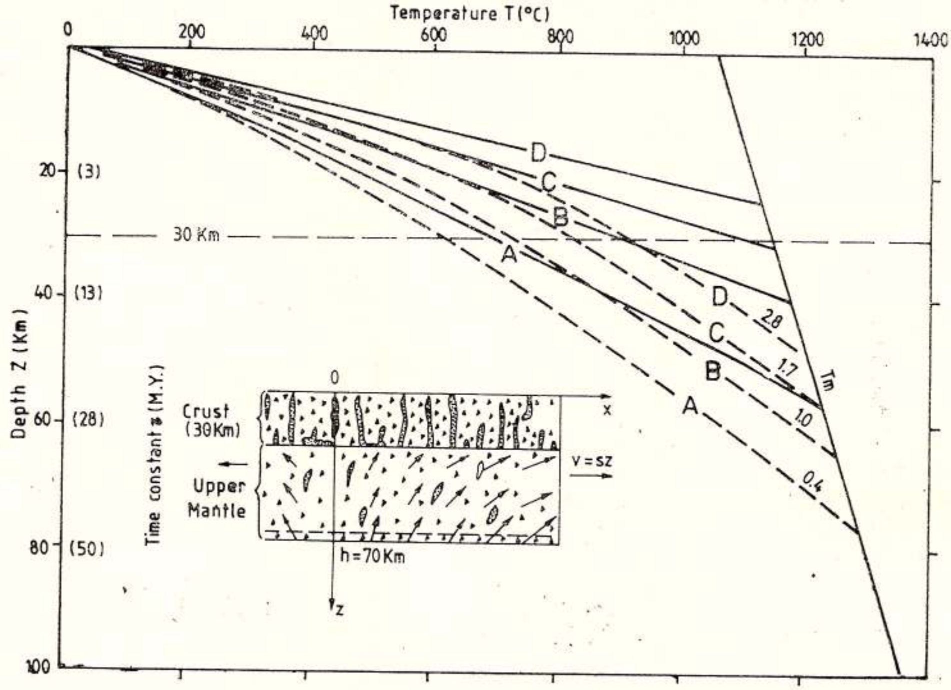
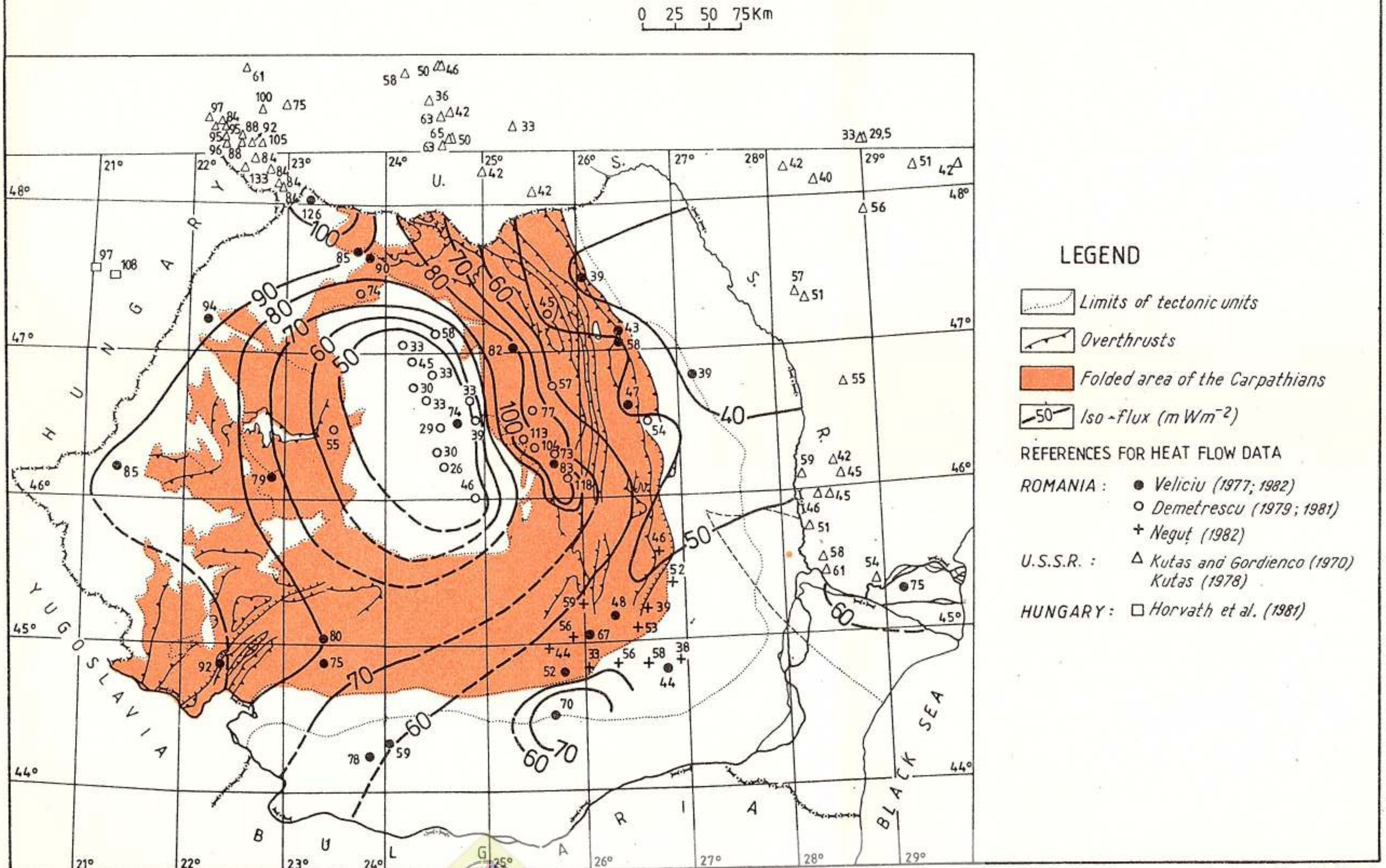


Fig. 11. Temperatures *vs.* depth diagram. Solid curves correspond to an extending lithosphere in the Pannonian area. Dashed curves correspond to geotherms for static case ($s = 0$). Numbers on solid curves are extension rates in % per m. y. Curves labeled A, B. C, D, correspond to surface heat flow values of 50, 70, 85 and respectively 110 mWm^{-2}. *Tm* is the mantle solidus.

Horváth, Royden (1981) have interpreted the Pannonian Basin as the result of Badenian extension (the "thermal" phase of subsidence). The extension became conspicuous during the Sarmatian time (asthenospheric attenuation or "mantle diapir") and manifested itself at a reduced rate during the Pannonian (passive cooling of the lithosphere). The cited authors assumed 50 to 100% extension in the Pannonian Basin and a total E—W extension was estimated as 75 to 100 km. Comparing this result to approximate 120 km Miocene crustal shortening estimated from palinspastic restorations for the East Carpathians (Săndulescu 1980; Ştefănescu 1980), the magnitudes are seen in fair agreement. However, it should be noticed that the palinspastic restoration of thrusts and folds provides a minimum estimate of crustal shortening. Further-

TERRESTRIAL HEAT FLOW IN ROMANIA

COMPILED BY S.VELICIU



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more, the above figures refer to the total Miocene shortening in the East Carpathians whereas the extension in the Pannonian Basin occurred only in the Badenian and post-Badenian time.

Geothermics of the Transylvanian Basin

From a geological point of view the Transylvanian Basin, located in the inner part of the Carpathians, was defined as a structural post-tectonic element corresponding to a homogeneous and young (Neogene) area which was subjected to molasse sedimentation (Dumitrescu, Săndulescu 1968). The molasse deposits overlie a folded basement and its post-tectonic cover.

From the measurements reported by Demetrescu (1973) and Veliciu et al. (1977) and from the Heat Flow Map (Plate I) a surprisingly low terrestrial heat flow appears typical of Transylvania. On the basis of the geothermal model elaborated by Veliciu, Visarion (1981), the present contribution tries to give a more detailed interpretation of this observation.

Reviewing the main geological and geophysical characteristics of the post-tectonic intermountain basins (intra- and inter-arc basins), Stegena et al. (1975) find the following common features : (1) thinner lithosphere, HCL and LVZ at higher position ; lower density in comparison with the average, (2) thinner crust, (3) synorogenic and particularly post-orogenic sediments affected less or not at all by tectonic movements, (4) low seismic activity for already developed basins, (5) sialic basins exhibiting andesitic volcanic activity of compressional type during the subducting process in the related areas, and "interarc-spreading" basaltic volcanism of extensional type after the subduction ceased, (6) high heat flow.

Regarding the last characteristic, the average heat flow in the Transylvanian Basin is only of 45 mWm^{-2}, indicating that the geothermal activity of this region is low. The low heat flow is outlined by the measurements performed in the surrounding tectonic units (Plate I), where values of 83—126 mWm^{-2} for the neo-volcanic chain of the East Carpathians, of 80 mWm^{-2} for the South Apuseni Mts and of 85—100 mWm^{-2} for the eastern part of the Pannonian Basin were determined (Veliciu, Demetrescu 1979).

From the pre-basin evolution of the Transylvanian Basin has been concluded that the main tectogeneses responsible for the structure of its folded basement took place between the Middle Cretaceous and the Senonian.

A complex correlation of geological, geophysical and drilling data reveals that the area occupied by the Neogene molasse of the Transylvanian Basin covers the junction of structural elements belonging to the major tectonic units of the Romanian Carpathians (Săndulescu, Visarion 1978). In the center of the depression both gravity and magnetic heights are partly related to the ophiolitic zone which is bilaterally overthrust on the Apuseni crystalline-bearing basement units westward and on the Central East Carpathians nappe system (the "root" of the nappes) eastward (Fig. 6-B). The folded basement is overlapped by a faulted and folded cover of the Neocretaceous, Paleogene and Lower Miocene age.

The Middle Mio-Pliocene molasse, which is approximately 4 km thick, fills this intramountain depression.

DSS data indicate that the crustal thickness is 34 km in the eastern part of Transylvania and decreases to 30 km in the center of the basin. Westwards the crust thickens out again to nearly 38 km. The crust is thin relative to the East Carpathians and Foredeep (40—50 km) but thick as compared to the Pannonian Basin (24—28 km). It is an interesting feature that the upper crust is only about 10 km thick, excluding the sedimentary complex.

Compressional wave velocities show the characteristics of continental type crust even under the ophiolitic zone : 5.5—6.2 km s^{-1} for the upper crust, 6.8—7.0 km s^{-1} for the lower crust and 8.2—9.0 km s^{-1} beneath Moho were recorded (Rădulescu 1979). The last figure may indicate that the density of the upper mantle could be greater than the average.

As for the lithospheric thickness, no direct investigation has been carried out so far. Nevertheless, some inferences can be made from the position of HCL in the eastern part of the Pannonian Basin (Adam 1965) and in the Vrancea region (Stănică, 1981) where depths of 60 km and more than 150 km respectively were recorded. A depth of about 100 km for HCL seems to be reasonable under the Transylvanian Basin.

Heat generation models were established separately for the Southern Apuseni crystalline-bearing basement units, the ophiolitic zone and the Central East Carpathians nappe system (Fig. 12). The calculated mantle heat flow varies from 20 mWm^{-2} to 30 % mWm^{-2}. The mantle heat flow

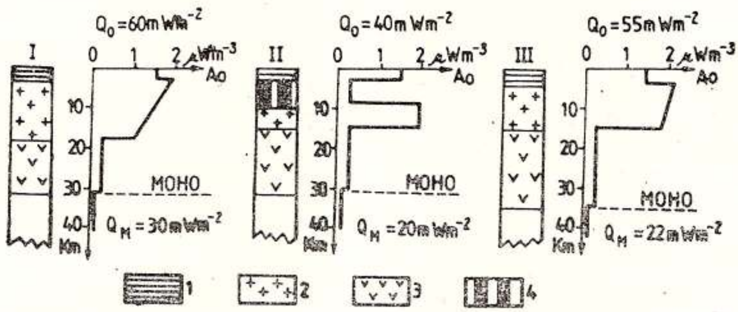

Fig. 12. Heat generation models A (z) for the crustal structure of the Transylvanian Basin after Veliciu, Visarion (1982). Model I for the Southern Apuseni crystalline-bearing basement units ; Model II for the ophiolitic zone ; Model III for the Central East Carpathians nappes system. (1-molasse and post-tectonic cover ; 2-upper crust ; 3-lower crust ; 4-ophiolites). Q_0 is surface heat flow ; Q_m is mantle heat flow.

has two components : one is due to radioactive heat sources in the subcrustal lithosphere, the other to the deeper contribution which arises from the asthenosphere and enters the lithosphere at its base. It is clear that the lower crust and upper mantle are less endowed with heat-producing radioactive sources than the upper crust ; on the other hand it is not known whether a decrement of surface enrichment is accompanied by a proportional decrement or increment in the lower crust (Pollack, Chapman 1977). Petrological arguments are consistent with either situation. The problem is open to discussion especially for the ophiolitic zone in the central part of Transylvania, where the heat generated in the upper crust is lowered by a factor of 6 or so, because of the content of basic rocks in the folded basement (Model II in Fig. 12). Fig. 13 shows geotherms

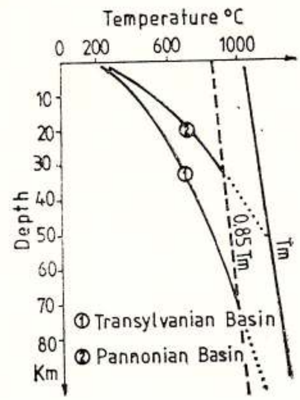

Fig. 13. Temperatures vs. depth diagram for the Transylvanian (1) and Pannonian (2) basins. T_m is the mantle solidus.

corresponding to the average surface heat flow observed in the Transylvanian Basin and the Pannonian area. The temperature calculated at the base of the crust in Transylvania (~600°C) is lower, as compared to that from the Pannonian Basin (800–900°C).

The geotherms were extended to a depth at which they intersect the mantle solidus (T_m) but were dotted to indicate provisionality above 0.85 T_m. The geotherms are characterized by nearsurface curvature due to the crustal heat generation and a nearly linear gradient through the depleted zone.

Various authors have suggested that the lithosphere-asthenosphere transition might begin at a temperature less than the solidus. Pollack, Chapman (1977) adopted as the depth of the base of the lithosphere the depth at which the geotherms reach 0.85 T_m. The geotherm for the Transylvanian Basin intersects 0.85 T_m at a depth of approximately 80 km while in the Pannonian Basin it reaches the same temperature at a depth of only 40 km. Consequently, the geothermal data clearly indicate a thicker and cooler lithosphere for Transylvania.

The model of Horváth, Stegena (1977) which explains the Late Cenozoic history and geophysical features of the Pannonian Basin in terms of updoming of the asthenosphere ("Mantle diapir") does not fit the characteristics of the Transylvanian Basin. During the Alpine subduction process, the Carpathian and Transylvanian area was subjected to compressions from the Middle Cretaceous to the Middle Miocene. After the subduction ceased no evidence indicates extension as occurred in the Pannonian Basin. It is shown by a thicker and cooler lithosphere, just a relatively thin crust, the scarcity of tensional faults, discontinuous subsidence history and lack of contemporaneous basic magmatic activity within the basin. Consequently, lithospheric stretching as described by Lachenbruch, Sass (1978) for extensional basins, is not very likely to be present here. A descending convection current in the mantle could be taken into consideration under Transylvania (Constantinescu et al. 1976; Visarion, Săndulescu 1979) which was probably related to the formation of the mantle "diapir" below the Pannonian Basin.

The author wishes to express his gratitude to professors Sabba Stefănescu, Liviu Constantinescu, Dan Rădulescu and Radu Botezatu for stimulating the interest in geothermal studies, for their encouragement and for suggesting some improvements to this work. It is also a pleasure to acknowledge Drs. Mircea Săndulescu and Marius Visarion for valuable discussions regarding the geology and geophysics of the Carpathians. Thanks are due to Dr. Marcian Bleahu for his kind support for this work to be published.

APPENDIX I

Temperature gradient, thermal conductivity and heat flow values for the Romanian territory

Coordinates lat. N	Coordinates long. E	Elevation a.s.l. (m)	Min. depth (m)	Max. depth (m)	Temp. gradient ($m°Km^{-1}$)	Therm. conduc. ($Wm^{-1}°Km^{-1}$)	Heat flow (mWm^{-2})	Ref.
1	2	3	4	5	6	7	8	9
47°28′	26°05′	424	2183	2284	18	2.20	39	1
47°15′	25°42′	823	40	320	15	3.30	45	4
47°13′	22°13′	148	2304	2502	51	1.80	94	1
47°08′	26°25′	440	2412	2915	26	1.70	43	1
47°07′	24°30′	400	100	3200	24	2.40	58	4
47°04′	24°10′	350	100	3500	20	1.70	33	4
47°03′	26°25′	409	3816	4038	26	2.60	58	1
46°56′	24°17′	374	80	980	32	1.50	45	4
46°49′	27°09	80	1731	1883	18	2.20	39	1
46°40′	24°53′	580	80	980	31	1.10	33	4
46°40′	25°47′	1018	100	530	22	3.60	57	4
46°37′	26°29′	501	1520	1610	20	2.30	47	1
46°36′	25°30′	1250	120	270	37	2.00	77	4
46°34′	24°54′	500	80	980	32	1.10	39	4
46°31′	24°45′	380	2050	2354	28	2.60	74	1
46°30′	26°40′	480	850	1012	23	2.30	54	1
46°24′	25°26′	1048	120	520	37	3.20	113	4
46°21′	25°31′	950	30	220	63	1.60	104	4
46°18′	25°44′	1240	40	200	41	1.80	73	4
46°15′	25°44′	525	314	510	50	1.70	83	1
46°12′	21°20′	120	193	377	45	1.90	85	1
46°09′	22°53′	712	215	402	35	2.20	79	1
46°09′	25°52′	600	20	540	70	1.50	118	4
45°11′	26°19′	315	5511	5653	21	2.30	48	1
45°03′	26°03′	290	1848	2100	30	2.30	67	1
45°02′	23°25′	298	1123	1250	37	2.20	80	1

APPENDIX I (continued)

1	2	3	4	5	6	7	8	9
44°53′	23°25′	220	2930	3440	29	2.60	75	1
44°51′	22°24′	311	610	785	40	2.30	92	1
44°48′	25°48′	168	5009	6255	18	3.00	52	1
44°47′	26°49′	64	2500	2724	24	1.80	44	1
44°31′	25°42′	144	1223	1404	37	1.90	70	1
44°20′	24°03′	204	2092	2422	35	1.80	59	1
44°14′	23°53′	182	1546	1700	45	1.80	78	1
45°36′	26°46′						46	2
45°17′	26°01′						59	2
45°13′	26°38′						39	2
45°06′	26°31′						53	2
45°03′	25°53′						56	2
45°03′	26°02′						45	2
44°59′	25°39′						44	2
44°53′	26°39′						58	2
44°52′	26°21′						56	2
44°52′	26°56′						38	2
44°50′	26°02′						33	2
45°16′	29°11′						75	3
48°08′	23°24′	380			45	2.80	126	5
48°38′	23°49′	410			56	1.60	90	5
47°40′	24°44′	470			35	2.40	85	5
47°02′	25°20′	750			38	2.20	82	5

References: (1) Veliciu et al. (1977); (2) Neguț (1982); (3) Veliciu (1978); (4) Demetrescu (1979); (5) Veliciu, Visarion (1984). Thermal conductivity from measurements by transient method for (1), (3) and (5), any by divided bar for (2) and (4).

REFERENCES

Adam A. (1965) Einige Hypothesen über den Aufbau des oberen Erdmantels in Ungarn. *Gerlands Beiträge zur. Geophysik*, 74 p. 20—40, Leipzig.

Airinei S. (1981) Potențialul geotermic al subsolului României. Ed. Științifică și Enciclopedică, București.

Birch F. (1954) Heat from radioactivity. In : H. Faul (ed) *Nuclear Geology*, p. 148—175, New York.

Bleahu M., Bocaletti M., Manetti P., Peltz S. (1973) Neogene Carpathian Arc : A continetal arc displaying the features of an "Island Arc". *J. Geophys. Res.*, 78, p. 5025—5032.

Bodri L. (1976) Deep temperature and heat flow in the Pannonian Basin. Ph. D. Dissertation, Eötvös University, Budapest.

Carsllaw H. S., Jaeger J. C. (1959) Conduction of heat in solids. 2nd edit., Calderon Press, Oxford.

Chapman D. S., Furlong K. (1977) Continental heat flow-age relationship. *EOS Trans. Am. Geophys. Union*, 58; p. 1240—1251, Washington.

Constantinescu L., Constantinescu P., Cornea I., Lăzărescu V. (1976) Recent seismic information on the lithosphere in Romania. *Rev. Roum. Géol. Géophys. Géogr., sér. Géophysique* 20, p. 33—41, Bucarest.

Cristescu T. (1977) Report Arch. IGPSMS, Bucureşti.

Cristian M., Dogaru L., Mocuţa S. (1969) Cu privire la variaţia temperaturii rocilor din principalele provincii petrogazeifere din R.S.R. *Petr. Gaze* 20 (8), p. 578–681, Bucureşti.

— Dogaru L., Mocuţa S. (1971) Consideraţii asupra regimului termic al sondelor de mare adîncime din R.S.R. *Petr. Gaze* 22 (9), p. 522–527, Bucureşti.

Demetrescu C. (1973) Valori preliminare ale fluxului termic în Transilvania. *St. Cercet. Geol. Geofiz. Geogr.* ser. *Geofizică* 11 (1) p. 13–21, Bucureşti.

— (1979) Valori ale fluxului termic al Pămîntului în unele unităţi tectonice din R. S. România. *St. Cercet. Geol. Geofiz. Geogr.* ser. *Geofizică*, 17, p. 35–46, Bucureşti.

— , Ene M., Andreescu M. (1981) Asupra regimului termic al Depresiunii Transilvaniei. *St. Cerc. Geol. Geofiz. Geogr.* ser. *Geofizică* 19, p. 61–71, Bucureşti.

Dewey J. F., Bird J. M. (1970) Mountains belts and the new global tectonics. *Theophrastus Publ.* S. A., Athens.

Dumitrescu I., Săndulescu M. (1968) Problèmes structuraux fondamentaux des Carpathes Roumaines et de leurs avant-pays. *Ann. Com. Géol.*, 36, p. 195–281 Bucarest.

— , Săndulescu M., Lăzărescu V., Mirăuţă S., Pauliuc S., Georgescu C. (1962) Mémoire pour la carte tectonique de la Roumanie. *Ann. Com. Géol.*, 32, p. 5–96, Bucarest.

Haenel R. (1979) Determination of subsurface temperatures in the Federal Republic of Germany on the basis of heat flow values. *Geol. Jb.*, E 15, Hannover.

Hamza V. M., Beck A. E. (1972) Terrestrial heat flow, the neutrino problem and a possible energy source in the core. *Nature*, 240, p. 343–344.

Hasebe K., Fujii N., Uyeda S. (1970) Thermal process under island arcs. *Tectonophysics*, 10, p. 335.

Herz N., Savu H. (1974) Plate tectonics history of Romania. *Am. Geol. Soc. Bull.*, 85, p. 1429–1440, Denver.

Horváth F., Stegena L. (1977) The Pannonian Basin : A Mediterranean interarc basin. In : B. Biju-Duval, L. Montadert (eds) Structural history of the Mediterranean basins. Éditions Techniq, p. 333–340, Paris.

— , Royden L. (1981) Mechanism for the formation of the Intra-Carpathian basins : a review. *Earth Evol. Sci.* 3–4, p. 307–315, München.

— , Döveniy P., Liebe P. (1981) Geothermics of the Pannonian Basin. *Earth Evol. Sci.* 3–4, p. 285–291, München.

Kappelmayer C., Haenel R. (1974) Geothermics with special reference to application. *Gebr. Borntraeger*, p. 1–238, Berlin–Stuttgart.

Kutas R. I. (1978) Pole teplovix potokov i termiceskaia modeli zemnoi korî. *Nauk. Dumka*, p. 1–140, Kiev.

— Gordienko V. V. (1970) Teplovie polie Ukrainu. *Nauk. Dumka*, p. 1–115, Kiev.

Lachenbruch A. H., Sass J. H. (1978) Models of an extending lithosphere and heat flow in the Basin and Range Province. *Geol. Soc. Am. Memoir* 152, p. 34–69, Washington.

Le Pichon X. (1968) Sea-floor spreading and continental drift. *J. Geophys. Res.* 73 (12), p. 3661–3697.

Lubimova E. A., Nikitina E. (1978) Thermal models of arcs and ridges. A "source-span-sink" model. *Tectonophysics* 45, p. 341–362, Amsterdam.

Mc Kenzie D. (1978) Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.* 40, p. 25–32.

— , Parker L. (1967) The North Pacific, an example of tectonics on a sphere. *Nature* 216 (5122), p. 1276–1280.

Mihăilă N., Peltz S. (1977) Contribuţii la cunoaşterea aparatului vulcanic Hegheş (Racoşu de Jos, Mţii Perşani). *D. S. Inst. Geol. Geofiz.* LXII 5, Bucureşti.

Morgan W. J. (1968) Rises, trenches, great faults and crustal blocks. *J. Geophys. Res.* 73 (6), p. 1959–1982.

Negoiță V. (1970) Etude sur la distribution des températures en Roumanie. *Rev. roum. Géol. Géophys. Géogr.* sér. *Géophysique* 14, 1, p. 25–37, Bucarest.

Neguț A. (1982) Estimarea parametrilor ce caracterizează regimul termic al formațiunilor geologice în Muntenia și Oltenia. *Ph. D. Dissertation*, Univ. of Bucharest.

Oxburg E. R., Turcotte D. L. (1971) Origin of paired metamorphic belts and crustal dilatation in island arc regions. *J. Geophys. Res.* 76, p. 1315–1327.

Palmason G. (1973) Kinematics and heat flow in a volcanic rift zone with application to Iceland. *Geophys. J. Roy. Astron. Soc.* 33, p. 451–481, London.

Paraschiv D., Cristian M. (1973) Asupra particularităților regimului geotermic în NE Depresiunii Panonice. *Rev. Petr. Gaze* 24 (11), p. 655–660, București.

— , Cristian M. (1976) Cu privire la regimul geotermic al unităților structurale de interes pentru hidrocarburi din România. *Stud. Cerc. Geol. Geofiz. Geogr. Geofizică* 14 p. 65–75, București.

Pollack N. H. (1980) The heat flow from the earth : a review. In : P. A. Davies, S. K. Runcorn (eds) Mechanism of continental drift and plate tectonics. Academic Press, London-N. Y.

— , Chapman D. S. (1977) On the regional variation of heat flow, geotherms and lithospheric thickness. *Tectonophysics* 38, p. 279–296, Amsterdam.

Polyak B. G., Smirnov Y. A. (1968) Relationship between terrestrial heat flow and the tectonics of continents. *Geotectonics* 4, p. 205–213.

Rădulescu D. P. (1973) Considerații asupra cronologiei proceselor vulcanice neogene din Mții Călimani, Gurghiu și Harghita *D. S. Inst. Geol.* LIX 4, București.

— , Borcoș M. (1969) Aperçu général sur l'évolution du volcanisme néogène en Roumanie. *Ann. Com. Géol.* 36, p. 177–182, Bucarest.

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

— , Dimitrescu R. (1982) Petrologia endogenă a teritoriului R. S. România. Ed. Univ. București.

— , Săndulescu M., Veliciu S. (1983) A geodynamic model of the Eastern Carpathians and the thermal field in the lithosphere. *Ann. Inst. Géol. Géophys.* LXIII, p. 136–144, Bucarest.

Rădulescu F. (1979) Cercetări seismice privind structura crustei terestre în România. *Ph. D. Dissertation*, Univ. Bucharest.

Rybach L. (1976) Radioactive heat production in rocks and its relation to other petrophysical, parameters. *Pageoph.* 114, p. 310–317.

Săndulescu M. (1975) Essai de synthèse structurale des Carpates. *Bull. Soc. Géol. France* 17 p. 299–358, Paris.

— , Visarion M. (1978) Considération sur la structure tectonique du soubassement de la Dépression de Transylvanie. *D. S. Inst. Géol. Géophys.* LXIV, p. 153–173, Bucarest.

Sclater J. G., Jaupart C., Galson D. (1980) The heat flow through oceanic and continental crust and the heat loss from the earth. *Rev. Geophys. Space Phys.* 18, p. 269–311.

Shaw H. R. (1974) Diffusion of H_2O in granitic liquids : Part II Mass transfer in magma chambers. In : A. W. Hofman, B. J. Yoder, R. A. Yound (eds) Geochemical transport and kinetics. *Carnegie Inst. Washington Publ.* 634, p. 139–170, Washington.

Smith R. L., Shaw H. R. (1975) Igneous-related geothermal systems. In : D. E. White, D. L. Williams (eds) Assessment of geothermal resources of the United States *U.S. Geol. Survey Circ.* 726, p. 58–83, Washington.

Socolescu M., Popovici D., Visarion M., Roşca V. (1964) Structure of the earth's crust in Romania as based on the gravimetric data. *Rev. roum. Géol. Géophys. Géogr.* sér. *Géophysique* 8, p. 35–49, Bucarest.

Stănică M., Stănică D. (1981) Utilizarea cîmpului electromagnetic natural al Pămîntului la elaborarea unui model structural în zona de curbură a Carpaţilor Orientali. *St. Cerc. Geol. Geofiz. Geogr. Geofizică* 19, p. 41–53, Bucureşti.

Stegena L., Geczy B., Horváth F. (1975) A Pannon medence késökainozóos fejlödése. *Bull. Hungarian Geol. Soc.* 105, p. 101–123, Budapest.

Suceavă M. (1974) Contributions gravimétriques à la connaissance des structures eruptives du SE des Monts de Gurghiu, Roumanie. *Bull. Volcan* 34, p. 1205–1222.

Toksöz M. N., Minear J. W., Julian B. R. (1971) Temperature field and geophysical effects of a downgoing slab. *J. Geophys. Res.* 76 (3), p. 5951–5975.

Veliciu S. (1977) Some results on the mantle heat flow investigation in Romania. *Acta Geol. Acad. Sci. Hungaricae* 21, p. 265–268. Budapest

— , Demetrescu C. (1979) Heat flow in Romania and some relations to geological and geophysical features. In : V. Cermak, Rybach L. (eds) Terrestrial heat flow in Europe. Springer-Verlag, p. 253–260, Berlin–Stuttgart–New York.

— , Visarion M. (1982) On the low heat flow in the Transylvanian Basin. In : V. Cermak, R. Haenel (eds) Geothermics and geothermal energy. Schweizerbartische Verlagbuchhandlung, p. 91–100, Stuttgart.

— , Visarion M. (1984) Geothermal models for the East Carpathians. *Tectonophysics* 103, p. 157–165, Amsterdam.

— , Cristian M., Paraschiv D., Visarion M. (1977) Preliminary data of heat flow distribution in Romania. *Geothermics* 6, p. 95–98, Pergamon Press; Great Britain.

Visarion M., Săndulescu M. (1979) Structura subasmentului Depresiunii Panonice în România. *St. Cerc. Geol. Geofiz. Geogr.* ser. *Geofizică* 17 (2), p. 191–201, Bucureşti.

Vitorello I., Pollack N. H. (1980) On the variation of continental heat flow with age and the thermal evolution of continents. *J. Geophys. Res.*, 85, p. 983–995.

Von Herzen R. P. (1967) Surface heat flow and Some implications for the mantle. In: "The Earth's Mantle" Academic Press — London.

Wyllie P. J. (1971) Experimental limits for melting in the earth's crust and upper mantle. In : The structure and physical properties of the earth's crust. *Am. Geophys. Union Geophys. Mon.*, 14, p. 279–301, Washington.

REGIMUL GEOTERMIC AL ARIEI CARPATICE

(Rezumat)

În cadrul lucrării este prezentat regimul geotermic care caracterizează aria carpatică, avînd drept bază de discuţie valorile observate ale fluxului geotermic, conductivităţii termice şi generării radiogene de căldură în roci. De asemenea, au fost stabilite corelaţii între fluxul geotermic din interiorul orogenului carpatic (40–120 mWm^{-2} incluzînd şi depresiunile post-tectonice), Platforma Moldovenească (39–55 mWm^{-2}), Platforma Moesică (39–78 mWm^{-2}) şi particularităţile geotectonice.

Analiza modelelor structurale şi geodinamice actuale – ce au încercat încadrarea ariei carpatice în conceptul tectonicii globale – la care s-au asociat datele obţinute de autor referitoare la distribuţia geo-

grafică a fluxului geotermic, conductivitatea termică a rocilor şi generarea radiogenă de căldură, a făcut posibilă stabilirea condiţiilor impuse de observaţiile geologice şi geofizice variaţiei temperaturii cu adîncimea, în crustă, pe teritoriul României. Pentru Carpaţii Orientali, Bazinul Panonic şi Bazinul Transilvaniei, s-au elaborat modele geotermice în regim dinamic, luîndu-se în consideraţie efectele termice ale procesului de subducţie alpină presupus în Carpaţi.

La elaborarea modelului geotermic al Carpaţilor Orientali s-a ţinut seama de sursele radiogene de căldură, de conducţia termică în regim staţionar şi de procesul de subducţie din timpul Miocenului. Drept concluzii asupra surselor anomaliei pozitive de flux geotermic, localizată la partea internă a Carpaţilor Orientali, au rezultat următoarele : (1) 20—25% din fluxul termic observat la suprafaţă ar putea proveni din căldura produsă de mecanica procesului de subducţie miocenă, căldură conservată încă în litosfera subcrustală datorită constantei de relaxare a perturbaţiei termice mai mare ca 25 mil. ani, pentru o crustă cu grosimea de 40 km ; (2) aproximativ 70% din fluxul la suprafaţă ar prezenta căldura generată radiogen, din care 2/3 în crustă ; (3) numai cîteva procente sînt probabil datorate conţinutului în căldură din părţile superficiale ale crustei, unde magmele calcoalcaline au staţionat în cursul ascensiunii lor şi unde echilibrul termic nu a fost încă atins.

Analiza datelor geotermice arată că anomalia de flux geotermic şi punerea în loc a lanţului neo-vulcanic din Carpaţii Orientali, par a fi o consecinţă a subducţiei litosferice alpine produsă în zona paleo-planului extern de consum al crustei din aria carpatică.

Atît variaţia temperaturii cu adîncimea, cît şi valorile fluxului geotermic, au demonstrat că Bazinului Panonic îi este asociată o anomalie geotermică pozitivă avînd o amplitudine de 96 mWm^{-2}.

Pornindu-se de la observaţia că în bazinele sedimentare post-tectonice, unde există o seamă de evidenţe clare geologice şi geofizice pentru un proces de extensie a litosferei (crusta şi litosfera subţiate ; prezenţa fracturilor de tensiune distribuite într-un anumit mod ; magmatism bazic recent la interiorul bazinului) fluxul geotermic la suprafaţă este anormal de ridicat, s-a studiat un model geotermic în regim dinamic. Modelul adoptat pentru Bazinul Panonic evaluează efectul termic al unei extensii orizontale continue, cu o viteză constantă (aproximativ 1 — 3% per mil. ani), care are drept rezultat o subţiere relativ uniformă a litosferei ; subţierea litosferei tinde să fie compensată la suprafaţă de fenomenul de subsidenţă şi sedimentare molasică, iar la baza litosferei subţierea este compensată de adăugarea de material bazaltic topit din astenosferă (aşa numitul fenomen de "atenuare astenosferică" sau diapirismul mantalei").

Pe baza modelului studiat se explică majorarea substanţială a gradientului geotermic în crustă, comparativ cu simpla conducţie în regim staţionar, formarea Bazinului Panonic fiind interpretată ca un rezultat al extensiei (sau mai corect „distensiei") începută în timpul Badenianului (faza „termală" a subsidenţei), cu o accelerare în timpul Sarmaţianului (faza „atenuării astenosferice") şi urmată de o reducere a ratei extensiei în timpul Pannonianului (faza răcirii „pasive" a litosferei).

Modelul propus pentru a explica, în termenii „atenuării astenosferei", particularitățile geotermice și istoria neogenă a Bazinului Panonic nu se potrivește caracteristicilor Bazinului Transilvaniei. Pentru Transilvania, media fluxului geotermic apare — surprinzător pentru un bazin post-tectonic — ca fiind de 40—45 mWm^{-2}. Curbele medii temperatură — adîncime calculate (geotermele), corespunzătoare mediei fluxului geotermic măsurat la suprafață în Bazinul Transilvaniei, respectiv aria panonică, indică existența la baza crustei a unor temperaturi de cca 600°C și respectiv 800—900°C.

Geoterma pentru Transilvania intersectează temperatura de tranziție în faza „solidus" a mantalei la o adîncime de aproximativ 80 km, în timp ce în Bazinul Panonic atinge aceeași temperatură la numai 40 km. În consecință, datele geotermice sugerează prezența unei litosfere mai reci și mai groase sub Bazinul Transilvaniei. Datorită subducției alpine din Carpați, aria transilvană a fost continuu obiectul compresiunii. După ce subducția a încetat, nu există nici un argument geologic sau geofizic în favoarea unui proces de extensie a litosferei, similar celui produs în Bazinul Panonic.

L'Annuaire de l'Institut de Géologie et de Géophysique a été publié le long des années sous les titres suivants:

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Anuarul Comitetului Geologic (Annuaire du Comité Géologique) t. XXIII-XXXIV (1950-1964)

Anuarul Comitetului de Stat al Geologiei (Annuaire du Comité d'Etat pour la Géologie) t. XXXV-XXXVII (1966-1969)

Anuarul Institutului Geologic (Annuaire de l'Institut Géologique) t. XXXVIII-XLII (1970-1974)

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