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MÉTAMORPHISME
METAMORPHISM
МЕТАМОРФИЗМ

UNTERSUCHUNG DER METAMORPHEN FAZIESGÜRTEL IN TRANSDANUBIEN[1]

VON

ENDRE BALÁZS[2]

Auf Grund der mehr als 1000 in den westlichen Bechengebieten Ungarns niedergebrachten Tiefbohrungen kann man heute schon mit ziemlicher Sicherheit die verschiedenen metamorphen Faziesgebiete angeben. Wenn im Aufbau der Gebirge Ungarns weniger als 10 % metamorphe Gesteine vorkommen, sind im Sockel der grossen Neogensenken mehr als 50 % zu finden. Eine ausführliche mineralogisch-petrographische Untersuchung der letzteren wird im Ungarischen Kohlenwasserstoff Forschungsinstitut durchgefürt. Das Ziel ist dabei natürlich in erster Linie eine auf praktischen Nutzen orientierte Strukturforschung. Dabei bildet — wie es sich schon oft bewiesen hat — die petrogenetische Analyse der metamorphen Gesteine eine nützliche Hilfe.

Im folgenden sollen nun die wichtigsten Forschungsergebnisse der vergangenen Jahre kurz zusammengefasst werden. Wegen des zusammenfassenden Charakters des Vortrages kann auf Teilergebnisse nicht eingegangen werden. Die Schlussfolgerungen, welche sich ans den mineralogischen und den speziellen optischen und Röntgenuntersuchungen von etwa 1000 Gesteinsproben ergaben, werden im weiteren skizziert. In den schwachmetamorphen Gesteinen führten wir auch eine genauere Bestimmung der Phyllosilikate durch. Diese betraf zum Teil die Bestimmung des Illit-Serizit Kristallisationsgrades sowie die Bestimmung des Gitterparameters bo von Muskovit. Dadurch konnte erstens eine Einordnung der Phyllitgesteine in die Anchizone — oder Grünschieferfazies erreicht werden, sund zweitens durch das Muskovit-Fengit verhältnis Schlussfolgerungen auf den während der Kristallisation herrschenden Druck gezogen werden.

Zwei Faktoren beeinflussen die Zuverlässigkeit der Methode in den untersuchten Gebieten nachteilig, einerseits ist die Streuung der Verteilung der Proben auserordentlich gross, so dass eine statistische Auswertung in vielen Fällen unmöglich wird, der andere Faktor ist im bedeutenden Karbonatgehalt der meisten Proben zu sehen.

In Transdanubien kann man das ursprüngliche, vor der Metamorphose befindliche Gestensmaterial auf Grund der Druck-temperaturver-

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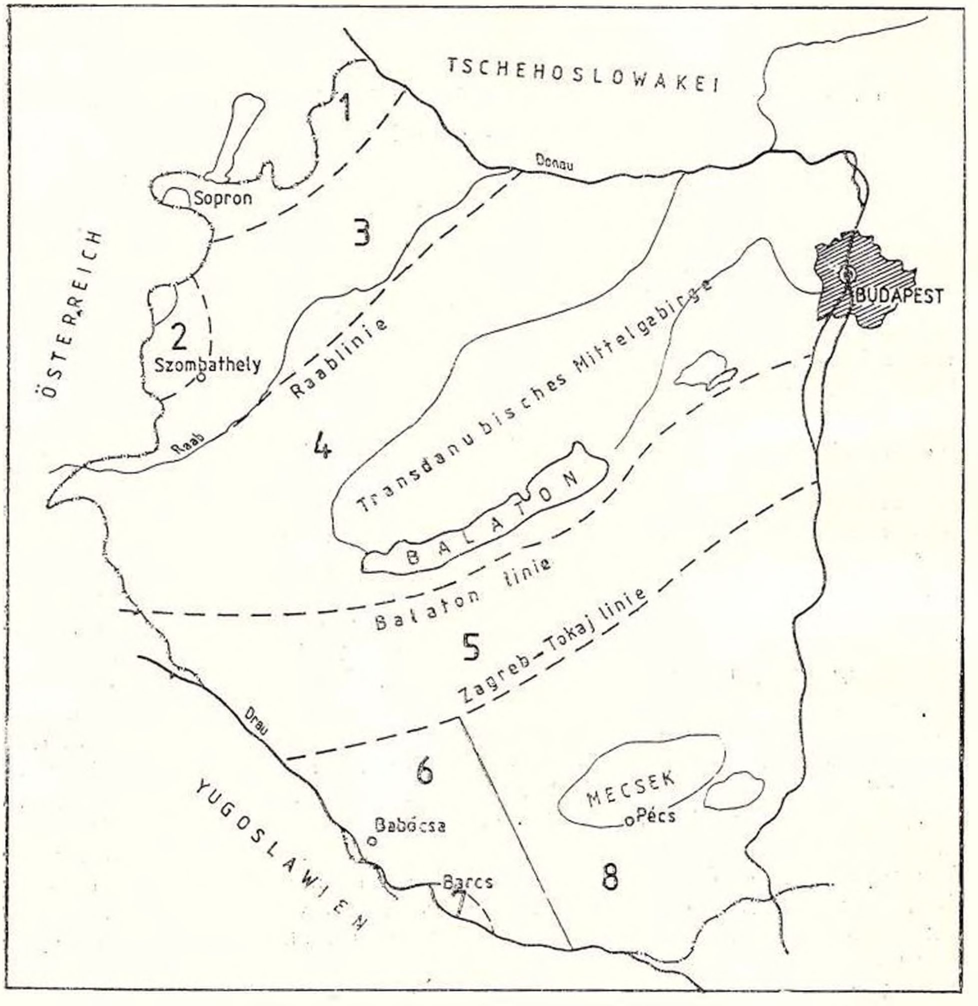
[1] Vorgetragen am 12. Kongress der Karpato-Balkanischen Geologischen Gesellschaft, 8--13 September, 1981, Bukarest, Rumänien.

[2] 1021 Budapest, II., Szajkó u. 9/A. fzs. 3.

hältnisse der Metamorphose, der Anzahl der metamorphen Wirkungen und seines progressiven oder retrograden Charakters in 8 voneinander abweichende Faziesgebiete einteilen (fig.) :

1. Soproner Gebirge und dessen SO-Ausläufer



Metamorphe Faziesgebiete in Transdanubien.

2. Köszeger Gebirge und Vashegy
3. Beckensockel der Kleinen ungarischen Tiefebene nordwestlich der Raablinie
4. Sockel des Transdanubischen Mittelgebirges
5. Mittleres transdanubisches Faziesgebiet (Bükk-Dinar Typ)
6. Draubecken (SW-Transdanubien)
7. Westlicher Rand des Draubechens (Barcs-Molve)
8. SO-Transdanubien, Mecsekgebirge und Mórágyer Scholle.

Im folgenden werden die genannten Faziesgebiete kurz charakterisiert. Dabei wird auf die von uns untersuchten Formationen des Beckensockels ein wenig ausführlicher eingegangen.

1. Soproner Gebirge und dessen SO-Ausläufer

Die Gesteine der Schieferinsel von Fertőrákos und des Soproner Gebirges sind als identisch mit den Gesteinen der östlichen Alpen betrachtet worden. Das Soproner Gebirge gehört zur „Grobgneisserie" und die Schieferinsel von Fertőrákos zur „Wechselserie" (P. Kisházy, 1976; L. Kósa, 1978).

Die Gesteine der ersten Generation von Metamorphiten des Soproner Gebirges sind Andalusit-Sillimanit-Biotit Schiefer und Zweiglimmerparagneise. Diese wurden nachträglich von Granit Intrusionen durchdrungen, Spater, wärend der Bildung neuerer Strukturen, wurde der Granit Gneis, in vewandelt und die älteren Metamorphite durch eine mannigfaltige retrograden Metamorphose überprägt. Entlang den Decken uberschiebungs flächen entstanden Leuchtenbergitschiefer, Leuchtenbergit-Disten-Quarzite usw. In dem in Richtung der Kleinen ungarischen Tiefebene plötzlich abfallenden Beckelsockel wurde in den Tiefbohrungen bei Pinnye schon in einer Tiefe von 1000 m zu dieser Serie gehörige Granat-Glimmerschiefer gefunden.

Die Gesteine der Schieferinsel von Fertőrákos sind stärker diaphtorisiert; Im Osten bei Mosonszolnok, Mosonszentjános und im süden bei Csapod wurden in Tiefborungen metamorphe Gesteine der Wechselserie gefunden. Die ursprünglichen Gesteine (Gneis, Amphibolit, Glimmerschiefer, Kalkschiefer) gehörten zur Almandin-amphibolit Fazies und wurden nachtraglich in Serizit-Chlorit-Karbonatschiefer umgewandelt (E. Balázs, 1975).

2. Das Köszeger Gebirge und der Vashegy

Die metamorphen Gesteine vom Wechseltyp des Csapoder Grabens treunen die penninische mesozoisch metamorphe Masse des Kőszeg-Rechnitz Gebirges und des Vashegy von den altpaläozoischen Metamorphiten des Beckensockels der Kleinen ungarischen Tiefebene. Die vom Obertrias bis zur Unterkreide reichenden metamorphierten mesozoischen Sedimente sind in ihren gegenwärtigen Zustand Kalkschiefer, metaanthrazitische Phyllite, Grünschiefer und Metakonglomerate. Diese Gesteinsgruppe setzt sich entsprechend unseres momentanen Konntnisse vom Csapoder Graben in Richtung Osten im Sockel nicht fort (T. Kotsis, 1976; P. Kisházy, 1978; E. Balázs, 1979).

3. Beckensockel der Kleinen ungarischen Tiefebene nordwestlich der Raablinie

In Gebiet zwischen dem Ostrand der Alpen und der Raab-Linie wurden im Beckensockel dem Grazer Paläozoikum analoge, bzw. sehr wenig metamorphe Gesteine angebohrt.

Die Schichtenfolge ist lithostratigraphisch in 5 Formationen einteilbar (E. Balázs, 1979):

Szentgotthárder Phyllitformation
Nemeskoltaer Sandsteinformation
Sótonyer Metabasitformation
Mihályer Phyllitformation
Büker Dolomitformation

Die zum Grazer Altpaläozoikum auch teritoriel am nächsten stehenden metamorphen Gesteine erbrachten die Bohrungen bei Szentgotthárd. Auf Grund des Kristallisationsgrades von Serizit-Illit befinden sich die aus Tonen und Tonmergeln entstandenen Gesteine an der Anchi-Epizonengrenze im niedrigsten Druckbereich der Grünschiefer fazies. Die Phyllite und dolomitischen Kalkphyllite sind örtlich stark chlorithaltig. Es kann demnach eine mit der Sedimentation gleichzeitig stattfindende basisch-intermediäre vukanische Einwirkung angenommen werden.

Der in der Nemeskoltaer Formation vorherrschende Gesteinstyp ist ein anchimetamorpher Sandsteinschiefer. Die ursprüngliche Struktur des Karbonatarmen Gesteins ist gut erhalten, die Textur ist schwach geschiefert. Die Oberfläche der klastischen Körnchen ist charakteristisch aufgelöst und grätenartig verwachsen. Ein Teil der klastischen Feldspat Körnchen stammt aus Gneis, der andere Teil ist pyroklastisch. Daneben kommt auch etwas durch Metamorphose entstandenes Albit vor.

Die Sótonyer Metabasitformation besteht aus submarinen Vulcaniten. Die gesteine entstanden aus Diabasen und Diabastuffen, die im Verlaufe der Anchimetamorphose zu Metadiabasen und epidotischen Chloritschiefern umgewandelt wurden.

Die Mihályer Phyllitformation ist die flächenmässig verbreiteste und auch lithologisch vielfältigste Einheit der Metamorphitschichtenfolge nordwestlich der Raab-Linie. Die Ursache der Vielfältigkeit ist grundsätzlich im ursprüglichen, ausserordentlich vielfältigen Sedimentgesteinsmaterial zu suchen. Die wichtigsten, in der Anchizone bzw. in der kleinsten Druckzone der Grünschieferfazies, entstandenen Gesteine sind: Sandsteinschiefer, Aleurolitschiefer, Serizitquarzit, Kieselschiefer, Serizitschiefer, Phyllit, Chloritschiefer, Kalkschiefer, Dolomitschiefer, Dolomit — bzw. an ein, zwei Orten metadiabas.

Der Büker Dolomit wurde als eine selbständige Formation betrachtet, obwohl er aller Wahrscheinlichkeit nach mit den jüngsten Gliedern der Mihályer Phyllitformation korreliert werden kann. Der Büker Dolomit entspricht litologisch dem Kirchfidischen Dolomit, der Fossilien aus dem Devon enthält.

4. Der Sockel des Transdanubischen Mittelgebirges

Der Sockel der nicht-metamorphen Oberkarbon-Permischen und Mesozoischen Sedimenschichten masse besteht aus schwach (anchimetamorphen Gesteinen deren kristallisierung wurde durch die varistische Orogenese verursacht wurde. Das Alter dieser Gesteine ist auf Grund von Fossilien und des Kristallisationsgrades des weissen Glimmers eindeutig altpaläozoisch. Zu dieser Schichtenfolge gehören der

Schiefergebirgszug des Balatonhochlandes, die Gesteine, welche durch Bohrungen an der Balatonlinie gefunden wurde, und der sockel des vom NW-Fusse des Bakonygebirges bis zur Raablinie sich erstreckende Beckengebiet. Letzterer wurde mit dem Namen Vaszarer Serizitschieferformation bezeichnet. Neben dem vorherrschenden Serizitschiefer erscheinen andere gesteinstypen wie Phyllit, Chloritschiefer, Sandsteinschiefer und Aleurolitschiefer nur untergeordnet.

In der sehr schwach metamorphen Schichtenfolge des Balatonhochlandes kann man mit Bestimmtkeit drei Faziesgebiete voneinander unterscheiden, ein quarzphyllitisches, ein quarzporphirisches und eines mit kalkigem Tonschiefer (I. Bubies, 1972; Gy. Majoros, 1977; E. Balázs, 1975, 1978).

5. Das mittlere Transdanubische (Bükk-Dinarischer Typ) Faziesgebiet

In dem sich von der Balatonlinie nach Süden erstreckenden Gebiet überwiegen Jungpaläozoisch-triasische Sedimente die in ihrer Entwicklung vom Mittelgebirgs-paläozoikum und Mesozoikum bedeutend abweichen. In dieser Zone Mitteltransdanubiens erfuhr ein Teil der durch Tiefbohrungen erreichten Perm-Triasgesteine eine charakteristische, sehr geringe Metamorphose. Diese Durchkristallisation ist in reinen Karbonatgesteinen weniger wahrnehmbar, in den mehr tonigen und tuffigen Sedimenten dagegen völlig eindeutig. Eine Durchkristallisation in der Zeolithfazies konnte durch Tiefbohrungen in der Umgebung vom Liszó nachgewiesen werden. Auf Grund des Serizit-Illit-Kristallisationsgrades befinden sich da die am wenigsten metamorphen Gesteine aus Transdanubien.

6. Das Draubecken (SW-Transdanubien)

Im Draubecken erschlossen sehr viele Tiefbohrungen den in Hinsicht auf seine Durchkristallisation im Wesentlichen einheitlichen, aus metamorphen Gesteinen der Almandin-Amphibolithfazies bestehenden Beckensockel. Hinsichtlich ihres Alters sind diese Gesteine wahrscheinlich die ältesten in Transdanubien. Ihre Streichrichtung verbindet sie mit den Dinarischen Strukturen. Die Diaphtorese ist nicht regional verbiertet, sondern nur auf tektonische Gürtel von kleinerer Bedeutung beschränkt. Diese Diaphtorese wurde im allgemeinen schon im Anfangsstadium beendet. Durch Isograden begrenzte metamorphe zonen vom Barrowtyp konnte nur im südöstlichen Teil erkannt werden. Ein vergleich mit den präkambrischen Gesteinen Südosttransdanubiens ergibt als sehr augenscheinlichen Unterschied, dass hier die an die Thermalachse gebundene selektive Umbildung zu Granit und die Migmatit-bildung nicht eintraten. Die Gesteimsfolge besteht aus biotitischen Muskovitglimmerschiefern und Plagioklasparagneisen der Almandin-Amphibolithfazies Lokal erscheinen kleinere Amphibolitheinlagerungen.

7. Der westlicher Rand des Draubeckens (Barcs-Molve)

Im Draubechen ist in verhältnismässig geringer Riefe eine schwachmetamorphe Schichtenfolge zu finden, welche tektonisch auf der westlichen Seite eines Grundgebirgsrückens liegt der aus metamorphen Gesteinen der Almandin-Amphibolithfazies besteht. In der bisher bekannten etwa 300 m mächtigen Schicktenfolge kann man oben eine dolomitisch-diabasische, in der Mitte ein phyllitische und im unteren Teil eine Sandsteinschieferfolge abgrenzen. Die Gesteine dieser drei Folgen können auf Grund ihrer Mineralassoziation in die schwach und sehr schwach metamorphen Gesteine (Anchizone-Grünschieferfazies) eingeordnet werden. Nach den Muskovitkristallisationsdaten (Kübler-Weberinder) kann eine Zugehörigkeit zur Anchizone als sicher bestätigt werden.

8. SO-Transdanubien, das Mecsekgebirge und die Mórágyer Scholle

In SO-Transdanubien ist im Sockel der aus metamorphen Gesteine der Amphibolithfazies besteht eine Granitisierung festzusellen welche entlang einer etwa SW-NO gerichteten Thermalachse verläuft, Hier sind die verschiedensten Gesteintypen des Migmatitbildungsprozesses zu finden, unter anderen geschichtate Migmatite, Agmatite, Diktionite, Diatexite, porphyroblastische, nebulithische und skialitische Granite (B . I a u t s h y, 1975).

Im Südteil des Migmatitgranits liegen magmatisch hochgepresste Serpentinitdiapire (T . S z e d e r k é n y i , 1975).

Schliesslich sollen, obwohl nur in indirekter Verbindung mit den metamorphen Gesteinen stehend, als neues Forschungsergebnis die Oberkreide-Paläogenen Tonalithe, erwähnt werden, welche entlang der Balatonlinie erbohrt wurden und die Mittelgebirgs — und Mitteltransdanubischen Faziesgürtel voneinander trennendon Alterbestimmung an frischen Biotiten aus feinkörnigen Quarzdioriten weisen auf Unter bis Mittlel Eozäns hin.

In der durch junge tektonische Bewegungen aufs änsserste komplizierten Mosaikstruktur Transdanubiens gelang es folgenden Ablauf der Metamorphosen nachzuweisen. Eine prävaristische/vermutlich präkambrische/ Metamorphose in der regionalen Almandin-Amphibolithfazies; lokal eine sich stüfenweise verstärkende Metamorphose und Ultrametamorphose vom Barrow Type; eine varistische Metamorphose in der Anchizone — bzw. geringen Grünschieferfazies und eine alpidische Metamorphose in der Grünschieferfazies.

THE CRYSTALLINE BASEMENT OF THE DANUBIAN UNITS IN THE CENTRAL SOUTH CARPATHIANS : CONSTITUTION AND METAMORPHIC HISTORY[1]

BY

TUDOR BERZA[2], ANTONETA SEGHEDI[2]

Introduction

This paper represents a possible interpretation of the data concerning the basement in the central part of the South Carpathians' Danubian Units, between the Jiu and Strei Rivers to the east and the Timiş River to the west (Muntele Mic, Petreanu, Ţarcu, Retezat, Culmea Cernei — Mehedinţi and Vîlcan Mountains). In the previously unitary Danubian Autochthon (Codarcea, 1940), various structural models have successively been proposed by Stănoiu (1973), Kräutner et al. (1978), Năstăseanu (1980) and Morariu and Morariu (1982). Strengthening Stănoiu's picture with new data, Berza et al. (1983) propose two main Alpine divisions : the Upper Danubian Group of Units and the Lower Danubian Group of Units, both of them consisting of several pre-Alpine and Alpine units.

A cover and a basement were distinguished, by tradition, in the Danubian Units. The ages ascribed to these distinct structural levels suffered large fluctuations in time, but in recent years a Mesozoic-Upper Paleozoic age for the cover and a Middle-Lower Paleozoic and Upper Precambrian age for the basement were generally accepted. Since the importance of the alpine metamorphism in the Danubian of the South Carpathians was reconsidered (Gherasi et al., 1973; Dimitrescu, 1976) and since a Liassic flora was found in the upper part of a sequence previously ascribed to the basement (Stănoiu, 1982), it was obvious that the basement/cover distinction is not so sharp as it was claimed.

In the south-eastern part of the Danubian area, a Precambrian infrastructure and a Paleozoic suprastructure constitute the crystalline basement (Savu, 1970; Berza, 1978). The close association of the Precambrian and Paleozoic sequences in Variscan and Alpine structures,

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[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, September 8–13, 1981, Bucharest, Romania.

[2] Institute of Geology and Geophysics, str. Caransebeş 1, 78344 Bucharest, Romania.

and the extensive retrogressions of the Precambrian rocks, frequently make this distinction difficult, the opinions concerning a given outcrop or area being often conflicting. However, as a substantial progress has realized in the last decade in dating the Paleozoic formations and in understanding the metamorphic history of the Precambrian rocks, the situation is rapidly improving.

Paleozoic Low-grade Rocks

Recent reviews of the Paleozoic formations from the Danubian Units were presented by Năstăseanu (1975; Năstăseanu et al., 1978 and in Kräutner et al., 1981) and Stănoiu (1976; in Kräutner et al., 1981). According to these authors, the Paleozoic formations were alloted in the Upper Danubian Units to the Drencova-Rîul Alb Group and in the Lower Danubian Units to the Tulişa Group.

The Drencova-Rîul Alb Group comprises a mainly volcanic sequence (Ordovician-Silurian) and a sedimentary sequence (Upper Devonian-Lower Carboniferous) (Kräutner et al., 1981). In the absence of specific studies, the low-grade metamorphism (anchizone to chlorite zone) of the Drencova-Rîul Alb Group rocks may be connected either with the Variscan, or with the Alpine deformations.

West of the Jiu Gorges, the Tulişa Group is represented by the psamitic Ordovician (+ Silurian?) Valea Izvorului Formation, by the psephitic and pelitic Devonian Vidra, Tusu and Sgura Formations and by the carbonatic and pelitic Lower Carboniferous or Mesozoic Oslea Formation (Kräutner et al., 1981). The Ordovician and Devonian rocks are involved in and are intensely deformed by the Variscan overthrust of the Drăgşan Group (Retezat-Parîng Unit) on the Lainici-Păiuş Group (Vîlcan-Pilugu Unit). The Oslea Formation overlies with angular unconformity the older formations of the Tulişa Group, or directly the Precambrian Drăgşan or Lainici-Păiuş rocks.

The data regarding the metamorphism of the Tulişa rocks are still scarce. As the low-grade index minerals quoted here (chlorite, chloritoid, fine grained white mica) are also found in the overlying Mesozoic rocks, it has not been proved so far that a Variscan metamorphism has occurred in the Lower Danubian Units.

Precambrian Polymetamorphic and Granitoid Rocks

In both groups of the Danubian Units, the low-grade Ordovician formations overlie older polymetamorphic and granitoid rocks. These rocks crop out in restricted zones, due to the complex structure of the Danubian area and to the overlying younger deposits. Therefore, a lot of local names were given by various authors to the polymetamorphic sequences and the correlation of the main lithologic units is still controversial. Pavelescu (1959) proposed the grouping of these metamorphic rocks into a few lithologic units: the Drăgşan, Lainici-Păiuş and Rîuşorul Series. We share this opinion, but as some sequences are difficult to be alloted to these units, and as the correlation between sequences belonging to the two Groups of the Danubian Units is evidently hypothetical, we propose a slightly different scheme. In the basement of the Upper Danubian Units of the Țarcu, Muntele Mic and the North Petreanu Mts, Gherasi et al. (1968, 1973, 1974) have developed a detailed lithostratigraphic terminology for a pile of low-grade looking rocks, supposed to represent Upper Precambrian and Lower Cambrian deposits, metamorphosed during the Baikalian and Caledonian orogenies. Kräutner et al. (1978) have remarked the retrogressed nature of some of these rocks, Solomon and Pop (1973) have identified staurolite and Savu et al. (1983 a) have reported the presence of kyanite. The main lithologies involved in this sequence are amphibolites, augen gneisses and mica gneisses, with the same mineralogical, structural and geochemical features (see Savu et al., 1983 a, versus Savu et al., 1983 b) as the corresponding rocks of the Drăgșan Group in the Lower Danubian Units. In this case, it would be possible to ascribe the polymetamorphic basement rocks of the Upper Danubian Units in the Muntele Mic, Țarcu, North Petreanu and North Retezat Mts to the Drăgșan Group. This has already been done by Pavelescu (1959, 1963), and Gherasi and Dimitrescu (1968) have used the name Zeicani-Drăgșan Series ; however, since Berza et al. (1983) confirmed the existence of two main groups of tectonic units in the Danubian area, and since in the Țarcu-Muntele Mic-Petreanu region the Zeicani Series is a well known name, we maintain this usage for the Upper Danubian Units. It is highly probable that in the future one of these names (Zeicani or Drăgșan) will prevail.

Various sequences of the Zeicani Group represent the basement of several Upper Danubian Units ; therefore, a reconstitution of its lithostratigraphy is impossible for the moment. If the equivalence with the Drăgșan Group is real, it is probable that the same lithologic units as in the Drăgșan Group are present within the Zeicani Group, too. Amphibolites, leptynites and mica gneisses are found in the Măru Unit, augen gneisses and amphibolites are dominant in the Muntele Mic Unit and amphibolites are the main constituent of the Poiana Mărului Unit.

The metamorphic history of the Zeicani Group is very complex and not clarified yet. An almandine amphibolite facies phase (probably the staurolite-almandine subfacies, according to the data of Solomon, Pop, 1973 and Savu et al., 1983 a) was followed by the intrusion of the Muntele Mic, Șucu and Rîul Șes granitoid plutons, inducing thermal and metasomatic action (Gherasi, Savu, 1969 ; Gherasi et al., 1974) in the surrounding rocks. Extensive retrogressive events may be connected with the postulated Variscan low-grade metamorphism of the Paleozoic Drencova-Rîul Alb Group and are surely linked to the Alpine overthrusts. These diaphtoreses are extremely evident in the proximity of the Getic and Upper Danubian overthrust planes, and next to the tectonic planes between various Upper Danubian Units, conferring a general greenschist aspect to the Zeicani rocks.

There are no accurate data yet concerning the age of the Zeicani Group. The palynological investigations carried out by Visarion (see Savu et al., 1973 ; Gherasi et al., 1973) have provided only large spectrum forms, from the Precambrian-Lower Cambrian interval. K/Ar ages between 94 and 456 m.y. found by Soroiu et al. (1970, 1972), Mînzatu et al. (1975) and Grünenfelder et al. (1983) show only

Alpine resetting of older ages. Based on speculative correlations to other metamorphic sequences of the Carpathians (K r ä u t n e r et al., 1981), we think that the Zeicani Group belongs to a non specified interval of the Upper Precambrian.

The pre-Ordovician basement of the Lower Danubian Units has a more complex constitution than that of the Upper Danubian Units. Two areas may be distinguished within the Lower Danubian Units, separated by a NE—SW directed and SE dipping tectonic plane — the Rîul Mare Fault. East of this plane, the Paleozoic and Mesozoic formations overlie two types of metamorphic basements : Drăgșan Group or Lainici-Păiuș Group, the former being overthrust (in the Retezat, Culmea Cernei and Northern Vîlcan Mts) on the latter, during a post-Devonian and pre-Jurassic (possibly even pre-Namurian) phase (B e r z a et al., 1983).

The Drăgșan Group rocks are found in two distinct areas : in the central Retezat Mts and in the Culmea Cernei-Northern Vîlcan Mts. Within this group (the Drăgșan Series of P a v e l e s c u , 1953 a) three formations — the Făgețel Gneisses, the Amphibolite Formation and the Dobrota Mica Gneisses — may be distinguished.

The Făgețel Gneisses outcrop in the Retezat Mts, and consist mainly of augen gneisses and a few amphibole gneisses. Due to extensive retrogressions, these rocks have a general aspect of low-grade rocks, but a careful examination reveals their polymetamorphic nature. The Amphibolite Formation is exposed in the Retezat Mts (Rîul Bărbat Formation) and in the Culmea Cernei-Northern Vîlcan Mts (Straja Formation). In both areas, it is represented by amphibolites, amphibole gneisses, biotite gneisses, but frequently retrograde events gave these rocks a greenschist appearence (P a v e l e s c u , 1953 a ; B e r z a , 1975). The Dobrota Mica Gneisses overlie in Culmea Cernei the Straja Formation and consist of mica gneisses, staurolite micaschists (I a n c u , 1974), biotite gneisses and amphibolites.

Despite the influence of the Culmea Cernei and Retezat granitoid plutons and of several retrogressive events, there are still places where critical minerals from the oldest metamorphism of the Drăgșan Group are preserved. I a n c u (1974) has reported staurolite and B e r z a , S e g h e d i (1975 a) have identified kyanite and staurolite ; these minerals, and the frequent andesine + hornblende + almandine association, point to the staurolite-almandine subfacies of the almandine-amphibolite facies conditions. From place to place, the subsequent retrogressions may be ascribed to a possible Variscan regional metamorphism of the Tulișa Group, to the Variscan overthrusting of the Drăgșan Group on the Lainici-Păiuș Group, to Alpine events, or to a combination of these processes.

The volcanic activity (P a v e l e s c u , P a v e l e s c u , 1962, 1964 ; S a v u et al., 1976, 1983 b ; B e r z a , 1978 ; S c h u s t e r , 1980 have considered the amphibolites as metamorphosed basic tuffs, volcanics and intrusions), the sedimentation, metamorphism and related plutonism of the Drăgșan Group are surely pre-Ordovician, as it supports the Valea Izvorului Formation. K-Ar ages between 97 and 325 m.y. (G r ü n e n f e l d e r et al., 1983) show Alpine resetting, but general correlations have enabled K r ä u t n e r et al. (1981) to state that the Drăgşan Group may correspond to the middle part of the Upper Precambrian sequence known in the Carpathians.

In the region considered in this paper, the Lainici-Păiuş Group rocks are also found in two distinct areas, in the Southern Vîlcan and Southern Retezat Mts. Separated as Lainici-Păiuş Series by M a n o l e s c u (1937), these rocks have a prominent individuality among the metamorphic rocks of the South Carpathians. In both areas, the Lainici-Păiuş Group consists of two subdivisions recognized by B e r z a (1978) and S c h u s t e r (1980): the Carbonate-Graphitic Formation and the Quartzite-Biotite Gneisses. Although these main lithologic units are easily mapped in the field, the superposition order is not known, due to extensive small scale folding and new schistosities. Another feature of the Lainici-Păiuş Group is the extensive penetration by granitic material, as various migmatites, countless small bodies and several large plutons.

The Carbonate-Graphitic Formation consists of crystalline limestones and dolomites, sillimanite-andalusite-cordierite metapelites, amphibolites and calcsilicate gneisses. The Quartzitic and Biotite Gneiss Formation contains various quartzites, in close alternation with biotite ± garnet gneisses. The successive metamorphic events affecting the Lainici-Păiuş Group have been interpreted by S a v u (1970) as low-pressure „Danubian" metamorphism, „static autoretrogression" induced by late fluids from granitoids, Variscan "alloretrogression" and Alpine diaphtoresis. Contact metamorphism was noticed (M r a z e c , 1898 ; I o n e s c u-B u j o r, 1911 ; M a n o l e s c u , 1937 ; S t a n , 1977) and a dynamic retrogression was observed near the tectonic contact with the Drăgşan Group (B e r z a , 1975 ; I a n c u , 1977 ; S a v u et al., 1983 b). This complex evolution is not well understood yet, but the polymetamorphic nature of the Lainici-Păiuş rocks is obvious. In the absence of relict Barrovian minerals, it is conceivable that the extensive arteritic migmatization (S a v u et al., 1972 ; S t a n , 1977 ; B e r z a , 1978 ; S c h u s t e r, 1980), the intruded Buta, Buseşti, Tismana, Frumosu, Bîlta, Suseni and Şuşiţa plutons and countless smaller granitoid bodies induced, at moderate pressures (permitting the coexistence of almandine, andalusite, sillimanite and cordierite) a heat flow sufficient to allow the regional occurrence of sillimanite. Several subsequent retrogressive events, which created more or less penetrative schistosities, brought frequently this sequence to low-grade appearance.

The low-pressure metamorphism and granitic intrusions are pre-Ordovician, since the fossiliferous Valea Izvorului Formation (S t ă n o i u , 1972) unconformably overlies both the Lainici-Păiuş rocks and granitoid bodies. K-Ar ages of metamorphic and granitoid rocks generally show Alpine resetting, but values as high as 665 m.y. were still found, and U-Pb ages on zircons from some granites are 610 ± 35 m.y. (G r ü n e n f e l d e r et al., 1983). These data clearly point to the Precambrian age of the sedimentation, metamorphism and granitization of the Lainici-Păiuş Group. In a subsequent, but still pre-Ordovician interval, the Lainici-Păiuş rocks and the granitoid bodies have been cut by countless hypabyssal dykes of porphyric microdiorites, microgranodiorites or micro-granites, showing only the schistosities and mineral alterations induced by Paleozoic or Mesozoic events (B e r z a , S e g h e d i , 1975 b).

The Rîuşoru Formation (Rîuşoru Series — P a v e l e s c u , 1953 b) constitutes the Nucşoara Unit and is mainly represented by biotite-garnet gneisses or quartzites. Detailed studies are still lacking, but it seems (D i m i t r e s c u , personal communication, 1980) to include two subdivisions : a carbonatic and a graphitic member (with some serpentinites) and a biotite-garnet quartzitic member. This constitution reminds of the Lainici-Păiuş sequence, but the scarcity of granitic intrusions differentiate it from the mentioned group. The Rîuşoru Formation has also a complex metamorphic history. The garnets indicate at least an upper greenschist facies metamorphism, but the presence of amphibolites points to the amphibolite facies. Subsequent retrograde events produced penetrative new schistosities and created a phyllitic appearence. This feature is extremely obvious in the southern part of the area, where the Rîuşoru rocks have been considered as a distinct formation— "the Rîul Mare biotite phyllites" — (G h e r a s i , D i m i t r e s c u , 1968 ; G h e r a s i et al., 1973, 1974). The age of the Rîuşoru Formation is uncertain, but by the mentioned correlation it is possible to infer an Upper Precambrian age.

West of the Rîul Mare Fault, in the Petreanu Mts, several polymetamorphic formations and gneissic bodies are exposed ; tectonic contacts between some of these lithologic entities were proposed by C o d a r c e a and G h e r a s i (1945) and B e r z a et al. (1983). From bottom to top, the superposition order is the following : Rof Formation in the Rof Unit, Furcătura Gneiss and Nisipoasa Formation in the Furcătura Unit and Petreanu Gneiss with Bodu Formation in the Petreanu Unit.

The Rof Formation consists of almandine micaschists, amphibolites, leptynites, biotite gneisses and quartzites (G h e r a s i , D i m i t r e s c u, 1968). The polymetamorphic nature of these rocks is striking, a staurolite zone metamorphism (G h e r a s i et al., 1974) being followed by an event in which hornblende was replaced by biotite; a later, distinct retrogression led to the development of low-grade minerals.

The Furcătura Gneiss is a microcline-biotite gneiss, cut by subsequent aplites (G h e r a s i , D i m i t r e s c u , 1968). Mineralogic and fabric aspects led D i m i t r e s c u (personal communication, 1980) to interpret this gneiss as a metamorphosed granitic body.

The Nisipoasa Formation (D i m i t r e s c u , personal communication, 1981), formerly described as Rîuşorul Series (G h e r a s i , D i m i t r e s c u , 1970, 1978), as Măgura Series (G h e r a s i et al., 1974), or as Biotite-Quartz Schists Formation (K r ä u t n e r , 1980), is a few hundred meters thick strip, sandwiched between the Furcătura and the Petreanu Gneisses. These monotonous biotite-quartz ($\pm$ garnet) schists resemble blastomylonites.

The Petreanu Gneiss consists of augen or banded gneisses with aplitic, amphibolitic, biotite gneissic or quartzitic layers (G h e r a s i , D i m i t r e s c u , 1968), and seems to have a lithostratigraphic control (G h e r a s i, D i m i t r e s c u , 1970).

The Bodu Formation (K r ä u t n e r , 1980) comprises quartzites, almandine micaschists, amphibolites, serpentinites and crystalline limestones, with frequent stromatic or ophtalmitic migmatites. This formation is intruded by the Vîrful Pietrii Granite (G h e r a s i et al., 1968).

The ages of these lithologic entities within the Petreanu Mts are not well known. The Bodu Formation and the Vîrful Pietrii Granite are overlain by the Devonian Vidra Formation (G h e r a s i et al., 1975), and K-Ar ages as high as 667 m.y. were found for biotite gneisses interlayered in the Petreanu Gneiss (G r ü n e n f e l d e r et al., 1983). The Bodu Formation reminds very well of the Lainici-Păiuş Group, and the Rof Formation resembles the Drăgşan Group. It is highly probable that these lithologic units of the Petreanu Mts are also Upper Precambrian.

From the above statements, we can say that in the lower polymetamorphic part of the Danubian Units basement, the most extended sequence is represented by the Drăgşan = Zeicani Groups.The Lainici-Păiuş Group is a different sequence, and several less extended suites in the Petreanu Mts may either correspond to parts of the above mentioned groups, or represent other levels of the Upper Precambrian.

SELECTIVE REFERENCES

G h e r a s i N., Z i m m e r m a n n V., Z i m m e r m a n n P. (1968) Structura şi petrografia şisturilor cristaline din partea de nord a muntelui Ţarcu (Banatul de est). *D. S. Inst. Geol.* LIV, 1, 55, Bucureşti.

— D i m i t r e s c u R. (1970) Anticlinalul Rof şi rolul lui în structura părţii de nord a Munţilor Retezat şi Petreanu. *Anal. St. Univ. "Al. I. Cuza" Iaşi, serie nouă,* II, b *(Geol.-Geogr.)*XVI, 55, Iaşi.

— V i s a r i o n A., Z i m m e r m a n n P. (1973) Considérations sur l'âge de certains schistes cristallins et dépôts sédimentaires de l'autochtone danubien situés au N des Monts Godeanu (Carpathes Méridionales). *Stud. cerc. geol. geof. geogr., ser. Geol.*, 18/2, 303—310, Bucureşti.

— Z i m m e r m a n n P., Z i m m e r m a n n V. (1974) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

G r ü n e n f e l d e r M., P o p e s c u Gh., S o r o i u M., A r s e n e s c u V., B e r z a T. (1984) K-Ar dating of metamorphic formations and associated igneous bodies in the Central South Carpathians. *Carp. Balc. Asoc. Geol. Congr.* XII, 1981, *An. Inst. geol. geofiz.* LXI, Bucureşti.

K r ä u t n e r H. (1980) Lithostratigraphic correlation of the Precambrian of the Romanian Carpathians. *An. Inst. geol. geofiz.*, LVII, 229, Bucureşti.

— M a i e r O., S t a n N., B e r z a T., M ă r u n ţ i u M., H â r t o p a n u I., H a n n H., G h e u c a I. (1978) Report , the archives of the Institute of Geology and Geophysics, Bucharest.

— N ă s t ă s e a n u S., B e r z a T., S t ă n o i u I., I a n c u V. (1981) Metamorphosed Paleozoic in the South Carpathians and Its Relations with the Pre-Paleozoic Basement. *Ghidul excursiei A1, Asoc. Geol. Carp. Balc. Congr.* XII, 1981, Bucureşti.

M î n z a t u S., L e m n e M., V î j d e a E., T ă n ă s e s c u A., I o n c i c ă M., T i e p a c I. (1975) Date geocronologice obţinute pentru formaţiuni cristalofiliene şi masive eruptive din România. *D. S. Inst. geol. geofiz.*, LX, 1/5,85, Bucureşti.

Năstăseanu S., Morariu S., Visarion A. (1978) Metamorphosed Paleozoic formations from the central part of the Southern Carpathians. *Rev. roum. géol. géophys. géogr., sér. Géol.*, 22.53—60, București.

Pavelescu L. (1953 a) Studiul geologic și petrografic al regiunii centrale și de SE a munților Retezatului. *An. Com. Geol.* XXV, 119—210, București.

— (1953 b) Cercetări geologice in Munții Retezat. *D. S. Com. Geol.* XXXVII, 105—144, București.

— (1959) Geologia Carpaților Meridionali. *Anal. Rom. Sov. secția St. Geol.* I—II, 5—25, București.

Savu H. (1970) Structura plutonului granitoid de Sușița și relațiile sale cu formațiunile autohtonului danubian (Carpații Meridionali). *D.S. Inst. Geol.* LVI, 5, 123—153, București.

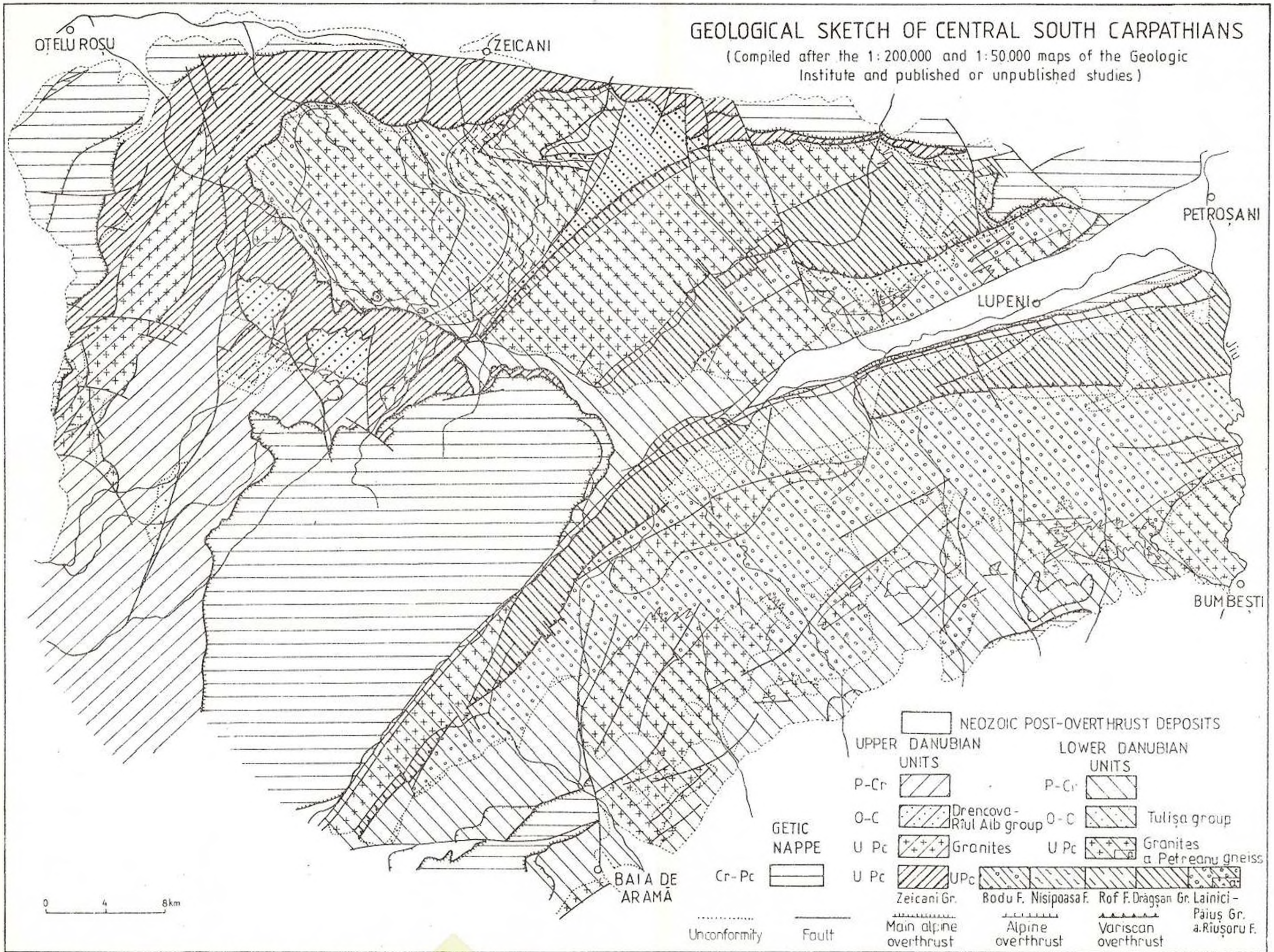
Schuster A. K. (1980) Geologische und petrographische Untersuchungen im Danubikum der Südkarpaten, Paring Gebitge, Rumänien. Dissertation. Technische Universität Clausthal.

Soroiu M., Popescu Gh., Gherasi N., Arsenescu V., Zimmermann P. (1970) K-Ar dating by Neutron Activization of some igneous and metamorphic rocks from the southern branch of the Romanian Carpathians. *Eclog. Geol. Helv.* 63/1. Basel.

Stănoiu I. (1973) Zona Mehedinți-Retezat, o unitate paleogeografică și tectonică distinctă a Carpaților Meridionali. *D. S. Inst. Geol.* LIX, 5, 127—171, București.

— (1976) Contribuții la stratigrafia formațiunilor paleozoice din versantul nordic al munților Vîlcan (Carpații Meridionali), cu implicații asupra părții externe a autohtonului danubian. *D. S. Inst. geol. geofiz.*, LXII/5, 219—256, București.

— (1982) Prezența fosilelor mezozoice și metaantracitului in cadrul șisturilor cristaline ale ,,grupului Tulișa" din versantul nordic al Munților Vîlcan și Paring. *D. S. Inst. geol. geofiz.*, LXVII/3, București.



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MIGMATITE BELTS IN THE BASEMENT COMPLEX OF THE REGION BETWEEN DANUBE AND TISZA[1]

BY

B. CSEREPES[2]

In my paper I should like to give a picture of the metamorphic rocks known from the hydrocarbon exploration wells drilled in the N and central parts of the region between the rivers Danube and Tisza. My aim was to study the NE-continuity of granitoids of the Mecsek Mountain and to trace the migmatite types described by B. J a n t s k y in his Mecsek monography (Fig. 1).

I had 556 rock samples at my disposal coming from 226 wells in 12 hydrocarbon exploration regions. Unfortunately, the oil wells are not continuously core drilled, only one or two samples are taken from a well (sometimes many hundred meters from each other). Because of this, it was impossible to construct a petrographical spot-map. Based on identical petrogenesis, 2 minor area units could have been distinguished, a N, and a S one (Fig. 2).

From the minerals of the granitogene rocks, the feldspars react with the highest sensitivity at the temperature and pressure changes taking place in the earth's crust, by altering the physical parameters. This fact can be used for the description of the circumstances of the genesis. The determination of the feldspar structure was made by optical measurements on the universal stage of Fedorow and by studying the accurate values "d" calculated from X-ray diffractograms. I have made 337 feldspar Fedorow measurements and 122 X-ray evaluations.

On Fedorow's stage the angle between the crystallographical axes of the feldspars and the cleavage or twin plane elucidating the twin type and the An % for the plagioclases, and triclinicity for K feldspar as well as axial angle, the 2V, referring among others, to the genesis temperature, can be determined. At the X-ray examinations, the angular difference between the peaks of the individual reflexion pairs is characteristic of the triclinicity grade. The factual values of triclinicity were calculated by the L a v e s (1954) method. As for the plagioclases, the change of

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[2] Hungary.

the Si/Al order grade and of the An percent is best reflected by the 20° value difference between the peak pairs 131 and 1$\bar{3}$1 as well as $\bar{2}$41 and $\bar{2}\bar{4}$1. To determine the An percent, the methods of B a m b a u e r, C o r l e t t et al. were used (1967).

An abbreviated scheme of the metamorphosis and the migmatization is as follows: from geosynclinal sediments and tuffs, under the in-

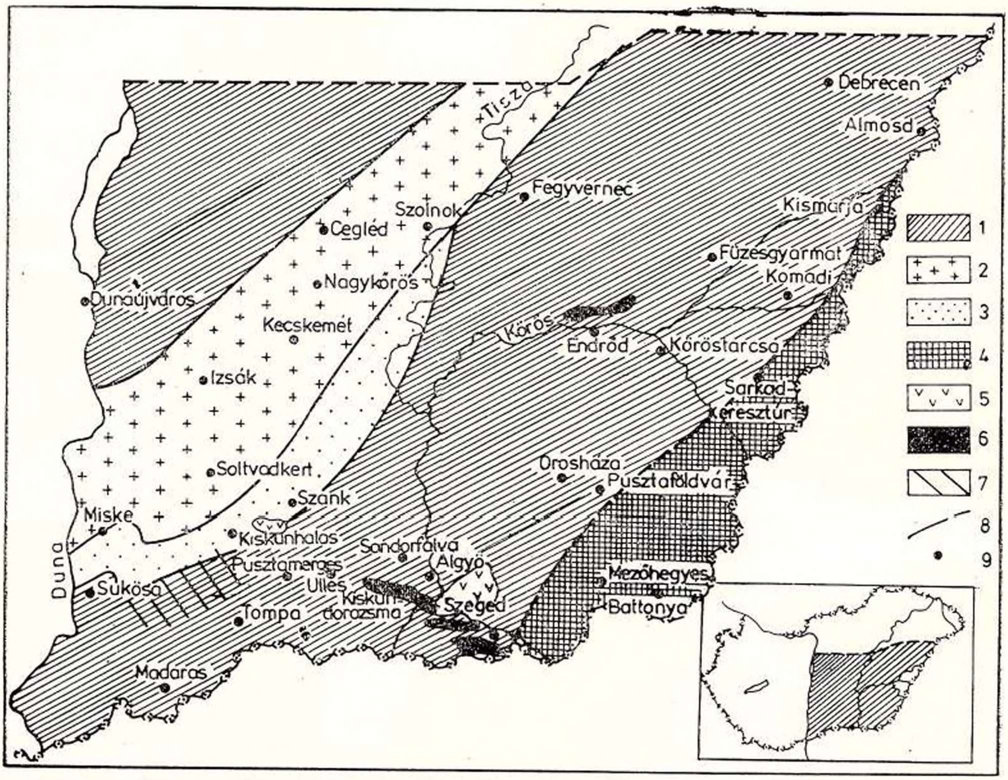

Fig. 1. — Metamorphic rocks in the Great Hungarian Plain. 1, regional metamorphic gneiss-micaschist-amphibolite; 2, I type migmatite; 3, II type migmatite; 4, III type migmatite; 5, low metamorphic formations; 6, singular granites; 7, diaphtorites; 8, boundaries of the area units; 9, locality.

fluence of the metamorphosis of the amphibolite facies, gneiss, amphibolite and micaschist were formed. Later on, sunk into the depth in a synorogenic phase, under the influence of higher temperatures, partly by transcrystallization in a solid phase, partly by a partial fusion and recrystallization, at first granular gneiss with porphyroblasts then various compound migmatites, laminar migmatite, agmatite etc. constructed by the leucosome and melanosome parts of various sizes and forms were formed; then, in the course of a further homogenization, diatexite of granodiorite and granite composition and, at length, the granite itself was formed.

In the first area unit, a total of 26 wells were drilled. Thus, the phases of the migmatization cannot be traced so well as in the following, i.e. the second area unit. Three rock types are known ; gneiss and micaschist were obtained exclusively from the Soltszentimre wells, transitional migmatites from Soltszentimre and Ujszilvás and granite were obtained in large mass from Cegléd, Kecskemét and Nagykőrös. All the three rock types present themselves in isolated units ; clarification of the connection to each other is not possible because of the low number of rock samples.

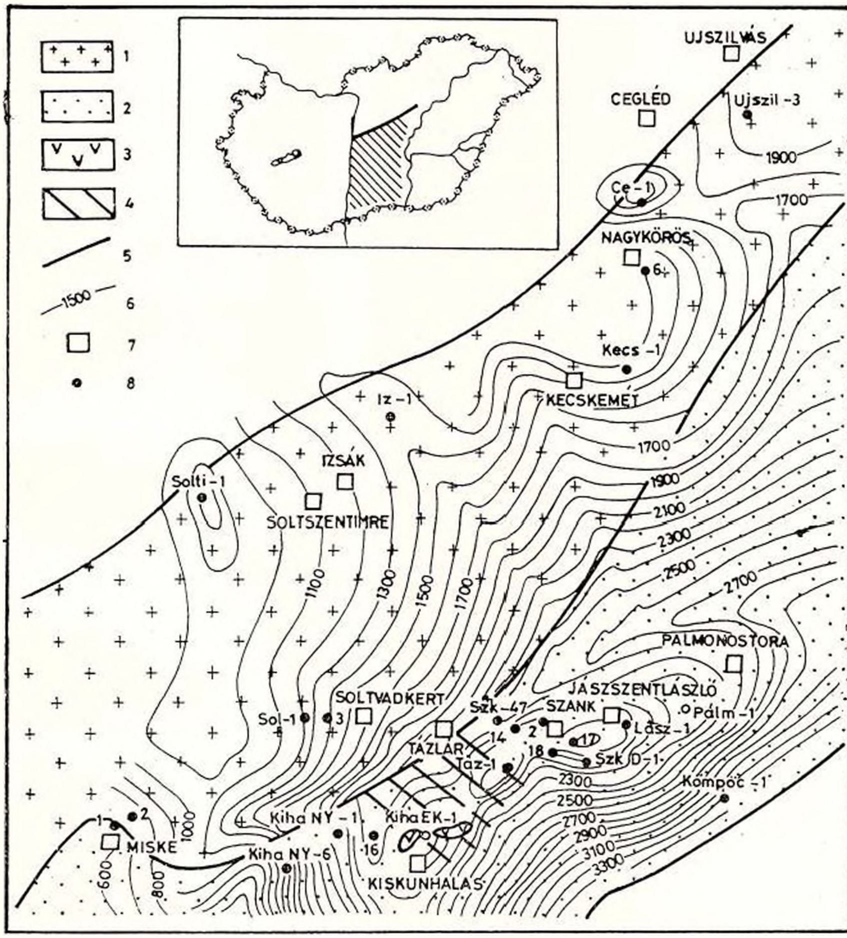

Fig. 2. — Migmatite Belts of the Region between Danube and Tisza.
1, N / I type migmatite ; 2, S/II type migmatite ; 3, low metamorphic formations ; 4, diaphtorites ; 5, boundaries of the area units ; 6, morphology of the metamorphic basement ; 7, locality ; 8, drill-site.

The *gneiss* is fine-grained, containing no porphyroblasts, it is strongly shaly, often folded, consisting of plagioclase, quartz and mica. From the micas, biotite is the most frequent one and it is often chloritized. The plagioclase is a low-temperature oligoclase of 18-32 An percent. As accessories, garnet, zoisite and apatite can be observed. It is not possible to denominate the variations of *the transitional migmatites* since the leucosome and melanosome forms cannot be examined in core pieces of about 10 cm in diameter. The samples consist of 1—2 mm and cm dark and light streaks or lenses, bulbs. These are partly parallel, partly tortuous encrusting each other but they touch each other always along sharp border lines. Ptygmatic streaks frequently occur. The dark melanosome parts are, from the view-point of structure and mineral composition, identical to the above mentioned gneisses, while the leucosome differentiates are of granodiorite and granite composition; they differ only in the frequency of the individual minerals.

Most wells gave granite in a *diatexite* condition according to M e h n e r t's nomenclature. This rock is generally red-and white-spotted or gray, not well homogenized having strongly varied grain size; it consists of red- and white-spotted plagioclase and purplish-gray K feldspar in the Cegléd and Kecskemét region, of bright red K feldspars and white plagioclase in the Soltvadkert region and of biotite showing some order in some places.

From the main minerals, the feldspars have been examined in detail. In my work, I made use of a paper of dr. B u d a . The following generations can be distinguished : With the *plagioclase* there are two generations : *the first one* is isometric or of irregular form, untwinned in 45 percent, twinned in 55 percent; from these, the complex and parallel twins are frequent, especially pericline, albite (ala, albite) Carlsbad ; the normal albite twins are less frequent. In the mineral, the round or raindrop-like quartz inclusions are a frequent occurrence (this is the old quartz generation) and the mirmekite structure is characteristic. The sericitization is very significant in some places. The An % is 18-31, determined by the X-ray method and optically, the 2V varies between 82-90°. *The second generation* is more sodic than this, of 6-12 An percent; the crystal is always fresh, generally twinned, constituting chiefly albite twins; it is more idiomorphous than the oligoclase but often it can only be observed as a thin streak formed around the border of the oligoclase. The 2 V is + 76 — + 88°. Both plagioclase generations are of low temperature.

Three types of *K feldspar* are known; presumably, the *oldest one* is that of 68—78% triclinicity; it is of — 72 — -76 ° 2V; it is not twinned but densely cleaved parallel with 001 and 110; it is hypidiomorphous or isometric, containing thick perthite strings, it is an orthoclase — microcline mixed crystal with the overweight of the orthoclase. *The second K-fp. type* is microcline of 85% triclinicity, fully xenomorphous; it is of undulating and spotted extinction and contains mass plagioclase and biotite inclusions, showing on the border of the grains, the characteristic albite and pericline twin-formation. The microcline is often perthiteous. The microcline has different 2V in the individual domains ranging from —76 to —86°. The *third type* is microcline of maximum triclinicity consisting fully of twin plates; it is much smaller than the second type only a few ten mm with a —86° 2V. Here, the perthite strings are frequent, too. This type is closely interwoven with the young quartz.

In the migmatites of transitional type, the plagioclase of the first generation and, if any, the K feldspar of the first and second type are frequent; the third one does not occur. With the diatexite, the two kinds of plagioclase are to be found in near identical amounts, the orthoclasic microcline being less frequent; the second and third type K feldspar prevails.

I have the following idea as to *the genetical development* of the area unit: as the first phase of migmatization, a certain segregation, a disintegration into melanosome and leucosome streaks have started accompanied by the growth of the grainsize. The intensive alkali metasomatism can be considered as the second phase in the course of which albite (encrusting also the oligoclase) and then the orthoclasic microcline and the two microcline types formed.

There were a lot more wells drilled in the *second area unit*. Only at Szank, 124 wells were drilled; the rocks are more varied, too; a full picture of migmatization can be obtained.

Here too, the starting rock is chiefly *the gneiss* corresponding to the above mentioned gneisses, the An% of the plagioclase ranges from 10 to 28 and the 2V from -86 to $+86°$. The biotite is rarely fresh, it is generally muscovitized, penninized. As accessories, garnet, zoisite, clinozoisite, titanite (often leucoxenized), zircon and apatite can be observed.

Besides the gneiss, *amphibolite* plays a significant role as a paleosome. It can be observed either in segregated units edged in 10—20 cm or 1—2 m streaks in between the gneiss, or mixed with the gneiss as an amphybolic gneiss. The prevailing mineral aggregates of rock samples showing no migmatic effect are the hornblende-biotite-plagioclase-quartz-accessories (titanite, garnet, epidote, zoisite, in some places mass clinozoisite and apatite). The amount of quartz is also considerable, reaching 20% in some places. The plagioclase of non-migmatitic amphibolites corresponds to those contained in the gneisses, with 18—25 An%; it is generally not twinned, and is isometric. In the neighbourhood of parts rich in biotite, a weak zonality can be observed, too.

The start of migmatization is marked by *the appearance* of 5—10 mm round or spindle-shape feldspar-quartz *porphyroblasts* that can be observed sporadically both in the gneiss and in the amphibolite. These are generally in the plane of the schistosity surrounded by the paleosome in a wreath-like form. With the progress of migmatization, migmatites of different types came into being; the porphyroblasts increased in number and size, originating from the fusion of several porphyroblast lenses and small streaks then, the streaks became thicker and were isolated from the melasome. The direction of the leucosome always corresponds to the original schistosity of the melanosome. Later, the parallel streaks get disintegrated and the homogenization begins; various agmatite types, and at last, diatexite and well homogenized granite come into being.

In addition to the minerals described in the gneiss and amphibolite, the following generations formed. *The zonal plagioclase* frequently occurs in every migmatite type. The angle value of cleavage or twin-planes measured from the refractivity directions is characteristic neither of the low temperature level nor of the high temperature level plagioclases; they probably represent an intermediary state. 2V varies irregularly from the interior of the crystal outwards. It is presumed that more calcic and more sodic members have segregated alternately. According to M e h n e r t, the inner core of the plagioclase is more sodic outwards, the An% increased until it reached an idiomorphous zone. From this, outwards from the idiomorphous zone, the An-content diminishes again and an extraordinary xenomorphous albite edge forms. The inner part of the crystal with its increasing An-content can be explained by a low degree temperature increase and with a change taking place in the stability of the plagioclase corresponding to it. Around the zone with maximum An-content, the pressure and temperature conditions decrease rapidly leading to the formation of the albite edge. The zonal plagioclases of the second area unit do not follow this empirical principle. That is why it can be said that the rock got warm and cold many times, the identically acid zones of the various crystals did not form at the same time.

As to the non-zonal crystals, three generations can unambiguously be recognized. *The first one* is probably identical to the oligoclase of the gneiss, therefore, it still belongs to the paleosome; it is strongly sericitized, and can be badly determined and it can be observed almost as an inclusion. The *second type* is fresh; it is twinned or untwinned in 50—50%, consisting mainly of thin albite twins, subordinately, of albite / ala twins; it contains mass quartz, amphibole, biotite and rutile inclusions, it is very often mirmekitic and of low-temperature; on the biotitic parts, i.e. on the parts of gneiss origin, its An% value ranges from 6 to 32, therefore, it is more sodic; on the amphibolic parts, however, its An % value ranges from 34 to 44, therefore, it is more calcic. The same habit is revealed by crystals with —84 to 60° 2V and with transitional temperature the An% value of which cannot be determined; but such crystals are evidently *older* than the previous one, since it presumably could form even at higher temperatures.

The amount of the transitional temperature plagioclase is, for all migmatite types, about one third of the quantity of the low temperature one.

Three types of *the k-feldspar* can be separated too. The *oldest* type is a simple normal twinned, hypidiomorphous orthoclase, cleft generally parallel with 001 and 110. According to X-ray diffractograms, the triclinicity is equal to zero. This fact is proved by the 2V value, too, ranging from —54 to —74°. The —54° 2V value marks a higher temperature and a smaller Si/Al order. In the orthoclase, zonal plagioclase crystals can often be observed where the last zone is idiomorphous as compared to the inner core. This fact is in accordance with M e h n e r t ' s observation according to which a higher temperature is necessary for the formation of a zone with maximum An-content. The *following* type is the untwinned, only cleft, also perthitic orthoclase with 2V ranging from —72 to —76, containing a very large number of inclusions; this orthoclase is fully xenomorphous and it often can be observed only in the cavities and fissures. The *third* type has a habit similar to the previous one, but it is of spotted and undulatory extinction; on the edges of the grain it is a cross-hatched microcline. The triclinicity ranges from 79 to 89% and the 2V value amounts to —82°.

In all phases of the migmatization, the old, simple twin orthoclase is the most frequent; after this, the 2V-72° orthoclase with the porphyroblastic gneisses and the transitional migmatites, while with the diatexite, the microcline can be observed.

I have the following idea as to the rock genetics of this area unit: the oldest rocks are the gneiss and the amphibolite that were formed from the clayey and sandy sediments of an ancient geosynclinal mixed with basic tuffs with a metamorphosis of amphibolite facies. The beginning of the migmatization is marked by the separation and growth of the plagioclase crystals, i.e. the porphyroblasts. As a first step, low-temperature albite — oligoclase formed obtaining various zones with temperature increase. Crystals called intermediary plagioclase and the twinned orthoclase segregated at this "high" temperature, then, with the decrease of temperature the cleft orthoclase and at last the microcline.

The difference between the migmatites found in the first and second area unit is that, in the second unit the amount of amphibole is significant, it is the orthoclase that prevails instead of the microcline and the temperature fluctuation can be proved.

CORRELATION OF PRE-MESOZOIC UNITS ALONG THE GEOTRAVERSE DUBROVNIK-NOVI SAD-BÜKK MOUNTAINS AND HIGH TATRA MOUNTAINS[1]

BY

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Introduction

One of the activities of the IGCP Project No. 5 "Correlation of Pre-Variscan and Variscan events in the Alpine-Mediterranean mountain belts" is the study of the Tethian evolution during the Proterozoic and Paleozoic along seven geotraverses across this region. At the meeting held in Pisa 1978 the presentation of such geotraverses was decided, which would be a basis for further and final correlations.

The Geotraverse C crosses the Dinarides, the Great Hungarian Plain and the Western Carpathians along the line of the Dubrovnik--Kaniža-Bükk Mts-High Tatra Mts. Due to different levels of geological knowledge and different experience along the Geotraverse, the correlation work requires the great and permanent effort of the participants to establish common and uniform ideas, principles and working-methods which may eventually lead to a successful correlation. The presented manner seems to fulfil such a requirement.

Using the stratigraphic, lithological and tectonic etc. principles elaborated by stimulators of the Project No. 5 and discussed at its meeting, on the basis of careful analysis of available data, these were plotted for correlation purposes. The Geotraverse C is illustrated by several graphic plots (strips) (plate):

1. extension of the geological units

2. a 50 km wide geological stripe-map, which indicates the situation either on the basement surface or surficial position

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3. a generalized geological profile depicting the crustal thickness and general tectonic setting of units and
4. characteristic lithological correlation forms unifying developments in greater units on the basis of several detailed charts presented elsewhere (Newsletter of IGCP Project No. 5, Vol. 2, and 3).

This material shows the present state of our knowledge. Further planned plots (and strips) should give palynspastic restoration to different chronological levels of the Pre-Mesozoic evolution. However, the recent state -due to the unsolved Alpine tectonism in many parts of the Geotraverse are regarded as preliminary solutions (Karamata, 1980; Szederkényi, 1980). Nevertheless, the presented plot allows some important deductions for the correlation.

Short Description of Presented Units

Nine units are distinguished along the Geotraverse C containing Pre-Mesozoic developments. They are the following from the south to the north :

The Praca-Lim Unit consists mainly of arenaceous to pelitic rocks with basaltic to keratophyric volcanics and less carbonates of Ordovician to Carboniferous age that suffered very-low grade or low grade metamorphism.

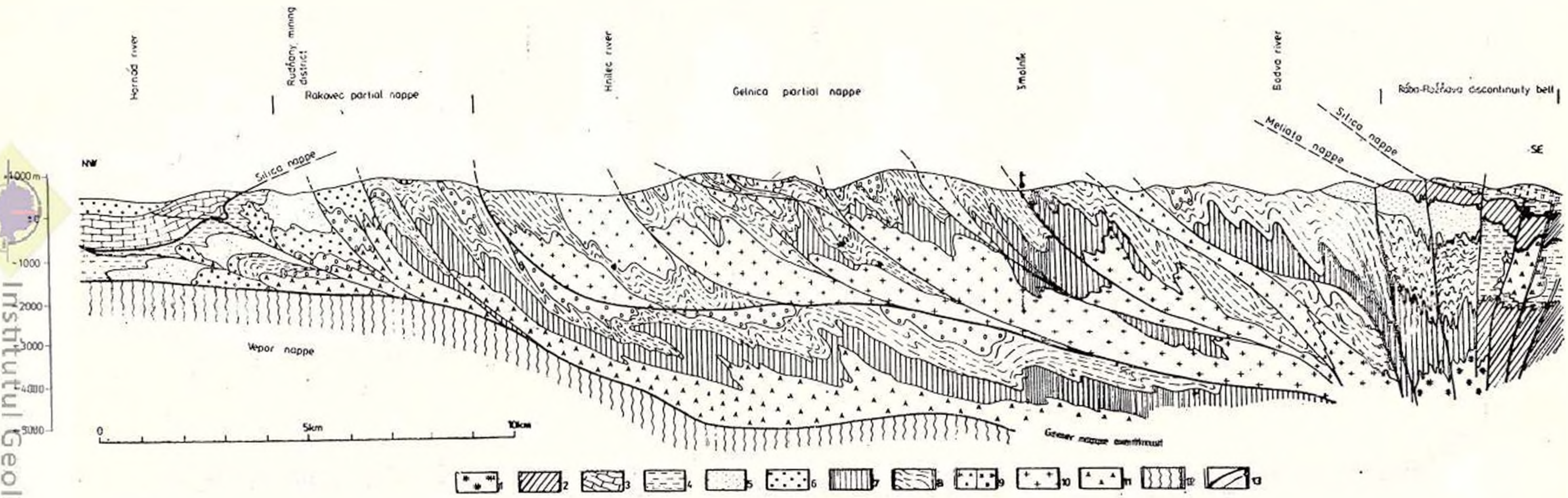
The Drina-Golija Unit is very similar to the preceding unit. It contains a little more igneous rocks and metamorphic sequences having higher grade. High pressure and low temperature metamorphism occur along the NE margin (probable Mesozoic overprint).

The Jadar Unit is built up of Devonian to Carboniferous psammites and pelites with limestone intercalations showing very-low grade and low grade metamorphism.

The Bačka (Bácska) Unit occurs below a thick Mesozoic and younger sequence of the Pannonian Basin. It contains at least two sub-units : (1) Barrow-type medium grade gneiss, micaschist, amphibolite and marble with palyngenous granite of probable Precambrian age and (2) greenschists, phyllite, metapsammite and carbonate schist of greenschist to epidoteamphibolite facies. Both sub-units are penetrated by aplite and microgranite veins of Carboniferous age as well as by some Lower Permian rhyolite vents.

The Bükk-Uppony-Szendrö Unit consists of calcareous and subordinated pelitic rocks of Devonian to Triassic age that suffered a very-low grade Variscan burial metamorphism overprinted by a slightly higher Alpine recrystallization (Árkay, 1979). Due to the permanent sinking (from Ordovician to at least Upper Triassic) this sequence represents a unique longliving sedimentary basin filling.

The Gemer Unit is composed by two partial structures (fig.) : (1) Huge layers of acidic volcanics (rhyolites), clastics and less frequent gra-



Geological profile across the Spišsko-Gemerské Rudohorie Mountains area (by P. G r e c u l a and I. V a r g a, 1979).

1, intrusives of Eocene to Miocene age, inferred; 2, Meliata-Bükk Mountains, Mesozoic facies of Early Paleozoic to Mesozoic age, undifferentiated; 3, mainly carbonate development of Middle to Upper Triassic and Jurassic in the Silica nappe. Gemer nappe: 4, sandstone, quartzite, arenaceous shale, Lower Triassic; 5, conglomerate, sandstone, arenaceous shale and rare evaporite layers (mainly anhydrite and gypsum), rhyolite to dacite and tuff, Permian; 6, conglomerate, sandstone, pelitic and arenaceous graphitic shale, basalt and tuff, all metamorphosed to greenschist facies and partly to amphibolite facies, Carboniferous; 7, predominating graphite-sericite phyllite; 8, predominating chlorite-sericite, phyllite; 9, predominating volcanite (a, acidic, b, basic varieties) (7 – 9 Early Paleozoic development in the Gemer nappe); 10, granitoids of the Variscan and Alpine cycles, undifferentiated; 11, basic rocks and crystalline schist in amphibolite facies metamorphism; 12, undivided Vepor nappe; 13, overthrusts of various order.

nite bodies of Variscan and Alpine age create the Gelnica sub-unit, whereas (2) considerable volume of ophiolite-type basic magmatites and graphite-bearing pelites are in the Rakovec sub-unit (and group). Supposed differences in age have been removed by the discovery of similar sporomorph assemblages in both sub-units (S n o p k o v á, 1973, 1980). Late Paleozoic developments differ on the northern and southern side. The Upper Carboniferous joins the Rakovec Group on the north by conglomerate, graphiteous metapsammite and basic volcanics, whereas the Upper Permian red conglomerate and rhyolite with local evaporite facies terminate the Paleozoic. In the south, a Lower Permian transgressive sequence (V o z á r o v á, 1980) with rhyolite bodies proceeds upwards into Upper Permian clastics and subordinated carbonatic lenses (Triassic beds are probable as well).

The Vepor Unit is the lower structure of the Central West Carpathians composed by two sub-units: (1) the Hron complex (K l i n e c, 1966) of slightly metamorphosed Early Paleozoic sediments and huge acidic and basic volcanic masses and (2) the Králova hola complex of Variscan granitoids and their polymetamorphic mantle. Late Paleozoic to Mesozoic cover of the Variscan folded structure consists of detrital rhyolite-bearing Permian and shallow-marine Mesozoic beds (V o z á r o v á-V o z á r, 1978, 1979).

The Tatric Unit comprises the outermost Variscan elements in the West Carpathians. High-grade Variscan crystallines and granitoids participate on the unit. The presence of older granitoids is supported by radiometric data (460 m.y.). The Post-Variscan cover consists of (but subordinated) probable Permian coarse grained clastics and younger Mesozoic sequences.

The Hronic Unit is a completely allochthonous Alpine nappe sheet. It contains Upper Paleozoic sequences (Upper Westphalian-Upper Permian to Lower Triassic) detached and far removed from its original basement. This unit comprises the most complete Variscan molasse basin developed along deep-faults reaching to sources of typical linear tholeiitic magmatism (V o z á r, 1978) with red clastics in its upper part (V o z á r o v á-V o z á r, 1978, 1980).

Value of the Presented Material for Correlation Purposes

Using accepted rules and legends for maps, profiles and lithostratigraphic correlation forms (the latter are prepared by F e n n i n g e r-O b r a d o v i c, 1980), the draft has been used as a basis for a first attempt of correlation along the Geotraverse. It should be stressed that numerous problems remain included in the plot, therefore it is regarded as preliminary work. The main problems are the following:

1. Considerable differences occur in the density and value of the geological data along the Geotraverse (e.g. the Carpathian and Dinaric part is plotted from surficial data, whereas only deep drillings reached the Pre-Mesozoic basement in the Great Hungarian Plain).

2. Undoubtedly Alpine nappe structure disturbs Paleozoic sequences in the West Carpathians. Similar pattern cannot be interpreted in the Dinarids.

3. Due to the lack of sufficient data, it is impossible to draw lithostratigraphic correlation forms for some crucial parts of the Geotraverse (e.g. between the outcropping Jadar and the covered Bačka Units or between the latter and the Bükk Unit).

Conclusions

Presented data indicate a conspicuous difference between the northern and southern part of the Geotraverse, which has already existed from the Pre-Mesozoic time. Proterozoic rocks may be expected in the basement of the Pannonian Basin or — considering the Variscan development — from the West Carpathians where they occur on the surface. However it is not possible to make any clear distinction between really. Proterozoic sequences and possible Paleozoic ones.

Paleozoic formations in the Dinarids — considering the lithology magmatism and metamorphic grade — probably represent parts of a single large basin, which had persisted from the Ordovician up to the Carboniferous. The Prača-Lim Unit represents the marginal portion whereas the Jadar Unit the central part of this basin. Anyhow relative, mostly lateral displacements among both units (or parts of the ancient basin) took place after the Paleozoic time. Zones of such movements are indicated probably by Mesozoic ophiolite belts. Similar displacements occurred along both margins of the Bačka Unit indicated by the South Bačka and Szolnok ophiolites.

In the Western Carpathians deeper crustal levels of the Variscan orogenic structure (or internal parts of the original Variscan orogenic belt) occur on the surface. Similar position of the Variscan crustal segment is valid for the Pre-Neogene surface in the majority of the basement of the Great Hungarian Plain. To the contrary, internal (or deeper crustal?) structural elements are not known on the surface within the Dinaric part of the Geotraverse. This portion has either not reached such a crustal state during the Variscan evolution or Post-Variscan erosion has not yet uncovered them.

In spite of the numerous mentioned problems, several further deductions may be performed from the presented pattern and distribution of the Pre-Alpine elements :

1. The Prača-Lim and Drina-Golija Units are mutually well correlable and (based on lithology, stratigraphic span and paleobiogeographic affinity) the Bükk-Uppony-Szendrő Unit can also be ranged into this group of units.

2. Due to their volcanogene and volcano-sedimentary sequences, the Gemer and Vepor Units differ from the previous ones. Mainly the Devonian ophiolites reflect here- a priori — an extreme geotectonic

and evolutionary situation, which is not repeated elsewhere in the Geotraverse during the Variscan evolution.

3. Composed by probably Proterozoic and high-grade metamorphics and large palyngenic granite masses, the Bačka and Tatra Units have much in common. A weak Variscan magmatic activity produced some aplite and microgranite veins in the Bačka Unit. A conspicuous phenomenon is the lack or subordinated existence of Variscan molasse in both units.

REFERENCES

All literature data are contained in the following papers :

Sassi F. P., Zirpoli G./eds/ (1980) Newsletter of IGCP Project 5, Univ. Padova, 2.

— Karamata S. /eds/ (1981) Newsletter of IGCP Project 5, Univ. Beograd and Univ. Padova, 3.

Imprim. Atel. Inst. Geol., Geol.

K-AR AND U-PB DATING OF THE METAMORPHIC FORMATIONS AND THE ASSOCIATED IGNEOUS BODIES OF THE CENTRAL SOUTH CARPATHIANS[1]

BY

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Introduction

In the central part of the South Carpathians, K-Ar geochronological investigations were already carried out by Soroiu et al. (1970, 1972) and Mînzatu et al. (1975). The aim of this paper is to present old and new K-Ar data obtained at the Atomic Physics Institute in Bucharest, U-Pb ages recorded at the Swiss Federal Institute of Technology, and to discuss the benefit brought by this information to the timing of the geological evolution of the mentioned area. The K-Ar ages were obtained by using fast neutron activation for potassium determinations and thermal neutron activation for radiogenic argon measurements. The $\lambda_{\beta} = 4.962 \cdot 10^{-10}$ yr^{-1}, $\lambda_e + \lambda'_e = 0.581 \cdot 10^{-10}$ and ^{40}K $= 0.01167$ atom % were used.

The U-Pb ages were obtained on zircon fractions separated from two rocks. U and Pb were determined by the isotopic dilution method. The used constants are given in Table 4 and the concordia diagram was drawn with an IBM 370/135 computer from IFIN. The individual ages are within 2% correct (1 σ), and the upper interception in the concordia diagram refers to a 5% error.

Geological Constitution of the Central South Carpathians

New refinements of the well known nappe structure of this part of the Carpathian Belt (Kräutner et al., 1981; Berza et al., 1983) show it as a pile of several overthrust nappes with thousands or tens of

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thousands square kilometers areas, kilometric thickness and observable net slips of several tens of kilometers. From top to bottom, these nappes are: Poiana Ruscă Supragetic Unit, Getic Nappe, Upper Danubian Group of Units and Lower Danubian Group of Units (Pl.).

The Poiana Ruscă Supragetic Unit is represented by the low-grade metamorphosed Paleozoic formations from the central and northern part of the Poiana Ruscă Mts and will not be further discussed here.

The Getic Nappe comprises a metamorphic basement and a sedimentary cover of Upper Carboniferous to Upper Cretaceous deposits. In the metamorphic basement, K r ä u t n e r (1980) recognized an angular unconformity and a metamorphic gap between an older (Middle Proterozoic) Sebeș-Lotru Group, in amphibolite facies, and a younger (Vendian-Cambrian) Cibin Group, in greenschist facies.

The Upper Danubian Group of Units also consists of a metamorphic basement and a sedimentary cover of Upper Carboniferous to Upper Cretaceous deposits but sometimes the latter shows a very low-grade metamorphism. The metamorphic basement presents again an unconformity between an older polymetamorphic Zeicani Group and a low-grade Ordovician to Lower Carboniferous Drencova-Rîul Alb Group (K r ä u t n e r et al., 1981). The internal structure of this nappe is complex, tectonic planes dividing it into several Alpine and pre-Alpine tectonic units, representing individual nappes.

The Lower Danubian Group of Units is the lowest tectonic element known in the South Carpathians, but it also probably overthrusts either unknown Carpathian units, or directly the Moesian platform block of the underthrust foreland . The metamorphic basement of the Lower Danubian Units consists of the polymetamorphic Lainici-Păiuș and Drăgșan Groups, and Bodu, Rof, Nisipoasa and Rîușorul Formations, crossed by several plutons and countless minor bodies of granitic or granodioritic composition, and unconformably overlain by the low-grade Ordovician to Lower Carboniferous Tulișa Group (K r ä u t n e r et al., 1981).The sedimentary cover is represented by mainly Jurassic and Cretaceous deposits, sometimes exhibiting a low-grade metamorphism. The internal structure of this group of units is also complex, several Alpine and pre-Alpine tectonic units being recognized by B e r z a et al. (1983).

Isotopic Ages

K-Ar ages. In tables 1, 2 and 3, 117 samples of various metamorphic and igneous rocks from the Getic, Upper Danubian and Lower Danubian Units are listed. The Sebeș-Lotru Group medium-grade metamorphic rocks from the Getic Nappe show K-Ar ages between 124 and 336 m.y. (Tab. 1). In the Upper Danubian Units, polymetamorphic rocks from the Zeicani Group give K-Ar ages between 94 and 456 m.y., and the intruding Muntele Mic and Rîul Șes granitoid bodies yielded K-Ar ages between 245 and 271 m.y., and 162 and 280 m.y., respectively. A Mesozoic slate from the same units has a 97 m.y. K-Ar age.

TABLE 1

K-Ar ages of rocks from the Getic Nappe

No	Locality	Rock type	Mineral	K %	$^{40}Ar_{rad}$ (10^{-9} moles/g)	K-Ar Age (m.y.)
1	Godeanu Mts	Amphibolite — Sebeş-Lotru Gr.	A	0.303	0.194	336±20
2	Godeanu Mts	Pegmatite — Sebeş-Lotru Gr.	B	7.62	1.70	124±4
3	Godeanu Mts	Pegmatite — Sebeş-Lotru Gr.	B	7.96	1.85	129±4
4	Godeanu Mts	Migmatite — Sebeş-Lotru Gr.	B	7.28	2.10	159±5
5	Godeanu Mts	Migmatite — Sebeş-Lotru Gr.	M	7.63	2.25	162±5
6	Godeanu Mts	Migmatite — Sebeş-Lotru Gr.	M	7.80	2.30	162±5
7	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	7.64	2.40	171±6
8	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	7.72	2.48	177±5
9	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	8.04	3.13	211±7
10	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	8.76	2.55	160±5
11	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	B	8.16	1.73	118±3
12	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	7.80	2.38	167±5
13	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	B	7.34	1.85	140±5
14	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	8.48	2.65	171±6
15	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	8.88	2.40	150±5
16	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	7.66	3.28	232±9
17	Lotrului Mts	Pegmatite — Sebeş-Lotru Gr.	M	8.10	2.98	200±6
18	Lotrului Mts	Micaschist — Sebeş-Lotru Gr.	B	4.00	1.13	155±6
19	Lotrului Mts	Micaschist — Sebeş-Lotru Gr.	M	5.59	1.58	155±5
20	Lotrului Mts	Micaschist — Sebeş-Lotru Gr.	M	6.63	1.83	152±6
21	Căpăţînii Mts	Pegmatite — Sebeş-Lotru Gr.	M	9.25	4.38	254±7
22	Căpăţînii Mts	Biotite gneisse — Sebeş-Lotru Gr	B	5.70	2.95	276±5
23	Mehedinţi Pl	Biotite gneisse-Sebeş-Lotru Gr.	B	3.20	0.810	140±4

TABLE 2

K-Ar ages of rocks from the Upper Danubian Units

No	Locality	Rock type	Mineral	K %	$^{40}Ar_{rad}$ (10^{-9} moles/g)	K-Ar Age (m.y.)
1	Petreanu Mts	Mica gneiss — Zeicani Gr.	WR	2.62	0.445	95±4
2	Petreanu Mts	Mica gneiss — Zeicani Gr.	M	8.55	1.93	126±4
3	Petreanu Mts	Mica gneiss — Zeicani Gr.	M	7.41	1.73	130±4
4	Petreanu Mts	Amphibolite — Zeicani Gr.	A	0.845	0.310	200±10
5	Petreanu Mts	Biotite gneiss — Zeicani Gr.	B	7.00	1.20	96±4
6	Petreanu Mts	Biotite shell of a — Zeicani Gr. serpentinite lens	B	6.96	1.37	110±4
7	Muntele Mic Mts	Amphibolite — Zeicani Gr.	A+B	1.10	0.685	328±13
8	Muntele Mic Mts	Amphibolite — Zeicani Gr.	A+B	1.50	0.815	289±11
9	Muntele Mic Mts	Amphibolite — Zeicani Gr.	A	0.533	0.480	456±18
10	Muntele Mic Mts	Amphibolite — Zeicani Gr.	B+A	2.66	0.765	159±6
11	Muntele Mic Mts	Biotite amphibolite — Zeicani Gr.	B	4.35	1.50	188±7
12	Muntele Mic Mts	Mica gneiss — Zeicani Gr.	WR	6.25	1.05	94±4
13	Muntele Mic Mts	Granodiorite — Muntele Mic B	B	7.42	3.75	271±10
14	Muntele Mic Mts	Granodiorite Muntele Mic Gr.	B	5.94	2.83	255±10
15	Muntele Mic Mts	Granodiorite Muntele Mic. Gr.	B	4.98	2.28	245±10
16	Ţarcu Mts	Granite-Rîul Şes Body	B+C	1.84	0.653	194±4
17	Ţarcu Mts	Granite — Rîul Şes Body	B+C	1.96	0.575	162±5
18	Ţarcu Mts	Granite — Rîul Şes Body	B	5.33	2.80	280±6
19	Ţarcu Mts	Mesozoic slate	WR	5.43	0.935	97±4

TABLE 3

K — Ar ages of rocks from the Lower Danubian Units

No	Locality	Rock type	Mine-ral	K %	$^{40}Ar_{rad}$ (10^{-9} moles/g)	K-Ar Age (m.y.)
1	Petreanu Mts	Biotite gneiss — Bodu F.	B	5.35	0.935	98±4
2	Petreanu Mts	Biotite gneiss — Bodu F.	B	3.95	0.685	97±4
3	Petreanu Mts	Biotite gneiss — Bodu F.	B	6.02	1.43	133±5
4	Petreanu Mts	Biotite gneiss — Bodu F.	B	5.66	1.20	118±5
5	Petreanu Mts	Biotite gneiss — Bodu F.	B	5.56	7.78	667±20
6	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	6.19	1.32	119±4
7	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	4.28	2.50	309±12
8	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	5.17	2.40	250±9
9	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	3.42	1.23	197±8
10	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	2.44	2.65	538±20
11	Petreanu Mts	Augen gneiss — Petreanu Gneiss	B	6.76	4.25	330±10
12	Petreanu Mts	Granite — Vîrful Pietrei Body	M	7.00	3.78	287±9
13	Petreanu Mts	Biotite gneiss Nisipioasa F.*	B	4.96	0.925	106±6
14	Petreanu Mts	Biotite gneiss — Furcătura Gn	B	6.07	3.13	276±8
15	Petreanu Mts	Amphibolite — Rof Formation	A	0.530	0.100	106±10
16	Petreanu Mts	Leptynite — Rof Formation	M	8.20	3.18	211±8
17	Petreanu Mts	Mica gneiss — Rof Formation	B+M	2.11	0.648	169±6
18	Petreanu Mts	Biotite gneiss — Rof Formation	B	7,02	3.75	285±9
19	Petreanu Mts	Biotite gneiss — Rof Formation	B	6.68	3.85	305±12
20	Petreanu Mts	Biotite gneiss — Rof Formation	B	3.78	2.60	359±13
21	Petreanu Mts	Biotite gneiss — Rîuşorul F.	WR	2.40	0.445	104±4
22	Retezat Mts	Granodiorite — Retezat Body	B	6.43	0.443	97±5
23	Retezat Mts	Granite — Retezat Body	M	6.59	1.00	207±8
24	Retezat Mts	Granite — Retezat Body	M	6.96	0.800	159±6
25	Retezat Mts	Granite — Retezat Body	M	6.61	0.602	126±5
26	Retezat Mts	Granite — Retezat Body	M	7.94	0.930	161±6
27	Paring Mts	Granite — Paring Body	B	3.20	1.30	220±8
28	Paring Mts	Amphibolite — Drăgşan Group	A	0.281	0.077	152±11
29	Paring Mts	Amphibolite — Drăgşan Group	A	1.04	0.450	234±8
30	Paring Mts	Amphibolite — Drăgşan Group	A	0.584	0.338	307±8
31	Paring Mts	Amphibolite — Drăgşan Group	A	0.760	0.470	325±9
32	Paring Mts	Amphibolite — Drăgşan Group	A	0.421	0.257	321±10
33	Paring Mts	Mica gneiss-Lainici-Păiuş Gr.	M	3.26	0.545	94±4
34	Paring Mts	Migmatite — Lainici-Păiuş Gr.	M	8.25	8.73	524±10
35	Paring Mts	Migmatite — Lainici-Păiuş Gr.	B	5.49	5.70	518±15
36	Paring Mts	Migmatite — Lainici-Păiuş Gr.	M	8.83	8.40	479±10
37	Paring Mts	Migmatite — Lainici-Păiuş Gr.	M	8.48	9.98	577±12
38	Vilcan Mts	Migmatite — Lainici-Păiuş Gr.	M	8.92	10.60	581±11
39	Vilcan Mts	Granodiorite-Suşiţa Body	B+C	0.578	0.233	218±8
40	Vilcan Mts	Granodiorite — Suşiţa Body	B+C	0.860	0.280	179±6
41	Paring Mts	Pegmatite — Suşiţa Body	M	8.90	9.83	547±10
42	Paring Mts	Granodiorite — Suşiţa Body	B	3.77	3.16	428±8
43	Paring Mts	Granodiorite — Suşiţa Body	B	2.84	2.08	380±10
44	Paring Mts	Granodiorite — Suşiţa Body	B	4.92	2.37	258±5
45	Paring Mts	Granite — Suşiţa Body	M	8.71	7.40	433±12
46	Paring Mts	Granite — Suşiţa Body	B	3.44	1.97	303±12
47	Paring Mts	Granite — Suşiţa Body	M	6 44	0.833	73±3
48	Paring Mts	Granite — Suşiţa Body	M	8.62	7.45	440±9
49	Paring Mts	Granodiorite — Novaci Body	A+B	1.16	1.17	503±15
50	Paring Mts	Granodiorite — Novaci Body	B+C	1.38	0.625	244±8
51	Paring Mts	Granodiorite — Novaci Body	B+A	3.20	2.55	410±14
52	Paring Mts	Granodiorite — Novaci Body	B	3.04	1.55	272±10

No	Locality	Rock type	Mineral	K %	$^{40}Ar_{rad}$ (10^{-9} moles/g)	K-Ar Age (m.y.)
53	Paring Mts	Granite — Novaci Body	A+B	0.620	0.530	436±17
54	Paring Mts	Granite — Novaci Body	A+B	0.848	0.430	271±10
55	Paring Mts	Granite — Novaci Body	A+B	2.28	2.30	505±18
56	Paring Mts	Granite — Novaci Body	A+B	2.44	2.49	510±18
57	Paring Mts	Granodiorite — Olteț Body	B	3.30	4.60	665±13
58	Paring Mts	Granodiorite — Olteț Body	B	5.28	6.63	609±18
59	Paring Mts	Granodiorite — Olteț Body	B+C	1.81	1.05	307±9
60	Paring Mts	Granodiorite — Olteț Body	B+C	1.76	1.16	345±7
61	Vîlcan Mts	Granite — Tismana Body	B+C	1.66	0.893	287±6
62	Vîlcan Mts	Granite — Tismana Body	B+C	4.59	5.55	590±12
63	Vîlcan Mts	Granite — Tismana Body	F	8.83	4.80	289±9
64	Vîlcan Mts	Granite — Tismana Body	B	5.15	5.93	565±20
65	Vîlcan Mts	Granite — Tismana Body	B	5.04	5.60	548±20
66	Vîlcan Mts	Granodiorite — Tismana Body	B	5.03	5.10	507±15
67	Vîlcan Mts	Granite — Tismana Body	F	9.08	4.30	254±7
68	Vîlcan Mts	Granodiorite — Tismana Body	B	5.79	6.20	530±16
69	Mehedinți Pl	Granite — Tismana Body	B+C	2.02	1.21	316±6
70	Mehedinți Pl	Granite — Tismana Body	B+C	2.21	1.63	382±8
71	Paring Mts	Mesozoic or Paleozoic slate	WR	3.01	0.565	105±4
72	Godeanu Mts	Mesozoic slate	WR	1.29	0.303	131±10
73	Godeanu Mts	Mesozoic slate	WR	3.72	0.600	90±6
74	Godeanu Mts	Mesozoic slate	WR	1.34	0.258	108±6
75	Mehedinți Pl	Mesozoic peridotite (Severin Nappe)	WR	0.543	0.091	94±5

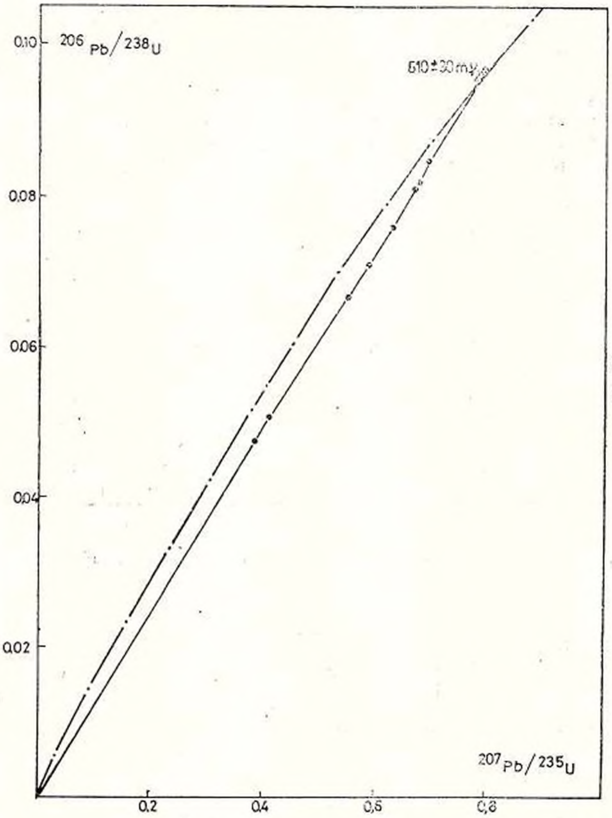
A — amphibole; M — muscovite; B — biotite; C — chlorite; F—K — feldspar; WR — whole rock

The samples from the Lower Danubian Units are grouped according to the tectonic units named by B e r z a et al. (1983). In the Petreanu Unit, the polymetamorphic rocks of the Bodu Formation and the connected Petreanu Gneiss have K-Ar ages between 97 and 667 m.y. and 119 and 538 m.y., respectively. The intruding Vîrful Pietrei granitic body gave a 287 m.y. K-Ar age. In the underlying units, a sample of the Furcătura Gneiss has a 276 m.y. K-Ar age, and polymetamorphic rocks from the Rof Formation shw K-Ar ages between 106 and 359 m.y.

Going eastwards, in the Nucșoara Unit, a polymetamorphic rock from the Rîușorul Formation has a 104 m.y. K-Ar age. Still eastwards and southwards, in the Retezat-Parîng Unit, polymetamorphic rocks from the Drăgșan Group have K-Ar ages between 152 and 325 m.y., and a sample from the intruding Parîng granite shows a 220 m.y. K-Ar age. The Retezat granitoid rocks gave K-Ar ages between 97 and 207 m.y. Finally, in the Vîlcan-Pilugu and Schela Units, polymetamorphic and migmatic rocks from the Lainici-Păiuș Group have K-Ar ages between 94 and 577 m.y., and the intruding Sușița, Novaci, Olteț and Tismana granitoid bodies have K-Ar ages between 73 and 547, 271 and 510, 307 and 665, and 254 and 590 m.y., respectively. Mesozoic slates from the Lower

Danubian Units show K-Ar ages between 90 and 131 m.y., and an Alpine Mesozoic peridotite from the Severin Nappe has a 94 m.y. K-Ar age.

U-Pb ages. Zircon crystals from two granitic rocks intruded in the metasedimentary Lainici-Păiuş Group were separated and the resultant $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ratios (Tab. 4) are plotted in the figure.



Concordia diagram for zircons from granitoid rocks (Paring Mts).

The first sample is a granite with K-feldspar megacrysts from the Novaci pluton, collected in the Gilort Valley. The second sample represents a similar-looking granite with K-feldspar megacrysts from a little body in the Jiu Gorges. All the eight zircon fractions from both granites fit a line discordant in respect of the concordia curve, with an upper interception age at 610 m.y. and the lower interception in the origin.

Geologic significance of the isotopic ages. Excepting a few samples, the data listed in tables 1, 2 and 3 come from metamorphic rocks ascribed to several groups and formations, from the Getic, Upper Danubian and Lower Danubian Units, or from igneous bodies intruding them.

Based on the typical lithological sequences, Kräutner et al. (1981) have proposed the following correlation scheme : the Drăgşan and

TABLE 4

U—Pb ages of granitic rocks of the Lower Danubian Units

Sample			Concentration		Total Pb isotopic composition				Age (m.y.)		
Location	No	Grain size (μ)	U %	Pb (Total) %	204 %	206 %	207 %	208 %	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
Zircon from Gilort Valley granite	1	42—53 n.m.	0.0522	0.0040	0.0532	80.511	5.534	13.901	447	469	592
	2	+125 n.m.	0.0503	0.0036	0.0621	80.807	5.684	13.447	420	446	596
	3	53—75 m.	0.0866	0.0048	0.0777	80.336	5.860	13.737	319	352	588
	4	42—53 m.	0.1069	0.0055	0.0796	80.638	5.892	13.390	297	333	582
Zircon from Jiu Valley granite	5	100—125n.m.	0.1921	0.0153	0.0318	91.891	5.905	2.172	526	536	593
	6	42—53 n.m.	0.2084	0.0160	0.0230	92.169	5.830	1.977	511	526	606
	7	100 m.	0.2011	0.0153	0.0229	92.297	5.822	1.858	509	523	601
	8	75—100 m.	0.2007	0.0145	0.0353	91.587	5.954	2.424	478	497	600

n.m. = nonmagnetic fraction
m. = magnetic fraction
Common Pb correction
204 = 1.409 %
206 = 24.887 %
207 = 21.505 %
208 = 52.198 %

Constants
$\lambda_{(^{238}U)} = 1.55125 \cdot 10^{-10}$ yr^{-1}
$\lambda_{(^{235}U)} = 9.8485 \cdot 10^{-10}$ yr^{-1}
$\frac{^{238}U}{^{235}U}$ present = 137.88

Zeicani Groups, together with the Rof Formation, correspond to the middle part of the Sebeș-Lotru Group, while the Lainici-Păiuș Group, Bodu Formation and Rîușoru Formation are equivalents of the upper part of the Sebeș-Lotru Group. The subsequent evolution of the Getic and Danubian areas was different. So, the Sebeș-Lotru Group of the Getic Nappe suffered a Barrovian followed by a low-pressure type metamorphism (Bercia and Hârtopanu, 1980), and, in restricted areas, an early Caledonian retrogression (Kräutner, 1980), its oldest cover deposits being Upper Carboniferous conglomerates. The basement rocks of the Danubian Units were subjected to Barrovian medium grade metamorphism and/or low-pressure high-temperature metamorphism connected with extensive granite emplacement, followed by prominent regressions. They support low-grade formations dated by macrofauna, macroflora and palynomorphs as Upper Ordovician to Lower Carboniferous (Kräutner et al., 1981). Considering these facts, and remembering that the South Carpathians represent a pile of Alpine nappes, some of them with complex pre-Alpine internal structures, the interpretation of the isotopic data points to the following ideas:

1. Young K-Ar ages (up to 200 m.y.) recorded from many mineral or whole rock samples of metamorphic or igneous rocks of the basements of the mentioned units show a partial loss of radiogenic Ar in the process of Alpine overthrusting, known in this region to be subsequent to the Upper Cretaceous, but earlier than the Paleogene, i.e. Laramic ($\approx 60-70$ m.y.). Only one 73 m.y. K-Ar age for a sample of Sușița granite points to a nearly complete resetting of the K-Ar clock. These young ages are clearly connected either with the overthrust planes at the bottom of the Getic Nappe and Upper Danubian Group of Units, or with other tectonic planes between various Danubian Units (Pl.). The loss of the radiogenic Ar may be ascribed either to the structural and mineralogical reorganizations of the rocks of these areas, due to mechanical and thermal effects of the nappe gliding (similar to a situation described by Lee et al., 1970), or to the Alpine metamorphism. The persistence of a lot of high ages in the unit with the lowest geometrical position favours the first assumption.

2. Intermediate K-Ar ages (200 to 500 m.y.) are also found for samples from the basements of the Getic and Upper Danubian Nappes (where they are in minority in respect of young ages), but are prevailing in the Lower Danubian basement rocks. Again, the mineralogical nature of the sample is not critical. These intermediate ages may either reflect Variscan events (Soroiu et al., 1972), or are the result of a lesser Alpine radiogenic Ar loss than in the case of the young ages. In the Danubian Nappes, tectonic (and probably also metamorphic) events are known at the Devonian Carboniferous and/or Lower/Middle Carboniferous boundaries, that is at 345 and 300 m.y., respectively; however there is no noticeable grouping of the intermediate ages around these values. This continuous age pattern, which may also be extended to the range of the younger and older ages, favours the interpretation of the intermediate ages as partial Alpine resettings. In this case, the postulated Variscan metamorphism was not very active in the removing of the

radiogenic Ar from the basement rocks, at least at the scale of the collected samples.

3. Old K-Ar ages (500 to 650 m.y.) are found for 16 samples, all from the basement of the Lower Danubian Group of Units, especially from its southernmost part, where Lainici-Păiuş rocks cut by various granites are exposed. There is again a complete range of values between 500 and 650 m.y. so that, almost certainly, the highest figures are also subjected to a partial resetting. These K-Ar ages demonstrate the Precambrian age of the Bodu Formation (and associated Petreanu Gneiss) and Lainici-Păiuş Group, but also of the Olteţ and Tismana granitoid bodies, emplaced within the Lainici-Păiuş rocks. If the correlation proposed by Kräutner et al. (1981) is real, this conclusion is also valid for the Sebeş-Lotru, Zeicani and Drăgşan Groups, and the Rof and Rîuşoru Formations. From geologic relations, the emplacement age of various Danubian granitoid bodies is ante-Silurian, but from the presented K-Ar data all that can be said is that the Şuşiţa Pluton is older than 547 m.y., and the Novaci Pluton is older than 510 m.y.

The age of the latter pluton, as well as that of a similar smaller body from the Jiu Gorges, was also checked with the U-Pb method. However, the U-Pb concordia diagram (Fig.) must be handled with care. From the excellent linear fitting of the eight discordant points, a cogenetic relationship between the two granites results. Theoretically, the upper interception age (610 ± 30 m.y.) may represent either the average age of the crustal rocks which have generated these granitic melts by anatexis, or the true emplacement age, if there are no significant amounts of inherited radiogenic lead, that is the granites are primary. The first possibility is ruled out because the Lainici-Păiuş host rocks (and evidently the underlying ones) are older than 665 m.y., the oldest K-Ar age of a granite crosscutting them. The second possibility is favoured by the 0 m.y. lower interception age, by the perfect euhedral habitus of the zircon crystals, and by the high petrographic resemblances of the analysed granites with the Tismana granites, for which trace elements and mineralogical data enabled Berza (1978) to advocate a mantle origin.

4. A few samples of Mesozoic low-grade metasedimentary rocks from the Danubian Nappes show K-Ar ages between 90 and 131 m.y. In the slates from the northern Godeanu Mts (Branul River), Jurassic palynomorphs were reported by Gherasi et al. (1973), so extraneous $^{40}Ar_{rad}$ must be invoked for sample 72 (table 3), if the low-grade metamorphism of the rocks is Austrian or Laramian. The 94 m.y. age of the Alpine peridotite sample from the Severin Nappe in the Mehedinţi Plateau is in good agreement with its Upper Jurassic or Lower Cretaceous intrusion age and Laramic underthrusting.

REFERENCES

Bercia I., Hârtopanu I. (1980) Domaines à basse pression dans la série de Sebeş-Lotru (Précambrien de la nappe gétique). *An. Inst. geol. geofiz.*, LVII, 297–304, Bucureşti.

B e r z a T. (1978) Masivul granitoid Tismana. *An. Inst. geol. geofiz.*, LIII, 5—176, București.

— K r ä u t n e r H., D i m i t r e s c u R. (1983) Nappe Structure in the Danubian Window of the Central South Carpathians. XII *Carp.-Balk. Geol. Congr.*, București.

G h e r a s i N., V i s a r i o n A., Z i m m e r m a n n P. (1973) Considerații asupra vîrstei unor șisturi cristaline și depozite sedimentare din autohtonul danubian, situate la marginea de nord a munților Godeanu (Carpații Meridionali). *Stud. cerc. geol., geofiz., geogr., seria Geologie*, 18, 2, 303—310, București.

K r ä u t n e r H. (1980) Precambrian Unconformity in the Getic Nappe (South Carpathians). *An. Inst. geol. geofiz.*, LVII, 305—324, București.

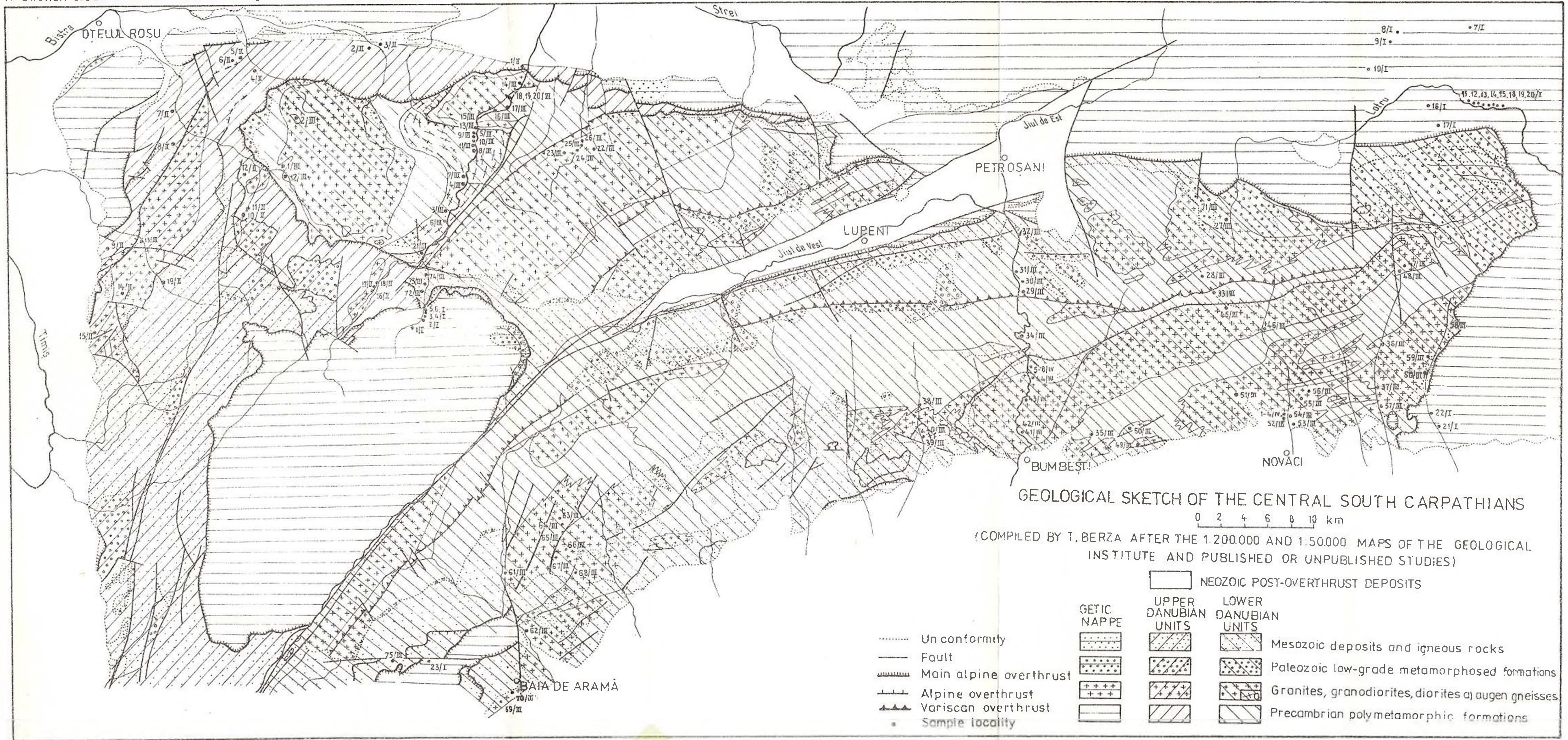
— N ă s t ă s e a n u S., B e r z a T., S t ă n o i u I., I a n c u V. (1981) Guide-book t. the A 1 Excursion. XII *Carp.-Balk. Geol. Congr.*, București.

L e e D. E., M a r v i n R. F., S t e r n T. W., P e t e r m a n Z. E. (1970) Modifications of Potassium-Argon ages by Tertiary thrusting in the Snake Range, White Pine County, Nevada. *U. S. Geological Survey Prof. Pap.* 700-D, 92—102.

M i n z a t u S., L e m n e M., V i j d e a E., T ă n ă s e s c u, A., I o n c i c ă M., T i e p a c I. (1975) Date geocronologice obținute pentru formațiuni cristalofiliene și masive eruptive din România. *D. S. Inst. geol. geofiz.*, LXI, 5, 85—111, București.

S o r o i u M., P o p e s c u G., G h e r a s i N., A r s e n e s c u V., Z i m m e r m a n n Po (1970) K-Ar Dating by Neutron Activation of some Igneous and Metamorphic Rocks from the Southern Branch of the Romanian Carpathians. *Ecl. Geol. Helv.*, 63, 1, 323—334.

— P o p e s c u Gh., K a s p e r U., D i m i t r e s c u R. (1972) Notă preliminară asupra geocronologiei cristalinului danubian. *Anal. St. Univ. Iași (serie nouă) Sect. II, b, geologie*, XVIII, 135—137, Iași.



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THE STRUCTURE – METAMORPHISM RELATION AND ITS METALLOGENETIC SIGNIFICANCE IN THE SOUTH CARPATHIANS [1]

BY

ANDREI GURĂU[2]

This paper is a synthesis dealing with the results of the microtectonic studies obtained by the author during several years on the crystalline schists in the South Carpathians.

Lately the idea of the succession of the tectonic and metamorphic phases of the crystalline schists in the South Carpathians has been largely acknowledged. The metamorphism and superposed structural phenomena are characteristic of the petrogenesis and structogenesis of the structural units of the South Carpathians. The superposition of the tectonometamorphic phases brought about essential modifications also within the internal structure of the rocks. These modifications manifest through the tendency of finding new mineralogical-petrographic equilibria accompanied by younger structural elements (foliations and lineations); the latter coexist with the older ones which tend to be obliterated.

The index mineral assemblage belonging to the various metamorphic generations together with the superposed structural elements (foliations and secondary lineations) constitute a clear evidence of the succession of the tectonometamorphic phenomena. The neoformation minerals as well as the secondary structural elements show characteristics which distinguish them from the primary generations of minerals and structures. Thus, neoformation minerals, such as muscovite, chlorite, chloritoid, epidote, zoisite and the recrystallized quartz, show an euhedral to anhedral outline and are reoriented in the secondary foliation plane. In the sections perpendicular to the secondary foliation plane the neoformation minerals are oriented at random. These characteristics of the neoformation minerals and of the secondary foliations clearly prove the existence of a direct relation between structure and metamorphism.

Microtectonic studies demonstrated that the dispersion phenomena of the L_1 mineral lineations, of the B_1 intersections and of the stratification

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schistosities correspond to the retromorphism of the index-primary minerals (biotite, almandine, staurolite, kyanite, sillimanite) (G u r ă u, 1980, 1981).

Secondary structural elements. The cleavage schistosities, the shearing cleavages and the fracture cleavages are characterized by their plane-parallel morphology on regional distances, while the B_2 intersections of these foliations exhibit a coaxiality with a homogeneous dispersion grade round the π pole.

The dispersion of the L_1 mineral lineations is usually achieved in two principal planes coinciding with the S_2 shearing cleavages or with the S_3 fracture cleavages. Field measurements and projections in the stereographic network with equal surfaces of L_1 lineations from various crystalline series of the Getic Domain and the Autochthon indicate that the dispersion phenomenon is characteristic of all the series of the crystalline massifs, showing a greater or smaller development ; this fact leads to the conclusion that this phenomenon manifests on a regional scale (Pl.), being probably caused by the Alpine diastrophism (G u r ă u, 1980 ; 1981). This conclusion is strengthened by the fact that the position of the migration planes of the L_1 primary lineations coincides with that of the S_2 shearing cleavages or of the S_3 fracture cleavages.

The cleavages are in their turn parallel to the surface of the Alpine overthrust nappes and other Alpine tectonic lines (Godeanu, Poiana Ruscă, Semenic, Lotru-Sebeș, Căpățînii, Făgăraș, Iezer-Păpușa).

The migration of the L_1 mineral lineations and of the S_1-S_2 intersection lineations confirms the block structure of the crystalline schists of the South Carpathians that have been rotated with respect to one another owing to the movement on the horizontal, subhorizontal or inclined faults. The K-Ar age determinations also account for the Alpine age of the neoformation minerals of the metamorphic units in the South Carpathians. Thus S o r o i u et al. (1970) obtained ages for the series of the Getic metamorphic rocks from the Godeanu Mountains that decrease to values between 120-160 m.y., 104–190 m.y. for the gneissic Petreanu Series ; 115-128 m.y. for the Rîușoru Series ; 103-278 m.y. for the Rof Series ; 91-283 m.y. for the amphibolites of the Măru Series ; 92-126 m.y. for the Zeicani Series.

Microscopic, microtectonic and radiometric age data plead for the simultaneity of the secondary foliations, the migration of the L_1 primary lineations and of the intersection lineations as well as for the Alpine retromorphism of the crystalline schists. The Alpine metamorphism included not only the crystalline schists but also the Mesozoic sedimentary formations (M r a z e c, 1904 ; S t r e c k e i s e n, 1934 ; G h e r a s i, 1937 ; S o l o m o n, 1968 ; G u r ă u, Ș e r b ă n e s c u , 1972 ; G h e r a s i et al., 1973), which reached the metamorphism grade of the chloritoid and chlorite zones (the Schela Formation, the Paroșeni Formation).

Based on systematic structural researches the following morphologic types of Alpine superposed structures could be separated :

1. *Superposed structures of the order I.* These structures are related with deformations leading to the secondary folding of the B_1 primary folds and are of the transversal ($B_2 \perp B_1$) or oblique ($B_2 \wedge B_1$) fold type.

The transversal folds ($B_2 \perp B_1$) are characterized by the presence of some sharp folds of Chevron type on the S_1 surfaces, parallel to the flow lineations of the B_1 primary folds. The position of the deformation ellipsoid of the transversal folds is changed.

These structures show the tectonograms of the S_1 (S_0) foliations forming a continuous belt with two superposed transversal dispersions corresponding to the two limbs of the longitudinal fold.

The L_1 mineral lineations are plotted on a small circle or ellipse.

Oblique folds ($B_2 \wedge B_1$). These folds are characterized by the fact that they form an angle smaller than 90° to the L_1 lineation. The folds exhibit a concentric and sinusoidal, sometimes, slightly sharp shape. These folds are typical of the overthrust zones and form as a result of the slip at various rates of the tectonic blocks parallel to overthrust surfaces. They occur on either side of the overthrust zone in the incompetent formations. The axes of these folds are disposed perpendicular to the slipping direction of the tectonic blocks.

The L_1 mineral lineations are plotted on a large circle.

2. *Superposed structures of the order II*. These structures are related with superposed cleavage foliations and the corresponding faults, intersecting the direction of the primary folds. These folds can be of the a-type — S_2 axial cleavage folds or fracture folds and of the b-type — S_3 transversal cleavage folds. They can be pointed out by the statistic study of the L_1 mineral lineations and the migration of these lineations to the plane of fracture systems parallel to the S_2 or S_3 cleavages.

Statistic analyses of these lineations revealed that their stereographic projections are plotted parallel to the projection of the cleavage planes, which confirms the idea that the migration of lineations is closely related to the formation of cleavages.

The migration of the L_1 mineral lineations to a system of subhorizontal shearing planes is characteristic of the epimetamorphic rocks in the north-eastern part of the Poiana Ruscă massif, the Getic metamorphic rocks from the Sebeș-Lotru Mountains as well as the Getic metamorphic rocks from the Godeanu Mountain, where the S_3 transposition foliations develop on a regional scale.

The migration of the mineral lineations in the plane of the S_3 foliations is typical of the metamorphic rocks of the Semenic Mountains and of the mesometamorphic rocks of the Tîncova-Nădrag Series (Strîmba Bach Valley).

Another element characteristic of the superposed structures of the order II, typical of the structure of the metamorphic rocks in the South Carpathians, is a system of younger faults trending EW and intersecting the direction of the overthrust and of the migration planes of the metamorphic lineations in the plane of the S_2 and S_3 cleavages. The tectonic compartments exhibited a rotation movement with respect to these faults, determining the change of the position of the migration planes of the lineations on either side of the faults.

3. *Superposed structures of the order III.* These structures represent a plicative deformation of the S_2 or S_3 secondary foliations after the formation of the latter. The folds generated as a result of these deformations can be of the secondary non-cylindrical conic folds type.

Owing to the conic position of the flanks of these folds, the poles of the secondary foliations are plotted more or less uniformly within a small circle, situated in the centre of the tectonograms. If the projection circle of the S_3 poles closes it indicates a complete secondary conic structure, and if it appears only as parts of an arc of a circle it represents only a fold fragment.

A peculiar feature of the crystalline schists in the South Carpathians consists in the fragmentary presence of some longitudinal primary folds, which are the consequence of the B_1 tectonics—bending with concentric slipping of the initial beds.

The primary folds are involved in the B_2 subsequent tectonics ; they are difficult to distinguish, appearing only as relics. Within the Getic metamorphic rocks the amplitude of these folds is of 4—5 m and the height is of 2—3 m. The axes of the B_1 folds are coaxial with the lineations of the B_1 intersection planes and migrate either as a result of a secondary plicative tectonics or due to the disjunctive deformations superposed on the primary and secondary faults.

Owing to this fact there is a preferential direction of regional character of the B_1 and even B_2 structures only within the limits of some subhorizontal blocks of reduced dimensions.

The very pronounced disjunctive tectonics in the Getic metamorphic rocks showing the S_2 shearing cleavages and the S_3 fracture cleavages gave rise to secondary cleavage folds disposed in virgation. The morphology of such folds was described by S a v u (1968) in the Getic metamorphic rocks of the Semenic Mountains, but it was assigned to the primary plicative structures.

The younger age of these folds is demonstrated by the coincidence of their "limbs" with the position of the migration planes of the L_1 lineations, parallel to the cleavage direction, and of the transposition foliation (G u r ă u, 1980 ; G r i d a n, 1980).

B_1 relict folds within the Ogradena Granitoids

The observations made on the Ieșelnița Valley revealed that the Ogradena Granitoids consist of a stratigraphic sequence of folded layers, consisting mainly of garnet micaschists, alternating with layers of quartz-feldspathic rocks of granitic or granodioritic composition ranging between 0.5-10 m in thickness. The careful observations on the structural elements of these rocks show that these layers have also undergone two plicative deformation phases, B_1 and B_2. The B_1 folds are obliterated by the B_2 folds which appear more clearly. These folds are sinusoidal, their amplitude being of 7—10 m, or even 20 m, reaching about 3—5 m in height. These rocks exhibit all the structural elements characteristic of the regionally metamorphosed sedimentary rocks.

The synmetamorphic structural elements are plotted in several planes. Also, the B_2 folds show various positions, lying in the shearing planes to which the L_1 lineations migrate as a result of the Alpine deformations. The B_2 folds are at angles of various sizes with the direction of lineations (5—10°, seldom to 90°), pointing to both their reduced plasticity and the various directions of the action of the tectonic forces.

Structural Facies

The diversity of the microtectonic structural elements in the crystalline schists of the South Carpathians enables the mapping of the latter according to the type of structural facies. Such a classification is also justified by the fact that the dynamic and regional metamorphism is manifested by complex plicative and disjunctive deformations, in addition to the progressive mineralogical facies, which are often accompanied by retromorphism phenomena.

The concept of "structural facies" should be as comprehensive as possible in order to suit the purpose of separating cartographically zones of structural facies specific for a certain dominant deformation type. Thus the classification of the structural facies should reflect the plastic or ruptural deformation stages of the crystalline schist terrains preserving the most characteristic forms of movements (homogeneous or nonhomogeneous) undergone by the rocks.

Under the classification we propose the structural facies include rocks of various metamorphic grades, while the mineralogical composition may correspond to one or several known mineralogical facies which underwent a predominant deformation type.

The name of "structural facies" designates an ensemble of synmetamorphic and postmetamorphic deformations, formed at certain structural levels under special geotectonic conditions, which are now in an equilibrium relation with the Alpine metamorphism grade of the deformed rocks.

Taking into account the genesis and morphological variety of the structural elements within the crystalline schists of the South Carpathians we made the following classification of the structural facies :

1. structural facies of the rocks showing a compact texture (unfoliated or slightly foliated);

2. structural facies of the metamorphic foliations :

a) — structural subfacies of the stratification schistosity ;

b) — structural subfacies of axial cleavage schistosity ;

3. ruptural structural facies of the (Alpine) postmetamorphic foliations :

a) — structural subfacies of the shearing cleavages ;

b) — structural subfacies of the fracture cleavages.

Structural facies of the rocks showing a compact texture

The rocks of this facies are marked by a large-grained size and are compact. The microtectonic structural elements, such as the L_1 lineations and the S_1 foliations, are partially or totally obliterated. They can be observed in some places at a close examination of the microstratigraphic boundaries which were preserved as relics.

The rocks of the Poneasca and Ogradena Granitoids might be of this facies, being generated by a complex anatexis and palyngenesis process (S a v u, 1971; A n a s t a s i u, 1976).

Within the Ogradena Granitoids the S_1 surfaces constitute the boundaries between the micaschists and granogneisses with which they alternate. At the same time the L_1 mineral lineations can be noticed especially on the surfaces of the S_2 shearing cleavages and, less clearly, on the S_1 surfaces. The microtectonic investigations carried out on these granitoids pointed out superposed structures manifested by the presence of B_2 folds and B_2 intersections which do not coincide with the position of the L_1 mineral lineations. The S_2 shearing cleavages of the B_2 folds show a high frequency which may lead to errors in the measurements of positions of the S_1 surfaces.

Structural facies of the synmetamorphic foliations

a) *Structural subfacies of the stratification schistosity.* The formations of this facies are characterized by the fact that the minerals formed during the regional metamorphism are oriented and flattened parallel to the S_0 initial stratigraphic plane, while from the tectonic point of view they show a development of the flexure-slip folds. This facies characterizes the mesometamorphic series of the Danubian Domain in the Cernei Mountains.

b) *Structural subfacies of the axial cleavage schistosity.* This subfacies is characteristic of the gneisses belonging to the Cumpăna Series from the Cumpăna-Holbav zone in the Făgăraș Mountains. The formations in this zone make up shearing folds, the limbs of which coincide with the axial cleavage schistosity. According to the definition of the structural levels proposed by D i m i t r e s c u (1967) these formations were assigned to the lower structural level by this author.

Ruptural structural facies of the Alpine foliations

The metamorphic formations belonging to this structural facies generally develop in the Alpine zones with superposed structures of the order I and II. These structures were achieved as a result of the mechanism of subvertical shearing and subhorizontal slipping, caused by the overthrusting and as a consequence of the movements leading to the uplift of the Carpathians at the boundaries of the tectonic microplates (S a v u, 1971; A i r i n e i, 1977). From the microtectonic point of view these structures are marked by a pronounced dispersion of the L_1 mineral lineations and of the S_1-S_2 intersection lineations in the plane of the shearing and fracture cleavages.

a) *Structural subfacies of the S_2 shearing cleavages.* This subfacies develops mainly in the metamorphic formations on the northern slope of the Făgăraș Mountains. It was found, from the west to the east, on the Tătarului and Neamțului Valleys, on the Scărișoara Mountain, on the Porumbacu, Bîlea, Albota, Arpașul Mare, Dejanilor Valleys. The frequency of this cleavage varies from 10 to 100 fissures per metre.

Another petrographic province that can be considered of this structural facies includes the Semenic Mountains, the Tincova-Nădrag Series as well as the metamorphic rocks of the Danubian Domain.

We mention that the boundaries between the granitoid massifs and the surrounding petrographic series are of tectonic nature; they coincide with subvertical systems of faults (hol) parallel to the S_2 shearing cleavages. The general trend of these cleavages is north-east-south-west, the same as that of the boundaries among the large tectonic blocks. The crystallization schistosity coincides with the initial stratification of the rocks and is cut by the S_2 shearing cleavage planes at subcentimetric distances.

b) *Structural subfacies of the S_3 fracture cleavages*. This subfacies is marked by the presence of cleavages of subhorizontal transposition. These cleavages appear at distances of a few centimetres of decimetres. This subfacies is characteristic of the epimetamorphic rocks in the north-eastern part of the Poiana Ruscă massif, the mesometamorphic rocks of the Godeanu Mountains as well as the metamorphic rocks of the Leaota Series from the Iezer-Păpușa massif (G u r ă u, 1980, 1981).

The L_1 mineral lineations migrate to a horizontal or subhorizontal plane that coincides with the statistic mean of the S_3 transposition foliations. The projection of the L_1 mineral lineations covers almost entirely the periphery of the stereographic network. Also, the poles of the S_3 surfaces are projected in the centre of the tectonograms.

The great dispersion of the L_1 mineral lineations and of the S_1-S_2 intersection lineations can be explained by the presence of some systems of horizontal and subhorizontal faults which separate tectonic blocks; the latter rotated parallel to these fault systems. Genetically there formed also drag folds as a result of the rotation of blocks.

This structural subfacies is characteristic of the Getic metamorphic rocks from the Lotru-Sebeș massif, the Căpățînii Mountains and the Bahna outlier.

Unlike the Poiana Ruscă Mountains and the Godeanu massif, the transposition faults are less frequent in these units (10-20 cm between the subhorizontal foliations).

Also, the dispersion angle of the L_1 mineral lineations is smaller, ranging between 45-120°.

The Metallogenetic Significance of the Structure-Metamorphism Relation

The knowledge of the structure-metamorphism relation provides a first image on the prospects of useful mineral substances of a region. Thus, in the terrains showing a structural facies of the synmetamorphic recrystallization schistosity the prospecting of both some volcano-sedimentary deposits, if the tectono-magmatic conditions permit it, and of some vein epigenetic deposits is possible.

Within the formations in the ruptural structural facies and the subfacies of the axial cleavage schistosity there is little probability of prospecting some primary stratiform metalliferous deposits of dimensions suitable for exploitation. Due to the very pronounced disjunctive tectonics at this structural level the primary deposits got destroyed. That is why we think that there is not much probability of prospecting some volcano-sedimentary deposits within the metamorphic rocks of the South Carpathians, in the zones of pronounced structural facies (Godeanu, Lotru-Sebeș, Făgăraș). Instead, veins controlled by systems of Alpine fractures are found in these regions.

Therefore we consider the structural facies method with the measurement and recording on maps of the linear structural elements of every genetic type to be very useful for the geological mapping of the crystalline schists; thus the extent of the subsequent tectonic movements is established and the structural indices characteristic of the prospection of certain genetic types of deposits are derived.

REFERENCES

Bercia I. (1975) Metamorfitele din partea centrală și de sud a masivului Godeanu. *Inst. geol. geof. Stud. tehn. Rom., seria I*. 12, București.

Gherasi N., Dimitrescu R. (1970) Structure des schistes cristallins du massif de Leaota. *Anal. Științ. Univ. Al. I. Cuza*, Iași.

Gurău A. (1980) On the regional dispersion of the mineral lineation L. in the crystalline schists from the southern Carpathians. *Rev. Roum. Geol. Geoph. Geogr., serie Geol.* București.

— (1981) Structural Facies in the Crystalline Fields of the Southern Carpathians and Their Importance for Geological Prospecting.

Hirtopanu I. (1978) Cristalinul Getic: Metamorfism polifazic sau polimetamorfism?. *Stud. cerc. geol., geofiz., geogr. seria Geologie*. 25/2, *Edit. Acad. R.S.R.*, București. *Rev. Roum. Geol. Geoph. Geogr., serie Geol.* București.

Savu H. (1965) Structura in virgație a Cristalinului munților Semenic. *D. S. Com. Geol.*, LI/1 a, București.

Soroiu M., Popescu G., Gherasi N., Arsenescu V., Zimmermann P. (1970) K-Ar Dating by Neutron Activation of Some Igneous and Metamorphic Rocks from the Southern Branch of the Romanian Carpathians. *Ecl. Geol. Helv.* 63, 1, Switzerland.

QUESTIONS

N. P. Shtcherbak: 1. Will you tell us whether your investigations included also radiologic absolute age determinations of the mineral assemblages? What are the results if any?

Answer: The absolute age determinations were carried out by: Soroiu, Popescu, Arsenescu, Gherasi, Zimmermann (1970). The absolute age for the Godeanu Crystalline ranges between 120—160 m.y.: for the gneissic "Petreanu" Series between 104—190 m.y.; for the "Riușorul" Series between 115—128 m.y.; for the "Rof" Series between 103—278 m.y.; for the amphibolitic "Măru" Series between 91—283 m.y. and for the "Zeicani" Series between 92—120 m.y.

Imprim. Atel. Inst. Geol., Geof.

THE POLYCYCLIC CHARACTER OF THE SOMEŞ SERIES METAMORPHICS IN THE WEST CARPATHIANS (ROMANIA)[1]

BY

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The polymetamorphism of the intensely metamorphosed crystalline units in the Carpathians appears as a natural phenomenon in light of stratigraphic-protistologic and geochronological data. Nevertheless, petrological and microtectonic arguments have not yet been given proper attention.

The present paper proposes to account for the polycyclicity of the Someş Series metamorphism on the basis of (1) the microtextural relationships in which phaneroblastic mineral components are implied; (2) the chemical changes during the blastesis of some of these components; (3) the analysis of microtectonic elements, a.s.o. The rocks examined are mainly metapelites, which abound in the area.

The Someş Series is the most intensely metamorphosed section of the Gilău Crystalline. To the west and to the east it borders on the granite block of the Muntele Mare. Its lithology is relatively homogeneous, consisting mainly of quartz-micaceous schists, micaschists and quartzites, as well as quartz-feldspar gneisses. Amphibolites, graphitic quartzites, crystalline limestone and spessartite rocks can be also found, but they have a subsidiary importance. The lithostratigraphic succession that has been set up starts with a metapelitic complex and by a migmatised metapelitic complex at the top.

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The Someş Series has been estimated to date from the Upper Precambrian, and the Muntele Mare granites from the Paleozoic.

The formations belonging to the Someş Series have been studied by several authors, the best known among them being D i m i t r e s c u (1958, 1966), G i u ş c ă et al. (1968), M â r z a (1969), M u r e ş a n (1980). In the framework of their papers they mention the bimetamorphic nature of these formations, implying especially the retrogressive chloritization overlapping an older regional metamorphism. At the point of contact with the Muntele Mare granite they unanimously acknowledge a thermic metamorphism.

Textural Relationships

In the present paper the study of the metamorphic processes will be made by taking into consideration the stages of deformation materialized in S planes and linear elements correlated with stages of mineral neoformations, or with relict minerals, and in connection with parageneses or more recent deformations.

The oldest mineral association includes kyanite, staurolite and garnet, which are also found as minerals of neoformations, albeit they have different characteristics. The first generations of kyanite and staurolite (d_1 and st_1) are represented by very minute crystals. They are randomly oriented with respect to any of the S planes noticeable at present, being chaotically piled up within some highly flat lenses. The xenomorphic nature and the deformed aspect of the kyanite and staurolite crystals are characteristic features, as well as the chemical distinctions between the central and marginal zone of individual crystals (at least in so far as the content of inclusions is concerned) (Fig. 1).

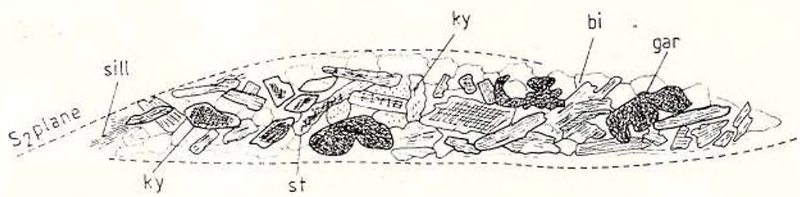

Fig. 1. — Polycrystals lens of kyanite, staurolite, garnet ± sillimanite in S_2 plane. sill—sillimanite; ky— kyanite; st— staurolite; bi— biotite; gar— garnet.

The first generation garnet is very clear or has very fine inclusions in the central zone, displaying a marked idiomorphism and suggesting a very slow growth toward the end of its formation. It has preserved its idiomorphism, by supergrowth, owing to the superposition of a synkinematic garnet different chemically, during a later stage, marked by an increased rate of growth (Fig. 2).

The deformation stage D_2, which individualizes the first generation of staurolite, kyanite and garnet with respect to time, materializes in an S plane with a highly penetrating character. At its level, massive neoformation of micaceous minerals and lens-like segregations occur. These lenses vary as to thickness (millimeters-decimeters) and their lengths are noticeable at outcrop, where they measure several meters (Fig. 3).

Quartz segregation can be correlated in time with the beginning of the crystallization of the second generation garnet. This can be determined on the basis of the great number of quartz inclusions with which its formation starts. In the same S plane, as a result of the D_2 deformation stage (S_2 plane), polymineral lenses measuring millimeters are also formed, which consist of kyanite and/or staurolite, quite often associated with

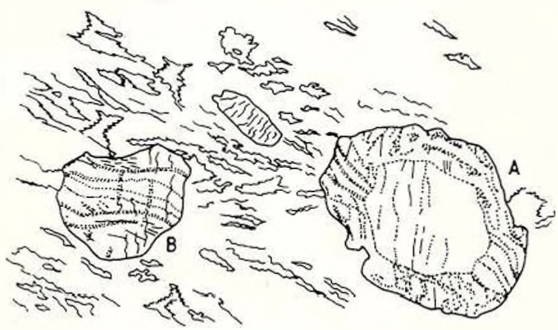

Fig. 2. — Optically zoned garnet (A) and younger unzoned garnet (B), from Someş Series.

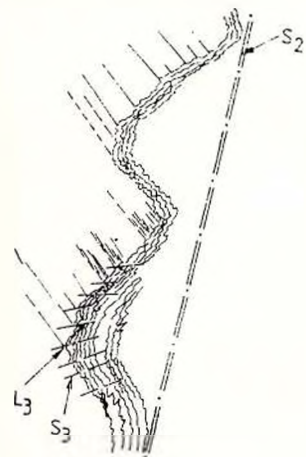

Fig. 3. — Superposed structures B_2 and B_3 in the Someş Series.

garnet. The S_2 plane also materializes by layering in quartz-fedspathic or amphibolite rocks, or by alternations of micaceous layers with quartzite or quartz-feldspathic ones in metapelitic rocks.

The pronounced penetrating nature of the S_2 deformation plane has had an especially strong deformation effect, giving rise to (1) notable discontinuities of initial stratification planes ($S_{0,1}$), (2) discontinuities of some rock blocks with restricted dimensions (amphibolites, quartzites) and (3) the crushing up of minerals with lengthened prismatic habit (the kyanite and staurolite in the above mentioned lenses). The penetrating nature of S_2 plane has also left its trace in a transposition effect at the level of this plane and the apparent orientation of all the rock blocks, at mesoscopic level, after S_2.

The mineral neoformation related to stage D_2 is predominantly micaceous, but also garnet and quartz-like, as noticed above. It is likely that the D_2 metamorphic stage has a postkinematic static component too, represented by the supergrowth of kyanite and staurolite from the lenses containing these minerals and, possibly, by individual crystals of kyanite and staurolite (Fig. 4).

A next deformation stage, D_3, materializes through a S_3 crenulation plane present everywhere in the area, characterized by a penetrating nature of variable intensity. In the zones where this characteristic is especially well represented, the crenulation schistosity is also very visible macroscopically, obliterating the foliation S_2. This is the reason why S_2 and S_3

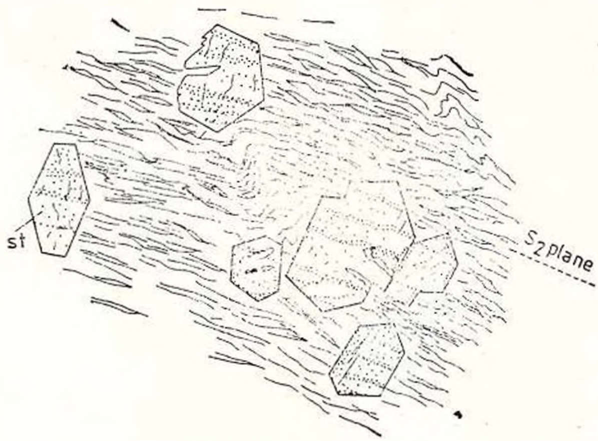

Fig. 4. — Staurolite postkinematic porphyroblasts grown post D_2-deformation. st-staurolite.

planes can be mistaken for each other, especially in the richly micaceous metapelites. The mineral neoformation is mainly muscovite-biotitic, but simultaneously, it can be chloritic, S_3 plane materializing through muscovite (quite often $\pm$ chloritized biotite) and/or chlorite. These minerals visibly intersect S_2 plane within the framework of S_3 plane (Fig. 5) which

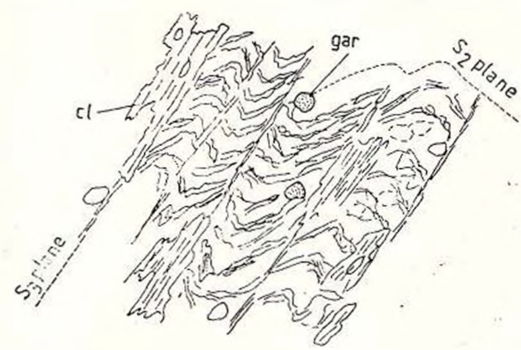

Fig. 5. — Crenulation S_3 plane, cutting S_2 plane. cl — chlorite; gar — garnet.

points to their subsequent character. A mineral neoformation stage related to D_3 stage, but having a postkinematic, static nature is represented by the random orientation of biotite and especially of chlorite in the rock texture.

Microtectonic Considerations

The measurements of mesoscopic planar and linear elements concerning exclusively S_2, S_3, L_2 and L_3. $S_{0.1}$, even when the measurements are taken at the point of contact between the contrasting lithological elements, represent the transposition effect of $S_{0.1}$ plane by S_2 only. The possibility of confusing S_2 and S_3 planes has led us to consider only those outcrops in which both planes were noticeable.

The statistical processing of the measurements of planar and linear elements was carried out through stereographic projection on Smith's network with equal surfaces, in the lower hemisphere.

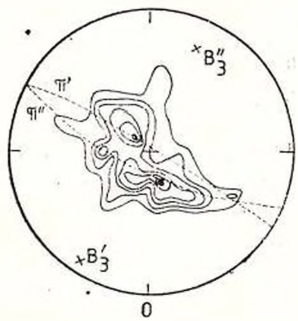

Fig. 6. — Stereogram representing planar elements S_3.12—10.5—7.5—4.5—3—1.5%

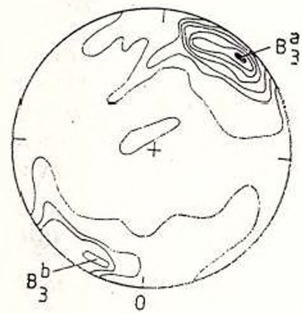

Fig. 7. — Stereogram of linear elements L_3. 14-10.5-8-5.5-4-2.5-1 %

The diagram of S_3 planes (Fig. 6) displays two maxima that indicate low inclination values of these planes. The manner in which the values of the spatial positions of S_3 planes center around the two maxima, seems to point out two distinct foliation populations represented in circles π' and π''; their poles are B_3' and B_3'', respectively. L_3 crenulation lineation diagram (Fig. 7) also indicates two main concentrations of these elements, likewise situated at very low inclination values. The positions of the two maxima (B_3^a and B_3^b) agree with the foliation poles of the π circles mentioned above (B_3', B_3'').

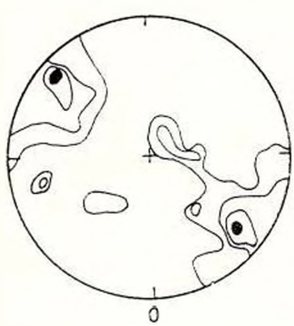

Fig. 8. — Stereogram representing planar elements S_2 10-6-4-2%

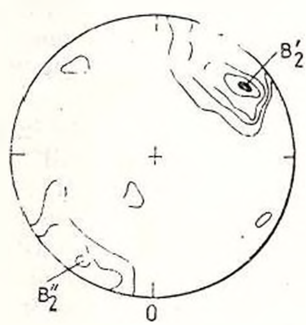

Fig. 9. — Stereogram of linear elements L_2. 15-11-6.5-4-2%

The diagram of S_2 planar elements (Fig.8) is characterized by high inclination values and constant direction of these planes. The linear elements L_2 (Fig. 9) concentrate primarily in the pole B_2' and to a lesser extent in B_2'', which coincide with B_3 poles. This coincidence is forseeable because the crenulation lineation is in fact an intersection lineation of crenulation plane S_3 with S_2 plane.

The analysis above reveals a sensible individuality of D_2 and D_3 deformational elements, manifest by the different position of poles of S_2 and S_3 planes. The subsequent character of structure B_3, as compared to B_2 also results from the fact that crenulation lineations L_3 affect the characteristic concentrical folds of deformational stage D_2. The existence of an angle, generally sharp, between the hinges of folds B_2 and lineations L_3, proves that they have formed as a result of a different strain. In contrast, the similar folds developed owing to deformation D_3 are perfectly identical in terms of spatial position with the crenulation lineations.

Chemical Zoning of Garnets

Microscopic observations have often indicated an optical zoning of garnets, as a result of the dissimilar frequency of inclusions along a transversal profile through the garnet crystal. From this point of view, it can be surmised that the global chemical structure of each zone changes at least with respect to the content of inclusions.

An electron microprobe analyser was implemented in determining the chemical analyses of regions within several of the garnets. It was determined that a cryptical chemical zoning corresponded to the observed optical zoning in these porphyroblasts. It was also shown that the chemical zoning is less pronounced when the metamorphism of garnet-bearing rocks is more intense, which is due to the homogenization of initially zoned garnets at the high temperature conditions.

We shall present below the analysis of a zoned garnet in the Someş Series. It consists of a clear central part with an idiomorphic contour. We consider it to belong to an initial stage of metamorphism. The beginning of the second zone is marked by quartz and indeterminate inclusions that border on the preceding zone. They are orderly disposed, S_1 displaying a sigmoidal shape. This arrangement testifies to the synkinematic character of the second zone.

The chemical analyses have been made with the help of an ETEC electronic microanalyser. All standards and standard chemical analyses are from the Mineral Constitution Laboratory at the Pennsylvania State University. The profile along which the analyses have been made is indicated in the adjacent figure (Fig. 10). Its length was 1.620 mm, and the step scan was 0.075 mm. The elements analysed and represented in the diagram as oxides are Fe, Ca, Mg, Mn and Ti.

All analyses are presented in terms of oxide mole percent with the FeO profile relying on a different ordinate scale. The PSU microprobe analyses with an approximate 3—4 micron diameter excitation area. Analytical uncertainties, were determined by X-ray counting statistics. Fluctuations in zoning profiles are probable artifacts of uncertainties, the interception of inclusion and/or possibly real. Nevertheless, general trends in chemical zoning can be determined. The large Ti content at 1.425 mm of the garnet in Figure 10 was produced by an ilmenite inclusion which was analysed with the garnet. The diagram of chemical content along the transversal profile reveals an obvious correspondence between the optical zoning and the one resulting from chemical fluctuations. The outer 0.3 mm of the garnet possesses abruptly reversed chemical profiles in the major cations, CaO displaying the most severe zoning. Manganese has a bell-shaped profile

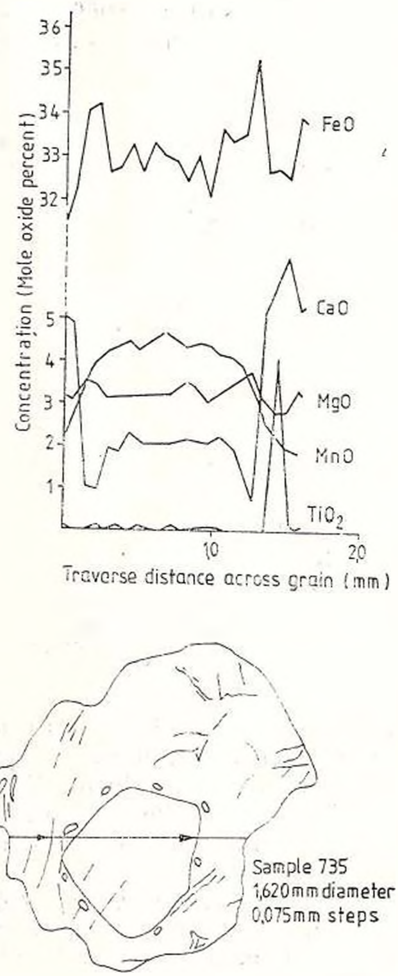

Fig. 10. — Compositional profiles in an optically zoned garnet

with simple symmetry as a result of a continuous drop in MnO during the garnet growth. But FeO, CaO and even MgO display complex symmetrical profiles, as a result of the discontinuous nature of the growth, which is commonly ascribed to polyphasic metamorphism (E d m u n d s and A t h e r t o n, 1971).

Along the transversal chemical profile one can notice a diminution in the Mn content in the same direction as the Fe increase, which produces a rise in the almandinic component to the detriment of the spessartinic one. This might indicate a general level of the rise in the physical conditions of metamorphism.

The diminution in the CaO content, or, in other words, the decline of the grossularite component, implies a temperature rise and a slight increase of pressure or a stationary level of the latter, a point on which we agree with Raheim and Green (1974).

Hollister (1969) has shown that the garnet MgO/FeO ratio rises in direct relationship with the metamorphic degree, suggesting

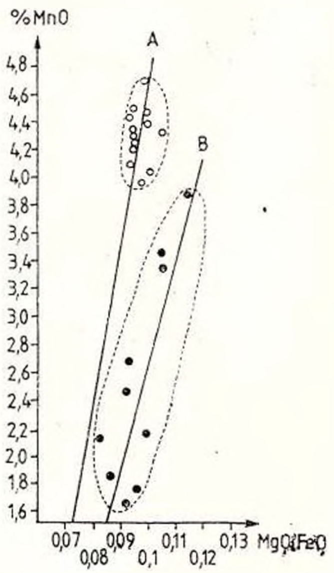

Fig. 11. — Plot of MnO versus MgO/FeO for zoned garnet (sample 735).

○ Values from core of garnet

● Values from edge of garnet

that direction A in Figure 11 represents high P and T conditions of regional metamorphism. Direction B represents the garnet growth under conditions of high temperature during the second stage of growth.

The garnet chemical profile is not complete, as it does not have the zone with fine inclusions and diffuse margins in the central part of the first stage garnet. It can be assumed that during the first period of its formation the rate of garnet growth was higher, but its growth conditions were static. The outer synkinematic growth zone, accompanied by a temperature rise, can be correlated with deformation D_2 and implicitly with the formation of foliation S_2. The rise of temperature in this period could be correlated with the thermic effect of Muntele Mare granite, which has a gneissic character in the marginal zone. The mineral neoformation in this stage was thus the result of the position with respect to the thermic source, with andalusite blastesis in the close vicinity of the granite, with sillimanite nucleation on the old kyanite and with the static rise of kyanite and staurolite in the more distant zones from the granite. It follows that the synkinematic sequence represents only a part of the thermic period generated by the granite or another source quasisynchronous with it.

Conclusions

The corroboration of the deformational elements with the textural relationships and with the chemical changes during the growth of the garnet have furnished us with arguments for distinguishing three important stages in the development of the Someş Series metamorphism. The first and oldest stage is characterized by the association of kyanite with staurolite and the first generation garnet. It has not visibly preserved linear and planar synchronous elements, these ones having been obliterated during stage D_2 in virtue of a marked transposition (mineral neoformation and reorientation) at the level of S_2 plane. The causes of the chemical changes in the garnet seem to be thermic as to nature. It is possible that this thermic surplus should represent an energetic effect of the strong deformation during stage D_2. Stage D_3 was characterized by the formation of S_3 planes with variably penetrating character, while the intensity of the physical factors of metamorphism and of the water excess diminished. This is the reason why mineral neoformation is represented especially by muscovite, biotite, chlorite — the last-named being the most common in the eastern part of the investigated area. The postkinematic crystallization of biotite and chlorite with respect to stage D_3, once again raises the question of a surplus energetic effect with respect to D_3 deformation.

The considerations concerning the evolution of physical conditions during these stages of metamorphism have been qualitative only. Unfortunately, we cannot ascribe the chemical changes to any temperature change, owing to the lack of the thermodynamic data available and the complex multicomponent zoning of the garnet.

REFERENCES

Dimitrescu R. (1958) Studiul geologic şi petrografic al regiunii dintre Girda şi Lupşa *An. Com. Geol.* XXXI, Bucureşti.

— (1966) Muntele Mare. Studiu geologic şi petrografic. *An. Com. Geol.* XXXV, Bucureşti.

Giuşcă D., Savu H., Borcoş M. (1967) Asupra stratigrafici şisturilor cristaline din Munţii Apuseni. *Stud. cerc. geol. geofiz. geogr., seria Geologie*, 12, 1, Bucureşti.

Edmunds W. M., Atherton M. P. (1971) Polymetamorphic evolution of garnet in the Fanad aureole, Donegal, Ireland. *Lithos*, 4.

Hollister L. S. (1969) Contact metamorphism in the Kwoiek area of British Columbia : an end member of two metamorphic processes. *Geol. Soc. Amer. Bull.*, 80.

Mârza I. (1969) Evoluţia unităţilor cristaline din sud-estul Muntelui Mare. Edit. Acad. RSR., Bucureşti.

Mureşan I. (1980) Geologia şi petrografia bordurii de nord-est a munţilor Gilău. Edit. Acad. RSR. Bucureşti.

Raheim A., Green D. H. (1974) Experimental determination of the temperature and pressure dependence of the Mg-Fe partition coefficient of garnet and clinopyroxene. *Contr. Mineral. Petrology*, 48.

ON THE AMPHIBOLITE FACIES METABASITES OF THE GEMERIDE PALEOZOIC (WEST CARPATHIANS)[1]

BY

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Introduction

The pre-Carboniferous complexes of the Central West Carpathians, regarded as Precambrian-Lower Paleozoic (?) are designated as the Tatrides and the Veporides (and/or Tatroveporides). Their metamorphic Variscan, and/or Assyntian (?) recrystallization reached the amphibolite facies conditions. Their intensity of metamorphic recrystallization is weaker towards the inner zones. Southwards of Margecany-Lubeník lineament, there occur Paleozoic complexes which form one of the innermost West Carpathian units-the Gemerides (Fig. 1).

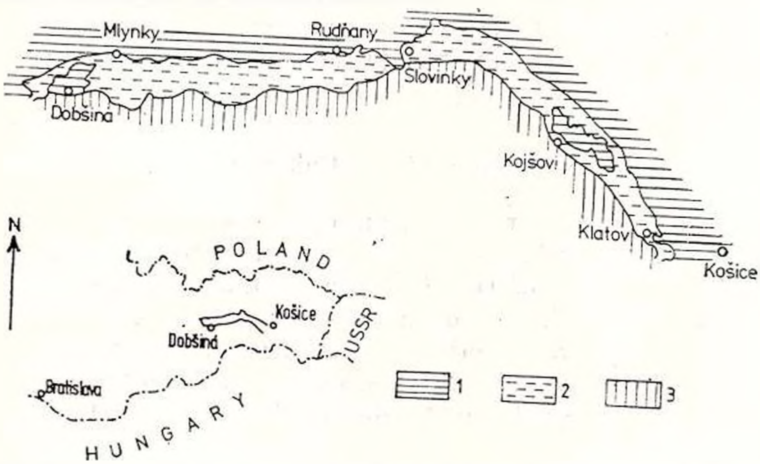

Fig. 1. — Scheme of the studied localities. 1, Upper Paleozoic and Mesozoic; 2, Rakovec Group; 3, Gelnica Group.

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The Gemeride Lower Paleozoic volcanic-sedimentary complexes are divided in the Gelnica and the Rakovec Groups. Opinions on the metamorphic recrystallization grade of these groups (the area northwards of Dobšiná is an exception to this ; R o z l o ž n i k, 1965 b) are so to speak identical — until now, all the authors have considered that it is likely that the mentioned groups belong to metamorphites of greenschist facies (K a m e n i c k ý, 1968, in M. M a h e í, T. B u d a y et al. 1968 ; V a r g a, 1973). However, esentially different appear the views on the age of the metamorphism. Some authors suggest the metamorphism to be Variscan (e.g. K a m e n i c k ý, l. c. ; R o z l o ž n i k, 1965 a ; B a j a n í k, 1968) others (e.g. V a r g a, 1973) assume the origin of metamorphic mineral assemblage to have been associated with the Alpine orogenesis.

Amphibolite bodies occurring in the Rakovec Group and partly in the Upper Carboniferous in the northern part of the Gemerides are the most appropriate for solving the age (Variscan, Alpine ?) and the grade of metamorphisms. The examination of the mentioned bodies was the objective and the results are summed up in this paper.

Metabasites : Geological Position and Characteristics

The first occurrences of amphibolite facies metabasites from the Gemeride Rakovec Group were described from Dobšiná by R o z l o ž n í k (1965 b), later on from the area of Klatov (D i a n i š k a, G r e c u l a, 1979). New occurrences have been discovered recently (B a j a n í k, H o v o r k a, in print). Concerning the Carboniferous, amphibolite facies metabasites can be traced only within the area of Rudňany (H o v o r k a et al., 1979 ; H o v o r k a, S p i š i a k, in print).

Amphibolites of the Rakovec Group

Within the west-east, trend, amphibolites are known to occur in the following areas : Dobšiná, Mlynky, Slovinky, Kojšov, Košická Belá-Bukovec. In the bodies of Dobšiná, Klatov and Rudňany also different types of gneisses were described together with amphibolites (biotite, amphibole-biotite, and garnet-biotite gneisses). Also crystalline limestones and metaquartzites were found in the mentioned areas. Within the metamorphites of the other areas only metabasites of different petrographic type can be traced.

Amphibolite facies metamorphites in the Gemeride of the Rakovec Group occur among the metamorphites of greenschist facies. The thickness of amphibolites is different, it varies from some ten meters to several hundred meters. The position of the amphibolite bodies is concordant with that of the surrounding metamorphites of the greenschist facies. In some cases tectonic contact seems probable.

The fine-grained (2 mm) foliated amphibolites with the mineral assemblage amphibole + plagioclase seem to be the prevailing amphibolite type. With increasing amphibole content they change into mela-amphibolites, locally also into foliated hornblendites. For amphibolites of all known bodies the presence of at least two generations of rock-forming minerals is characteristic: the first generation (plagioclase I + amphibole I) corresponds to amphibolite facies conditions; the second generation (epidote, amphibole II, plagioclase II, chlorite) originated in greenschist facies conditions. Amphiboles of different generation differ in composition — as confirmed by pleochroic colours; plagioclases II exhibit pronounced acidity (albites), whereas plagioclase I has the basicity An_{28-36}.

Upper Carboniferous Amphibolites

The bodies of amphibolite facies metabasites from the Carboniferous were discovered so far only in the area of Rudňany (H o v o r k a et al., 1979; H o v o r k a, S p i š i a k, in print). They appear to occur in concordant position with the neighbouring metapsamites and with schists containing large portion of organic material, and/or with metaconglomerates. Their concordant position with the mentioned layers, regarded as the Carboniferous only on the basis of their lithology, is confirmed by drilling and mining works. The thickness of the gneiss — amphibolite complex is 150-300 meters; direction length over 4 km. Fine-grained amphibolites are the dominant metabasite type of the locality under consideration; the garnet amphibolites are the characteristic types. Various types of paragneisses are intercalated in the amphibolites. Similarly as in other areas, apart from the Ist generation minerals, the presence of minerals — products of younger changes — can be observed. This mineral assemblage corresponds to pressure-temperature conditions of the greenschist facies (light-green long prismatic actinolitic amphibole a.o.).

Composition of Amphiboles

The analyses of coexisting amphiboles and garnets in metabasites were carried out using the electron microprobe ARL SEMQ. The results are listed in Table 1.

Viewing the criteria for classification of amphiboles (L e a k e, 1978), the amphiboles of metabasites under consideration are members of the calcic amphibole group. The diagram of the mentioned author (l.c.; Fig. 2) shows all projection point of analyzed amphiboles in the field VII "magnesio hornblende". A composition like this indicates higher pressure-temperature conditions than those of the greenschist facies. The latter ones provide the possibility for the following amphibole types to originate in metabasites: "actinolite", "tremolite", and/or "tremolite-actinolitic hornblende", and "ferroactinolitic hornblende" types.

The composition of metabasite amphiboles of the Rakovec Group and that of metabasite amphiboles of Upper Carboniferous of Rudňany area seem identical.

TABLE 1

Composition of amphiboles

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	45.74	44.47	44.61	47.05	45.82	45.46	46.01	44.74	43.34	42.74	45.80
TiO_2	1.31	1.34	1.41	1.05	1.23	1.58	1.60	1.37	1.95	1.78	0.77
Al_2O_3	10.86	11.20	8.89	10.92	9.56	9.86	8.66	11.58	10.39	10.69	9.88
FeO	16.65	16.17	17.37	13.55	15.62	17.96	15.82	14.04	15.92	16.60	14.29
MnO	0.26	0.27	0.31	0.21	0.24	0.26	0.17	0.07	0.30	0.24	0.14
MgO	11.71	11.70	10.82	12.88	11.19	10.29	11.40	12.23	11.42	9.77	12.26
CaO	10.06	10.33	10.74	10.81	10.87	10.15	10.10	10.11	10.39	10.55	10.05
Na_2O	1.29	1.66	1.33	1.70	1.36	1.44	1.30	1.78	1.70	1.35	1.45
K_2O	0.17	0.17	0.22	0.22	0.24	0.48	0.15	0.55	0.24	0.64	0.17
	98.05	97.31	95.50	98.39	96.13	97.48	95.21	96.46	95.61	94.36	94.80
Structural formulae on basis of 23 (0)											
Si	6.73	6.61	6.86	6.81	6.87	6.80	6.96	6.64	6.58	6.60	6.89
Al^{IV}	1.27	1.39	1.14	1.19	1.13	1.20	1.04	1.36	1.42	1.40	1.11
Al^{VI}	0.61	0.53	0.31	0.66	0.56	0.53	0.49	0.67	0.44	0.57	0.64
Ti	0.14	0.15	0.16	0.11	0·14	0.18	0.18	0.15	0.22	0.21	0.09
Mg	2.57	2.59	2.48	2.74	2.50	2.29	2.57	2.70	2.59	2.24	2.75
Fe^{2+}	1.68	1.73	2.03	1.45	1.80	2.02	1.76	1.48	1.75	1.98	1.52
Fe^{3+}	0.37	0.29	0.20	0.23	0.16	0.24	0.24	0.27	0.25	0.17	0.28
Mn	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.01	0.04	0.03	0.02
B Ca	1.59	1.64	1.77	1.66	1.74	1.62	1.62	1.61	1.67	1.74	1.62
B Na	0.01	0.04	0.02	0.12	0.07	0.09	0.11	0.12	0.04	0.06	0.08
A Na	0.36	0.35	0.39	0.36	0.33	0.32	0.27	0.39	0.46	0.35	0.34
A K	0.03	0.03	0.04	0.03	0.05	0.09	0.03	0.10	0.04	0.13	0.04
A Mg	0.61	0.60	0.55	0.66	0.58	0.53	0.59	0.65	0.60	0.53	0.64
$Mg + Fe^{2+}$	/n=1/	/n=5/	/n=4/	/n=7/	/n=4/	/n=3/	/n=3/	/n=2/	/n=5/	/n=6/	/n=3/

1–6 = amphiboles of the Rakovec Group metabasites;
7–11 = amphiboles of the Upper Carboniferous metabasites

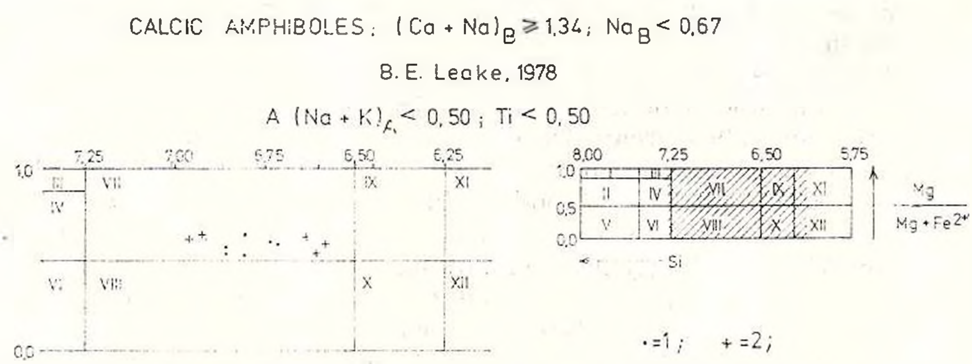
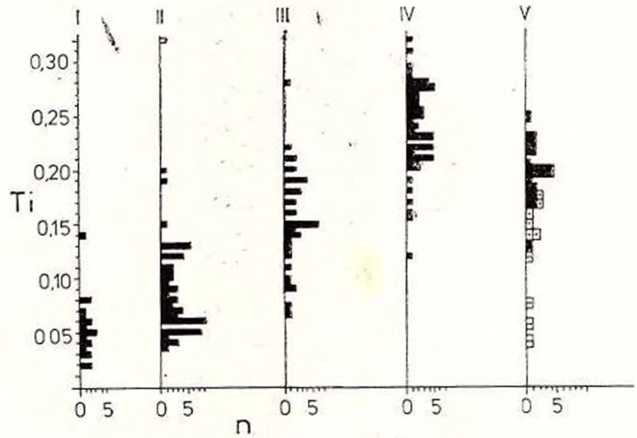


Fig. 2. — Projection of analysed amphiboles in B. E. L e a k e ' s (1978) diagram. 1, amphiboles of Rakovec Group metabasites (Š. B a j a n í k, D. H o v o r k a, in print); 2, amphiboles of Upper Carboniferous (D. H o v o r k a, J. S p i š i a k, in print).

The higher the temperature of the amphibole origin, the higher the titanium portion in amphiboles. R a a s e (1974) applied this phenomenon to graphic projection of Ti contents in metabasite amphiboles belonging to different facies. Analyses of Gemeride amphiboles correlate the Ti contents with those of the amphibolite facies (Fig. 3)

Fig. 3. — P. R a a s e's (1974) diagramatic representation of Ti contents of amphiboles.
I. greenschist facies; II. epidote amphibolite facies; III. amphibolite facies; IV. granulite facies; V. studied amphiboles: those of Rakovec Group (full squares) and of Upper Carboniferous (empty squares).



Composition of Plagioclases

The An molecule content in plagioclases appears to be one of the fundamental discriminatory criteria when ranging the metamorphosed rocks into certain metamorphic facies. Despite the intensive retrograde recrystallization of the metamorphic amphibole-plagioclase assemblage (products of the process are represented by light-green amphibole and plagioclase assemblage of the IInd generation) we have succeeded in finding plagioclases grains of the Ist generation appropriate for the optic study on universal stage. Simple, non-zonal structure is characteristic for plagioclase I. The grains show sporadically twinning according to the albite law. They are intensively pseudomorphosed by minerals of the epidote group and by other products of the saussuritization. The basicity obtained by means of standard optic methods varies within the range An_{28-36}. Except for plagioclases I, almost in every thin section there can be seen pure, small, frequently platy plagioclase II. Apart from grains in the rock matrix, they also fill veinlets and irregular nests. The basicity of this plagioclase is An_{4-8}.

Composition of Garnets

A homogeneous (non-zonal) chemical composition appears to be the characteristic feature of the analyzed garnets. The almandine molecule is dominant (Tab. 2). The projection points of the analyzed garnets from assumed Carboniferous metabasites of the Rudňany area (H o v o r-

TABLE 2

Composition of garnets (Rudňany area)

	I	II	V	I	II	V	
SiO_2/wt. %/	37.67	38.63	37.48	3.04	3.03	2.95	Si
TiO_2	0.11	0.07	0.17	0.01	0.005	0.01	Ti
Al_2O_3	21.03	21.77	21.55	2.00	2.01	1.99	Al
FeO	23.59	24.99	26.01	1.62	1.64	1.71	Fe^{2}
MnO	2.26	0.32	0.24	0.15	0.02	0.02	Mn
MgO	2.49	4.95	5.74	0.30	0.58	0.67	Mg
CaO	9.75	8.03	8.29	0.84	0.67	0.70	Ca
	96.90	98.76	99.43				
end members proportions :							
alm	55,50	56.30	55,53				
pyr	10.27	19.83	21.26				
spes	5,31	0.71	0,50				
gros	28,92	23,16	22,71				

Analyses of garnets from the paper D. H o v o r k a, J. S p i š i a k (in print).

k a, S p i š i a k, in print) form a field, which corresponds with fields introduced for metabasite garnets of amphibolite facies elsewhere in literature.

Metamorphic Conditions

Different criteria were applied to determine pressure-temperature conditions of metamorphic recrystallization, the rock textures, spatial relationship to adjacent rocks, composition of the principal metabasite minerals and some of their physical properties as well as the distribution coefficients of coexisting mineral pairs. The study of Mg, Fe and Mn distribution in the pairs garnet — amphibole and amphibole — plagioclase in amphibolites of the Rudňany area in concordance with procedures given by P e r c h u k, R y a b c h i k o v (1976) provided the following temperatures ranges :

$$K_{gar}^{Mg} : K_{hbl}^{Mg} = 510-630^{0}C$$

$$K_{hbl}^{Ca} : K_{plg}^{Ca} = 550^{0}C$$

Conclusions

Except for strong prevalence of greenschist facies metamorphites, some amphibolite facies metamorphites have been recognized in the Gemeride Paleozoic. They occur along the tectonic contact of this geological unit with northern zones of the Central West Carpathians. The occurrences form an intermittent belt. Field study and laboratory analyses revealed various types of paragneisses, crystalline limestones, metaquartzites and amphibolites. The latter ones were discussed in detail in this paper.

Apart from the coexisting minerals of the Ist generation (plagioclase I, amphibole I, garnet) almost in all amphibolite bodies under consideration the IInd generation minerals are present in non-equilibrium relation with those of the Ist generation. The IInd generation minerals concern strongly acid plagioclase (An_{4-8}), actinolitic light-green calcic amphibole, chlorite and others. They are the product of superposed (hydrothermal and metamorphic recrystallization) processes.

Analyzed amphiboles of the Ist generation belong to the calcic amphibole group, the "magnesio hornblende" type (L e a k e, 1978). Coexisting plagioclases show the basicity An_{28-36}. Garnet porphyroblasts (to 7 mm) are non-zonal and generally the almandine molecule is prevailing. These garnets have been found only in rocks of the Carboniferous from the Rudňany area.

The calculated distribution coefficients of coexisting pairs amphibole — garnet and plagioclase — amphibole together with the composition of minerals mentioned above, provide the evidence for their origin in amphibolite facies conditions by metamorphic recrystallization of basic volcanics and their volcanoclastics.

The presence of amphibolite facies amphibolites in the Rakovec Group (Middle Devonian — Lower Carboniferous?) suggests that zones of elevated heat flow have probably existed locally during the late Variscan folding and metamorphic processes. Nevertheless, we admit the schists with elevated content of "graphitic" substance could function as barrier in zones of elevated heat flow in the assumed Upper Carboniferous (?) pile. The metamorphic events are most probably of Permian (?) age.

REFERENCES

B a j a n i k S. (1978) Tektonická analyza rakoveckej série. *Zborn. geol. vied, Západné Karpaty*, 9,49—70, Bratislava.

— H o v o r k a D. (in print) Amphibolite facies metabasites of the Rakovec group (Lower Paleozoic, the Western Carpathians). *Geol. zbor. — geologica carpathica*, Bratislava.

D i a n i š k a I., G r e c u l a P. (1979) Amfibolitovo-rulový komplex ako súčesť ofiolitové suity rakovechého prikrovu. *Mineralia slovaca*, 11, 405—427, Bratislava.

H o v o r k a D., M i h a l o v J., O n d r e j k o v i č K. (1979) Metamorfity amfibolitovej fácie z oblasti Rudnian. *Mineralia slovaca*, 11, 481—504, Bratislava.

— S p i š i a k J. (in print) Coexisting garnets and amphiboles in metabasites of Rudňany area (Paleozoic, Spišskogemerské rudohorie Mts, the Western Carpathians). *Mineralia slovaca*. Bratislava.

K a m e n i c k ý J. (1968) in : M. Mahel, T. Buday et al., 1968 : Regional geology of Czechoslovakia. Part II. The West Carpathians. 3—723, Praha.

L e a k e B. E. (1978) Nomenclature of amphiboles. Min. Mag., 43, 533 — 563, London.

P e r c h u k L.L., R y a b c h i k o v I. D. (1976) Fazovyje sootnoschenija v mineralnych sistemach. 3—287, Moskva.

R a a s e P. (1974) Al and Ti contents of hornblende, indicators of pressure and temperature of regional metamorphism. *Contr. Mineral. Petrol.*, 45, 231—236, Berlin-New York.

R o z l o ž n i k L. (1965 a) Analýza štructúrno-metalogenetických elementov medzi Dobšinou a Mlynkami. *Zborn. geol. vied, rad ZK*, 4. 29—93, Bratislava.

— (1965 b) Petrografia granitizovanych hornin rakoveckej série v okoli Dobšinej. *Zborn. geol. vied. rad ZK*, 4, 95—144. Bratislava.

V a r g a I. (1973) Mineralnyje associaciji metamorfizma i ich zonalnost' v Spišsko-gemerskom rudogorie. *Mineralia slovaca*, 5, 115—134, Bratislava.

POLYCYCLIC DEFORMATIONS AND METAMORPHISM OF SOME CRYSTALLINE ROCKS OF THE SOUTH CARPATHIANS[1]

BY

VIORICA IANCU[2]

The study of the polymetamorphic formations raises several questions with regard to the possibilities of division, following the space distribution and correlation of the main lithostratigraphic units.

The establishing of the lithostratigraphic succession in the metamorphic fields meets with real difficulties due to the modification of the initial boundaries and unconformities by metamorphic differentiation and transposition. It is more difficult in case of the polycyclic formations, which present the cumulated effects of several major events, spatially superposed, of metamorphism, deformation, migmatization, magmatic intrusion.

When palaeontological and micropalaeontological data as well as conclusive radiogenic ages are missing the systematic approach of the elements of metamorphism and deformation can lead to important specifications.

Moreover, if we take no care of the mineralogical and petrological evidence as well as of the interrelations between metamorphism and deformation, the data obtained on age may be incorrectly interpreted.

The detailed research of certain metamorphic formations in the South Carpathians allowed the differentiation of several generations of micro- and mesostructures, which may be correlated with major metamorphic events. The predominance of a set of structural elements on large distances and thicknesses gives a false impression of oneness, homogeneity and uniformity. The preservation of certain relict elements, with a regional statistical frequency, constitutes the proof of palaeostructures preserved inside the main Alpine structures. Even if these structures cannot be easily followed in the field, their systematic distribution may lead to the whole interpretation due to the qualitative information offered by it. The high frequency and the regional distribution of the micro- and meso-

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structures is a proof of the fact that they belong to structures of higher amplitude which exceed the scale of the outcrop. Such evidence might influence the scepticism of certain researchers in connection with the use of the micro- and mesostructural data in the regional structural interpretation and correlation.

We consider that the carrying out of lithostratigraphic correlations in the polycyclic fields implies a field research accompanied by differentiated measurements, which have to take into account the regional distribution as well as the peculiar character of certain elements or zones of metamorphism and deformation.

The comparative study of metamorphics implying difficulties of separation (both lithostratigraphic and of delimitation of their structural systems) may be tackled by the correlation of the elements, as follows: metamorphism (parageneses, index minerals, metamorphic zonations, etc.), deformation (*S* planes, mineral lineations, structures and textures, fold generations, etc.), migmatization and associated magmatism.

This paper presents schematically the observation data based on differentiated measurements, accumulated during several years. The Proterozoic and Paleozoic metamorphics, to which we shall refer further on, occupy limited areas in the External Danubian Unit (lower, according to B e r z a et al., 1981, unpublished data) as well as in the outliers belonging to the Getic Nappe and the Supragetic Units.

A first lithostratigraphic correlation of the metamorphics in the Romanian Carpathians has been made by K r ä u t n e r (1980).

In the present paper a general scheme has been proposed, which contains the intervals of sedimentation and the main metamorphic and deformational events, previous to the deposition of the Variscan molasse (Upper Carboniferous-Permian) (Plate).

The main intervals of sedimentation which raise important questions concerning the stratigraphic unconformities are presented as two variants (I and II).

The same difficulties occur in the differentiation of the main metamorphic events, if the metamorphic and structural peculiarities of certain important "piles" of metamorphic rocks are not taken into account.

If our published and unpublished data are considered, the main metamorphic events took place in the time interval indicated in variant II. It leads to the possibility of estimating the intervals of sedimentation also according to variant II.

On the basis of the metamorphic events and the deformational ones an attempt has been made to delimit the main orogenic events which have affected the Proterozoic formations (pre-Grenvillian, Grenvillian, Assynthic and the Paleozoic formations (Caledonian and Variscan).

There are individualized metamorphic events, characterized by the superposition of the thermal maximum with the deformational one (regional thermodynamic metamorphism), and events in which the thermal metamorphism and the deformation take place successively. The relationship metamorphism-migmatization-magmatism has a complex character and requires special studies. While the main processes of metatectic migmatization correlate in time and space with the high-grade metamorphism (in which the anatexis isograde can be exceeded), the emplacement of the large bodies of magmatic rocks (granitoids) takes place with a significant shifting as against the place of formation of the magmas, generating contact metamorphism, unconformable structures, arteritic and metasomatic migmatization.

As the initial contacts between the polycyclic metamorphic formations are hard to be observed and followed, we have tried to individualize lithostratigraphic units (which might be delimited as supergroups) with specific metamorphic and deformational characters.

Supergroup A is represented by the oldest metamorphics, whose main characteristic feature is the presence of the effects of two events of thermodynamic metamorphism of the Barrovian type (medium pressure facial series). This group includes the metamorphics of the Drăgșan Group (Danubian Units), of the Sebeș-Lotru Group (Getic outliers), the Tilva Drenii Formation (augen gneisses), Valeapai Formation, and probably the mesometamorphic rocks in the Buziaș Region (Timiș Units).

The first metamorphism is proved by : relict minerals and paragenesis, relict S_1 planes and B_1 folds, relict mineral lineation (with E–W trend in the Bahna Outlier), deformed metatectic migmatites and associated, remetamorphosed eclogitoid (pyroxenitic) rocks.

The second metamorphism (M_2) represents the regional isofacial reorganization, with a different orientation of maximum stress. It is characterized by an advanced transposition generated by B_2, regional folding, associated with very penetrative S_2 foliation. Mineral adaptation to these new conditions occurred under the amphibole-almandine facies. A metamorphic zonality is known in the Semenic Mts (Savu, 1970) and the Godeanu Getic Outlier (Bercia, 1975), which can represent the first (M_1) or the second (M_2) metamorphism. We consider that the ages determined by Rb/Sr and U/Pb methods (850–1000 m.y.), cf. Bagdasarian, 1972, Pavelescu et al., 1979 (unpublished data) can represent this second metamorphism. Mineralogical evidence (Hârtopanu, 1978) and structural elements (Iancu, Hârtopanu, 1979, 1981, unpublished) show that the second metamorphism (M_2) had a regional character.

Supergroup A shows superposed, areal, repeated remetamorphism : Assynthic (low-pressure), Caledonian and Variscan.

M_3 (low-pressure) metamorphism occupies areas with regional extent, characterized by new isogrades related to a thermal dome (Hârtopanu, 1975), in Godeanu and Bahna Getic outliers.

The adaptation to these new P–T conditions was differentiated in relation whith the interception of new isogrades.

This metamorphism can represent the effects of the Assynthic cycle and shows a higher geothermal gradient. It can be correlated in time with the Assynthic (synkinematic) metamorphism of the Lainici-Păiuș Group.

No important retrogressive, regional transformations were observed in the rocks of the Getic outliers and the Tîlva Drenii region. Only some areas of the Supragetic units present rocks of Supergroup A with two main phases of retrogressive character, in greenschist facies conditions. Thus, the rocks of the Valeapai Formation show advanced transposition on new planes, marked by the greenschist facies paragenesis (Caledonian overprint). Later, they are, however, affected by deformations and recrystallizations, which can correlate with the Variscan events. In some cases areas with higher temperature paragenesis, related to the Paleozoic granitoid rocks (Sicheviţa, Cherbelezu, etc.), are known.

S u p e r g r o u p B represents the upper sequence of mesometamorphic rocks, with only one regional (prograde) thermodynamic metamorphism of the Barrovian type. Their M_1 event can be synchronous with the M_2 regional remetamorphism of Supergroup A.

This supergroup can include : the Lainici-Păiuş Group (belonging to an external, Lower Caledonian Unit, enclosed in the Danubian Alpine Units), the Jidoştiţa and Ivanu Formations (possibly synchronous with the Miniş Formation, in the Getic Units, and the Bîrzava Group, in the Supragetic Units.

Recognized metamorphic and deformational elements of these rocks show effects of a unique metamorphic event, of the Barrovian type.

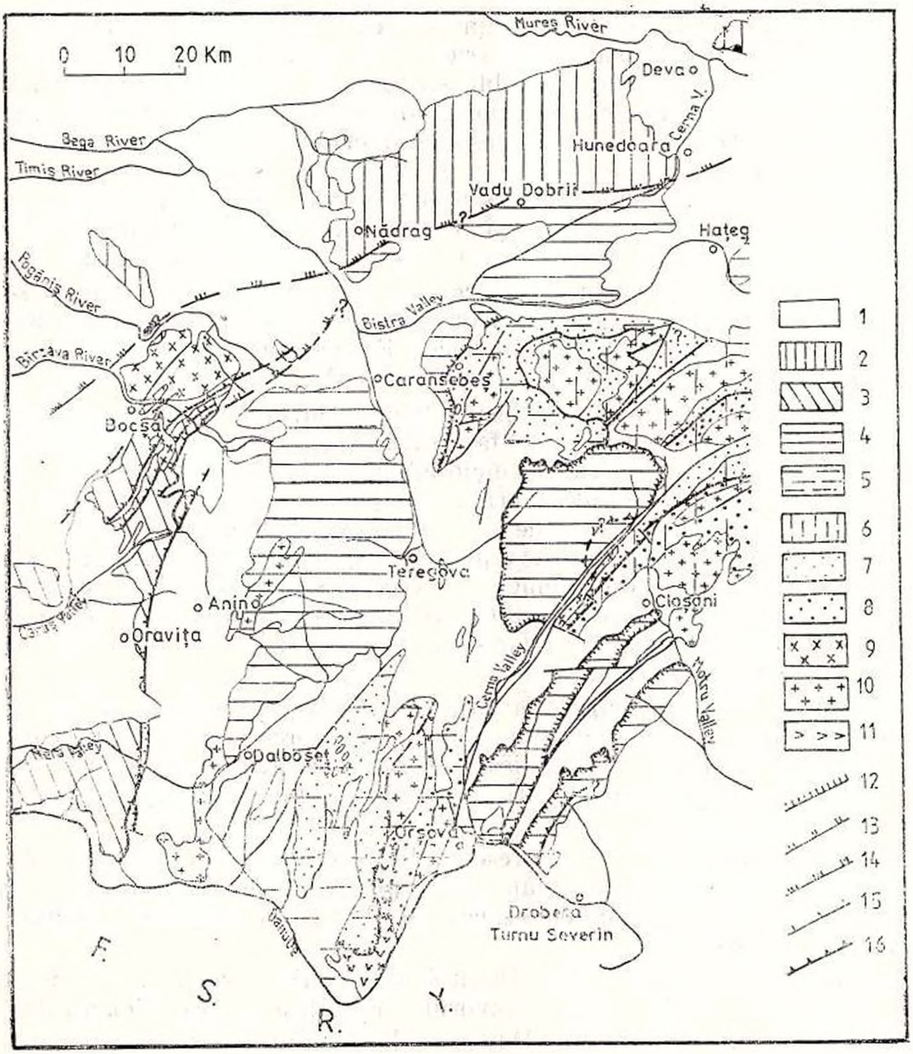
The first metamorphism (M_1) of Supergroup B can correlate with the regional metamorphism (M_2) of Supergroup A. The second metamorphism of Supergroup B is of low-pressure type, similar to the M_3 event of Supergroup A, probably representing the same main metamorphic event (Assynthic). The metamorphic-deformational conditions differ in the two main Caledonian Belts, separated by thrust (or cryptic suture) lines of the Caledonian System, from the Danubian Units (Fig.).

In the Upper Caledonian Unit (internal), the thermal maximum is not superposed on the deformational maximum being static or interkinematic metamorphism. In the Lower (external) Caledonian Unit the rocks of the Lainici-Păiuş Group exhibit M_1 relict metamorphism (I a n c u, unpublished data), followed by syndeformational low-pressure metamorphism (Danubian type) considered by S a v u (1972) and B e r z a (1978) as a unique event.

The radiogenic data indicate 610 m.y. for the associated Tismana type Granitoids (P a v e l e s c u et al., 1979, unpublished data). This age converges with the oldest ages obtained for the granitoids and the metamorphic rocks of the Danubian Realm (S o r o i u et al., 1970; M î n z a t u et al., 1975). Thus, the emplacement of synchronous granitoids and M_2 (low-pressure) metamorphism can be attributed to the Assynthic cycle. Some tardeorogenic granitoids, with contact metamorphic zones, can be considered younger (Early Caledonian).

The Caledonian overprint (M_3 of Supergroup B) produced areal retrogressive metamorphism of older rocks of the Danubian complex and prograde metamorphism (greenschist facies adaptation) of the system of basic dyke rocks.

The Bîrzava Group (Supragetic Units) has undergone regional remetamorphism, outlined by advanced, but incomplete transposition of new planes (marked by greenschist facies minerals). This event can be considered the effect of early Caledonian metamorphism on the basis of comparative studies.



Sketch map of SW part of the South Carpathians, after the geologic map of Romania, 1 : 1 000 000 modified.

1, Post-Lower Carboniferous cover; 2, Timiş (Upper Supragetic) Units; 3, Lower Supragetic Units; 4, Getic Units; 5, Internal (Upper) Danubian Units; 6, External (Lower) Danubian Units; Caledonian Units; 7, Caledonian (Internal) Unit; 8, Peri-Caledonian (External) Unit; 9, Alpine magmatic rocks; 10, Pre-Alpine granitoid rocks; 11, Pre-Alpine basic and ultrabasic rocks. Alpine tectonic lines; 12, Overthrust line of the Getic Nappe System; 13, Overthrust line of Lower Supragetic Units; 14, Overthrust line of Timiş Units (Senonian); 15. Undifferentiated Alpine thrust lines; 16, Pre-Alpine undifferentiated thrust lines.

The initial unconformities between rocks of Supergroups A and B are folded, transposed and obliterated by recrystallization (e.g. the contact between the Sebeş-Lotru and Jidoştiţa Formations).

S u p e r g r o u p C. The unique sequence that can represent the Lower Cambrian (by the geometric, lithostratigraphic position) is the Vodnic Group (I a n c u, unpublished data) in the Supragetic Unit of the Banat region. These rocks point to obvious stratigraphic and metamorphic discordance with the mesometamorphic rocks of the core of the Caledonian Vodnic Anticline. The "contact zone" is marked by commonly folded and metamorphosed rocks under greenschist facies conditions. An epimetamorphic sequence was partly separated (in this region) by C o n s t a n t i n o f (1980) as "Valea Caraşului" Ordovician-Silurian series.

In the Vodnic Group (Supergroup C) two formations can be separated: volcanogenous (greenschists with albite porphyroblasts), Rafnic Formation, and terrigenous, Dognecea Formation. They are overlain by epimetamorphic quartzites (probably of Ordovician age).

The prograde metamorphism (M_1) or Supergroup C is proved by oriented minerals of greenschist facies degree (chlorite, epidote, muscovite, albite, actinolite), sometimes enclosed in tardekinematic albite porphyroblasts. Internal (si) and external (se) planes are in this case in continuity and represented by the same minerals. Corresponding blastesis in mesometamorphic basement is represented syn-S_2 minerals (chlorite, albite, epidote). The older metamorphic porphyroclasts are enclosed in an oriented metamorphic matrix; relict minerals (garnet, biotite, plagioclase) are deformed by Caledonian and Variscan events. Plagioclase shows polystage zones of recrystallization.

The Variscan overprint of Supergroup C is represented by the deformation of first epimetamorphic minerals. The pre-existing albite porphyroblasts were rotated and the micaceous minerals were in places reoriented. Microfolds with partly penetrative new planes and static minerals can also be seen.

S u p e r g r o u p D represents Paleozoic, slightly metamorphosed rocks (post-Lower Cambrian and pre-Upper Carboniferous); the Ordovician-Silurian sequence requires special studies concerning the type and age of metamorphism.

In the External (Lower) Danubian Unit, these rocks were separated by S t ă n o i u (1972): Valea Izvorului Formation (Ordovician-Silurian) and Poiana Mică Formation (Devonian-Lower Carboniferous).

The sequences of very low-metamorphosed rocks can be separated on the Bîrzava Valley (Valea Satului volcano-sedimentary Formation and Cîrşie metapsephitic-metapelitic Formation). The age of these formations (Devonian and Lower Carboniferous, respectively) was determined by microfloristic associations (V i s a r i o n, unpublished data).

A general correlation can be made with the Lescoviţa Series (M a i e r, 1979), which is likely to include older rocks, too.

The rocks of supergroup D are characterized by the well-preserved bedding (S_0) and ŭnuniform, very low-grade metamorphism (low greenschist facies monometamorphism). They are associated with magmatic rocks (dykes) affected by Variscan metamorphism and deformation.

S_1 planes (metamorphic minerals) are generally perpendicular to the bedding and partly penetrative. The visible planes, at a mesoscopic scale, are S_0 (bedding) and S_1 (axial plane cleavages).

The rocks of the Devonian-Lower Carboniferous sequence preserve sedimentary and magmatic (premetamorphic) structures. Metamorphic adaptation has been differentiated according to : mineralogical and petrochemical constitution, relative thickness of beds, amplitude of folds, and "density" of the penetrative planes.

These rocks show an incipient, predominantly mechanic transposition, associated with folding and shear planes.

The Tilva Mare quartzites, with uncertain tectonic position present special metamorphic features. The quartzites of this formation show elements of two deformational events, pre-Variscan and Variscan and seem to overlie unconformably the rocks of the Vodnic Group. Their probable age may be Ordovician.

The zones with Paleozoic and Proterozoic metamorphics, for which deformational and metamorphic elements have been discussed, belong to the External (Lower) Danubian Unit, the Getic Nappe and the Supragetic Units.

The sketch on the map of SRR (scale 1 : 1,000,000) (Fig.) presents two Caledonian Units affected by Danubian Alpine Nappes. We consider that the blastomylonitic (shear) zone between the Lainici-Păiuș Group and the Drăgșan Group represents an old tectonic (cryptic suture line) at the contact of an internal Caledonian Belt and an external, peri-Caledonian Belt.

In front of this line there is a dyke-system of basic rocks related to a distensional phase. In the Upper Danubian Unit, a similar blastomylonitic zone was mentioned by Mărunțiu, Seghedi (unpublished data), associated with basic and ultra-basic pre-Alpine rocks, under the Mraconia (pre-Carboniferous) line. This blastomylonitic zone can represent the more internal correspondent of the blastomylonitic zone of the External (Lower) Danubian Unit.

The sketch also presents new data considering the Supragetic Units (nappes), separated firstly by Streckeisen (1934). The front of these nappes thrusts over the sedimentary (Upper Paleozoic-Mesozoic) cover of the Getic Nappe. The Banat Zone of the Supragetic Units constitutes an independent area with metamorphosed Paleozoic formations, different from the synchronous formations in the Poiana Ruscă Massif as concerns both the lithofacial features (internal and external facies, according to Maier, 1979) and the tectonic position. Further arguments for the individualization of a group of internal units (Timiș Units) in relation to the Supragetic Units (*sensū stricto*) are offered by the relations between the strongly laminated metamorphics thrusting over Mesozoic formations, Jurassic-Upper Cretaceous, in the right side the Pogăniș Valley.

The comparative study of the Proterozoic and Paleozoic formations, as well as the relations with the Paleo-Mesozoic formations led to the individualization of two large Supragetic Units (groups of units). The Supragetic Unit *(sensu stricto)*, which appears in the Banat region, also emerges again from under the Făgăraș Subunit on the Olt Valley (the Călimănești and probably the Cîineni Units, according to H a n n, S z a s z, unpublished data) and in the Leaota Unit.

The innermost group (Timiș Units) appears from under the Neogene and Miocene deposits (agglomerates, tuffs with a level of Lumachelle limestones) in the crystalline islands at Valeapai and Buziaș and continues probably with the "Poiana Ruscă Unit" (K r ä u t n e r et al., 1978, unpublished data). We consider that the Paleozoic in the Poiana Ruscă Massif, essentially different in lithostratigraphic respect from the Banat one, represents the Paleozoic of a large internal unit, which contains also the Timiș, Poiana Ruscă and Făgăraș "Units".

I should like to express my thanks to my colleagues and to Dr. K r ä u t n e r for the possibility to see several regional profiles in the South Carpathians within the Working Group for the Precambian.

REFERENCES

B a g d a s a r i a n P. G. (1972) Despre vîrsta absolută a unor roci eruptive și metamorfice din masivul Ditrău și Munții Banatului din România. *Stud. cerc. geol., geogr., seria Geologie,* 17, 1, p. 13—21, București.

B e r c i a I. (1975) Metamorfitele din partea centrală și de sud a masivului Godeanu. *Stud. tehn. econ., Inst. geol. geofiz., seria I,* 12, p. 3—159, București.

B e r z a T. (1978) Considerații privind vîrsta formațiunilor cristalofiliene și a masivelor granitoide din infrastructura părții externe a autohtonului danubian (Carpații Meridionali). *Acad. R.S.R., Stud. cerc. geol. geofiz., seria Geologie,* 23, 2, p. 173—184, București.

C o n s t a n t i n o f f D. (1980) Complexul banatitic de la Oravița-Ciclova (rezumatul tezei de doctorat). Univ. București.

H â r t o p a n u I. (1975) Le métamorphisme de basse pression dans les Monts Mehedinți. *D. S. Inst. geol. geofiz.,* LXI, 1, p. 217—238, București.

— (1978) Cristalinul getic : metamorfism polifazic sau polimetamorfism? *Acad. RSR. Stud. cerc. geol., geofiz., geogr., seria Geologie,* 23, 2, p. 180—193, București.

I a n c u V., H â r t o p a n u I. (1979) Successive deformations and superposed structures in the crystalline rocks of the Mehedinți Mountains. *Rev. roum. géol., géophys., géogr., série Géologie,* 23, 1, București.

K r ä u t n e r H. G. (1980) Lithostratigraphic correlation of Precambrian in the Romanian Carpathians. *An. Inst. geol. geofiz.,* LII, p. 229—296, București.

M a i e r O. W. (1979) The pre-Alpine metamorphosed formations from the Supragetic Unit of Banat (Romania). *Rev. roum. géol. géophys., géogr., série Géologie,* 23, 2, p. 137—147, București.

M î n z a t u S., L e m n e M., V î j d e a E., T ă n ă s e s c u A., I o n c i c ă M., T i e p a c I. (1978) Date geocronologice obținute pentru formațiuni cristalofiliene și masive eruptive din România. *D. S. Inst. geol. geofiz.,* LXI/5, p. 85, București.

Savu H. (1970) Stratigrafia și izogradele de metamorfism din provincia metamorfică prebaicaliană din Munții Semenic. *An. Inst. geol.*, XXXVIII, p. 223–311, București.

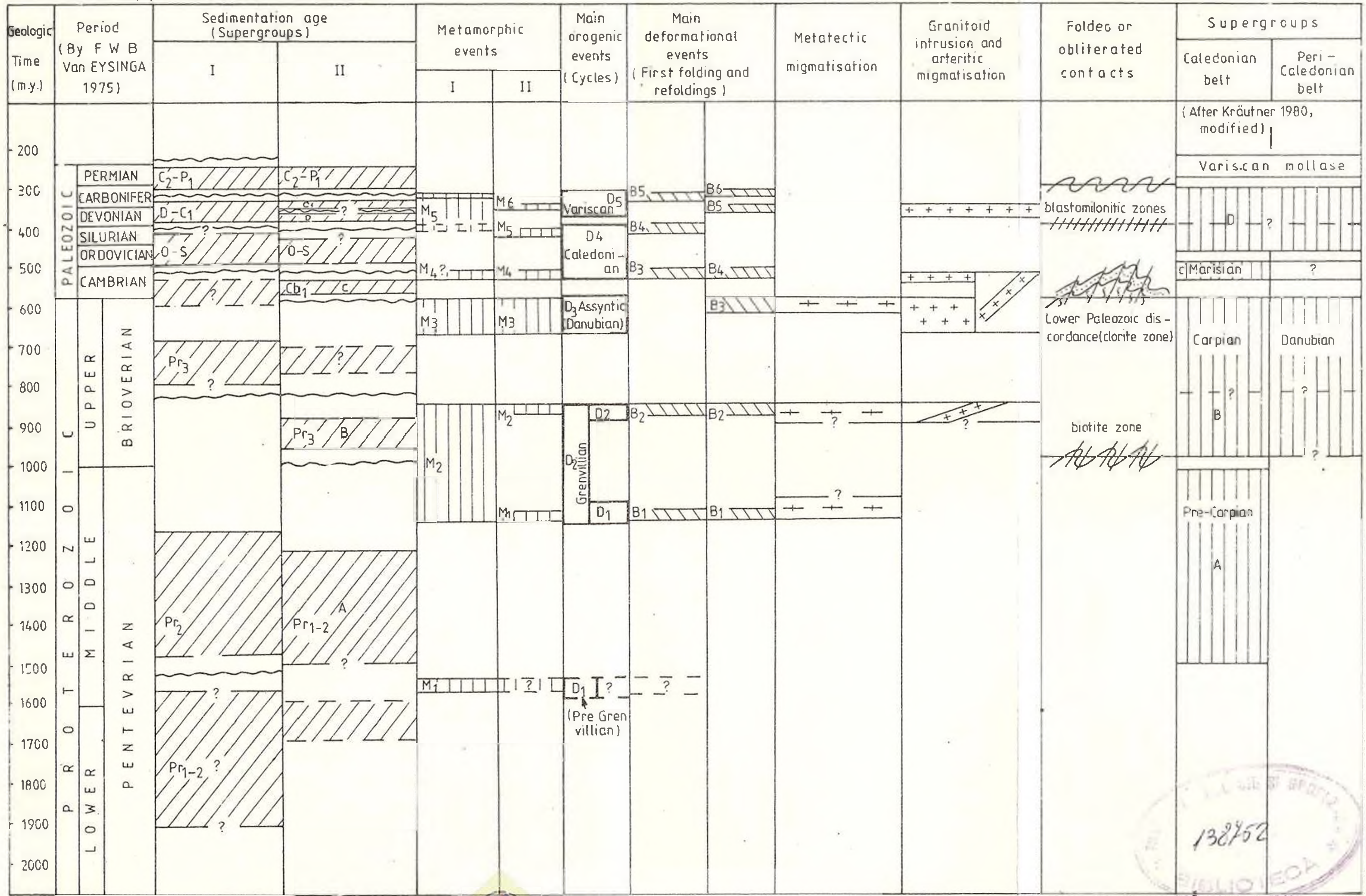
— (1972) Relația de timp și spațiu dintre cutarea seriei de Lainici-Păiuș și intruziunea granitoidelor tardeorogene din regiunea Cărpiniș-Novaci (Munții Paring). *D. S. Inst. geol.*, LVIII/5. p. 97–123, București.

Soroiu M., Popescu G., Gherasi N., Arsenescu V., Zimmermann P. (1970) K-Ar Dating by Neutron Activation of Some Igneous and Metamorphic Rocks from the Southern Branch of the Romanian Carpathians. *Ecl. Geol., Helv.*, 63/1, p. 323.

Stănoiu I. (1972) Încercare de reconstituire a succesiunii Paleozoicului din partea de est a autochthonului danubian, cu privire specială asupra regiunii de la obirșia văii Motru (Carpații Meridionali). *D. S. Inst. geol.*, LVIII/3, p. 57–71, București.

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

MAIN METAMORPHIC AND DEFORMATIONAL PRE-ALPINE EVENTS (PRESUMABLE SKETCH)



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PRECAMBRIAN CORRELATION IN THE CARPATHO-BALKAN BELT [1]

BY

H. G. KRÄUTNER [2]

In younger mobile zones the correlation of the Precambrian becomes more difficult owing to: 1) fragmentary exposures in nappe structures subsequently divided by fault systems and internal sedimentary basins; 2) overprint by younger orogeneses with metamorphic events. Therefore attempts of Precambrian Correlation in such areas can be only speculative and it is thus that the present model has to be considered.

This correlation attempt covers the Alpine belt including the West, East and South Carpathians, the Apuseni Mountains and the Balkan Zone. For the Romanian Carpathians the correlation model proposed by Kräutner (1980 a) was used, resulting from his own researches as well as from the compilation of literature data and field trips made with the Romanian Working Group for Precambrian. For the Soviet and Slovakian Carpathians and for the Balkan Zone data have been considered from the relevant literature, from the co-operation in the ambit of the IGCP Project 22, the PKG IX, as well as from field trips in 1971 in the West Carpathians under the guidance of Prof. Dr. B. Cambel, Dr. L. Kamenicky, Dr. I. Roikovici and in 1979 in Bulgaria under the leadership of Dr. D. Kozhoukharov, Dr. E. Kozhoukharova and Prof. Verghilov.

I would like to express my gratitude to the mentioned Bulgarian, Slovakian and to my Romanian colleagues for their highly competent guidance and useful discussions.

Principles

1) There are various ways of attempting the correlation of the Precambrian considering lithostratigraphy, biostratigraphy, magmatic, metallogenetic and metamorphic events, history of deformations, angular

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and metamorphic unconformities, geochemistry etc. The models which are in concordance with most of the mentioned ways and do not follow a single correlation line are more reliable.

2) Similar sequences of semblable lithostratigraphic rock assemblages were considered as the principal element of correlation. This means that single formations, even if considered as characteristic, were not taken into consideration for correlation; confidence was admitted only for identical sequences of two or more similar formations.

3) The present attempt refers to long distance correlations between the main tectonic units of the belt. It is considered that inside these structural units at least the partly known lithostratigraphic sequences can be taken as representative for the respective zones.

Tectonic Setting

Roughly following Săndulescu's (1975, 1980) geotectonic model and having in mind the zones in which Precambrian rocks are widely exposed, only the following main structural units will be considered in our correlation attempt (Fig. 1).

1. Tatro-Veporide Nappes without details referring to their complicated tectonic structure and constitution due to the lack of lithostratigraphic details for the Precambrian Jaraba *s. s.*, Kokava *s. s.*, "Series".
2. Gemeride Nappes including the Gelnica Group.
3. Codru and Biharia Nappe Systems with the Bihor Group (Muncel Formation, Biharia Formation) and the Baia de Arieș Group (including also the Vidolm-Lunca "Series", Mădrizești Formation).
4. Bihor Autochthon with references only to the Someș (Gilău) Group.
5. Central East Carpathian Nappes and the Făgăraș Nappe of the Eastern Supragetic Units including the Bretila (Belipotok), Rebra (Lower Delovetsk), Făgăraș and Cumpăna Groups as well as the Tulgheș (Upper Delovetsk) Group.
6. Getic Nappe with the Sebeș-Lotru Group and the Cibin Group (Sibișel, Dăbîca, Miniș, Buceava "Series").
7. Upper Danubian Nappes including the Zeicani Group, the Jelova "Series" and the Poiana Mraconia Formation).
8. Lower Danubian Units with the Drăgșan Group and the Lainici-Păiuș Group.
9. Rhodope Massif consisting of the Prarhodopian (Ograzdenian) Group, the Rhodopian Group and the Diabase-Phyllitoid Group.
10. Serbo-Macedonian Massif including the "Lower Metamorphic Complex" and the Vlasina Group.
11. Pelagonian Massif with references to the whole known sequence.

References for the lithostratigraphic sequences and the sense in which the mentioned lithostratigraphic terminology is used in this paper are indicated in the Plate.

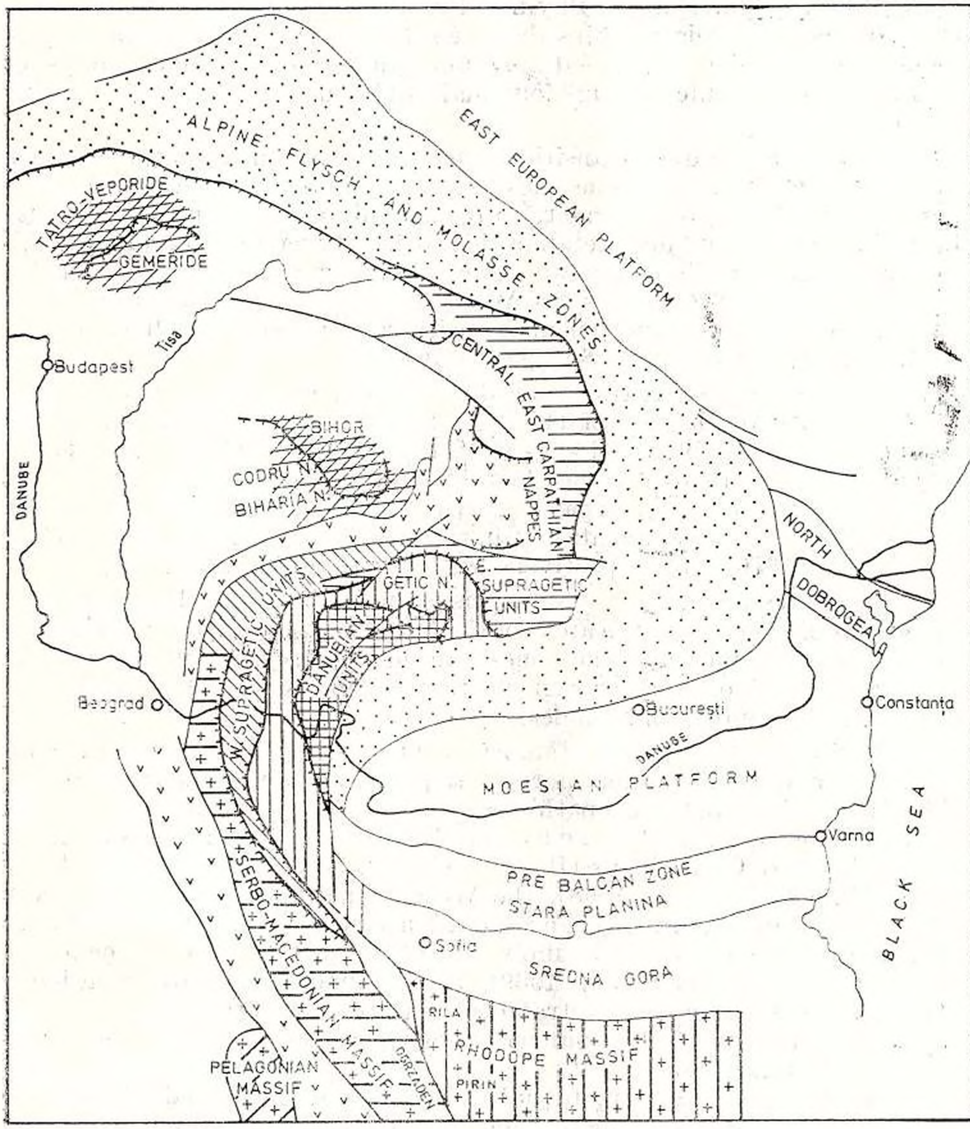

Fig 1. — Carptho-Balkan Belt.
The main structural units with Precambrian rocks.

Arguments

Over the whole area low and medium metamorphic rock assemblages do appear, deriving from sedimentary piles of quite different lithological sequences. Most of the low metamorphic sequences have been assigned to the Ordovician-Lower Carboniferous or to the Mesozoic by paleontological evidence in the last 15 years. We will therefore refer only to the pre-Variscan (pre-Ordovician and/or Silurian) sequences, namely the Gelnica, Tulgheş, Bihor, Cibin, Diabase-Phyllitoid and Vlasina Groups. As concerns the non-tectonic relationships between these low grade metamorphic sequences and the mentioned medium grade metamorphic groups or "series" (see tectonic setting) four main opinions may be found in literature.

1. Lateral gradual transition due to a single metamorphism with different intensity in a unique pile of sediments. This idea developed especially in the Alps (fide Exner, 1978) according to the supposition that both the medium and low metamorphic rocks derived from Paleozoic sedimentary educts. Indirectly it was admitted also for the East Carpathians (Kräutner, 1938) and for the West Carpathians by the assumption of a mainly Variscan metamorphism of the Tatro-Veporide crystalline (Cambel, Veselsky, 1980; Siegel, 1980).

2. Vertical gradual transition due to the decrease of the metamorphism intensity to the top of the pile. This model was admitted in the South Carpathians by Ghika-Budeşti (1940), Gherasi et al. (1966), Savu (1970), in the Apuseni Mountains by Dimitrescu (1966) and in the East Carpathians by Boiko (1970). Continuity or discontinuity of the sedimentation in the primary pile have been admitted by Savu (1970) and respectively Maier et al. (1975).

3. Regressive (poly) metamorphism due to a regional overprint in the greenschist facies. Large areas of the Carpathians, described in the past as prograde low metamorphic zones, are now considered by most of the authors as zones of younger regional retromorphism on rocks which were initially in the amphibolite facies.

4. Sedimentary and metamorphic unconformity. For the Precambrian low and medium metamorphic sequences this relationship was considered in the South Carpathians (Codarcea-Dessila, 1965; Giuşcă et al. , 1969; Bercia, Bercia, 1975; Kräutner, 1980), the East Carpathians (Bercia et al., 1976), the Apuseni Mountains (Giuşcă et al., 1968), the West Carpathians (Kamenicky, 1980), the Rhodope Massif (Kozhoukharov et al., 1978, 1980), the Pelagonian Massif (Stojanov, 1980) and the Serbo-Macedonian Massif (Dimitrijevici, 1969). In this model the above-mentioned (point 3) zones of regional retromorphism are considered as an effect of the metamorphism of the younger Precambrian sediments in their older metamorphic basement.

It can be proved that for all the four mentioned models there are examples of convenient application in the Carpatho-Balkan Belt, but as concerns the Cibin, Vlasina and the Diabase-Phyllitoid Groups only model 4 is in concordance with the available petrologic, radiometric, palynological data and the field evidence. (For the Gelnica, Tulgheş and Bihor Groups only tectonic contacts are known). This means that we have to suppose that Precambrian unconformities separate low and medium metamorphic piles formed by different metamorphic events on sedimentary educts of different ages.

Now the following problems arise : 1) whether these unconformities, accepted or supposed in different areas, represent fragments of a single wide extended plane of discontinuity or not and 2) whether the rock sequences situated under or over the mentioned unconformities correspond to the same cycles of sedimentation and metamorphism or not.

Low grade metamorphic sequences overlying the unconformities. In the six mentioned groups two main types of rock assemblages may be recognized :

a) Blastodetrital sequences with intercalations of rhyolitic metavolcanics, graphite formations with metalydites and limestones, subordinately greenschists. The Tulgheș Group (Tg), the Gelnica Group (Ge) and the Muncel Formation belong to this type. Possible correlations of these sequences have been mentioned by Visarion, Dimitrescu (1971), Kräutner (1980), Kamenicky (1980). This is proved by lithostratigraphy and metamorphism for Tg, Ge, Mu, by metallogeny for Tg, Ge (stratiform sulfide deposits linked to the rhyolitic volcanism and syngenetic manganese deposits associated with the metalydites), by Cambrian age, indicated by radiometric Pb—Pb, U—Pb data (Boiko et al., 1975 ; Vîjdea, Anastase, 1975) and palynology (Iliescu et al., 1972, 1981 ; Visarion, unpublished data ; Snopcova, Snopco fide Bajanic, 1980). Recently also Ordovician-Lower Devonian assemblages have been mentioned (fide Grecula et al., 1981) in rocks ascribed to the Gelnica Group. K/Ar ages reach maximum 475 m.y. (Kräutner et al., 1976), but in general the values are lower due to Ar loss in the Variscan and Alpine events.

b) Blastodetrital sequences with albite porphyroblast schists (metagreywacke and/or metatuffites) with intercalated greenschists and local thin beds of dolomites (and limestones). This type of rock assemblage includes the Biharia Formation (Bi), the Cibin (Cb), Diabase-Phyllitoid (DFF), Vlasina (Vl) Groups and probably the Lerești Formation (Le) of the Leaota-Iezer-Păpușa Mts (South Carpathians). Possible correlations of these groups have been partly mentioned by Dimitrijevici et al. (1967), Giușcă et al. (1969), Visarion, Dimitrescu (1971), Kamenicky (1980), Kräutner (1980). These correlations are in concordance with some lithostratigraphic details for Bi, Le, and scarce palynological data for Bi, Cb, Vl, suggesting an Upper Proterozoic age.

Normal relationships between rock assemblages of type *a* and *b* were reported only from the Apuseni Mountains, where Dimitrescu et al. (1974) described the Muncel "Series" overlying the Biharia "Series" without discontinuity. No other areas in which rock assemblages of types *a* and *b* appear together are known.

In conclusion the available data are consistent with the idea that the groups Bi, Cb, Vl, DFF and respectively the sequences Tg, Ge, Mu may be correlated. If we accept the normal relationships reported from the Apuseni Mountains all the six low metamorphic sequences over the mentioned unconformities may belong to the same sedimentary cycle, with two successive stages of development corresponding to the rock assemblages of types *a* and *b*. In any case the discontinuities separating the groups Cb, Vl, DFF from their higher metamorphic basement may be supposed to belong to the same wide extended unconformity.

Medium grade metamorphic sequences underlying the unconformities. Only lithostratigraphic and metamorphic correlation lines may be followed for these groups as palynological data are scarce and not confident for the age, metallogeny developed only locally and the radiometric ages (K/Ar, partly also Rb/Sr) evidently indicate younger events. Correlation attempts for the Romanian Carpathians (K r ä u t n e r, 1980; K r ä u t n e r et al., 1981) indicated that most of the medium metamorphic groups (Rebra, Bretila, Sebeș-Lotru, Zeicani, Drăgșan, Lainici-Păiuș, Ielova, Baia de Arieș, Someș) may be considered as fragmentary preserved parts of a single sequence (Carpian Supergroup) constituted of six formations.

In the Carpatho-Balkan Belt a similar sequence of formations was reported only from the Rhodope Massif (Rhodopian "complex") by K o z h o u k h a r o v et al. (1978, 1980) in spite of the fact that B e r c i a and M a i e r, M ă r u n ț i u (unpublished data) mentioned some petrographic similarities between the Prarhodopian and Sebeș-Lotru crystallines. Semblable palynological assemblages were reported from the Rhodopian and Carpian sequences by K o z h o u k h a r o v, T i m o f e e v (1980), I l i e s c u, M u r e ș a n (1972). The metamorphic similarities consist in the great thickness of the amphibolite facies assemblages (over 8,000 m) in the kyanite zone and in an increase of migmatization at the lower parts (K o z h o u k h a r o v, K o z h o u k h a r o v a, 1977; K r ä u t n e r, 1980). Polymetamorphic characters (B e r c i a, 1975; H â r t o p a n u, 1975, 1978; B a l i n t o n i, G h e u c a, 1977) cannot be considered as conclusive for lithostratigraphic correlations because they are linked probably to younger events in respect of the object of correlation.

Lithostratigraphic correlations of the Rhodopian Group of the Rhodope, Rila, Pirin and Ograzden Mountains were performed by K o z h o u k h a r o v et al. (1978, 1980). The extension of this Rhodopian sequence to the Serbo-Macedonian and Pelagonian Massifs was supposed by K o z h o u k h a r o v and S t o j a n o v (oral communication). Correlations of the Pelagonian sequence with the Serbo-Macedonian and Rhodopian piles were discussed by A r s o v s k i et al. (1977) and D i m i t r i j e v i c i (1969).

Concerning the Prarhodopian Group, it is considered as a lower lithostratigraphic unit ("complex"), separated from the overlying Rhodopian Group by a sedimentary and metamorphic unconformity (K o z h o u k h a r o v et al., 1978, 1980). Neither similar sequences nor a possible equivalent of the mentioned unconformity have been reported so far from the Carpathians, but their possible existence cannot be excluded as for large areas no lithostratigraphic details are available (e.g., Sebeș-Lotru Mountains, Tatro-Veporide Nappes). D i m i t r i j e v i c i (1969) supposed that a part of the Serbo-Macedonian "Lower Complex" may represent an equivalent of the Prarhodopian Group.

For the Tatro-Veporide medium metamorphic and retromorphic rocks no detailed lithostratigraphic sequences were reported. The rock assemblages of the Jaraba *s.s.* and Kokava *s.s.* Groups are very similar to that of the Carpian Bretila, Cumpăna and Rebra, Făgăraş Groups of the Carpathians. This is why Kamenicky (1980) supposes a possible correlation of the Jaraba and Bretila "Series".

In conclusion there are no available data against a correlation of the Carpian, Rhodopian and Tatro-Veporide sequences. Therefore one can suppose that the mentioned piles were formed during the same cycle of sedimentation and metamorphism. According to the Bulgarian geologists the Prarhodopian sequence represents the product of an older Precambrian cycle.

A Correlation Model

Considering the mentioned possible correlations a three cycle model may be supposed for the pre-Variscan metamorphic sequences of the Carpatho-Balkan Belt. According to it the pre-Variscan metamorphic rocks may be included in three main lithostratigraphic sequences separated by two Precambrian sedimentary and metamorphic unconformities. The post-cycle history of metamorphism and deformation may be different in the same sequence occurring in different structural units.

The Prarhodopian (Ograzdenian) sequence was reported only from the Rhodope, Rila and Ograzden Massifs (Kozhoukharov et al., 1978, 1980). It consists of a lower gneissic formation (A_1), a median varied formation with amphibolites and gneisses (A_2) and an upper gneissic formation (A_3). The sequence is highly migmatized, at least two pre-Rhodopian migmatite phases are supposed (Kozhoukharov et al., 1978, 1980).

The Prarhodopian sequence was assigned to the Archean or Lower Proterozoic (Kozhoukharov et al., 1978, 1980). A Lower Proterozoic age seems probable as no evidence exists for the Archean in the European Alpine Belt and no lithologic similarities with the Archean of the foreland may be recognized.

The Rhodopian-Carpian sequence is widely exposed in the whole Belt. It probably includes also the Jaraba *s.s.* and Kokava *s.s.* Groups for which no lithostratigraphic data are available. According to the correlation suggested in Figure 2 and in the Plate — the following common sequence of formations may be accepted for the Carpian (Kräutner, 1980) and Rhodopian (Kozhoukharov et al., 1978, 1980).

1. Amphibolite formation with gneisses, limestones and thin beds of micaschists ($Pt_{1.1}$., Cp_1)[3]. Some small syngenetic iron ores are known in the Sebeş-Lotru Group and in the East Rhodopes.

2. Lower gneiss formation, prevailing biotite or biotite-muscovite plagiogneisses, partly with stronger migmatization in the Carpathians (Bretila, Cumpăna and Sebeş-Lotru Groups) ($Pt_{1.2}$; Cp_2).

3. Leptyno-amphibolite formation consisting of an alternation of amphibolites, white leptynitic gneisses and plagiogneisses with variable ratios between these three main lithologic components ($Pt_{1.3}$; Cp_3).

4. Gneiss-micaschist formation. In the Carpathians plagiogneisses (with migmatizations in the Bretila and Sebeş-Lotru Groups) prevail in

the lower part and micaschists (locally with amphibolites) in the upper part (Cp_4). In the Rhodopian sequence three lithostratigraphic units may be considered as equivalents of Cp_4: biotite and two mica gneisses ($Pt_{2.1}$), leptytoid gneisses of detrital provenance ($Pt_{2.2}$) and micaschists, with schistose gneisses and locally amphibolites in the upper part ($Pt_{2.3}$).

5. Carbonatic formation consisting of limestones (marbles) and dolomites with intercalations of micaschists, quartzites and amphibolites ($Pt_{3.1}$; Cp_5).

6. Quartz micaschist formation represented by quartz-micaschists with intercalations of quartzites, amphibolites and limestones ($Pt_{3.2}$, Cp_6).

This sequence of formations shows that at the lower part of the pile prevail gneissic rocks, locally associated with amphibolites, while the upper part is constituted of micaschists, limestones and dolomites associated with amphibolites. If the rank of supergroup is assigned to the sequence of a cycle (Zoubek, 1977) the mentioned lithologic contrast suggests a possible large scale subdivision in a lower group including Cp_{1-4}, Pt_{1-2} and an upper group corresponding to Cp_{5-6}, Pt_3. Different names are in use for these groups following geographic or tectonic restraints. Thus, in the Carpathians, the lower unit is represented, for example, by the Bretila, Cumpăna, Sebeș-Lotru and Jaraba Groups and the upper unit by the Rebra, Făgăraș, Baia de Arieș and Kokava Groups.

The rocks corresponding to the Rhodopian-Carpian sequences were assigned in the Romanian Carpathians to the Upper Precambrian A (1,650 ± 50–850 ± 50 m.y. (national maps 1 : 50,000, 1 : 1,000,000) or Middle Proterozoic (Kräutner, 1980), in Bulgaria to the whole Proterozoic (Kozhoukharov, Timofeev, in Kozhoukharov et al., 1980), in the Slovakian and Russian Carpathians either to the Precambrian (e.g. Kamenicky, 1980; Rudakov, 1980) or to the Paleozoic (e.g. Cambel, Veselsky, 1980; Siegel, 1980), in Yugoslavia to the Proterozoic (Dimitrijevici et al., 1967). At present the following data are available: 1. In the South Carpathians (Getic Nappe), the Rhodope and Serbo-Macedonian Massifs the Rhodopian-Carpian sequences are unconformably overlain by Vendian or Upper Riphean low grade metamorphic sequences, namely by the Cibin, Vlasina and Diabase-Phyllitoid Group (Karamata, Petcovic fide Stojanov, 1980; Kozhoukharov et al., 1978, 1980; Codarcea-Dessila, 1965; Giușcă et al., 1969; Kräutner, 1980 b). 2. Palynological data suggest the Precambrian age for the Rebra Group (Iliescu, Mureșan, 1972) and the Rhodopian Group (Kozhoukharov, Timofeev, in Kozhoukharov et al., 1980). 3. Rb/Sr whole rock and isochrone ages are in general influenced by younger events; the highest value (838 m.y.) was obtained by Bagdasarian (1972) on biotite of the Sebeș-Lotru Group. 4. K/Ar model ages reach accidentally 600–700 m.y. (Soroiu et al., 1970; Kräutner et al., 1976); generally they indicate Ar loss in younger events. Interpretations on K/Ar isochrones suggest 800–900 m.y. for the metamorphism of the Bretila Group (Kräutner et al., 1976).

According to these data it is probable that the Rhodopian-Carpian sequence corresponds to the Middle Proterozoic (sensu Zoubek, 1977) (Fig. 2).

The Marisian-Vlasina-Diabase-Phyllitoid sequence is represented in most of the structural units. According to the correlation proposed in the Plate two main sequences may be recognized.

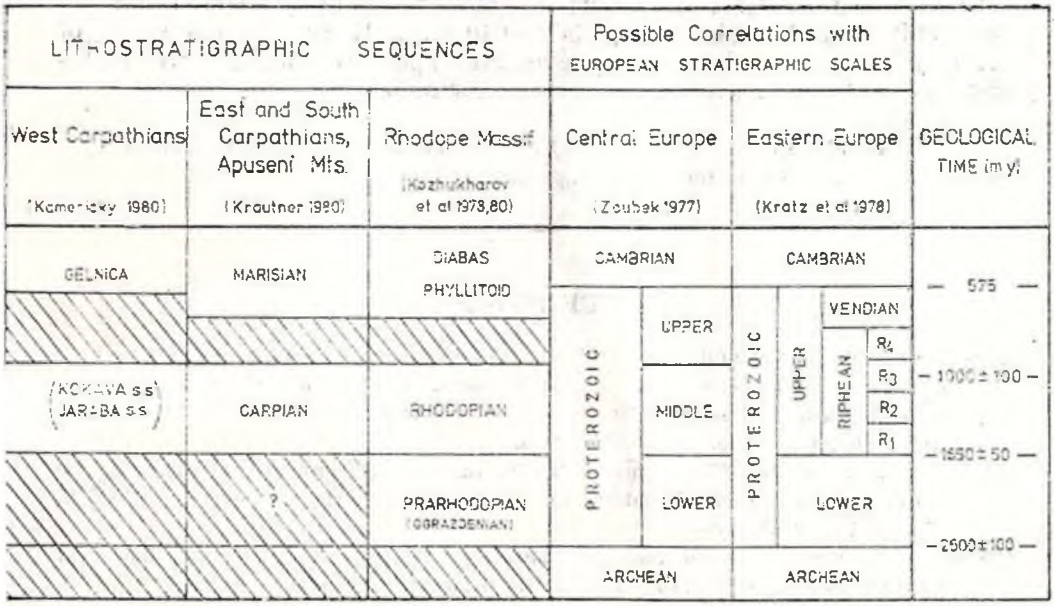

Fig. 2. — Suggested stratigraphic and lithostratigraphic correlation of Precambrian in the Carpatho-Balkan Belt.

1) A basic volcano-sedimentary sequence with greywacke and scarce thin beds of limestones and/or dolomites, corresponding to the Biharia Formation and the main parts of the Cibin Group (maybe also of the Lereşti Formation) in the Romanian Carpathians, to the Vlasina Group in the Serbo-Macedonian Massif and to the Diabase-Phyllitoid Group in the Rhodope Massif.

2) A blastodetrital sequence with rhyolitic metavolcanics and metalydites, represented by the Tulgheş Group (Tg), the Gelnica Group (Ge), the Muncel "Series" and the uppermost parts of the Cibin (Cb_3), Vlasina and Diabase-Phyllitoid Groups. The most complete sequence seems to be preserved in the Tulgheş Group; it consists of a blastodetrital quartzitic formation (Tg_1, Mu_1, Cb_3), a graphitic formation with metalydites (Tg_2) and syngenetic manganese ores in Tg and Ge, a metarhyolitic volcano-sedimentary formation (Tg_3) with syngenetic base metal sulfide ores in Tg and Ge, a blastodetrital formation with quartzites phyllites (Tg_4) and an upper graphitic formation with greenschists and limestones (Tg_5).

The available palynological data suggest for the basic volcano-sedimentary sequence a Vendian (maybe also uppermost Riphean) age (V i s a r i o n , D i m i t r e s c u, 1971; C o d a r c e a - D e s s i l l a, I l i e s c u, 1969; P a n t i c i et al., 1967 fide D i m i t r i j e v i c i , 1969), while the blastodetrital sequence with rhyolitic metavolcanics (Tulgheş

Group and Muncel Formation) may be assigned to the Cambrian (Iliescu, Mureşan, 1972; Iliescu et al., 1983; Visarion, unpublished data). These palynological data are in concordance with the radiometric ages reported from the Tulgheş Group (Pb-Pb= 540–600 m.y., U–Pb zircon 560–640 m.y., K/Ar maximal values of 472 m.y.). For the Gelnica Group even an Upper Ordovician-Lower Devonian age (Grecula et al., 1981) was assumed.

[3] Abbreviations used by Kozhoukharov et al. (1978, 1980) for the Rhodopian (Pt) and Kräutner (1980) for the Carpian (Cp) sequences.

REFERENCES

Arsovski M., Dumurdzanov N., Hristov S., Ivanov T., Ivanova V., Petkovski P., Stojanov R. (1977) Correlation of the Precambrian Complexes of the Pelagonian Massif, Vardar Zone and Serbo-Macedonian Massif. VII *Coll. Geol. Aegean Region*, II, p. 549–557, Athens.

Bagdasarian P. G. (1972) Despre vîrsta absolută a unor roci eruptive şi metamorfice din masivul Ditrău şi Munţii Banatului din România. *Stud. cerc. geol. geogr., seria Geologie*, 17, 1, p. 13–21, Bucureşti.

Bajanic S. (1980) Position of the early Paleozoic in the geotectonic history of the West Carpathians. In Sassi F. P. ed. I.G.C.P. No. 5, Newsletter 2, p. 88–91, Padova.

Balintoni I. (1969) Asupra caracterului retromorf al paragnaiselor biotitice cu clorit de pe Bisca Groşetului (Făgăraş). *Bul. Soc. şt. geol. România*, XI, p. 275–281, Bucureşti.

— Gheuca I. (1977) Metamorfism progresiv, metamorfism regresiv şi tectonica în regiunea Zugreni-Barnar (Carpaţii Orientali). *D. S. Inst. geol. geofiz.*, LXIII, 5, p. 11–38, Bucureşti.

Bercia I. (1975) Metamorfitele din partea centrală şi de sud a masivului Godeanu. *St. tehn. econ. seria I*, 12, p. 3–19, Bucureşti.

— Bercia Elvira (1975) Formaţiunile cristaline din sectorul românesc al Dunării (Banat-Carpaţii Meridionali). *An. Inst. geol. geofiz.*, XLIII, p. 5–63, Bucureşti.

— Kräutner H. G., Mureşan M. (1976) Pre-Mesozoic Metamorphites of the East Carpathians. *An. Inst. geol. geofiz.*, L, p. 37–70, Bucureşti.

Berza T. (1978) Studiul mineralogic şi petrografic al masivului granitoid de Tismana. *An. Inst. geol. geofiz.*, LIII, p. 5–176, Bucureşti.

Boiko K. A. (1970) Doverhne paleozoiskii complecs severozapadnogo oconciania marmaroşscogo massiva (Vostocinie Carpati). *Izd. vo Lvovskogo Universiteta*, Lvov.

Borcoş M., Hîrtopanu I., Dimitrescu R., Boştinescu S. (1979) Geological map of Romania, scale 1 : 50,000, sheet 57 a, Muntele Mare (unpublished). Inst. geol. geofiz., Bucureşti.

Cambel B., Veselsky I. (1980) Evaluation of the nuclear geochronology of West Carpathian Crystalline Complexes from the point of view of their stratigraphy. *An. Inst. geol. geofiz.*, LVII, p. 199–215, Bucureşti.

Codarcea-Dessila M. (1965) Studiul geologic şi petrografic al regiunii Răşinari-Cisnădioara-Sadu. *Mem. Inst. geol.*, VI, Bucureşti.

— Iliescu Violeta (1969) Noi date microfloristice asupra vîrstei complexului calcaros al seriei de Sibişel. *Stud. cerc. geol. geofiz., geogr., seria Geologie*, 14, 1, p. 279–282, Bucureşti.

D i m i t r e s c u R. (1966) Muntele Mare. Studiu geologic şi petrografic. *An. Com. geol.* XXXV, p. 165—239, Bucureşti.

— (1978) Structure géologique du massif cristallin Făgăraş-Ezer-Leaota. *Rev. roum. géol. géophys. géogr., sér. Géol.*, 22, p. 43—51, Bucureşti.

— B o r d e a J., B o r d e a S. (1974) Geological map of Romania 1 : 50,000. Sheet 57 c, Cîmpeni. Inst. geol. geofiz. Bucureşti.

D i m i t r i j e v i c i M. D. (1969) The metamorphic rocks of Yugoslavia. *Acta Geol. Acad. Sci. Hung.*, XIII *(Bull. IX Congr. Carp. Balkan Geol. Assoc.)*, 1, p. 41—53, Budapest.

— G r u b i ć A., P e t r o v i ć B., A l e k s i ć V., D i v l j a n S., B a b o v i ć M., K a l e n i ć M. (1967) Metamorphic complexes of the Carpatho-Balkanic arch and adjacent areas. *Acta Geol. Acad. Sci. Hung.*, 11, 1—3, p. 23—34, Budapest.

E x n e r C. (1978) Das Präkambrium-Problem in Oesterreich. Materials to the IGCP Project No 22, "Precambrian in Younger Fold Belts". Praha.

G h e r a s i N., M a n i l i c i V., D i m i t r e s c u R. (1966) Studiul geologic petrografic al masivului Ezer-Păpuşa. *An. Com. Stat Geol.*, XXXV, p. 47—104, Bucureşti.

G h i k a - B u d e ş t i S. (1940) Les Carpates Méridionales Centrales. Recherches pétrographiques et géologiques entre le Paring et le Negoi. *An. Inst. Geol. Rom.*, XX, p. 175—220, Bucureşti.

G i u ş c ă D., S a v u H., B o r c o ş M. (1968) La stratigraphie des schistes cristallins des Monts Apuseni. *Acad. R.S.R. Rev. roum. géol. géophys. géogr., sér. Géol.*, 12, 2, p. 143—159, Bucureşti.

— S a v u H., B e r c i a I., K r ä u t n e r H. G. (1969) Sequence of tectono-magmatic pre-Alpine cycles on the territory of Romania. *Acta Geol. Acad. Sci. Hung.*, 13, p. 221-234, Budapest.

G r e c u l a P., V o z á r o v á A., V o z á r J. (1981) Characteristics of lithotectonic profiles in geotraverse "C", West Carpathians (Czechoslovakia) . In Karamata S. Sassi F. P. ed. IGCP No. 5, Newsletter, 3, p. 77—79, Beograd.

H î r t o p a n u I. (1975) Le métamorphisme de basse pression dans les Monts Mehedinţi. *D. S. Inst. geol. geofiz.*, LXI/1, p. 217—238, Bucureşti.

— (1978) Cristalinul Getic : metamorfism polifazic sau polimetamorfism? *Stud. cerc. geofiz. geogr., ser. Geol.*, 23, 2, p. 185—193, Bucureşti.

I a n o v i c i V., B o r c o ş M., B l e a h u M., P a t r u l i u s D., L u p u M., D i m i t r e s c u R., S a v u H. (1976) Geologia Munţilor Apuseni. Edit. Acad. R.S.R., 631 p. Bucureşti.

I l i e s c u V., M u r e ş a n M. (1972) Asupra prezenţei Cambrianului inferior în Carpaţii Orientali — Seria epimetamorfică de Tulgheş. *D. S. Inst. Geol.* LVIII/4 (1971), p. 23—38, Bucureşti.

— K r ä u t n e r H. G., K r ä u t n e r F., H a n n H. (1983) New palynological proofs on the Cambrian age of the Tulgheş Series (East Carpathians). *An Inst. geol, geofiz.*, LIX, p. 7—11, Bucureşti.

K a m e n i c k y I. (1980) Problèmes actuels de la stratigraphie du cristallin des Carpates Tchécoslovaques. *An. Inst. geol. geofiz.*, LVII, p. 185—198, Bucureşti.

K o z h o u k h a r o v D., K o z h o u k h a r o v a E. (1977) Temperature and pressure during the metamorphism of the Precambrian Complexes from the Rhodope Massif. *Geologica Balcanica*, 7, 3, p. 103—116, Sofia.

— K o z h o u k h a r o v a E., Z a g o r c h e v I. (1978) The Precambrian in Bulgaria. Materials to the IGCP Project No. 22 "Precambrian in Younger Fold Belts". Brno.

— D a b o v s k i C. (1980) The Precambrian in South Bulgaria. *Geol. Inst. Bulg. Ac. Sci.* Sofia.

Kräutner H. G. (1980 a) Lithostratigraphic correlation of Precambrian in the Romanian Carpathians. *An. Inst. geol. geofiz.*, LVII, p. 229—296, București.

— (1980 b) Precambrian unconformity in the Getic Nappe (South Carpathians). *An. Inst. geol. geofiz.*, LVII, p. 305—324, București.

— Kräutner F., Tănăsescu A., Neacșu V. (1976) Interprétation des âges radiometriques K/Ar pour les roches métamorphiques régénérées. Un example — les Carpates Orientales. *An. Inst. geol. geofiz.*, L, p. 167—229, București.

— Năstăseanu S., Berza T., Stănoiu I., Iancu V. (1981) Metamorphosed Paleozoic in the South Carpathians and its Relation with the Pre-Paleozoic Basement. *Inst. geol. geofiz., Guidebook Series*, 16 *(XII Carp. Bal. Geol. Congr.)*. București.

Kräutner Th. (1938) Das kristalline Massiv von Rodna. *An. Inst. Geol. Rom.*, XIX, p. 164—287, 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. geofiz.*, XLIII, p. 65—189, București.

Rudakov S. G. (1980) Pre-Mesozoic history of the Marmarosh Massif, East Carpathians (Soviet Part). In Sassi F. P. ed. IGCP No. 5, 2, p. 49—52, Padova.

Săndulescu M. (1975) Essai de synthèse structurale des Carpathes. *Bull. Soc. Géol. Franc.*, 7, XVII, 3, p. 299—358, Paris.

— (1980) Analyse géotectonique des chaînes alpines situées autour de la Mer Noir Occidentale. *An. Inst. geol. geofiz.*, LVI, p. 5—54, București.

Savu H. (1970) Stratigrafia și izogradele de metamorfism din provincia metamorfică precambriană din Munții Semenic. *An. Inst. Geol.*, XXXVIII, p 223—311, București.

Schuster A. (1977) Geological map of Romania 1 : 50,000, sheet 109 a, Negoi (unpublished) Inst. geol. geofiz. București.

Siegel K. (1980) The West Carpathian Crystalline and the possibility of Precambrian occurrence. *An. Inst. geol. geofiz*, LVII, p 221—228, București.

Soroiu M., Popescu G., Gherasi N., Arsenescu V., Zimmermann P. (1970) K-Ar dating by neutron activation of some igneous and metamorphic rocks from the southern branch of the Romanian Carpathians. *Ecl. Geol. Helv.*, 63, 1, p. 323—334, Basel.

Stojanov R. (1974) Petrological characteristics of magmatic and metamorphic rocks of the Wider-Prilep environs (Macedonia/Yugoslavia). *Posebni Izdanija*, 4, 5, Geol. Zavod. Scopje.

— (1980) Correlation of Precambrian and Riphean-Cambrian rocks on the territory of Yugoslavia. *An Inst. geol. geofiz.*, LVII, p. 459—468, București.

Visarion A., Dimitrescu R. (1971) Contribuțiuni la determinarea vîrstei unor șisturi cristaline din Munții Apuseni. *An. St. Univ. "Al. Cuza" din Iași*, II, *Geologie*, XVII, p. 1—13, Iași.

Vijdea E., Anastase Ș. (1975) Opredelenia vosrasta metodom obiknovenovo svintsa v nekotorich mestorojdeniach svrasanich s cristaliceskich masivami vostocinich Karpat (Romania). *Proc. X Congr. CBGA*, *sect.* IV, p 253—260, Bratislava.

Zoubek V. (1977) Remarks to the problem of subdivision of the Precambrian in "*Correlation of Precambrian*", 1, p. 62—93, Moscow.

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CONSIDERATIONS REGARDING THE GEOLOGICAL STRUCTURE OF THE NORTHERN PART OF THE RETEZAT MOUNTAINS (SOUTH CARPATHIANS)[1]

BY

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The zone presented in this paper is situated on the northern limb of the Retezat Mountains, being delimited by the Nucșoara Valley on the west, the Hațeg sedimentary basin on the north, the Rîul Bărbat Valley on the east and by a line following the contact with the Retezat granodioritic massif on the south. The paper is meant to provide some details on the geological structure of this zone and especially on the relathionships among its various formations.

Historical Background

In a synthetic paper on the crystalline schists of the South Carpathians, Pavelescu and Răileanu (1963) separated two large groups of crystalline schists : the crystalline of Getic type (group 1), exhibiting a high crystallinity and the crystalline of Danubian type (group II), displaying a more reduced crystalinity, correspond ng to the Getic Domain (the Getic Nappe) and, respectively to the Danubian Domain (the Autochthon). Three series of crystalline schists were separated within the Danubian Autochthon : the Drăgșan Series, with a lower amphibolite complex and an upper chlorite-sericite one ; the Lainici-Păiuș Series ; and a slightly metamorphosed series — the Tulișa Series — consisting of three horizons. The Lainici-Păiuș Series and the amphibolite complex of the Drăgșan Series were generated during the Precambrian orogeneses, while the chlorite-sericite complex, probably during the Cambrian-Ordovician. As regards the Tulișa Series, the authors assign the lower conglomeratic horizon to the Silurian or Devonian, and the upper part , beginning with limestones, to the Dinantian. On the sketch accompanying the paper the "Tulișa Series" is plotted only next to the succession beginning with limestones, while the unconformable conglomeratic lower horizon is plotted within the chlorite-sericite complex of the Drăgșan Series.

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The geological constitution of the Retezat Mountains is in concordance with the general view of the South Carpathians crystalline.

P a v e l e s c u (1953 a, b) distributes the geological formations making up the Retezat Mountains to the Danubian Autochthon crystalline, with the overlying sedimentary rocks, and to the Getic Nappe crystalline, with the overlying sedimentary rocks. The crystalline of the Getic Nappe on the northern side of the Retezat Mountains, from Hobiţa to Nucşoara, consists of paragneisses with biotite and garnet, biotitic micaschists, amphibolites with biotite and biotitic quartzites. The Mesozoic deposits on the Getic Nappe crystalline are represented by Tithonic limestones. The crystalline schists of the Autochthon consist of biotite-muscovite quartzitic schists, garnet phyllite micaschists, graphite quartzitic schists etc. — the Rîuşorul Series, chlorite-amphibolite schists — the Drăgşan Series and a series underlying the Drăgşan Series — the Pilugu Series. Within the Drăgşan and Pilugu Series the Retezat and Buta granitoid rock massifs are separated. G h e r a s i and D i m i t r e s c u (1968, 1970) separate within the Danubian Autochthon in the northern part of the Retezat and Petreanu Mountains the Rof banded chlorite-biotitic and amphibolitic schists series, the Rîuşorul biotitic schists series, the Zeicani chlorite schists with muscovite schists and the Rîul Mare Series consisting of blackish or greyish graphite, quartz or slightly sericitized limy phyllites, bearing sometimes fine biotite lamellae. The authors show that the main structure of the whole north-western slope of the Retezat Mountains is the Rof anticline.

M i c u and P a r a s c h i v e s c u (1970) published the results of the researches carried out between Nucşoara and Rîul Alb, describing within the Danubian Autochthon the Drăgşan and Tulişa Series and showing that the mesometamorphic crystalline formations of the Getic Nappe are almost identical with those of the Sebeş Mountains. The Tulişa Series would be Upper Proterozoic-Cambrian in age, the Drăgşan Series being ante-Upper Proterozoic in age. From the tectonic point of view two overthrust lines are pointed out : one between the Getic Nappe and the Autochthon, and the other between the Drăgşan Series and the Tulişa Series.

G h e r a s i et al. (1974) extend the name of "Măgura Series" to most of the epimetamorphic rocks of the Danubian Autochthon, which develops between the Muntele Mic and Petrii massifs, continuing towards the east up to the Rîul Bărbat Valley. The Getic metamorphic rocks (the Sebeş-Lotru Series) are assigned to the Dalslandian cycle, while the metamorphic rocks of the Autochthon are assigned to the Cadomian cycle (the Rof, Rîuşorul and Măgura Series) and respectively to the Caledonian cycle (the Zeicani Series). The sedimentary formations of the Danubian Domain are represented by the Rîul Mare (Silurian) Formation and the (Devonian) Vidra Formation.

S o l o m o n et al. (1976) assign the crystalline schist and anchimetamorphic formations of the Danubian domain in the northern part of the Vîlcan Mountains and on the south-eastern side of the Retezat Mountains to the Infracambrian Amphibolitic Series, the Cambrian Series, the Muntele Coarnele Laminated Conglomerates Series, the Tulişa Series and the Paroşeni Formation.

Geological Considerations

The geological formations making up the constitution of the zone between the Nucșoara Valley and the Rîul Bărbat Valley belong to both the Getic Nappe and the Danubian Autochthon.

The Getic Nappe covers the northern part of the zone, being developed within a reduced width from the left bank of the Nucșoara Valley to the left bank of the Rîul Alb Valley, getting wider towards the east up to the Rîul Bărbat basin. Northwards it is overlain by the sedimentary rocks of the Hațeg Basin and southwards it contacts the metamorphic formations of the Autochthon. It consists of the crystalline formations exhibiting an advanced metamorphism grade and underlying Mesozoic sedimentary deposits over a reduced area. The crystalline formations belong to the Sebeș-Lotru Series and consist of an alternation of garnet biotitic gneisses, muscovite-biotitic gneisses, quartz-feldspathic gneisses, muscovite-biotitic micaschists, feldspathic quartzites with biotite and muscovite, and amphibolites. Throughout the contact with the formations of the Danubian Autochthon the metamorphic rocks of the Getic Domain are strongly crushed. The mineralogical associations, the crystallinity grade of rocks and the feldspar aspect show that all the types of rocks described within the Getic metamorphic rocks were generated by the metamorphosis under the almandine amphibolite facies, the kyanite-almandine-muscovite and sillimanite-almandine-muscovite subfacies conditions (Turner, Verhoogen, 1960) of some originally eruptive formations — the quartz-feldspathic gneisses and at least some of the amphibolites or sedimentary formations — the other gneisses and amphibolites and the micaschists. The sedimentary formations from the Getic Domain form a narrow band between the Paroș Valley and the Lazu Valley, which is interrupted in the Rîul Alb basin; they consist of strongly fissured white-greyish limestones. Drăghindă et al. (from Micu and Paraschivescu, 1970) show that these limestones are Lower Cretaceous in age.

The Danubian Autochthon between the Nucșoara Valley and the Rîul Bărbat Valley consists of crystalline formations, granitoid rocks and slightly metamorphosed sedimentary formations. Based on the detailed field observations the following lithostratigraphic units have been separated: the Drăgșan and Zeicani Series, the Valea Nucșoarei Formation, the Retezat granitoids and the Tulișa Series.

Drăgșan Series. We use this term (sensu Pavelescu, 1953) only for the metamorphic rocks that contact the Retezat granitoid rocks and underlie the rocks of the Tulișa Series or of the Valea Nucșoarei Formation in the northern part. From the Prislop Peak westwards the schists assigned to the Drăgșan Series correspond mainly to the complex of lower basic metatuffs of the Măgura Series (Gherasi et al., 1974). The rocks of this series show a great development in the eastern and southern parts of the Retezat granitoid massif. From the petrographic point of view one can describe albite chlorite-epidote schists, sericite-chlorite quartzitic schists ± epidote and amphibolic schists that formed by the metamorphosis under the greenschist facies conditions of some pelitic sediments and some basic tuffites and tuffs. The presence of biotite in most of the rocks described within this series may be the result of the contact with the mass of granitoid rocks; but the presence of garnet, although isolated, indicates that biotite may be at least partially the result of the regional metamorphism, which locally reached the almandine isograde.

Zeicani Series. Within this series Gherasi et al. (1967) describe the formations which appear at the upper part of the Autochthon and are generally overlain by the Getic metamorphic rocks. Three complexes are described: the complex of metatuffites and leptynites, the complex of metagreywackes and the complex of metavolcanics. Later on the lower complex was assigned to the Măgura Series, the Zeicani Series comprising the metagreywacke complex and the volcano-sedimentary complex (Gherasi et al., 1974). Between the Nucşoara Valley and the Rîul Bărbat Valley, we assigned to this series the epimetamorphic crystalline extending between the Tulişa Series and the Getic Nappe; it forms a band trending EW and seldom reaching 500 m in width between the Valea de Munte Brook and the Paroş Valley, getting wider towards the west and reaching almost 3,000 m on the Stirbina-Prislop Crest. Based on lithological criteria and the metamorphism grade a lower volcano-sedimentary complex and an upper detrital one can be separated.

The volcano-sedimentary complex covers most of the Zeicani Series in this zone and corresponds to the chlorite-sericite complex of the Drăgşan Series (Pavelescu, 1953) or to the complex of upper basic metatuffs of the Măgura Series (Gherasi et al., 1974). Also, on the southern side, it overlies the formations assigned by Micu and Paraschivescu (1970) to the Tulişa Series. From the petrographic point of view one can separate chlorite-albitic schists with epidote and calcite, sericite-chlorite quartzitic schists, sericite-chlorite albitic schists, sericite-chlorite schists and metagabbros. Between the Paroş Valley and Nucşoara several levels of metaconglomerates are intercalated within this complex, the best represented ones being those on the Sălăşel Brook and on the crest north of the Colţu Mare Peak. There are also a few white quartz lenses,of which that on the left side of the Paroş Valley crops out over an area of 50 m in length and 3-4 m in width. The rocks of this complex were formed by the metamorphosis of a volcano-sedimentary series under the greenschists facies conditions, some zones reaching the biotite isograde. The metagabbros represent a basic magmatic intrusion which graded partially to the greenschist facies.

The detrital complex is found at the top of the Zeicani Series in this zone, extending as an almost continuous band from the Nucşoara Valley to Valea de Munte, rarely exceeding some tens of metres in width. It is overlain by the formations of the Getic Nappe and corresponds to the complex with metagreywackes of the Zeicani Series (Gherasi et al., 1974). Micu and Paraschivescu (1970) assign these rocks to the top of the Drăgşan Series as a level of muscovite chlorite schists. The complete sequence of this complex crops out on the Paroş Valley, where a level of limy metapsammites with rare feldspar fragments lies in the base, being overlain by feldspathic metapsammites

± muscovite which underlie chlorite greenschists. The level with limy metapsammites also occurs on the Beuşi Brook, the remaining part of this complex being represented by feldspathic metapsammites ± muscovite. The "metagreywacke" aspect may be due to a strong cataclasis undergone by these rocks during the Getic overthrust.

Valea Nucşoarei Formation. In the Nucşoara Valley basin there develop some rocks over a large area, overlying the Drăgşan Series and tectonically underlying the Zeicani Series; they are assigned to the Tulişa Series by Pavelescu (1953 b) and Micu and Paraschivescu (1970), while Gherasi et al. (1974) assign them to the Rîuşorul Series. As they are different from both these series, especially as regards the metamorphic grade, we propose to call them the "Valea Nucşoarei Formation" consisting of sericite-chlorite quartzitic schists ± biotite, graphite schists and crystalline limestones. We consider the rocks of this formation as representing the youngest terms of the Rof anticlinal structure (Gherasi, Dimitrescu, 1970), being in an upper position with respect to the Rîuşorul Series.

The serpentinite outcrops in the Nucşoara Valley basin show a special situation. South of the Stirbina Peak, immediately under the crest line, there crops out a body of uralitized pyroxenites with decimetric aphanitic intercalations consisting of epidote actinolite schists. Rocks exhibiting a similar character occur also on the right slope of the Nucşoara Brook, where the epidote actinolitic schists show a great development; in addition, biotite occurs as porphyroblasts visible with the naked eye. Also on the right slope, and especially on the left slope of the Nucşoara Brook, there are several outcrops of talc-bearing schists with actinolite and tremolite; antigorite and chrysotile are frequently found. Taking into account also the serpentinites occurring farther west, on the left bank of the Rîuşorul Valley, they were assigned to the Tulişa Series (Micu, Paraschivescu, 1970) or to the Măgura, Rîuşorul and Zeicani Series (Gherasi et al., 1974). As the ultrabasic rocks in this zone are hosted in various lithostratigraphic units indeed, it seems plausible to state that they were emplaced on a fault system trending approximately east-west and reactivated at various time intervals. The talc-bearing schists and the actinolitic ones showing an aphanitic aspect might be much older than the pyroxenites on the right bank of the Nucşoara Brook and the antigoritic serpentinites on the left bank of the Rîuşorul Brook, the present aspect of which is due rather to the autometamorphism processes than to the regional metamorphism.

Granitoid rocks. They are represented by the marginal facies of the north-eastern extremity of the Retezat massif, contacting the crystalline formations of the Drăgşan Series in the eastern part and the formations of the Drăgşan Series or of the Tulişa Series in the northern part. As regards the composition, structure and texture gneissic granodiorites and laminated granodiorites can be described. Pavelescu (1953) considers the lamination of the marginal zones as the result of the consolidation of the granodioritic body under stress conditions, the contact with the crystalline cover being tectonic; this would also explain the lack of thermal contact phenomena.

Still, the above author shows that the absence of a typical contact zone could be also explained by the later transformation of some contact biotitic schists. Micu and Paraschivescu (1970) separate a zone of quartzitic schists with biotite at the contact with the granitoid rocks between the Mălăiești Brook and the Cirnic Brook, showing that the presence of biotite would be caused by the influence of the granitoid rocks. Between the Nucșoara Valley and the Riul Bărbat Valley, where the contact with the schists of the Drăgșan Series is not masked by the formations of the Tulișa Series, the two types of magmatic and crystalline rocks mix up within a thickness of 20-30 m. This phenomenon can be also observed microscopically through an increase of the crystalline schist granulation, accompanied by pronounced feldspathization and biotitization. As the distance from the contact increases, the biotite amount decreases, so that 200-300 m from the contact the normal rocks of the series are found. The mineralogical, structural and textural aspects indicate that the rocks of the cover underwent a thermal contact metamorphism which superposed on the regional metamorphism of the Drăgșan Series, changing into schistose hornfelses with biotite and epidote, or even hornblende in some places. The presence of chlorite and at least a part of epidote may be due to the retromorphism which took place during the tectonic movements subsequent to the intrusion of the massif.

Tulișa Series. From the right bank of the Nucșoara Valley to Valea de Munte a narrow zone (maximum 200 m) develops, consisting of slightly metamorphosed (anchimetamorphic) sedimentary deposits overlying transgressively either the Drăgșan Series or directly the granitoid rocks and underlying the formations of the Zeicani Series. By similarity with the surrounding zones, these rocks may be assigned to the Tulișa Series (sensu Pavelescu, 1953 a, b). The complete sequence in this zone begins with laminated conglomerates which occur only between the Preotesei Brook and the Serel Brook and continues with microconglomerates, sandstones and arkoses, quartz sericite $\pm$ graphite phyllites, calcarenites and sericite-graphite phyllites. The laminated conglomerates on the right branch of the Serel Brook overlie the schists of the Drăgșan Series, while those on the left branch (Apa Beuși Brook) underlie without any transition the limestones level. They consist of a psammitic matrix formed of fragments of granitoid rocks, feldspar and quartz grains and muscovite leaflets, including relatively small elements (1-2 cm) of granitoid rocks and larger quartz lenses that can reach 10 cm in length. Taking into account the fact that in the zone investigated by us the laminated conglomerates show a higher metamorphism grade than the other formations of the Tulișa Series and directly underlie the limestone level, one can state that there are unconformity relations among the two rock types. Thus the laminated conglomerates may be equivalent to the conglomerates of the lower complex of the Tulișa Series (Solomon et al., 1976) and to the Capu Plaiului conglomerates (Stănoiu, 1976), the main difference from the latter consisting in the petrographic nature of the elements and matrix. This is why we propose the name of "Apa Beuși conglomerates" for the conglomerates described by us, after the name of the left branch of the Serel Brook, where they are very frequent. The other formations of the sequence described would correspond to the upper part of the lower complex and to the middle complex of the Tulișa Series (Solomon, 1976) or to the Gîrbovu Formation (Stănoiu, 1976).

Tectonic Considerations

The major elements controlling the structure of the north-eastern part of the Retezat Mountains are the intrusion of the granodioritic massif, the Getic Nappe overthrust and an important overthrust within the crystalline formations of the Autochthon. The granodioritic massif is responsible to a great extent for the anticlinorium structure of the Retezat Mountains, the effect of this structure in the studied region being manifested by the divergent dips of the crystalline formations of the cover. Thus, on the whole, the metamorphic rocks between the Rîul Bărbat Valley and the left bank of the Nucșoara Brook can be regarded as a monocline with northern dips; only within the formation on the Nucșoara Valley there are several small anticlinal and synclinal folds. The abnormal relations between the formations of the Getic Domain and those of the Danubian Domain were first pointed out by Gh. Munteanu-Murgoci, who published in 1912 in Stockholm the map on which the nappe structure of the South Carpathians was first presented. In the region investigated the thrust of the Getic Nappe over the Autochthon is well pointed out by mapping and by the strong cataclasis, especially of the overlying rocks within a thickness that can exceed 100 m. The Getic metamorphic rocks overlie the formations of the Zeicani Series everywhere and, although we did not find clear outcrops, we estimate the average dip of the overthrust plane at 30-40°. Another important overthrust line is clearly pointed out within the Autochthon, setting the formations of the Zeicani Series in tectonic position over the Valea Nucșoarei Formation, the Tulișa Series or directly over the Drăgșan Series. The contact with the rocks of the Tulișa Series is the most typical, being followed as an almost continuous steep slope, reaching in some places several tens of metres, from the right bank of the Nucșoara Brook to the Rîul Bărbat hydrographic basin. Usually the overthrust took place over a phyllite level that enabled the advance. The rocks of the Zeicani Series over the overthrust plane are generally represented by slightly schistose epidote chlorite-albitic schists that were strongly retromorphosed, getting a pronounced schistosity. This overthrust line is mentioned by Micu and Paraschivescu (1970), although it is drawn much farther north on the map accompanying the paper, and by Gherasi et al. (1974) who draw it at about the same level as we draw it. The dip of the overthrust plane is greater than in the case of the Getic Nappe, reaching 60-70° in some places.

Considering this situation, between the Nucșoara Valley and the Rîul Bărbat Valley, a subunit displaying a Parautochthon role can be separated within the Danubian Autochthon; it consists entirely of the formations of the Zeicani Series, the Autochthon proper being formed of the granitoid rocks, the Drăgșan Series, the Valea Nucșoarei Formation and the Tulișa Series. The fact that the slightly metamorphosed formations of the Paleozoic Tulișa Series lie under the overthrust plane indicates that the dislocation took place during the Alpine movements, probably concomitantly with the achievement of the Getic overthrust. The most important of the disjunctive transversal structures is the fault on the right bank of the Nucşoara Valley (the Nucşoara Fault — Gherasi et al., 1974) trending approximately NNW-SSE; along it a lowering of the eastern compartment took place. Also, there are some faults of less importance which generally shift the limits of the Getic crystalline.

REFERENCES

Gherasi N., Dimitrescu R. (1968) Contribuţiuni petrotectonice la stractura cristalinului danubian din partea nordică a munţilor Retezat şi Petreanu. *Anal. şt. Univ. Al. I. Cuza — Iaşi, Sect.* II, b, XIV, Iaşi.

— Dimitrescu R. (1970) Anticlinalul Rof şi rolul lui în structuara părţii nordice a munţilor Retezat şi Petreanu. *Anal. şt. Univ. Al. I. Cuza — Iaşi, Sect.* II, b, XVI, Iaşi.

— Zimmermann P., Zimmermann V. (1974) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

Micu C., Paraschivescu C. (1970) Contribuţii la cunoaşterea geologiei părţii de nord a munţilor Retezat, între Rîul Alb-Rîul Nucşoara, cu privire specială asupra ivirilor de talc. *D. S. Inst. Geol.*, LVI (1968—1969), Bucureşti.

Pavelescu L. (1953 a) Studiul geologic şi petrografic al regiunii centrale şi de sud-est a munţilor Retezatului. *An. Com. Geol.*, XXV, Bucureşti.

— (1953 b) Cercetări geologice în munţii Retezat. *D. S. Com. Geol.*, XXXVII (1949—1950), Bucureşti.

— (1963) Contributions à l'étude du soubassement cristallin et l'aire de distribution des différentes formations cristallophylliennes dans la partie centrale et orientale de l'Autochtone Danubien. *Asoc. geol. Carp.- Balc., Congr.* V, 1961, II, Bucureşti.

— Răileanu Gr. (1963) Considérations générales sur l'âge des schistes cristallins de l'Autohtone des Carpates Méridionales. *Asoc. geol. Carp.-Balc., Congr.* V, 1961, II, Bucureşti.

Schuster C. A. (1972) Consideraţii asupra evoluţiei domeniului danubian (Carpaţii Meridionali). *D. S. Inst. Geol.*, LVIII (1971), Bucureşti.

Solomon I., Visarion A, Iordan M. (1976) Consideraţii asupra formaţiunilor cristalofiliene şi anchimetamorfice din munţii Vîlcan şi munţii Retezat (Carpaţii Meridionali). *D. S. Inst. geol. geofiz.*, LXII (1974-1975), Bucureşti.

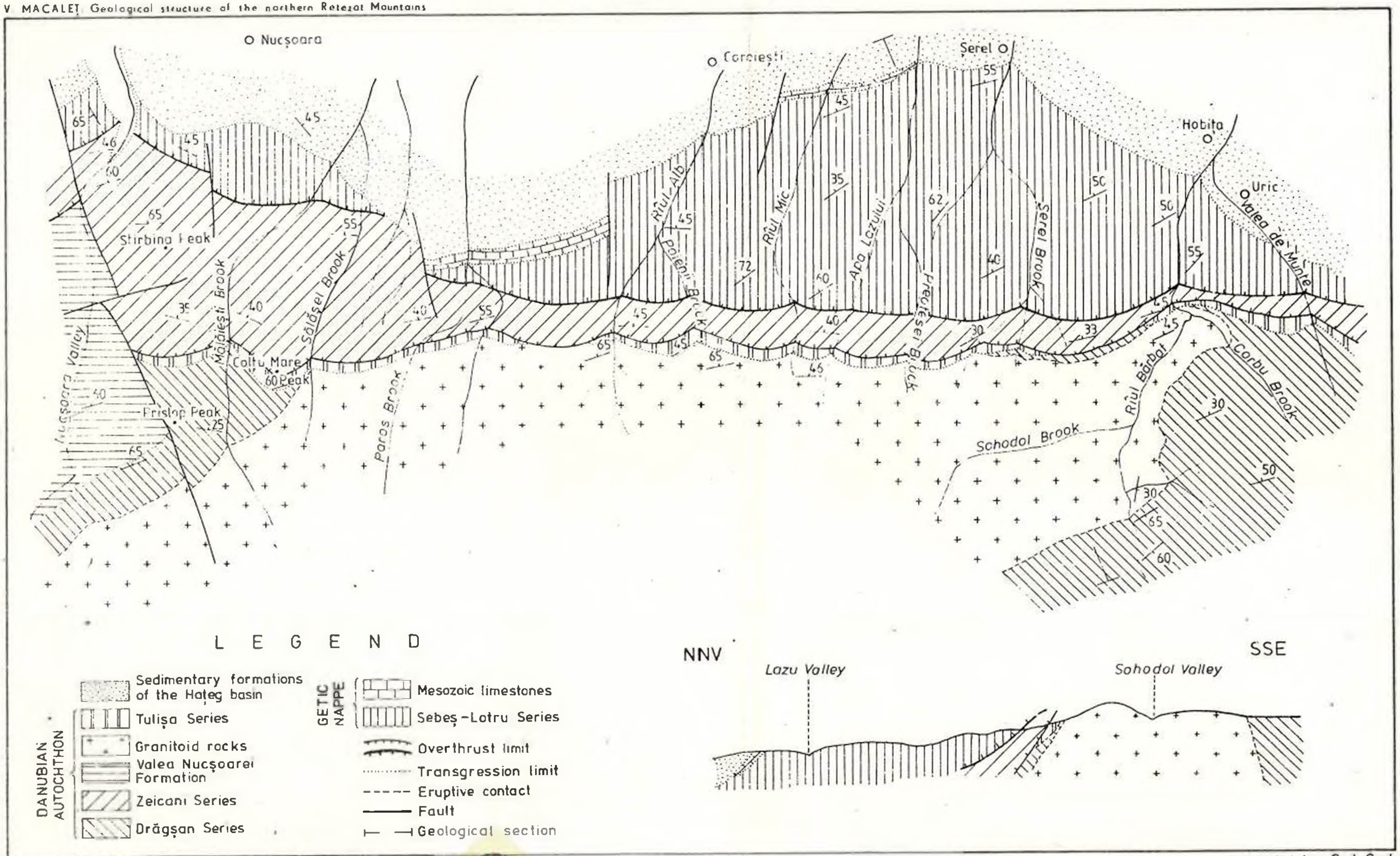
Stănoiu I. (1976) Contribuţii la stratigrafia formaţiunilor paleozoice din versantul nordic al munţilor Vîlcan (Carpaţii Meridionali) cu implicaţii asupra părţii externe a autohtonului danubian. *D. S. Inst. geol. geofiz.*, LXII (1974—1975), Bucureşti.

horta 1:200 000
25 Deva + 26 Orăștie

GEOLOGICAL SKETCH OF THE NUCȘOARA-RÎUL BĂRBAT ZONE

SOUTH CARPATHIANS

0 500 1000 1500 2000m



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CONSIDERATIONS ON THE THICKNESS OF THE REGIONALLY METAMORPHOSED FORMATIONS [1]

BY

MIRCEA MUREȘAN [2]

Structural investigations of most of the metamorphic formations of Romania revealed that their microfolding on a mesoscopic and microscopic scale represents a frequent and penetrative deformation in most of the regionally metamorphosed rocks. Our researches carried out on the metamorphics of Poiana Ruscă (South Carpathians) in the Crystalline-Mesozoic Zone of the East Carpathians and central Dobrogea as well as the observations made on samples from various collections of metamorphic rocks, correlated with some data from the relevant literature lead us to the conclusion that the transposition of the stratification schistosity represents also a widespread and penetrative deformation in the metamorphics on the Romanian territory, especially in the incompetent rocks. As a matter of fact the large distribution of the two types of deformation in the regionally metamorphosed formations is mentioned in the world relevant literature on the territories consisting of such rocks. Although both the numerous and various morphologic and genetic aspects and their consequences on the structural ensemble to which they belong were presented and discussed, the relations between these deformations and the thickness of the piles within which they occur were not pointed out, so that they are going to be presented in this paper.

Our considerations are based on proper observations on the following regionally metamorphosed formations from various zones, formed during various intervals of the Precambrian and Paleozoic.

a) *Bretila Series (East Carpathians) — Upper Precambrian* A^3; metamorphism in the almandine amphibolite facies, about 850±50 m.y. ago; micaschists and gneisses, amphibolite intercalations (B e r c i a et al., 1976; K r ä u t n e r et al., 1976; M u r e ș a n, 1980);

b) *Rebra Series (East Carpathians) — Upper Precambrian* A; metamorphism in the almandine amphibolite facies, about 850±50 m.y.

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ago; three regional lithostratigraphic complexes: the lower and upper complexes consisting of micaschists, paragneisses and biotite quartz schists with intercalations of amphibolites, limestones, quartz-feldspathic rocks, black quartzites: the middle complex consisting mainly of dolomites and limestones (Kräutner, 1968; Kräutner et al., 1976; Bercia et al., 1976; Mureşan, 1981);

c) *Negrişoara Series (East Carpathians) — Upper Precambrian A (?)*; metamorphism in the almandine amphibolite facies; biotite quartz schists, paragneisses, amphibolite intercalations (Balintoni, Gheuca, 1977; Mureşan, 1981);

d) *Altîn Tepe Series (Central Dobrogea) — Upper Precambrian A*; metamorphism in the almandine amphibolite facies, about 850±50 m.y. ago; micaschists, paragneisses, amphibolites (Mureşan, 1972);

e) *Tulgheş Series (East Carpathians) — Vendian (?)-Lower Cambrian*; metamorphism in the greenschists facies, about 500 m.y. ago; sericite-chlorite schists, quartz-feldspathic rocks; greenschists, quartzites, black quartzites with intercalations of limestones, metaconglomerates, syngenetic sulphide ores (Bercia et al., 1976; Kräutner, Popa, 1973; Kräutner et al., 1976; Mureşan, Mureşan, 1977);

f) *Poiana Ruscă Group (South Carpathians) — Silurian (?)-Devonian-Lower Carboniferous*; metamorphism in the greenschist facies, about 320 m.y. ago; sericite chlorite schists ± quartz schists, sericite-graphite schists, dolomites, limestones, greenschists, quartz-feldspathic rocks with intercalations of quartzite, black quartzites, syngenetic iron ores (carbonates and oxides) and sulphide (Kräutner et al., 1969; Mureşan, 1973).

Although the mentioned series were regionally metamorphosed in different periods and show various metamorphism grades, it is found that their synmetamorphic deformational evolutions are essentially similar, especially as regards the formation of the mesoscopic folds and the transposition of the stratification schistosity. Within the formations belonging to each series an ensemble of the oldest structural elements, formed during the regional metamorphism, can be distinguished — the B_1 tectonics. The deformations defining the B_1 tectonics are spatially congruent, being connected with the same regional movement and exhibiting a penetrative character. In several cases the B_1 tectonics elements were subsequently deformed during the Hercynian and Alpine movements, as it is noted for the Bretila, Rebra, Negrişoara and Tulgheş Series as well as the Poiana Ruscă Group (Bercia, 1967; Bercia, Bercia, 1970; Kräutner, 1968; Kräutner et al., 1969; Kräutner, Popa, 1973; Mureşan, 1964, 1973, 1980; Mureşan, Mureşan, 1977; Balintoni, Gheuca, 1977).

Owing to their microscopic and mesoscopic penetrative character, the B_1 tectonics deformations are the most important in the mentioned series as they cannot be completely effaced by the subsequent deformations. These latter deformations may often display a semipenetrative character and an unhomogeneous distribution ; they appeared during stages in which the blastesis of the respective formations was totally or almost totally terminated. It follows from the above statements that the thicknesses of the series considered were first and to the greatest extent influenced by microfolds and by the transposition of the stratification schistosity that occurred during the B_1 tectonics. We shall further present the main features of the most widespread deformations of the B_1 tectonics.

a) *The Ss stratification schistosity* formed during the synmetamorphic main blastesis, in its plane crystallizing most philosilicates and the elongation of many quartz, feldspar etc. grains being parallel to it. One can notice the disposition in bands of the mineralogical associations (for example the alternation of predominantly sericite-chlorite small beds with dominantly quartz bands, found in the quartz sericite-chlorite schists) which shows that the stratiform schistosity was generated by metamorphic differentiation during the differential movements among beds, determined by the folding of the formations through bending with concentric slipping. From this last point of view, the stratification schistosity may be considered to have originated by the shearing of the sedimentary material along some planes, generally parallel to the stratification of the initial sediments ; this fact is demonstrated by the Ss parallelism to the lithologic boundaries (which is observed in the sample and outcrop) and the lithostratigraphic ones (a situation that can be observed on a regional scale). As the formation of Ss started with the first differentiated movements between strata, that is with the beginning of the synmetamorphic folding movements, this tectonic element can be considered to have started its evolution earlier than the mesoscopic and microscopic folds (drag-folds type), formed somewhat later on the limbs of the megascopic folds. Indeed on a mesoscopic and microscopic scale one can notice that the minerals in the plane of the stratification schistosity followed the B_1 microfolds, which shows that the main process of metamorphic recrystallization took place especially before the formation of the B_1 folds. In fact Ss is affected by all the deformations of the B_1 tectonics.

b) *The B_1 folds* were synmetamorphically generated in the stage of plastic "flow" of the material subjected to folding. These deformations formed by flexural folding with concentric slipping (B e r c i a, B e r c i a, 1964, 1970 ; B e r c i a, 1967 ; K r ä u t n e r et al., 1969 ; M u r e ș a n, 1972, 1973 ; M u r e ș a n, M u r e ș a n, 1977) ; we think this is shown, among other things, by the folding disharmonies, the existence of the polygonal folds (with divergent axial planes), the thickness variations of each bed in directions parallel to the axial plane of the folds etc ; the petrotectonic data (D i m i t r e s c u, 1965 ; K r ä u t n e r, 1965) confirm this formation mode.

In cross section, the B_1 meso- and microscopic folds are characterized by the existence of a longer limb and of a shorter one, the same situation being found also for the successive folds and leading to the development of a scale (drag folds ; plis en escalier) on the limb of each fold of superior order on which they occur. The B_1 meso- and microscopic limbs generally show divergent limbs in the competent rocks (amphibolites, quartzites, dolomites, limestones, syngenetic sulphide ores); these folds are frequently sharp or isoclynal in the incompetent rocks (sericite-chlorite schists, sericite-graphite schists, micaschists etc). By fine micro-folding the l_1 lineations formed at the boundary between mesoscopic and microscopic, representing in fact a particular case of the B_1 folds. It is important for our considerations that the microfolds are seldom isolated on the same limb, succeeding one another, especially in the incompetent rocks, at relatively regular intervals, so that each bend is usually situated at a height of 1/3-2/3 of the length of the long limb of the next fold; this fact brings about the considerable shortening of the strata in cross section with respect to the folding direction. The frequency of microfolds and their amplitude are relatively constant within a level consisting of the same rock type. The decimetric mesoscopic folds prevail in the competent rocks, while the microscopic and centimetric and millimetric ones prevail in the incompetent rocks. It is worth mentioning that in the case of some very thick piles (of the order of some hundred and thousand metres) of dolomitic rocks (the Hunedoara and Luncani dolomites from the Poiana Ruscă Group) the mesoscopic folds are very rare, occurring only in zones of facial interfingering of the latter with the terrigenous rocks; instead, in the case of some thick limy beds (the Ruschiţa and Alun marble limestones from the Poiana Ruscă Group, some limestone horizons from the Rebra Series) the mesoscopic microfolds (usually centimetric and decimetric) are very frequent.

The formation process of the B_1 folds can be imagined like this: within a quasihorizontal pile, subject to the first tangential pressures, there form large folds, leading to the appearance by shearing of the stratification schistosity, in the plane of which the primary material recrystallizes — metamorphic differentiation; the continuous impact of pressures amplifies the incipient folds, augmenting the differential movements between strata, thus leading to the formation of the mesoscopic and microscopic drag folds (parasite microfolds; plis en escalier). Therefore, in our opinion, the B_1 micro- and mesoscopic folds appear later than the megascopic folds.

By deforming the stratification schistosity, the micro- and mesoscopic folds belong to the post-schistose folds (Bellière, 1958).

Points *a* and *b* represent the most widespread cases within the formations of the metamorphic series considered. But there are also other deformation modes, which are much less frequent than those already described, being on the whole contemporaneous with the typical deformations of the B_1 tectonics. Thus, for instance, within the metamorphic formations of the Poiana Ruscă Group (Mureşan, 1964, 1973; Dimitrescu, Maier, 1966) and of the Tulgheş Series (Dimitrescu, 1971; Mureşan, unpublished data) there occur "mixed folds", within which the competent beds folded by bending with concentric slipping and the intercalations of incompetent material formed similar folds; in other cases, within thick piles of terrigenous rocks, especially phyllites, similar folds formed (synschistose — as defined by Bellière, 1958).

c) *The axial Cl_1 cleavage and the schistosity of the Scl_1 axial cleavage* mark the ruptural stage of the B_1 tectonics, when the stresses exceed the plasticity threshold of the folded material. These structural elements formed by shearing along some planes parallel to the Pa_1 axial plane of the B_1 folds when the accentuation of the drag folds leads to the lamination and then breaking of the limbs of these folds, being thus connected with the same movement that brought about the formation of the B_1 folds. Indeed the sense of the movement observed along these shearing planes (Cl_1 and Scl_1) is the same as that of the movements leading to the formation of meso- and microscopic drag folds. The above mechanism explains why, on a fold limb, the sense of movement along the Cl_1 and Scl_1 planes is the same, a considerable shortening of the strata in a transversal sense with respect to the direction of the main folding resulting by totalling. The difference between Cl_1 and Scl_1 lies only in their frequency within the series described. Thus within the competent rocks there develops prevailingly Cl_1 with a centimetric-decimetric frequency, while within the uncompetent rocks (of the sericite-chlorite schist or micaschist type) there develops especially Scl_1 with a millimetric and submillimetric frequency. By close shearing in a constant sense, along Scl_1, the transposition of the Ss stratification schistosity appears. This is the most widespread case within the incompetent rocks, such as the sericite-chlorite schists, the sericite-graphite schists, the micaschists and the micaceous paragneisses. The same phenomena of transposition of the stratification schistosity are found also in some calcareous rocks with philosilicates (such as the Rușchița-Alun limestones) or with graphite pigment. Thus Scl_1 partially or totally obliterates the former Ss, becoming one of the main elements measurable during the field investigations, often erroneously confounded with Ss. At the intersection of the Ss and Scl_1 stratification schistosities there form l'_1 lineations by microshearing, parallel to the l'_1 lineations and the axes of the B_1 folds respectively; consequently the l'_1 lineations provide the same indications on the spatial position of the axes of the B_1 megascopic folds. In some rocks (for example quartz schists), due to the shearing of the set of planes of the Ss stratification schistosity by the set of planes Cl_1 (Scl), "pencil" type structures may arise. Sometimes, by the transposition and rotation of the old quartzes, extruded along the Ss (or of carbonates — e.g. from the tuffogenous albite chlorite-calcite schists of the Poiana Ruscă Group and the Tulgheș Series) boudines may be generated, "mullion" structures or "rods", all of them elongated in the direction of the B_1 structure.

d) When the shearing movement along Cl_1 (respectively Scl_1) stops the folding stresses continuing to act, bring about the "*modification by flattening*" (sensu R a m s a y, 1962) of the already formed folds and of the bend remnants within the rocks with transposed schistosity along Scl_1. Due to the modification by flattening many of the directly observable B_1 folds, although formed by bending with concentric slipping, exhibit morphological features resembling the "similar" folds. As during the processes of modification by flattening the upset of the material, also along the folds axes, may take place, we consider that the formation of the tension fissures of "ac" type is connected with these very processes ; the "ac" fissures being quasiparallel to the folds axes, they can indicate the directions of the B_1 plicative structures.

The facts presented at points *a-d* reveal that the successive formation of the Ss stratification schistosity, of the B_1 folds, of the Scl_1 cleavage schistosity, the transposition of Ss by Scl_1 as well as the modification by flattening are organically connected with the same movement, being found in the various rocks of the series considered. Taking into account the generality of this mechanism of complex deformation in various metamorphic series differing in the age of deposition and of metamorphism which we pointed out also in other metamorphic lithostratigraphic entities from the Carpathians and Dobrogea (observations made on samples from various collections — M u r e ş a n, 1964) we can state that the genetic considerations on the main elements of the B_1 tectonics can be applied to most regionally metamorphosed formations in Romania and other countries.

The lithostratigraphic investigations of the regionally metamorphosed formations of Romania led to, among other things, the assignment of some thicknesses for the identified entities (series, formations, complexes, horizons, levels etc). So far these thicknesses have been established through the use of the thicknesses found on various profiles, taking into consideration the length of the perpendicular linking the lower surface to the upper one of the measured lithostratigraphic term. Totalling the thicknesses thus obtained for each subdivision of a succession, the thicknesses of the various metamorphic piles of Romania have been obtained. Thus a thickness over 12,000 m was found for the Poiana Ruscă Group, 3,500—4,000 m for the Tulgheş Series, 4,500-5,000 m for the Rebra Series and 2,000 m for the Altîn Tepe Series.

It follows from the above statements that these thicknesses, calculated in the mentioned way, are much greater than the initial thicknesses of the initial sedimentary piles. Indeed the meso- and microscopic (often isoclynal or sharp) folds as well as the transposition of the stratification schistosity (usually both coexisting deformation types) are widely distributed in the regionally metamorphosed formations. Our observations show that the two deformation types leading to an accumulation of the material (Figs 1, 2) bring about the increase of the thickness of the affected beds, by the increase of the distance between the two enveloping surfaces[4] of the bed (in the case of the undeformed beds, the two enveloping surfaces coincide with the lower and respectively upper surface of the bed). Within the incompetent rocks (sericite-chlorite $\pm$ graphite schists, micaschists etc) the modification of the initial thickness by deformation is much greater as compared to that within the competent rocks (dolomites, limestones, quartzites etc). In the course of mappings, mining works, drillings etc the geologist encounters such thicknesses (called by us "deformed thicknesses"), augmented by the mentioned deformations. These modified thicknesses (which were not acknowledged as such) have been assigned to the metamorphosed piles of Romania or other countries so far. A careful study of several outcrops and samples from one region allows the statistic evaluation for each rock type of the value of the ratio existing between the thickness of the undeformed parts of beds or laminae and the thickness modified by their deformation. With the aid of the coefficient obtained one can deduce (approximately) the thickness before deformation of the thicker sets of beds formed of the same rock type (therefore of the same competence). Then totalling the thicknesses thus obtained, the thickness of the series comprising these sets of beds is obtained. As the thicknesses

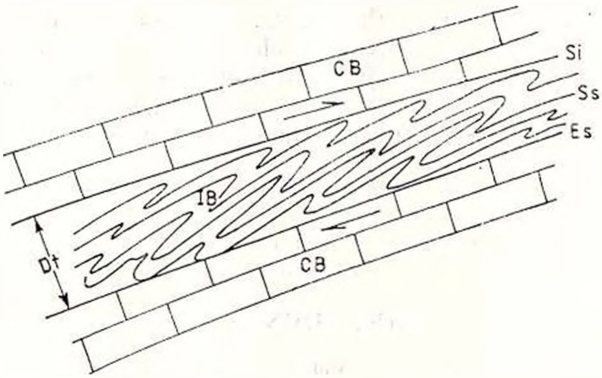

Fig. 1. — Deformation of thickness by mesoscopic folding.
The distance between the two competent beds (CB) increased after the deformation, resulting in the deformed thickness (Dt) for the incompetent bed (IB). Es, the enveloping surface is represented by the boundary between the competent bed and the incompetent bed; Ss, stratification schistosity.

thus obtained for the metamorphosed formations are much closer to the thicknesses of the sedimentary rocks from which they proceed, as compared to the deformed thicknesses used so far, they can be used for

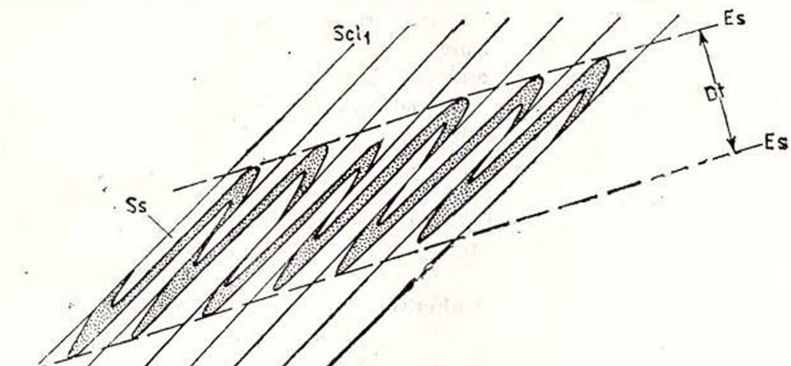

Fig. 2. — Deformation of thickness by transposition of the stratification schistosity (Ss) after the cleavage schistosity (Scl_1).
Es, enveloping surface of the transposed bed; Dt, deformed thickness.

the palaeogeographic evaluations, the reconstitution of the subsidence and sedimentation processes, the more real approximation of the burying depths etc.

The calculation method proposed by us is in fact valid also in the cases when the B_1 tectonics deformations are followed by other subsequent deformations. Naturally the initial thickness is modified during the regional metamorphism also by other processes, such as the volume decrease taking place during the metamorphic recrystallization of the sedimentary material, the modification by flattening, the upset of the material in the direction of the B_1 fold axes. Nevertheless we think that these modifications (as a matter of fact, they are difficult to estimate quantitatively) are not so important as those determined by the meso- and microscopic foldings and by the transposition of the stratiform schistosity respectively.

[3] Subdivision adopted for the geological map of Romania, scale 1:50,000, the interval of 1600-850±50m.y.

[4] Surface envelope; enveloping surface; Faltenspiegel.

REFERENCES

Balintoni I., Gheuca I. (1977) Metamorfism progresiv, metamorpfism regresiv şi tectonică în regiunea Zugreni-Berner (Carpaţii Orientali). *D. S. Inst. geol. geofiz.*, LXIII, 5, p. 11–38, Bucureşti.

Bellière J. (1958) Contribution à l'étude pétrogénétique des schistes cristallins du massif des Aiguilles Rouges (Haute-Savoie). *Annal. Soc. Géol. Belgique*, 81, p. 1–198, Liège.

Bercia I. (1967) Studii microtectonice în şisturile cristaline din regiunea Ghelar-Vadu Dobrii (Munţii Poiana Ruscă). *Acad. R.S.R., Stud. cerc. geol., geofiz., geogr., seria Geol.*, 12, 1, p. 97–128, Bucureşti.

— Bercia E. (1970) Contribuţii la cunoaşterea geologiei regiunii Vatra Dornei-Iacobeni. *An. Inst. Geol.* XXXVIII, p. 7–49, Bucureşti.

— Kräutner H., Mureşan M. (1976) Pre-mesozoic metamorphites of the East Carpathians. *An. Inst. geol. geofiz.*, L, p. 37–70, Bucureşti.

Dimitrescu R. (1965) Notă asupra structurii cristalinului din regiunea Iacobeni. *D. S. Inst. Geol.* LI, 1, p. 15–30, Bucureşti.

— Maier O. (1966) Observaţii asupra fenomenului de şistuozitate în partea de nord a masivului cristalin Poiana Ruscă. *Acad. R.S.R., Stud. cerc. geol., geofiz., geogr., seria Geol.*, 11, 1, p. 257–258, Bucureşti.

— (1971) Notă asupra mecanismului de deformare a unor cuarţite negre din regiunea Iacobeni. *Anal. Şt. Univ. "Al. I. Cuza" Iaşi, Secţ.* II, *b, Geol.*, p. 117–118, Iaşi.

Kräutner H. (1965) Consideraţii genetice asupra zăcămintelor de sulfuri complexe din cristalinul Carpaţilor Orientali. *Acad. R.S.R., Stud. cerc. geol., geofiz., seria Geol.*, 10, 1, p. 195–228, Bucureşti.

— (1968) Vederi noi asupra masivului cristalin al Rodnei. *Acad. R.S.R., Stud. cerc. geol., geofiz., geogr., seria Geol.*, 13, 2, p. 337–355, Bucureşti.

— Kräutner F., Mureşan G., Mureşan M. (1969) Stratigrafia, evoluţia magmatismului, metamorfismul şi tectonica şisturilor cristaline din unitatea epimetamorfică a masivului Poiana Ruscă. *An. Com. Stat Geol.*, XXXVII, p. 179–264, Bucureşti.

— Popa Gh. (1973) Succesiunea litostratigrafică şi tectonica Cambrianului inferior epimetamorfic (seria de Tulgheş) din regiunea Bălan-Carpaţii Orientali. *D. S. Inst. geol.*, LIX, 1, p. 252–278, Bucureşti.

— Kräutner F., Tănăsescu A., Neacşu V. (1976) Interprétation des âges radiométriques K/Ar pour les roches métamorphiques régénérées. Un exemple—Les Carpates Orientales. *An. Inst. geol. geofiz*, L, p. 167-229, Bucureşti.

M u r e ş a n M. (1964) Asupra prezenţei unor elemente microtectonice în şisturile cristaline din R. P. Română. *Stud. cerc. geol., geogr., geofiz., seria Geol.*, 9, 2, p. 451—475, Bucureşti.

— (1972) Studii asupra zăcămîntului de pirită cu magnetit de la Altin Tepe (Dobrogea Centrală). II. Poziţia stratigrafică a mineralizaţiei. *D. S. Inst. geol.* LVIII, p. 26—51, Bucureşti.

— (1973) Formaţiunile epimetamorfice din partea de nord-vest a masivului Poiana Ruscă (Carpaţii Orientali). *An. Inst. geol.*, XLII, p. 7—337, Bucureşti.

— M u r e ş a n G. (1977) Mineralizaţiile de sulfuri vulcanogen-sedimentare metamorfozate din zona Hărlăgia în contextul litostratigrafic magmatogen şi tectonic al seriei epimetamorfice de Tulgheş-Cambrian inferior (Carpaţii Orientali). *Stud. tehn. econ. Inst. geol. geofiz., seria* A, 12, p. 77—124, Bucureşti.

— (1980) Replissements alpins régionales des metamorphites precambriennes et paléozoïques de la chaîne carpatique. Un exemple : Carpates Orientales. *An. Inst. geol. geofiz.*, LVI, p. 325—365, Bucureşti.

— (1984) Despre pînzele de şariaj cu metamorfite din partea sudică a zonei cristalino-mezozoică din Carpaţii Orientali. *D. S. Inst. geol. geofiz.*, LXVIII, 5, (1981). Bucureşti.

R a m s a y J. G. (1962) The geometry and mechanics of formation of "Similar" type folds. *Journ. Geol.* 70, 3, p. 309—327, Chicago.

S i t t e r L. U. (1969) Geologie structurală. Edit. tehnică, p. 1—499, Bucureşti.

T u r n e r F. S., W e i s s L. E. (1963) Structural analysis of metamorphic tectonites, p. 1— 545, *Mc. Graw-Hill Book Comp., Inc.*, New-York — San Francisco — Toronto—London.

DONNÉES GÉOLOGIQUES CONCERNANT LE CRISTALLIN DES CARPATHES ORIENTALES [1]

PAR

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L'ilot Cristallin du Nord, comme on a nommé le noyau cristallin autour duquel se sont formés les Monts du Bistrița, a constitué l'objet des recherches géologiques du XIX-ème siècle.

Concernant la pétrogenèse et la structure du noyau cristallin on a formulé plusieurs hypothèses, dont on peut retenir :

— le noyau cristallin s'est formé à la suite de plusieurs phases de métamorphisme régional, phases qui appartiennent à quelques orogenèses du Précambrian supérieur et du Paléozoïque ; la structure du noyau cristallin est caractérisée par la séparation de plusieurs systèmes superposés des nappes de charriage d'âge alpin ;

— les roches cristallines sont le résultat du métamorphisme dans la coupole d'un batholite granitique et la structure du massif est considérée comme normale ;

— les séries cristallines anciennes, métamorphisées régionalement, ont subi des processus de feldspatisation métablastique, potassique sur alignement, qui correspondent à l'actuelle série des gneiss œillés de Rarău. Les rapports structuraux anormales sont dûs aux déversements d'après des failles inverses.

Nous allons présenter ci-dessous une série de données géologiques concernant la série des gneiss œillés de Rarău, obtenues grâce à plusieurs travaux structuraux (forages et galeries), qui ont traversé la série de l'ouest à l'est entre la cote maximale de 1450 m (affleurements) et la cote 0 (fig. 1). Les données concernent les particularités pétrographiques des phases métamorphiques, leur âge et les éléments structuraux majeurs, présents dans l'actuelle position tectonique de la série, dans l'ensemble de noyau cristallin.

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La série des gneiss oeillés type Rarău, d'après les recherches géologiques effectuées, est constituée d'une alternance de micaschistes, quartzites feldspathiques faiblement micacés, amphibolites avec ou sans grenats, les roches constituent des bancs épais de centaines de mètres et bordent, sur la direction NO-SE du cote 0 et E, le complexe des gneiss oeillés développé dans la zone axiale de la série.

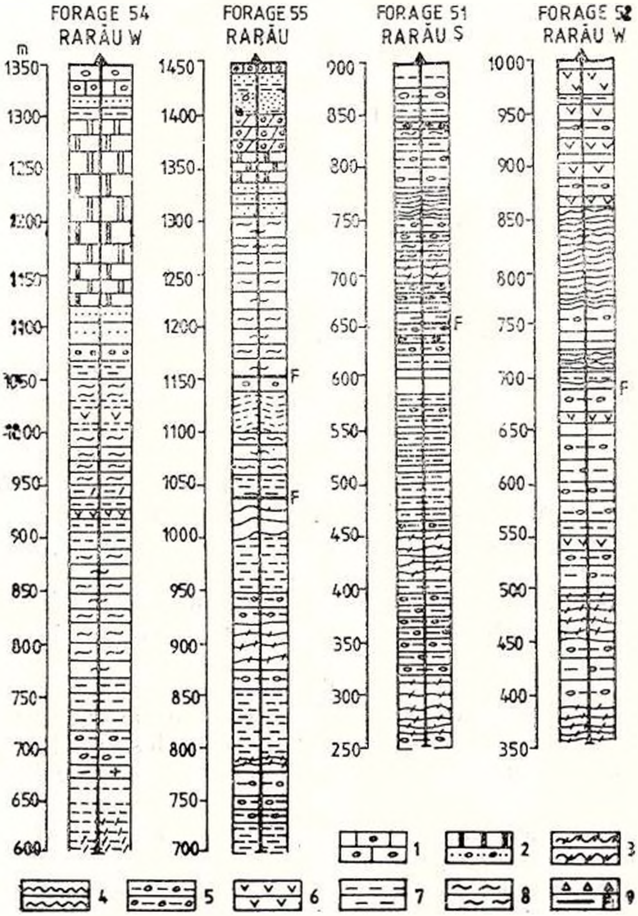

Fig. 1. — Forage 54. Rarău W, forage 55 Rarău, forage 51 Rarău, forage 51 Rarău S et forage 52 Rarău W.

Crétacé inférieur + moyen : 1, calcaires organogènes, schistes argilo-gréseux. Trias inférieur + moyen : 2, calcaires dolomitiques, conglomérats, grès microconglomératiques. La série des gneiss oeillés type Rarău ; 3, micaschistes entièrement chloritisés ; 4, micaschistes rétromorphes ± grenats ; 5, schistes quartzitiques micacés ; 6, amphibolites ± grenats ; 7, schistes quartzeux feldspathiques ; 8, gneiss type Rarău ; 9, brèche de faille. Faille.

Le complexe des gneiss oeillés est constitué d'une association intime de : gneiss oeillés à microcline rose-rouge ou gris-blanchâtre, des bancs épais de quelques mètres des roches porphyrogènes, des métagranites, pegmatites, micaschistes à rares yeux de microcline rose, amphibolites à rares yeux de microcline rose-rouge, des gneiss tonalitiques, métadacites, métadiorites, granites au schachbrettalbite. On doit mentionner que dans le cadre des bancs de micaschistes et d'amphibolites à rares yeux de microcline on constate à l'oeil nu l'augmentation de la fréquence des porphyroblastes de microcline jusqu'à ce que les schistes soient transformés dans de vrais gneiss oeillés (cote 980 Gemenea pour les amphibolites, cote

900 Runc Mic pour les micaschistes). Fréquemment, à la limite du passage, on peut observer des lits centimétriques, jusqu'à décimétriques de quartz et feldspath rouge-rose. A l'intérieur de ces lits on constate la présence des reliquats de quartz et de micas, surtout de muscovite.

Les principaux minéraux pétrogènes de la série sont : le plagioclase — présent dans la majorité des roches — offre une composition formée, en fonction du type pétrographique, à partir de l'albite et jusqu'à l'oligoclase-andésine. Elle est développée sous forme des porphyroblastes, mais aussi comme pétroblastes. Les porphyroblastes, fréquents dans les amphibolites, présentent d'habitude une composition oligoclase-andésine ; sur eux se forme secondairement l'épidote ; ils sont cataclasés et forment fréquemment des macles courbés.

Les pétroblastes représentent la forme caractéristique de blasthèse du plagioclase acide An 5-15 ; l'amphibole, représenté par la hornblende verte, rarement brune, est le principal constituant des amphibolites. La hornblende est partiellement actinolitisée ; le grenat (almandin) et fréquemment présent dans les micaschistes et les amphibolites et moins fréquent, dans les quartzites feldspathiques micacés et dans les gneiss oeillés. Il apparaît sous la forme des porphyroblastes idiomorphes des dimensions millimétriques. Très rarement, le grenat apparaît non alteré, comme dans l'horizon minier 940 m — Rarău, où il présente une couleur rose. Dans des cas pareils, il offre des structures rélictiques, englobant des micas. Dans bien des cas il est maintenant casé et substitué métasomatiquement par chlorite et carbonate. La substitution par chlorite peut être partielle jusqu'à totale et dans ce dernier cas, la chlorite conserve le contour du grenat. La chlorite est féreuse-pennine. Les substitutions par des carbonates se font d'habitude sur des fissures ; le disthène est présent sporadiquement. Il a été signalé dans deux sections minces des gneiss granitiques, prélevés du cours supérieur de la vallée Ungureni (Rarău) ; la biotite et le muscovite sont des substituents fréquents de tous les types pétrographiques des roches appartenant à la série des gneiss oeillés type Rarău. Dans quelques échantillons des micaschistes aux grenats, prélevés de l'horizon minier 780 m, dans le compartiment structural Rarău on a observé des reliquats de biotite en muscovite largement développés. Ce fait, ainsi que l'existence d'une biotite aux contours empâtés (Gemenea horizon minier 1050 m) peuvent suggérer l'existence d'une biotite formée dans une phase métamorphique plus ancienne, biotite transformée en muscovite dans les phases métamorphiques plus jeunes.

Le forage 50 Gemenea au mètre 530 (cote absolue 390 m) a mis en évidence, intercalé par des diabases, un banc des micaschistes, épais d'un mètre, contenant de la biotite transformée en fibrolite.

Les deux micas, dans le cadre des roches, sont disposés d'après les directions S_1 et S_2, qui forment entre eux un angle d'environ 30^c ; le microcline est le minéral qui définit les gneiss oeillés type Rarău. Elle apparaît comme porphyroblastes, pétroblastes, filonets et lentilles millimétriques. Comme pétroblastes elle participe à la constitution de la mésostase et elle a été formée par la substitution métasomatique du plagioclase acide et du quartz. Dans cette situation elle conserve un aspect frais.

Les porphyroblastes aux dimensions variables, de quelques millimètres à quelques centimètres, se sont formées par la substitution des porphyroblastes de plagioclase, oligoclase-andésine avec lesquelles elles forment des pertites, plus rarement des antipertites. Les porphyroblastes sont impurifiées par des oxydes de fer, ce qui confère au minéral une couleur rose-rougeâtre. Quand le minéral est pur, les porphyroblastes présentent une couleur blanche-grise (c'est le cas des gneiss blanc-gris sur la vallée Gemenea et Ceahlău-ouest) et elles sont beaucoup plus cataclasées que d'habitude. Dans quelques cas, assez rarement, on a signalé la disposition des porphyroblastes de microcline avec leur longue axe transversal sur la structure de la roche. Le pourcentage de sa participation dans la composition des gneiss oeillés est variable, tant sur la direction de la série, que son pendage.

Dans le forage 50 — vallée Gemenea — à la profondeur de 700 m (cote absolue 140 m), les porphyroblastes de microcline sont arrondies mécaniquement, prenant un aspect de galets ; dans ce cas les gneiss oeillés offrent l'aspect d'un phyllite aux porphyroblastes arrondies de microcline ; le quartz est le principal constituant des roches. Il apparaît sous forme de pétroblastes, filonets et lentilles, disposées au long de la schistosité de la roche. Avec le plagioclase il forme des concrétions mirmékitiques et, inclus dans le microcline, il est optiquement anomal. Par des recristallisations il corrode le microcline. Normalement il présente une extinction ondulatoire ; les ankérites présentent des corps lenticulaires, boudinés, intercalés entre les micaschistes aux grenats et formés par substitution métasomatique des micaschistes. Les corps ont des épaisseurs métriques et des longueurs jusqu'à quelques centaines de mètres. Ils sont disposés dans la base des gneiss oeillés tant à l'O qu'à l'E, formant des corps concordants dans les micaschistes.

La substitution des micaschistes par les carbonates type ankéritique est partielle ou bien totale ; les micaschistes sont conservés sous forme d'îlots reliquats dans la masse des carbonates, qui offrent ainsi un aspect brécheux. Il paraît que le développement maximal des carbonates se produit dans les compartiments structuraux aux processus de granitisation les plus intenses (par exemple les compartiments Ceahlău, Rarău-Gemenea).

Les principaux minéraux constituants des roches de la série des gneiss oeillés sont affectés par les processus de transformation rétromorphique.

La chloritisation dans laquelle la chlorite, d'habitude la pennine, substitue la biotite et le grenat. La substitution est partielle ou totale. Dans les cas du grenat, la chlorite substituante conserve le contour de celui-ci. La biotite chloritisée peut être reconnue d'après les reliquats, le contour du minéral substitué et les aiguilles de sagénite présentes dans la masse de la chlorite. La substitution par chlorite présente un caractère progressif avec la profondeur. Ainsi, dans les compartiments au développement maximal des processus de granitisation, en commençant par les horizons miniers 850 m, 750 m, et 700 m, la biotite, substituée presque en totalité, peut être reconnue d'après la fréquence des aiguilles de sagénite et chlorite. Dans des situations pareilles, des gneiss oeillés peuvent être diagnostiqués comme gneiss oeillés au chlorite. Parmi les autres processus de transformation rétromorphique on retient : la séricitisation, la fengitisation, l'albitisation et l'argilisation.

La cataclase des roches et des minéraux constituants de la série favorise les transformations physico-chimiques des minéraux.

Les datations radiométriques par les méthodes K-Ar et Rb-Sr, effectuées sur des minéraux et roches globales appartenant à la série des gneiss oeillés de Rarău, indiquent un âge qui correspond aux orogenèses calédonienne, hercynienne et alpine.

— L'orogenèse calédonienne a été mise en évidence par la datation d'un micaschiste aux grenats de Gemenea (cote 1050 m). La biotite aux contours empâtés et entouré des dépôts de séricite, a été datée par la méthode K-Ar-425$\pm$20 MA.

— L'orogenèse hercynienne est mise en évidence par les datations K-Ar effectuées sur muscovite, biotite, amphiboles et roches globales. Dans le cadre de cette orogenèse ont été mises en évidence les isochrones suivantes : 368$\pm$7 MA ; 275$\pm$3 MA.

Les datations par la méthode Rb-Sr, ont précisé les âges suivants : — micaschistes aux grenats (1 échantillon) isochrone 315$\pm$4 MA ; — gneiss oeillé type Rarău (1 échantillon) isochrone 275$\pm$7 MA.

Le rapport des isotopes de Sr indique l'origine crustale du gneiss oeillé type Rarău.

L'orogenèse alpine a été mise en évidence par les datations K-Ar sur la séricite, les minéraux argileux et un échantillon de stilpnomélane. Les âges sont échelonnés sur l'intervalle de 200–100 MA.

Dans la région on a exécuté quatre forages structuraux emplacés sur le flanc est de la série, à une distance d'approximativement 3 km du contact avec le Flysch.

Les forages 2 Ceahlău W_1 et 1 Ceahlău W_2, après avoir traversés les roches appartenant à la série des gneiss oeillés, ont intercepté la ,,Ligne Centrale Carpathique" aux profondeurs de 560 m et respectivemment de 900 m, c'est-à-dire aux cotes absolues de 100 m et 50 m.

Le forage 50 Gemenea, déplacé aussi à peu près à 3 km ouest du contact avec le Flysch, a été arrêté à la profondeur de 750 m dans les gneiss oeillés, complètement dynamiquement rétromorphisés.

Ces données mettent en évidence (fig. 2) que la ,,Ligne Centrale Carpathique" présente dans le compartiment sud (Ceahlău) des pendages de 20°–30° vers sud-ouest, pendant que dans le compartiment nord du noyau cristallin ce contact peut avoir des pendages d'environ 50°–60°.

L'analyse des principaux minéraux pétrogènes qui participent à la formation de la série de gneiss oeillés type Rarău nous a conduit à la séparation de quelques fases métamorphiques :

— fase I — la plus ancienne. Pendant cette fase se sont formés : l'almandin, le plagioclase, la hornblende verte (éventuellement brune), la biotite (actuellement aux bordures empâtées), le disthène (trés rare), le quartz. Après ces minéraux, le métamorphisme s'est déroulé dans les conditions du faciès almandin-amphibolite. D'après l'âge de la biotite aux contours empâtés, 425$\pm$20 MA, cette fase appartient à l'orogenèse calédonienne ;

— fase II — la suivante dans la succession. Comme minéraux indicateurs se sont formés : probablement le grenat frais, la biotite commune, le muscovite, l'actinote sur la hornblende, l'épidote sur le plagioclase, la chlorite sur le grenat et sur la biotite.

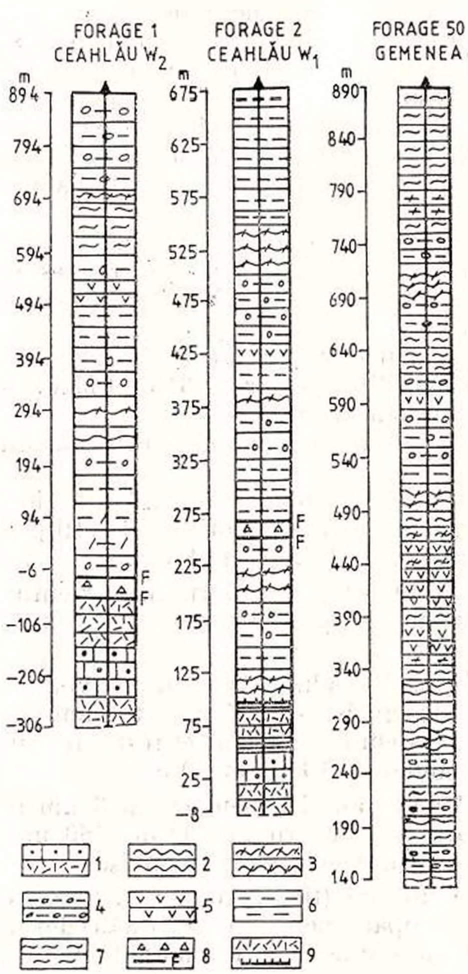

Fig. 2. — Forage 1 Ceahlău W_2, forage 2. Ceahlău W_1 et forage 50 Gemenea. Crétacé inférieur . 1, brèche marno-calcaire, marno-calcaire, grès calcaire ; 2. micaschistes rétromorphes ± grenats ; 3, micaschistes entièrement chloritisés ; 4. schistes quartzeux micacés ; 5, amphibolites ± grenats ; 6, schistes quartzeux feldspathiques ; 7, gneiss type Rarău ; 8 brèche de faille. Faille ; 9, brèche de faille. Ligne de charriage. 2–8 = la série des gneiss oeillés type Rarău.

Les substitutions physico-chimiques du développement des réactions métamorphiques peuvent être attribuées au faciès des schistes verts, d'isograde biotite.

Les datations radiométriques effectuées, indiquent pour ces recristallisations métamorphiques des âges appartenant à l'orogenèse hercynienne.

Pendant la deuxième étape de l'orogenèse hercynienne, dans les mêmes conditions, dans la série se sont déroulés des processus métasomatiques alcalins, qui ont conduit à une granitisation partielle ou bien totale des roches affectées, processus qui ont abouti à la formation du gneiss oeillé type Rarău (migmatites métablastiques). D'après les âges Rb-Sr (275±7 MA) et d'après l'isochrone K-Ar (275±3 MA) la formation du gneiss oeillé occupe une ultime étape de l'orogenèse hercynienne.

— fase III appartient au métamorphisme cataclastique, associé à la dynamique alpine. Pendant cette fase ont continué les processus de chloritisation et se sont formées : la fengite, la séricite, le stilpnomélane, les minéraux argileux.

Cette fase est définie par des processus dynamiques (très élevés) qui ont abouti à la cataclase des roches et des minéraux substituents et à leur transformation dans des minéraux, correspondants aux nouvelles conditions physico-chimiques. Les datations radiométriques par la méthode K-Ar sur ces minéraux indiquent une série des âges appartenant à l'orogenèse alpine.

Actuellement, la série des gneiss oeillés type Rarău est une série cataclasée de fases métamorphiques regressives, commençant avec le faciès almandin-amphibolite, continuant avec le faciès des schistes verts et avec le métamorphisme cataclastique.

L'origine crustale des gneiss oeillés, attestée par les isotopes de strontium, indique que dans la deuxième étape de l'orogenèse hercynienne le noyau cristallin faisait déjà partie de la composition de la croûte.

La présence d'une fase métamorphique prépaléozoïque ne peut pas être exclue, mais il n'y a pas de données pour pouvoir la soutenir.

Pendant l'orogenèse alpine, à part le métamorphisme cataclastique, les processus dynamiques, causés principalement par la subduction de la plate-forme est-européenne, ont déterminé aussi le charriage du noyau cristallin sur la zone du Flysch Carpathique crétacé inferieur.

BIBLIOGRAPHIE

Bercia I., Kräutner H., Mureșan M. (1976) Pre-Mesozoic Metamorphites of the East Carpathians. *An. Inst. geol. geofiz.*, L, p. 37–70, București.

Kräutner H., Kräutner F., Tănăsescu A., Neacșu V. (1976) Interprétation des âges radiométriques K/Ar pour les roches métamorphiques régénérées. Un exemple-les Carpathes Orientales. *An. Inst. geol. geofiz.*, L, p. 167–229, București.

Pavelescu L., Cristea I., Soroiu M. (1979) Evolution of the crystalline bassement of România in the light of radiometric dating. *Rev. roum. géol., géophys. géogr., Géologie*, 23, p. 23–31, București.

— Pop Gh., Aileni Gh., Cristea I., Soroiu M. (1976) Evoluția nucleului cristalin al Carpaților Orientali în lumina datărilor radiometrice. *Stud. cerc. geol. geofiz.*, 14,1, p. 39–49, București.

Pop Gh., Baciu Fl., Gâță Gh., Popescu M., Soroiu M. (1980) Rapport, les archives de l'I.M.R., Bucarest.

Rădulescu I. (1976) Cristalinul Bistriței. Stratigrafia și structura masivului cristalin din partea de nord a Carpaților Orientali. *Stud. cerc. geol. geofiz. geogr. seria Geologie*, 14, 4, București.

CONTRIBUTIONS TO THE KNOWLEDGE OF THE STRATIGRAPHY AND TECTONICS OF THE CRYSTALLINE SCHISTS SOUTH OF MUNTELE MARE, BETWEEN CÎMPENI AND SĂLCIUA (NORTHERN APUSENI MOUNTAINS)[1]

BY

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The area presented in this paper is situated in the southern part of the Northern Apuseni Mountains, namely south of the Muntele Mare granitic massif (fig. 1), where previous researches (Dimitrescu, 1958, 1962) pointed out the concept of nappe structure of this unit of the Romanian Carpathians.

In accordance with this concept, on the geological map of the S. R. Romania, sheet Cîmpeni, scale 1 : 50,000 (Dimitrescu et al.,

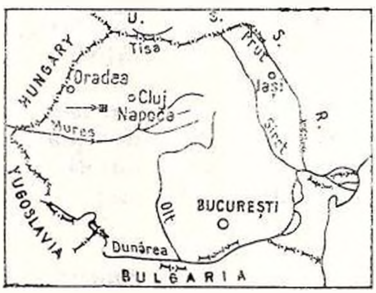

Fig. 1. — Location of the region on the map of the S. R. Romania.

1973) as well as in a previous paper on the geology of the Apuseni Mountains (Ianovici et al., 1976) four major tectonic units are separated :

— the Bihor Autochthon, comprising the Someş and Arada metamorphic series, the Muntele Mare granite and Werfenian sedimentary deposits ;

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— the Codru Nappe, comprising the Codru migmatite series, the Biharia Series, the Muncel Series and Permian sedimentary deposits;
— the Muncel-Lupşa Nappe, comprising the Biharia Series, the Muncel Series and the Marble Series;
— the Baia de Arieş Nappe, comprising the Baia de Arieş Series, the Marble Series and the Vincţa granites.

These tectonic units overlie one another from the south towards the north in the inverse order of the above enumeration, constituting a pile of superposed nappes which represent in fact the eastward prolongation of the structures known in the Bihor and Codru Moma Mountains.

The six metamorphic series making up these tectonic units are characterized as follows:

1. *The Someş Series* consists of an alternation of micaschists, paragneisses and amphibolites lying over anatectic migmatites. Its age is early Precambrian, while the metamorphism is pre-Hercynian in age.

2. *The Arada Series* consists of sericite-chlorite (dominantly) quartzitic schists with metaporphyry intercalations, chlorite schists with albite porphyroblasts, basic and acid metatuffs and, only accidentally, graphite schists and dolomitic limestones. Without following a certain stratigraphic position, rare levels of chloritized garnet schists showing a retromorphic character are separated. At the base of this series a thick set of beds of sericite-chlorite schists with biotite and muscovite lies directly over the mesometamorphics of the Someş Series. According to Visarion and Dimitrescu (1971) the age of this series is between terminal Precambrian and Lower Cambrian.

There were some contradictory discussions about its belonging to the Someş Series (as an upper retromorphosed part) or to an independent series, and the existence of a stratigraphic and metamorphism unconformity between the Arada and the Someş Series (Giuşcă et al., 1967).

3. *The Biharia Series* consists mainly of chlorite schists with albite porphyroblasts, including intercalations of albitic gneisses, amphibolic schists, orthoamphibolites, epidotites and dolomites. This series is Upper Precambrian-terminal Precambrian in age (Visarion, Dimitrescu, 1971).

4. *The Muncel Series* consists of an alternation of sericite-chlorite schists, quartzitic schists of phyllitic aspect, black quartzites, schists with albite prophyroblasts, metakeratophyres, metaporphyries, acid metatuffs and metatuffites, amphibolic schists etc. It is Lower Cambrian-Upper Cambrian in age (Visarion, Dimitrescu, 1971).

5. *The Baia de Arieş Series* is characterized by the predominance of the "garnet phyllites and garnet quartzitic schists" including intercalations of paragneisses, amphibolites and limestones. It is Lower Proterozoic in age (Visarion, Dimitrescu, 1971).

6. *The Marble Series* is represented by marble-like, white or greysh large-sized limestones that crop out in the western part of the area; they

are considered by Dimitrescu (in Ianovici et al., 1976) to overlie transgressively the Muncel Series. They are Cambrian-Ordovician in age (Visarion, Dimitrescu, 1971).

In the eastern part of the investigated region, between the Misernița Crest and the Sălciua Valley, there develops a sequence of terrigenous rocks consisting of conglomerates, quartzites, dolomites, ankeritic dolomites and white or rosy marble-like limestones known as *the Vulturese-Belioara Series.*

The analysis of all the known data raised the following problems:

— some common features of the Biharia and Muncel Series (the existence of the volcanic, basic and acid activities) pointed to the possibility of considering them as a single series;

— the boundaries between the tectonic units drawn on the previous maps do not look like overthrust planes of nappes;

— the "Arada Series" and its assignment to one of the major stratigraphic units known in the Northern Apuseni Mountains;

— the stratigraphic and chronological assignment of the Marble Series;

— the assignment of the "garnet phyllites" to the Muncel or Baia de Arieș Series;

— to establish the normal stratigraphic sequence and position of the Vulturese-Belioara Series with respect to the other metamorphic series of the region.

The very detailed researches carried out by us (Pitulea et al., 1976–1980) led to the following conclusions:

— the two already delimited Biharia and Muncel Series represent in fact one and the same series, the "Biharia-Muncel Series", consisting of five stratigraphic complexes disposed in normal superposition;

— the boundaries between the two mentioned series ("Biharia" and "Muncel") are normal transition boundaries, no stratigraphic or metamorphic unconformity existing between them and both series were assigned to the same age interval. At the boundary between the terrigenous and the basic volcanic complexes of these series there appears a transition zone, marked by the presence of some sets of beds of phyllites containing albite, a mineral omnipresent in the basic complexes (Fig. 2).

— within the investigated area what appears as the Arada Series on the previous maps is, in our opinion, the basal part of the Biharia-Muncel Series; thus the same sequence is likely to extend farther east within the so-called "northern epimetamorphic rocks", especially if we take into account the conclusions of Mureșan (1971). According to this author, at the basal part of the Biharia Series (considered a complex) there is another complex, called the volcano-sedimentary complex, showing similar characteristics to the basal part of the Biharia-Muncel Series in the region, respectively to the former Arada Series. Within the above-presented sequence only the lower set of beds consisting of biotite and muscovite sericite-chlorite schists overlying directly the Someș Series and representing the retromorphous "cover" should be assigned to the Arada Series. In this case it is possible that the idea put forward by Giușcă et al. (1967) concerning the existence of a stratigraphic and metamorphism unconformity between the Arada Series (considered by the above authors of sure epimetamorphic character, therefore like the Biharia-Muncel Series) and the Someş Series may be real.

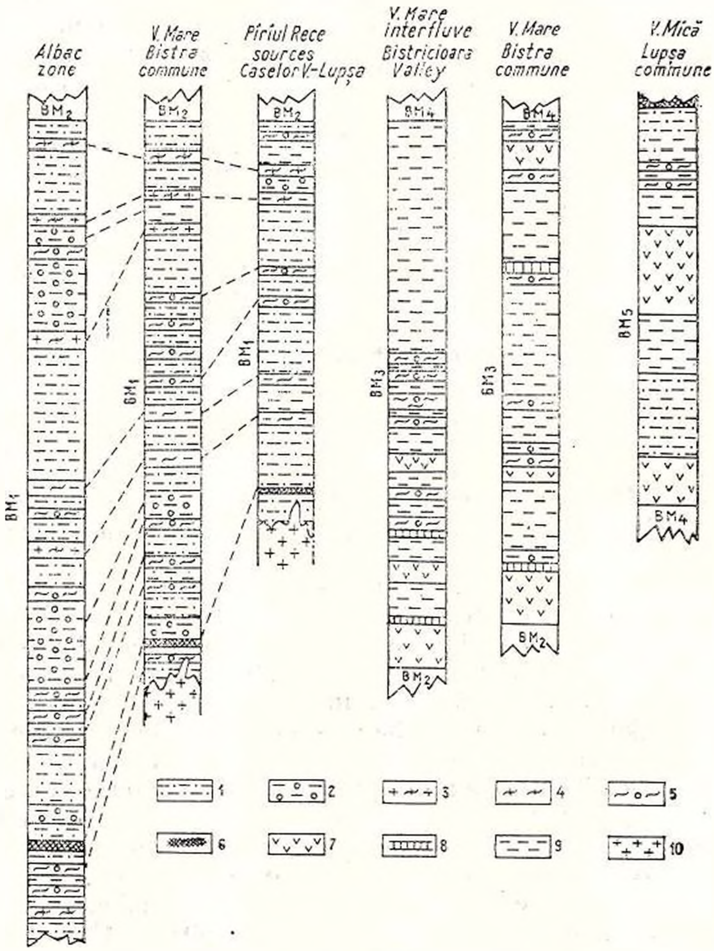


Fig. 2. — Stratigraphic columns within the Biharia-Muncel Series between Albac and Lupşa

1, quartzitic sericite-chlorite schists; 2, chlorite schists with albite porphyroblasts; 3, metarhyolites; 4, acid metatuffs; 5, basic metatuffs; 6, black quartzites; 7, albite phyllites; 8, crystalline limestones; 9, sericite-chlorite phyllites; 10, Muntele Mare granites.

The accidental presence of garnet, mostly chloritized, cannot be an argument for assigning the base of the Biharia-Muncel Series (in our acceptation) to the retromorphosed Someş Series (as the Arada Series is defined by many researchers).

The chloritization of garnets, which are present especially in the schists in the vicinity of the basic metavolcanics, is the effect of a dynamic, not regional retromorphism, so that the part of the Arada Series we assigned to the Biharia-Muncel Series does not represent a retromorphosed mesozone, but a sequence of rocks metamorphosed in the greenschist facies, the biotite isograde, like the Biharia-Muncel Series. An additional argument is the composition of the plagioclase feldspars within the basic complexes, determined both by us and P a n a i t e (1978) in the presented region and respectively in the Albac zone ; it is an albite containing below 8% anorthite.

The researches carried out on the Albac Valley, between Albac and Arada, considered as a classic outcropping zone, led us to the same conclusion. The comparison of the stratigraphic column on the Albac Valley with the stratigraphic columns on the Valea Mare a Bistrei and the upper course of the Caselor Valley of Lupşa, situated farther east, shows similarities as regards the lithological development and the stratigraphic sequence of the constituent formations (Fig. 2).

In conclusion, we think that the term "Arada Series" should comprise only a part of what was initially assigned to it, namely the thin retromorphous "cover" of the Someş Series, most of the sequence being represented in fact by the Biharia-Muncel Series as defined by us. D i m i t r e s c u (in I a n o v i c i et al., 1976) puts forward the idea that the whole Arada Series might represent an equivalent of the Biharia and Muncel Series, but in another tectonic unit ;

— the transgressive character of the "Marble Series" cannot be supported as there are intense brecciations in the outcrop areas ; also, these marbles cannot be considered to form an independent series, the more so as south of the Arieş Valley these rocks are clearly intercalated in the lower part of the Baia de Arieş Series. The sometimes unconformable position of the marble with respect to the underlying formations is due to tectonic causes, they representing in fact an upper digitation of the Baia de Arieş Nappe (P i t u l e a et al., 1975, 1980) ;

— the garnet phyllites occurring at the upper part of the Biharia-Muncel Series and constituting real stratigraphic levels cannot be an argument for supporting the existence of the Baia de Arieş Series in the Lupşa Valley, where the latter would exhibit retromorphous characters due to its thrust over the "Muncel Series". Unlike the garnets of the lower complex of the Biharia-Muncel Series, as defined by us , the garnets within these phyllites are seldom chloritized ;

— the Vulturese-Belioara Series, initially included in the Păiuşeni Series (M â r z a, 1969) is considered by D i m i t r e s c u (in I a n o v i c i et al., 1976), who gave its present name, as an independent series. Although this series has a less characteristic development in the presented zone, the sections carried out on the main valleys crossing the metamorphic formations, between the Sartăş Valley and the Ocolişul Valley, provided a new image on its relations with the Biharia-Muncel Series and on the real stratigraphic succession of the component formations. Thus, we agree with D i m i t r e s c u, considering this succession as an independent series and not a complex, as stated by S o l o m o n et al. (1978). The Vulturese-

Belioara series overlies transgressively the Biharia-Muncel Series and does not represent a complex of the Baia de Arieș Series, as stated by S o l o m o n et al. (1978). This later hypothesis cannot be accepted as the base of the series consisting of extremely slightly metamorphosed conglomerates and quartzites — resembling the Werfenian laminated conglomerates which occur between Cîmpeni and Albac (assigned by B o r c o ș and B o r c o ș, 1955, first to this Mesozoic unit) — attests the fact that the metamorphism that affected these deposits did not exceed the upper part

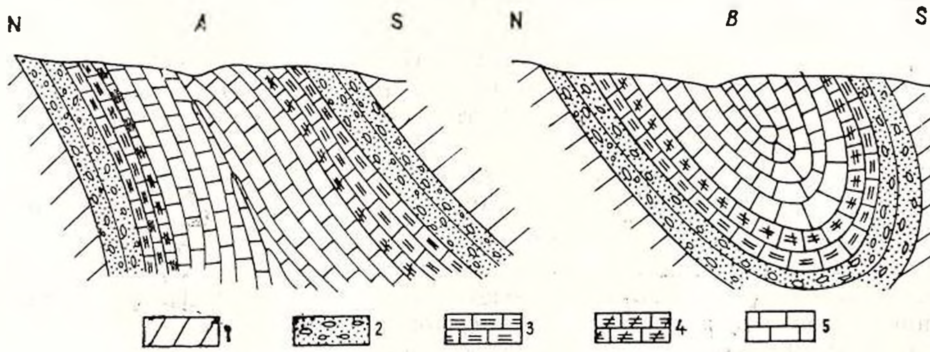


Fig. 3. — Geological section within the Vulturese-Belioara Series.
A, old interpretation (M â r z a , 1969); B, new interpretation (P i t u l e a et al., 1977). Biharia-Muncel Series — 1, phyllites; Vulturese-Belioara Series — 2, metaconglomerates and metaquartzites with phyllite intercalations; 3 graphite dolomites; 4 ankeritic dolomites; 5 white saccharoidal limestones.

of the greenschist facies. This does not allow to assign those rocks to the Baia de Arieș Series, metamorphosed at the amphibolite facies, the sillimanite zone (T r i f, 1952). M â r z a (1969) presents the stratigraphic sequence of this series (later generally accepted) : conglomerates and quartzites with phyllite intercalations, interlayered graphite dolomites, pink-yellow ankeritic dolomites and white, sometimes rosy, large crystallized, marble-like limestones; he concludes that the upper part of the series (marble-like limestones) belongs to the Baia de Arieș Series, penetrating the core of an anticline slightly overturned towards the north, overlain by conglomerates and dolomites that were considered as equivalents to the Păiușeni Series.

Our investigations indicated that there is a continuity of sedimentation between dolomites and the marble-like limestones, while from the structural point of view a syncline and not an anticline may be distinguished, the whole sequence representing the Paleozoic cover of the Biharia-Muncel Series (Fig. 3).

Stratigraphy of the Crystalline Schists

Our conclusions reveal that three crystalline series are present within the territory presented : the Baia de Arieș Series, the Biharia-Muncel Series and the Vulturese-Belioara Series.

The brevity of this paper does not allow a comment on the lithological composition and the stratigraphic succession of these series; instead, we shall present in detail their stratigraphic columns in the plate (Pl.). Still we remind that on this area the Baia de Arieș Series is only represented by its two lower complexes (BA_1 and BA_2), separated by P i t u l e a et al. (1975, 1980) within the Baia de Arieș Spur, where it shows the most characteristic development.

We separated five stratigraphic complexes in the Biharia-Muncel Series: the lower terrigenous complex (BM_1), representing most of the Arada Series, the lower basic volcanic complex (BM_2), the acid volcano-sedimentary complex (BM_3), the upper basic volcanic complex (BM_4) and the upper terrigenous complex (BM_5). The complexes BM_2 and BM_4 were previously considered to represent the Biharia Series, while the complexes BM_3 and BM_5, the Muncel Series. The lithostratigraphic sequence of the Vulturese-Belioara Series was concisely presented above.

Tectonics

One of the main problems concerning the tectonics of the region was to establish to what extent the superposed nappe structure existing farther west is also present here.

Indeed, in the western part of the region, between Mihoiești and Albac, the tectonic units figured on the previous maps were confirmed. But we point out the fact that the dip of the overthrust planes is great, resembling some reverse faults.

East of Cîmpeni, some of these tectonic lines continue by quasivertical axial fractures which are likely to disappear gradually. This is the case of the contact between the Codru Unit and the Bihor Autochthon, as there are normal transition relations between the "Biharia" and "Muncel" Series.

The existence of the dislocation between BM_4 and BM_3 (the Muncel-Lupșa Nappe — according to D i m i t r e s c u) was not confirmed, but tectonic relations were pointed out between the upper complex BM_5 (in "Muncel facies", therefore younger) and the complex BM_4 (in "Biharia facies", therefore older according to the previous researchers). This system of axial fractures is present throughout the Biharia-Muncel Series, bringing about also the chloritization of garnets within the complex BM_1, when the garnet bearing rocks were also affected.

In the eastern part of the region, between the Lupșa Valley and the Sălciuța Valley, within this tectonic system there took place the thrust of the Baia de Arieș Series over the Biharia-Muncel Series (the Baia de Arieș Nappe, confirmed also on the Baia de Arieș Spur, where the Biharia-Muncel Series occurs in a few tectonic windows) during the Austrian structogenesis as well as the thrust (like a scale) of the same series over the paleontologically dated Senonian deposits (P i t u l e a et al., 1975) during the Laramian movements.

Another feature of the disjunctive tectonics is due to the system of newer dislocations, perpendicular to the structure, which brought about slips or vertical shifts of variable amplitudes, so that various levels, complexes and even series contact one another tectonically. Along the planes of these fractures there occurred rotations by almost 90° of the structure.

Thus a block (ice packs) tectonics is pointed out, which moved and rotated differently, tectonic blocks in respect of the resistant Muntele Mare granite that functioned as an obstacle to the tectonic efforts directed from the south towards the north.

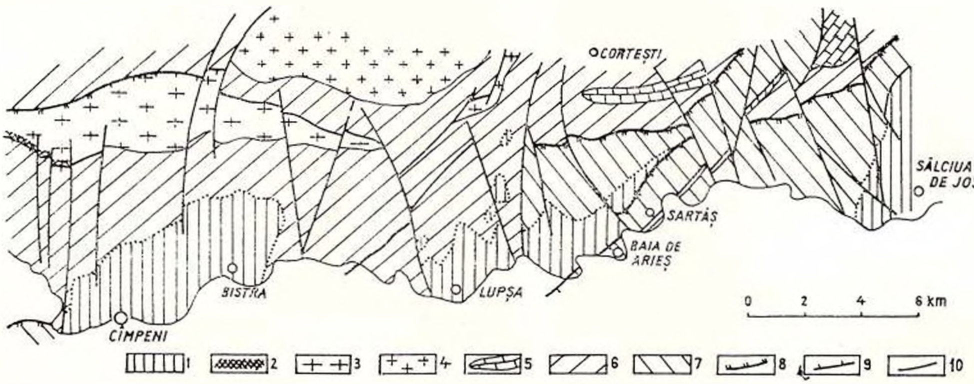


Fig. 4. — Tectonic sketch of the Cîmpeni-Sălciua region. 1 Upper Cretaceous deposits; 2. Permian deposits; 3 Codru granitoids and migmatites; 4. Muntele Mare granites; 5, Vulturese-Belioara Series; 6. Biharia-Muncel Series; 7, Baia de Arieș Series; 8, tectonic lines like a nappe; 9. tectonic lines like a scale; 10, normal and inverted faults.

Concerning the plicative tectonics we mention that, on the whole, the Baia de Arieș Series and the Biharia-Muncel Series form monoclines with northern vergency, disturbed by folds of reduced amplitude, which show ascending limbs towards the north.

The Vulturese-Belioara Series forms a syncline slightly overturned towards the north (Fig. 4).

Conclusions

The former Biharia and Muncel Series represent a single series (the Biharia-Muncel Series) consisting of five complexes among which there are no stratigraphic, tectonic or metamorphism unconformities. According to this conception the Muncel-Lupșa Nappe, figured on the previous nappes, is considered not to be present in the region.

On the whole the Arada Series represents the lower part of the Biharia-Muncel Series as defined by us.

The "Marble Series" represents in fact a carbonatic horizon intercalated within the lower complex of the Baia de Arieș Series.

There are overthrust relations between the Baia de Arieș Series and the Biharia-Muncel Series (the Baia de Arieș Nappe was generated during the Austrian diastrophism). Also, during the Laramian movements, a part of the Baia de Arieș Series was pushed towards the north, over the Senonian deposits.

All the deposits assigned to the Vulturese-Belioara Series form a single series, representing the Paleozoic cover of the Biharia-Muncel Series.

REFERENCES

Borcoş M., Borcoş E. (1962) Cercetări geologice şi petrografice în regiunea Runc-Segagea-Valea Ierii sat (bazinul văii Iara, Munţii Apuseni). *D. S. Inst. Geol.*, XLVII (1959–1960), p. 131–144, Bucureşti.

Dimitrescu R. (1958) Studiul geologic şi petrografic al regiunii dintre Gîrda şi Lupşa. *An. Com. Geol.*, XXXI, p. 51–119, Bucureşti.

— (1962) Studiul geologic şi petrografic al regiunii Arada-Muntele Mare. *D. S. Inst. Geol.*, XLVII (1959–1960), p. 275–279, Bucureşti.

— Bordea S., Bordea J. (1973) Harta geologică a României sc. 1 : 50 000, foaia Cîmpeni, Edit. Inst. Geol. Bucureşti.

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

Ianovici V., Borcoş M., Bleahu M., Patrulius D., Lupu M., Dimitrescu R., Savu H. (1976) Geologia Munţilor Apuseni. Ed. Acad. RSR, 631 p., Bucureşti.

Mârza I. (1969) Evoluţia unităţilor cristaline din sud-estul Muntelui Mare. Ed. Acad. RSR, 162 p., Bucureşti.

Mureşan I. (1971) Studiul mineralogic şi petrografic al formaţiunilor dintre Someşul Cald şi valea Căpuşului cu privire specială asupra substanţelor minerale utile. Rezumatul tezei de doctorat. 36 p., Bucureşti.

Pitulea G., Ghiţulescu T. P., Ghiţulescu I., Mărunţeanu Cr., Mustăţea E. (1975) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

— Ghiţulescu I., Mărunţeanu Cr., Florescu C., Berches S., Berches M. (1976) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

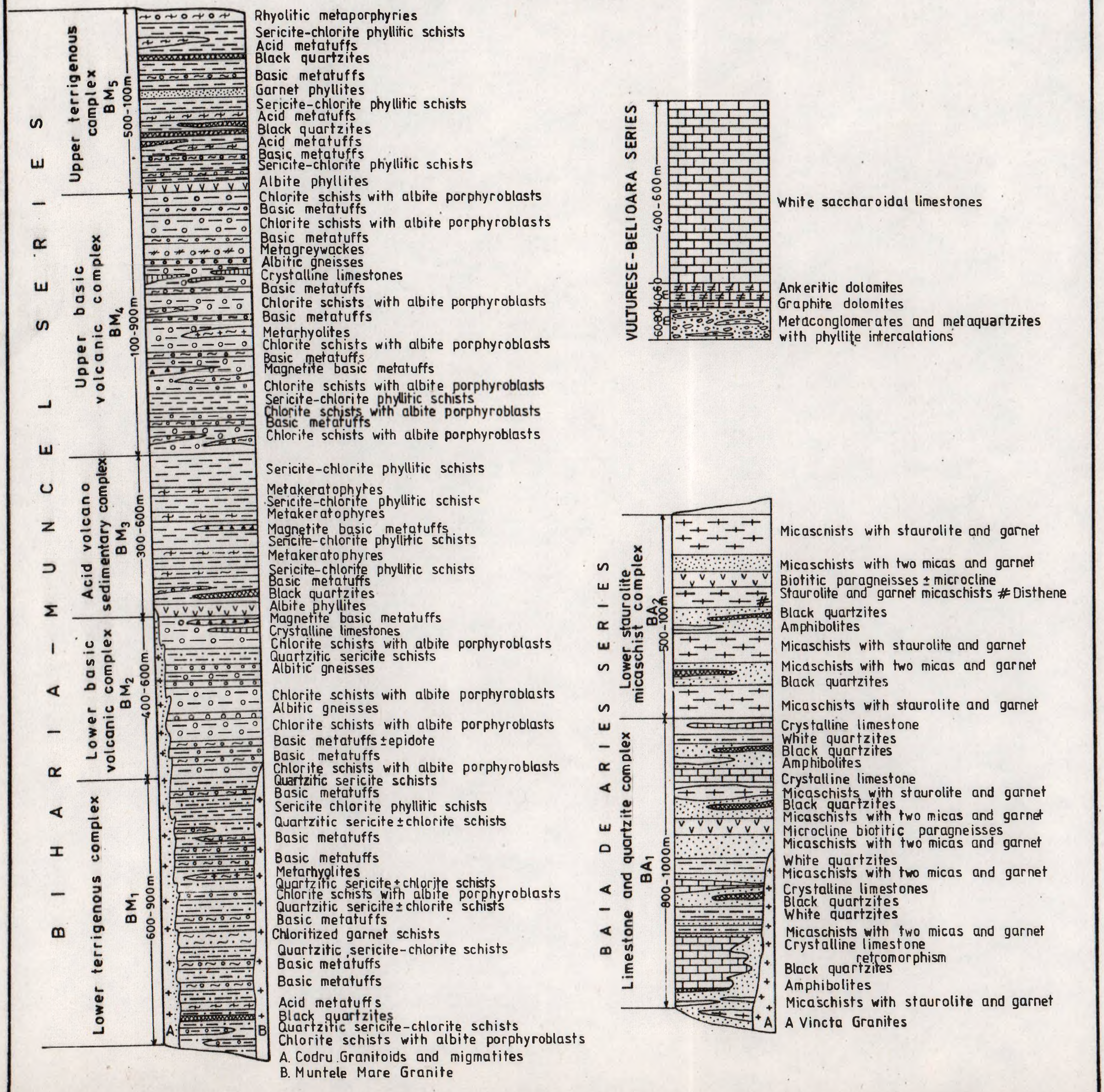
— Ghiţulescu I. (1980) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

Solomon I., Moţoi Gr., Moţoi A., Mărgărit Gh., Mărgărit M. (1978) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

Trif A. (1952) Silimanitul de pe valea Ierii. *Acad. R.P.R., Fil. Cluj, St. şt.*, III, 1–2, p. 142–153, Cluj.

Visarion A., Dimitrescu R. (1971) Contribuţii la determiarea vîrstei unor şisturi cristaline din Munţii Apuseni. *Anal. şt. Univ. Iaşi (Seria nouă), Secţia II-a, b*, XVII, 1–13, Iaşi.

GENERAL STRATIGRAPHIC COLUMNS OF THE BIHARIA-MUNCEL, BAIA DE ARIEȘ AND VULTURESE-BELIOARA SERIES BETWEEN CÎMPENI AND SĂLCIUA-APUSENI MOUNTAINS



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PRE-ALPINE STRUCTURAL ELEMENTS IN NORTH DOBROGEA[1]

BY

ANTONETA SEGHEDI[2]

Introduction

Most of the pre-Mesozoic rocks of North Dobrogea are exposed in its western part — the Măcin unit. This unit is a Kimmerian nappe (Mirăuță, in Patrulius et al., 1973) with a complex geological constitution, involving rocks ascribed to the Precambrian (polymetamorphic sequences of the Orliga and Megina Groups), Paleozoic sequences (low grade metamorphic rocks of the Boclugea Group, sometimes assigned to the Upper Precambrian; fossiliferous Silurian and Devonian deposits, Lower Carboniferous Carapelit Formation), various granites and some Mesozoic deposits.

The rocks of the Măcin Unit are involved in major NW-SE trending folds (Murgoci, 1914; Rotman, 1917; Cădere, 1925; Mirăuță, Mirăuță, 1962 a, b) with a southeasterly plunge (Mirăuță, 1966). Murgoci noticed a westward arching of these folds in the northwestern corner of the district. A Variscan age has been unanimously accepted for these major folds, Carboniferous deposits occurring in the syncline cores in the central part of the Măcin unit (Murgoci, 1914; Ianovici et al., 1961; Mirăuță, Mirăuță, 1962 a ; Mirăuță, 1966). However, Mesozoic deposits (Triassic and Jurassic rocks) lie in the same syncline cores to the south, at Camena (Cădere, 1925; Mirăuță, Mirăuță, 1964), pointing to the Alpine age (early Kimmerian phase — Mirăuță, Mirăuță, 1964), of the main structures of the Măcin unit (Cădere, 1925).

Structural Elements in Rocks of the Orliga Group

The Orliga Group is a polydeformed sequence of mainly psamitic and pelitic deposits, with subordinate basic and carbonatic layers. Recumbent isoclinal or tight folds (F_1 folds) with an axial planar schistosity (S_1) connected with kyanite, staurolite, garnet and sometimes silli-

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manite crystallisation occur in gneisses and micaschists of the Orliga Promontory. Upright isoclinal folds (F_2 folds) strongly deform these early structures and are responsible for the E-W lithological layering of the whole sequence. The most striking feature of the Orliga Group is their E-W trending, steeply southward dipping schistosity (S_2), axial planar to F_2 folds. Disrupted fold hinges within amphibolitic and gneissic layers (Fig. 1 a) suggest that the transposition of S_1 along S_2 planes has occurred.

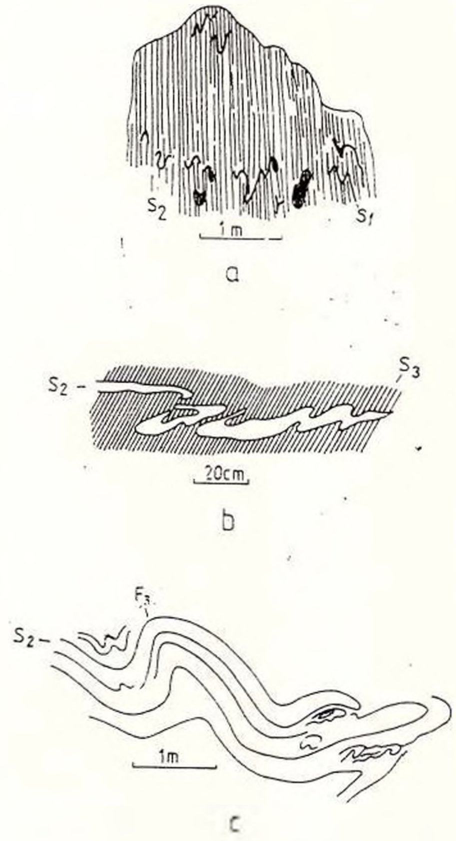

Fig. 1. — Structural elements in rocks of the Orliga Group.
a, Transposition of S_1 foliation on S_2 planes in the Sărărie Gneisses, Sărărie Hill (solid black-leucocratic layers); b, S_3 crenulation cleavage deforming S_2 foliation in the Vițelaru Gneisses, Vițelaru Hill (unornamented-microcline gneisses); c, F_3 fold affecting S_2 layering intrafolial fold, and F_2 in the Vițelaru calcsilicate gneisses Gorganu quarry.

This transposition is marked by biotite and muscovite recrystallisation, in large flakes, at the expense of amphibolite facies minerals. They form a flat lying or slightly eastward dipping lineation (L_2). A later event produced a NW-SE trending crenulation cleavage, axial planar to microfolds of kink type or to concentric open folds (F_3 folds) (Fig. 1 b, c). Microscopic kinking of phyllosilicate layers resulted during this folding episode, while in shear zones related to the same event quartz recrystallisation along S_3 planes occurs.

Superposed Deformations in the Megina and Boclugea Groups

Early structural elements in micaschists of the Megina Group are oriented trails of quartz inclusions in garnet porphyroblasts. The dominant schistosity of these rocks, marked by garnet and staurolite crystallisation, belongs to a later metamorphic event, followed by intense greenschist facies retrogression. In the Bugeac Hills, retrogressed Megina rocks are

Fig. 2. — Structural elements in rocks of the Boclugea and Megina Groups. a, S_1 foliation in the Bugeac Quartzites affected by S_2 crenulation cleavage, Crăcănel Hill; b, F_2 recumbent similar fold in the Bugeac Quartzites (unornamented) and phyllites North Bugeac Hills; c, F_2 recumbent fold refolded by F_3 concentric fold in the Bugeac Quartzites, Bugeac Hill; d, F_3 folds folding S_2 foliation in Priopcea Quartzites, Priopcea Hill (solid black-Megina amphibolites); e, f, F_4 folds in the Megina Group amphibolites. Priopcea Hill; g, F_4 antiformal concentric fold and related cleavages overprinting F_3 tight synformal fold and associated crenulation foliation in the Megina Group quartzo-feldspathic rocks and the Boclugea Group phyllites, La Cadin Hill.

involved in the same sequence of deformational events as the Boclugea Group rocks. Since the spatial orientation of the structural elements connected to the early metamorphic phases in the Megina Group rocks is obliterated by later foldings, the sequence of the same deformations affecting both retrogressed Megina rocks and the Boclugea Group rocks will be described later, using the same notation :

In the Buzeac Hills (north of Măcin Unit), amphibolites and quartz-feldspar schists of the Megina Group and metapelitic and metapsamitic rocks of the Boclugea Group (Bugeac graphitic Quartzites, Priopcea Quartzites, Piatra Cernei Phyllites) are complexly folded together. The earlier structural element preserved in the Bugeac Quartzites is a strongly crenulated foliation — S_1 (Fig. 2 a). The crenulation cleavages (S_2) are axial planar to tight recumbent similar folds (F_2 folds) (Fig. 2 b). Mesoscopic F_2 folds, usually refolded, have been found in several outcrops (Fig. 2 c, d). Relics of this deformational phase occur in several places within rocks of the Boclugea and Megina Groups. F_2 folds are deformed by two sets of upright folds and their interference results in a complex pattern of double plunging mesoscopic folds, well exposed both in the Bugeac Quartzites and in amphibolites in the Megina Hills. F_3 folds are responsible for the E-W lithological layering in the north Bugeac Hills. They are tight concentric folds in quartzites, amphibolites and quartz-feldspar schists (Fig. 2 d, g), while S_3 crenulation cleavages occur in metapelites (Fig. 2 e, g). Carbonatic schists of the Silurian Cerna Formation are also involved in this folding phase. NW-SE trending F_4 folds deform F_3 folds in north Bugeac Hills and highly obliterate them in the southern part of the Bugeac area and in the main outcrop area of the Megina Group, in the central part of the district. Mesoscopic F_4 folds are large folds with rounded hinges or kink folds of chevron type (Fig. 2 e, f, g). The heterogeneous nature of this folding, illustrated by rapid changes in attitude of fold axial planes, indicates a high crustal level of development.

Structural Elements in Unmetamorphosed Paleozoic Sequences

Silurian deposits have a well developed slaty cleavage (S_1), axial planar to tight isoclinal folds in slates (Fig. 3 a) and to concentric folds in grits (F_1 folds) (Fig. 3 b). In the vicinity of thrust zones, the slaty cleavage is deformed by small scale, asymmetric kink folds (Fig. 3 a).

In Lower Devonian deposits (Bujoare Formation) at Iglița, the bedding is deformed by open or tight folds (F_1 folds), with a steeply dipping axial planar slaty cleavage (Fig. 3 d-h). A similar penetrative slaty cleavage is typical in conglomerates and siltstones of the Carapelit Formation.

Conclusions

There are few geochronological age determinations in metamorphic rocks of the Măcin unit, and they lack in Paleozoic sequences. Timing of deformational and metamorphic events encountered within pre-Mesozoic rocks of this unit is a difficult problem since two phases of Variscan orogeny (A t a n a s i u , 1940 ; M i r ă u ț ă, M i r ă u ț ă, 1962 a) affected them, followed by early and late Kimmerian movements (M i r ă u ț ă, M i r ă u ț ă, 1964). The low grade metamorphism of the Boclugea Group was considered the result of the Taconic phase (Ştefan, 1966) and the amphibolite facies metamorphism of the Orliga Group was also ascribed to the Caledonian cycle (Mirăuţă, Mirăuţă, 1965).

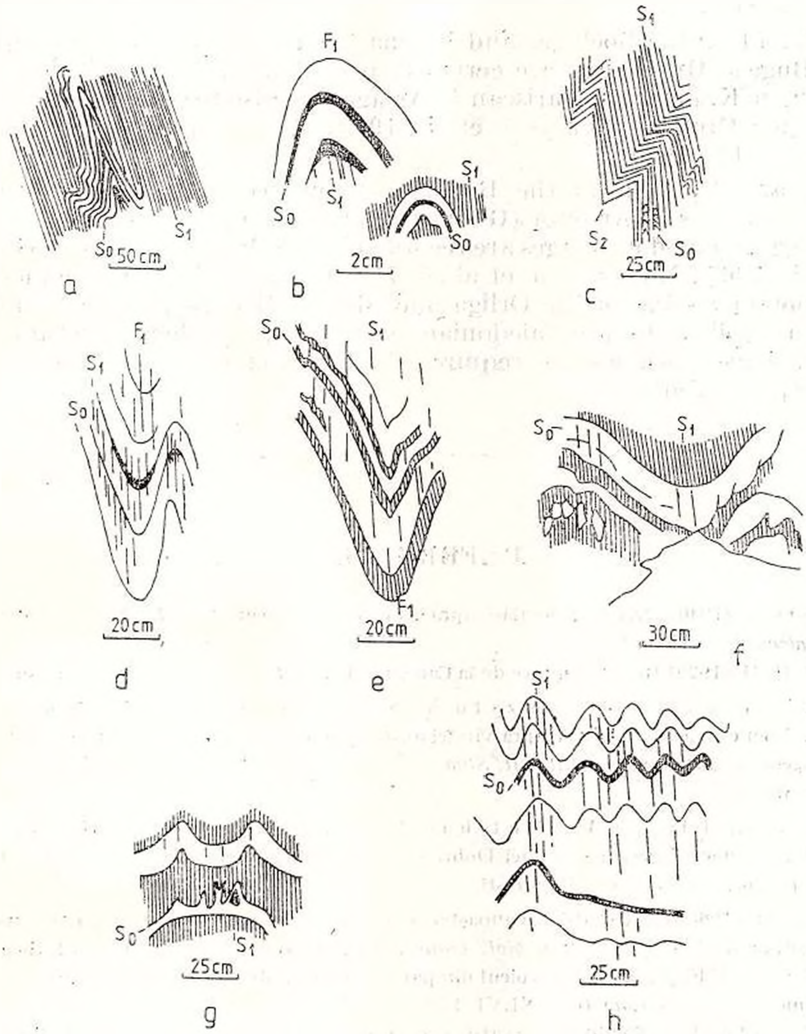

Fig. 3. — Structural elements in unmetamorphosed Paleozoic sequences.
a. F_1 intrafolial isoclinal fold in Silurian grey slates with brown carbonatic layers, Piatra Cernei Hill; b, F_1 tight fold and related slaty cleavage in Silurian grits (unornamented and slates, Priopcea Hill; c, S_2 cleavages deforming S_1 slaty cleavage of Silurian grey slates, Priopcea Hill; d-h. S_1 slaty cleavages related to F_1 folds affecting S_0 in Devonian limestones (white) and slates, Igliţa.

The main late events revealed by structural studies are F_1 folds in Paleozoic suites, F_4 folds in the Megina and Boclugea Groups and F_3 folds in Orliga Group, all with NW-SE trends. No metamorphism is related to these structures, which seem to be related to Kimmerian events, as K-Ar ages recorded in the Megina and Orliga Groups (G i u ș c ă et al., 1967; Mînzatu et al., 1975; R o m a n e s c u, V î j d e a, in S e g h e d i, 1980) indicate.

F_3 folds in the Boclugea and Megina Groups involve Silurian deposits in the Bugeac Hills; they are certainly post Silurian and could be either Variscan, or Kimmerian. Variscan K-Ar ages are also recorded in the Orliga and Megina Groups (G i u ș c ă et al., 1967; R o m a n e s c u, V î j d e a, in S e g h e d i, 1980).

F_1 and F_2 folds in the Boclugea Group could be related to Caledonian events, as K-Ar data (G i u ș c ă et al., 1967; M î n z a t u et al., 1975) suggest. Caledonian ages are recorded in the Orliga Group too (G i u ș c ă et al., 1967; M î n z a t u et al., 1975). It is possible that the medium grade metamorphism of the Orliga and Megina Groups is related to Caledonian as well as to pre-Caledonian events, but further structural and geochronological studies are required for a better dating of folding and metamorphic events.

REFERENCES

A t a n a s i u I. (1940) Privire generală asupra geologiei Dobrogei, Iași, *Lucr. Soc. Geogr. "D. Cantemir"*, III, Iași.

C ă d e r e D. M. (1925) Rocile eruptive de la Camena. *An. Inst. Geol. Rom.* VI (1912), București.

G i u ș c ă D., I a n o v i c i V., M î n z a t u S., S o r o i u M., L e m n e M., T ă n ă s e s c u A., I o n c i c ă M. (1967) Asupra vîrstei absolute a formațiunilor cristaline din vorlandul orogenului carpatic. *Acad. R.S.R. Stud. cerc. geol. geof. geogr., ser. Geol.*, 12, 2, 287–297, București.

I a n o v i c i V., G i u ș c ă D., M u t i h a c V., M i r ă u ț ă O., C h i r i a c M. (1961) Privire generală asupra geologiei Dobrogei. *Asoc. Geol. Carp.–Balc. Congr. V., Ghidul excursiilor, D-Dobrogea*, București.

M i r ă u ț ă O. (1966) Contribuții la cunoașterea formațiunilor paleozoice din partea sudică a munților Măcinului. *Rev. roum. géol. géoph. géogr. ser. Géologie* II, 2, 498–512, București.

— M i r ă u ț ă E. (1962 a) Paleozoicul din partea sudică a Munților Măcin (regiunea Cerna-Hamcearca). *D. S. Com. Geol.* XLVI, 129–142, București.

— M i r ă u ț ă E. (1962 b) Observații asupra structurii geologice a regiunii Bașpunar-Camena-Ceamurlia de Sus (Dobrogea). *D. S. Com. Geol.* XLIV, 84–93, București.

— M i r ă u ț ă E. (1964) Cretacicul superior și fundamentul bazinului Babadag. *An. Com. Geol.* XXXIII, 343–380, București.

— M i r ă u ț ă E. (1965) Le Paléozoïque de la zone de Tulcea (Dobrogea Septentrionale). *Carp.-Balk. Geol. Assoc., VII Congr, Sofia, Rep., part* II, I, Sofia.

M î n z a t u S., L e m n e M., V î j d e a E., T ă n ă s e s c u A., I o n c i c ă M., T i e p a c I. (1975) Date geocronologice obținute pentru formațiunile cristalofiliene și masive eruptive din România. *D. S. Inst. geol. geofiz.*, LXI/5,85–111, București.

M u r g o c i G. M. (1914) Cercetări geologice în Dobrogea nordică cu privire specială asupra paleozoicului și tectonicei. *An. Inst. Geol. Rom.*, V, 2, București.

Patrulius D., Mirăuță E., Mureșan M., Iordan M. (1973) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Rotman D. (1917) Masivul eruptiv de la Greci. *An. Inst. Geol., Rom.* VII, București.

Seghedi A. (1980 b) Metamorphic formations of North Dobrogea. *An. Inst. geol. geofiz.*, LVII, 469–477, București.

Ștefan R. (1966) Report, the archives of the Enterprise of Prospection for Solid Mineral Substances, Bucharest.

ПЕТРОХИМИЧЕСКИЕ ОСОБЕННОСТИ МЕТАМОРФИЧЕСКИХ ПОРОД УКРАИНСКИХ КАРПАТ [1]

Ю. Р. ДАНИЛОВИЧ[2]

Кристаллические породы в Украинских Карпатах развиты в районе Раховских и Чивчинских гор, которые являются северо-западным окончанием Марамурешского массива. Многообразие и неоднократность метаморфических процессов определяют современный облик кристаллических сланцев и свидетельствуют о сложной и длительной истории геологического развития района.

Предлагаемую здесь стратиграфическую схему (Т к а ч у к, Д а н и л о в и ч, 1965) домезозойских кристаллических сланцев можно представить в следующем виде. Нижний комплекс сложен самыми древними метаморфическими породами белопотокской, или гнейсово-сланцевой, свиты. На территории Румынии такие породы отнесены к ,,мезометаморфической'' серии, или серии Бретила. Верхний комплекс сложен породами деловецкой, или карбонатно-сланцевой, свиты с двумя подсвитами — нижней и верхней. В Румынских Карпатах эти породы сопоставляются с ,,эпиметаморфической'' осадочно-вулканогенной серией, или серией Тульгеш. Считают, что породы белопотокской свиты можно отнести к верхнему протерозою, а породы деловецкой свиты — к нижнепалеозойским образованиям.

Среди пород белопотокской, или гнейсово-сланцевой свиты выделяются следующие разновидности между которыми существуют постепенные и взаимные переходы: крупнозернистые узловатые двуслюдяные ставролит-гранатовые сланцы, биотитовые плагиогнейсы, микроклиновые очковые гнейсы, кварциты и кварцитовые сланцы. Здесь развиты дайкообразные и штокообразные тела амфиболитов и амфиболитовых сланцев, а также катаклазированные граниты яворникового типа. Породы белопотокской свиты несут следы метаморфизма средних ступеней и повторного регрессивного метаморфизма — диафтореза. Породы смяты в макро-и микроскладки и часто катаклазированы.

Среди пород деловецкой, или карбонатно-сланцевой, свиты выделяются следующие разновидности: различные по составу сланцы — био-

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тит-мусковитовые, хлорит-мусковитовые, хлоритовые, кварц-мусковитовые, кварциты и кварцитовые сланцы, мраморы и известково-слюдистые сланцы, кварцевые и полевошпатово-кварцевые порфироиды, туфоиды и туффитоиды. К областям развития свиты приурочены амфиболиты и амфиболитовые сланцы, ортогнейсы менчульского типа, в Чивчинских горах — гранито-гнейсы и плагиограниты. Породы деловецкой свиты также несут следы метаморфической переработки с последующими диафторическими преобразованиями.

Работ по петрохимической реконструкции первичного состава метаморфического комплекса Украинских Карпат совершенно недостаточно, хотя характеристики химического состава кристаллических сланцев приводятся многими авторами. Здесь для петрохимической характеристики пород использовано около 120 химических анализов (40 химических анализов пород белопотокской свиты и 80 химических анализов пород деловецкой свиты). Эти анализы пересчитаны на химические коэффициенты по методу Семененко (1963) и нанесены на АС (FM) — диаграммы.

Согласно химической систематике Семененко (1963) среди кристаллических пород фундамента исследованного района четко можно выделить породы следующих изохимических рядов: алюмосиликатного (собственно подгруппы алюмосиликатных, железисто-магнезиально-алюмосиликатных и щелочноземельно-алюмосиликатных пород), щелочноземельно-глиноземистых основных пород орторяда и известково-карбонатной подгруппы щелочноземельно-известкового ряда, то есть это соответственно породы: различные слюдистые сланцы и кварциты с кварцитовыми сланцами, метаэффузивные породы, амфиболиты и амфиболовые сланцы и метаморфические карбонатные породы. Преимущественным развитием пользуются породы алюмосиликатного ряда как среди белопотокских, так и среди деловецких пород. Поля алюмосиликатного ряда в значительной степени перекрываются, и по химическому составу, таким образом, не могут быть выделены алюмосиликатные ортопороды и парапороды.

Алюмосиликатные породы белопотокской свиты образуют на диаграмме АС(FM) довольно четкие поля. Это, прежде всего, наиболее представительное по количеству химанализов и четко выраженное поле крупнозернистых двуслюдяных сланцев к которому тяготеют также слюдянокварцитовые и кварцитовые сланцы. Они образуют поле подгруппы железисто-магнезиально-алюмосиликатных пород и тяготеют при этом к собственно алюмосиликатным породам. Обращает на себя внимание несколько повышенная глиноземистость этих пород, в которых железо преобладает над магнием или они находятся в примерно равных соотношениях. Такие сланцы могли образоваться за счет глинистых или песчано-глинистых осадочных пород, а несколько завышенная глиноземистость их свидетельствует о том, что не весь глинозем связан в полевых шпатах, а присутствует в слюдистых минералах (и ставролите), которые всегда находятся в сланцах в достаточных количествах.

Биотитовые плагиогнейсы образуют поле железисто-магнезиально-алюмосиликатных пород. Наличие реликтовых псаммитовых структур, окатанных акцессорных минералов и большое количество кварца (по данным химического состава в том числе) подтверждает факт образования их за счет кварцевых и кварц-полевошпатовых песчаников с большей или меньшей примесью глинистого материала. Более тяготеют к собственно алюмосиликатным породам очковые микроклиновые гнейсы, которые от биотитовых плагиогнейсов несколько отличаются избытком калия, а в остальном очень к ним генетически близки и также, по-видимому, образовались за счет кварц-полевошпатовых песчаников с глинистым цементом.

Совершенно четкое и обособленное поле щелочноземельно-глиноземистых основных пород орторяда образуют амфиболиты и амфиболитовые сланцы. При этом образование части амфиболитовых сланцев за счет осадочных пород мергелистого состава в настоящее время нельзя исключить полностью.

Среди пород деловецкой свиты, к алюмосиликатным породам относятся все разновидности слюдистых, слюдисто-хлоритовых, биотит-мусковитовых и других сланцев, кварцитов и кварцитовых сланцев, полевошпатово-кварцевых и кварцевых порфироидов, туфоидов и туффитоидов, плагиогнейсов. Большое количество химических анализов этих пород ложится в поля подгруппы железисто-магнезиально-алюмосиликатных и щелочноземельно-алюмосиликатных пород, некоторое количество попадает в поле глиноземисто-магнезиально-железисто-кремнистых пород и некоторые хлоритовые сланцы (которые развились по амфиболитам и амфиболитовым сланцам) в поле щелочноземельно-глиноземистых пород. Совершенно обособленное и четкое поле известково-карбонатной подгруппы щелочноземельно-известкового ряда образуют карбонатные породы — кристаллические известняки, слюдисто-карбонатные и карбонатно-тальковые сланцы.

Полевошпатово-кварцевые и кварцевые порфироиды и туфоиды попадают в поле пород щелочноземельно-алюмосиликатного орторяда, для которого исходными при метаморфизме породами служат вулканогенные продукты кислого и среднего состава. При этом некоторые анализы ложатся в поле собственно алюмосиликатных пород, что вполне допустимо для лейкократовых разновидностей метаморфизованных кислых вулканогенных продуктов.

Химические коэффициенты слюдисто-карбонатных и карбонатно-хлоритовых сланцев указывают на принадлежность этих пород к щелочноземельно-известковому ряду в котором магний преобладает над железом или иногда железо над магнием, а иногда они тяготеют к известково-карбонатным породам. Исходными для метаморфизма таких пород могли явиться песчано-глинисто-карбонатные породы, в которых глинистое и карбонатное вещество находилось в разных количественных соотношениях. Образование кристаллических известняков-мраморов за счет карбоналитов, содержавших обломочные зерна кварца, полевых шпатов и некоторой примеси глинистого вещества не вызывает сомнений. Карбонатно-тальковые сланцы по химическим коэффициентам можно отнести к подгруппе известково-карбонатных пород щелочноземельно-известкового ряда, а наиболее вероятными производными их считать кварц-доломитовые породы.

Слюдистые сланцы различные по составу, кварциты и кварцитовые сланцы, биотитовые плагиогнейсы относятся к типичным алюмосиликатным породам параряда, при этом некоторое количество анализов попадает в поля железисто-магнезиально-алюмосиликатных и щелочноземельно-алюмосиликатных пород, или в поле глиноземисто-магнезиально-железисто-кремнистых пород. Такие сланцы могли образоваться за счет первичных песчано-глинистых или глинистых пород с разным количеством псаммитового материала. Повышенное значение коэффициента избыточного глинозема свидетельствует о том, что не весь глинозем связан в полевых шпатах, а присутствует в слюдистых минералах, что вполне согласуется с минералогическим составом сланцев. При этом некоторый избыток кремнезема в сланцах и плагиогнейсах вполне очевиден. Плагиогнейсы произошли за счет полевошпатово-кварцевых песчаников с глинистым цементом, а при избытке глинистого материала получались слюдистые разности. Кварциты и кварцитовые сланцы можно отнести к алюмосиликатному изохимическому ряду метаморфических пород. Это несомненные парапороды, возникшие за счет преобразования олигомиктовых и даже мономинеральных кварцевых песчаников, содержавших некоторую примесь глинистого материала. Нередко глинистый материал в большей или меньшей степени был обогащен органическим веществом, что дало начало субграфитовым разновидностям.

Химические составы амфиболитов и амфиболитовых сланцев, актинолитовых сланцев указывают на принадлежность их к щелочноземельно-глиноземистому изохимическому ряду метаморфических пород, образующихся за счет основных вулканогенных продуктов. Актинолитовые сланцы представляют собой метаморфизованные эффузувы андезито-базальтового состава, о чем говорят их химический состав и линзообразно-пластовая форма залегания. Альбит-хлоритовые сланцы (по данным Лавренко, 1972) в одном случае можно отнести к щелочноземельно-глиноземистому изохимическому ряду метаморфических вулканогенных продуктов, в другом — обнаруживается сходство химического состава с андезито-базальтами. Особенности химического состава, близкая к миндалекаменной текстура альбит-хлоритовых сланцев, обедненность кварцем и связь с актинолитовыми сланцами свидетельствует о том, что альбит-хлоритовые сланцы могут представлять собой метаморфизованную пирокластическую породу генетически связанную со средне-основными эффузивными продуктами.

Многие анализы пород деловецкой свиты, выделенных по минералого-петрографическим признакам как ортопороды, не попадают на диаграмме Семененко в подгруппу орторяда. Но, как отмечает Семененко, это может быть связано с лейкократовым характером метаморфизованных кислых изверженных пород, которые при этом попадают в поле собственно алюмосиликатных пород. Таким образом, картина геохимической реконструкции деловецких сланцев еще более усложняется и не исключено, что некоторые разновидности алюмосиликатных пород деловецкой свиты отнесены к первичноизверженным породам ошибочно или они являются породами смешанного состава.

Биотитовые плагиогнейсы характеризуются довольно низким отношением глинозема к сумме щелочей при сравнительно высоком отношении натрия к калию. В то же время в двуслюдяных ставролит-гранатовых сланцах и различных других слюдистых сланцах первое отношение возрастает, а второе резко понижается. Так как в биотитовых плагиогнейсах биотит является единственной фазой содержащей калий, то можно предположить, что эти наиболее высокотемпературные преобразования происходили в условиях его недостатка. Количественное соотношение щелочей в разных по степени метаморфизма породах (от биотитовых плагиогнейсов, двуслюдяных ставролит-гранатовых сланцев до различных слюдистых сланцев деловецкой свиты) свидетельствует о том, что при метаморфизме парапород происходил привнос натрия, а калий практически оставался инертным.

В породах белопотокской свиты (двуслюдяные ставролит-гранатовые сланцы) наблюдается два минерала (биотит и мусковит), содержащие калий, поэтому можно предположить, что метаморфизм эпидот-амфиболитовой ступени происходил в условиях насыщенности калия и, возможно, выносом натрия.

В породах деловецкой свиты характерно снижение щелочей и резкое увеличение гидратации. Здесь биотит-мусковитовые сланцы редки, зато преобладают сланцы мусковит-хлоритовые. Значит метаморфизм зеленосланцевой ступени происходил с выносом щелочей: калий фиксируется здесь только в мусковите, а вместо биотита образуется хлорит и все это происходит на фоне интенсивной гидратации. Здесь породы отличаются хлоритсодержащими парагенезисами.

Таким образом, метаморфизм алюмосиликатных пород отличался различным поведением щелочей: от условий насыщенности калием и низкого значения натрия он сменился уменьшением роли щелочей вообще, что привело к широкому развитию гидратации и широкому развитию хлорита. Увеличение активности калия способствовало интенсивной мусковитизации и привело к широкому развитию различных по составу сланцев с мусковитом.

Можно заключить, что кристаллические породы исследованного района являются производными осадочных и изверженных пород различного минералого-петрографического и химического состава. Осадочные породы представили собой песчаные, алевролитовые и глинистые отложения, иногда это были типичные карбонатные осадки. При этом большинство парапород метапелитового и метапсаммитового состава представляли собой полимиктовые или олигомиктовые, редко аркозовые песчаники, в которых глинистый цемент присутствовал в большем или меньшем количестве. Мраморы образовались за счет типичных карбонатолитов с разной примесью глинистого или мергелистого материала. Среди ортопород преобладают кислые вулканогенные продукты, представленные порфироидами и туфоидами, менее развиты образования основного состава — амфиболиты, актинолитовые, кварц-хлоритовые и другие сланцы. Первичная природа ортопород подтверждается не только химическим составом, но и структурно-минералогическими особенностями. Реже встречаются породы смешанного осадочно-вулканогенного образования, — туффитоиды. Распределение метаморфических фаций в Раховских и Чивчинских горах показывает, что господствующими типами метаморфизма являются прогрессивный метаморфизм в широком термодинамическом интервале и регрессивный полиметаморфизм. Площади распространения пород различных фаций метаморфизма связаны со стратиграфическим разрезом кристаллических сланцев, но подчинены при этом тектонической обстановке в различных структурно-тектонических зонах. Элементы полифациального метаморфизма весьма характерны для исследованного района, где породы амфиболитовой фации сменяются породами зеленосланцевой фации, а в зонах перехода перемежаются породы разной степени метаморфизма. Регрессивные полиметаморфические преобразования происходили не при понижении температуры, а при повторном ее повышении вследствие складчатости нового этапа и прогрессивного метаморфизма осадочных образований.

Современный структурный план кристаллического массива обусловлен последовательной сменой процессов первичного осадконакопления, вулканических процессов, регионального метаморфизма, диафтореза и тектонических процессов. Новые данные по геологии исследованного района позволяют выделить несколько четко очерченных этапов в накоплении исходных пород. Эти этапы следовали непосредственно один за другим, и между ними нет перерывов в осадконакоплении. Отложения первого этапа представлены породами белопотокской, или гнейсово-сланцевой, свиты, второго — породами деловецкой, или карбонатно-сланцевой , свиты.

Максимальный метаморфизм, отвечающий эпидот-амфиболитовой (и местами амфиболитовой) фации, проявился в центральной части Раховских гор. Здесь породы представлены различными плагиогнейсами, очковыми гнейсами и сланцами, амфиболитами связанными с белопотокской свитой и слагающими ядро крупной антиклинальной структуры. Как уже отмечалось при описании пород белопотокской, или гнейсово-сланцевой, свиты биотитовые плагиогнейсы могли образоваться за счет кварц-полевошпатовых псаммитовых пород, слюдяные и слюдяно-кварцитовые, кварцитовые сланцы являются производными существенно глинистых, песчано-глинистых или песчаных пород, амфиболиты — изверженных пород основного состава. Таким образом, в дометаморфическое время в белопотокской свите происходило накопление кварц-полевошпатовых псаммитовых (с переменным количеством плагиоклаза и калишпата) и глинистых отложений и внедрение продуктов основного (диабазы и габбро-диабазы) вулканизма. Развитие получили алюмосиликатные умеренно-пересыщенные глиноземом породы, реже породы высокоглиноземистые, или щелочноземельно-глиноземистые породы орторяда. Парагенетические минеральные ассоциации являются среднетемпературными, и характеризуют эпидот-амфиболитовую (а иногда даже амфиболитовую) ступень метаморфизма.

В деловецкое время осадконакопление продолжается и образуются мощные толщи песчано-глинистых и известковых пород деловецкой, или карбонатно-сланцевой, свиты. Исходными породами для плагиогнейсов послужили кварц-полевошпатовые песчаники, которые вместе с глинистыми породами, имевшими примесь алевритового и псаммитового материала положили начало различным по составу сланцам. Кварциты и кварцитовые сланцы образовались при метаморфизме кварцевых псаммитовых пород, которые содержали некоторую примесь глинистого материала. Образование кристаллических известняков-мраморов за счет карбоналитов, содержащих обломочные зерна кварца, полевых шпатов и некоторой примеси глинистого вещества, не вызывает сомнений. При смешивании пирокластического материала кислого состава с терригенными осадками образовались туффиты, которые при метаморфизме превращались в туффитоиды. Толща деловецких пород испытывала внедрение основных (диабазы и габбро-диабазы) и кислых (липариты и липарито-дациты, дациты) вулканогенных пород. За счет эффузивов основного состава образовались амфиболиты и некоторые актинолитовые сланцы, а из эффузивов кислого состава произошли кварц-полевошпатовые порфироиды и туффоиды. Биотитовые ортогнейсы менчульского типа могли образоваться за счет кислых или промежуточных между кислыми и средними пород. Катаклазированные граниты яворниковсго типа внедрялись, видимо, к окончанию действия метаморфических агентов, поэтому оказались мало измененными в условиях первого прогрессивного этапа метаморфизма, зато претерпели значительные катакластические воздействия. Таким образом, деловецкую свиту в дометаморфическом состоянии можно представить как вулканогенно-осадочную толщу, в которой происходило накопление терригенных, глинистых и карбонатных пород. При этом седиментация сопровождалась значительным проявлением вулканической деятельности и внедрением в вулканогенно-осадочную толщу интрузий магмы не только основного, но и кислого состава, которые зафиксированы соответственно в виде амфиболитов и гнейсов, катаклазированных гранитов. Развитие получили алюмосиликатные, щелочноземельно-известковые и щелочноземельно-глиноземистые породы. Парагенетические минеральные ассоциации являются низкотемпературными и характеризуют зеленосланцевую (или эпидот-амфиболитовую) ступень метаморфизма.

ЛИТЕРАТУРА

Лавренко Е. И. (1972) Основные эффузивы в метаморфическом комплексе Украинских Карпат. В сб.:,,*Вопросы литологии и петрографии*'', кн. 2, изд. Львов. ун-та, с. 170—174.

Семененко Н. П. (1963) Метаморфизм подвижных зон. Изд.-во АН УССР, 258 с. Киев.

Ткачук Л. Г., Данилович Ю. Р. (1965) Метаморфизм кристалличних сланцив Украиньских Схидних'Карпат. *Геол. журнал*, 6.

MAGMATISME
MAGMATISM
МАГМАТИЗМ

CHRONOLOGY OF MIOCENE VOLCANISM IN NORTH-EAST HUNGARY [1]

BY

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The results of a K/Ar chronologic study carried out on the Miocene volcanic rocks in NE Hungary will be reviewed. A part of the dated rocks is sampled from the surface, others are cores of deep drillings. The investigated area is shown on the sketch map on Figure. Petrographic study and geological evaluation were made in the Department of Mineralogy and Geology of the Kossuth University and in the Hungarian Geological Institute; dating was done in the Institute of Nuclear Research of the Hungarian Academy of Sciences.

K/Ar chronologic work in this area started in 1974 (Balogh, Rakovits, 1976) and since then the continental investigations (Széky-Fux et al., 1980) have been accompanied by a systematic development of the experimental methods and instruments. From 1979 a statistically operated microprocessor controlled magnetic mass spectrometer has been used for argon determination (Balogh, Mórik, 1978; Molnár, Paál, 1980).

Earlier determinations were regularly controlled by the more advanced method; the control measurements usually confirmed the previously obtained ages. In a small number of cases, however, a deviation exceeding the analytical error was experienced, therefore only the results obtained or confirmed by the most precise method are included in this review.

The chronology of the Miocene volcanism in East-Slovakia (Slanske vrchy Mts, Vihorlat-Popričny Mts) and in Soviet Transcarpathia are studied in detail (Vass et al., 1978; Ďurica et al., 1978; Славик и др., 1976; Лазаренко и др., 1968; Мерлич, Спитковская, 1974).

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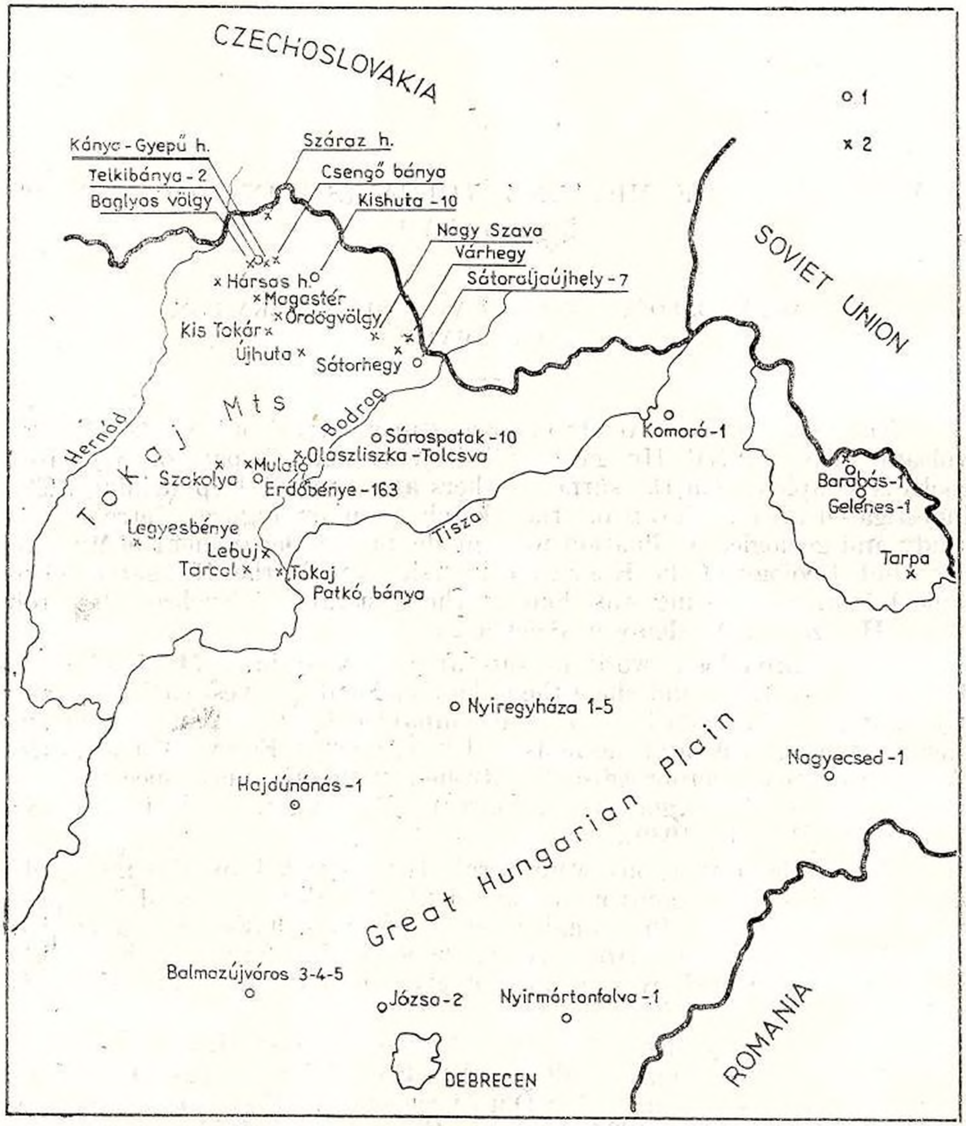
The data obtained on the neighbouring Slovakian and Soviet territories were taken into consideration during the evaluation of our results and, we think that the chronology of NE Hungary will similarly promote



Locality of dated rocks in NE Hungary.
1, sample from borehole; 2, sample from the surface.

a deeper understanding of the regional connection of the Miocene volcanism in East Slovakia and Transcarpathia.

Our data are summarized in Tables 1 and 2. For the sake of brevity only a short petrographic description of the rocks, the used mineral fraction, the potassium content, the apparent K/Ar ages and (in parentheses) the number of argon analyses are given.

In the Tokaj Mts Miocene volcanism started in the Upper Badenian (P a n t ó, 1966; G y a r m a t i, 1977). The radiometric dating of this volcanic phase is uncertain in some cases, since the extensive Sarmatian volcanism, metasomatism and hydrothermal alterations might have resulted in too young apparent ages.

The borehole Telkibánya-2 reached the Upper Badenian rocks at a depth of 790 m (S z é k y - F u x, 1970). The K/Ar age of these rocks, as determined on andesitogene propylite and potash trachyte (No 190, 191), is 13.3±0.8 m. y. and 13.8±0.5 m.y. which is in agreement with the stratigraphic position. The age of the Lower Sarmatian rocks (No 189, 211, 192) from the upper part of this borehole is in the 13.1—11.8 m.y. interval, which is also in accordance with the stratigraphy.

These radiometric data show that in this area in the Upper Badenian and also in the Lower Sarmatian time there is no age difference exceeding our analytical error between the time of volcanic activity propylitization and potash metasomatism.

The age of the Sarmatian andesites around the Telkibánya village (No 283, 286) agrees with the age of the Sarmatian rocks penetrated by borehole Telkibánya-2, while the age of andesite dykes (No 360, 641) is remarkably younger. The pyroxene andesite collected from the Hársas hill contacts the Lower Sarmatian sediments, its radiometric age is in accordance with this stratigraphic position.

The mean age of pyroxene andesites sampled around the Sátoraljaújhely town (No 286, 485) is 11.8±0.4 m.y. A metasomatized variety of this rock from the hill near Nagy Szava (No 477) is 11.8±0.5 m.y. Rhyolite and rhyodacite flood tuffs reached by borehole Sátoraljaújhely-7 (No 731, 735) resulted also in an average age of 11.8±0.5 m.y. On the basis of indirect stratigraphic data these rocks are regarded as Upper Badenian. This stratigraphic classification, however, is not fully convincing, therefore further detailed investigations are needed on the volcanic rocks which are considered at present as Upper Badenian.

There is a very good internal agreement among the ages of rocks coming from the vicinity of the Erdőbénye village (No 713, 483, 716, 719). A mean age of 11.1±0.3 m.y. indicates that the andesites in this area were formed in a geologically short interval. Therefore a similar age can be expected for the andesites, which form the basement of the limnic basin at Erdőbénye. These andesites, which have not been dated up to now, are very promising for the direct determination of the Sarmatian-Pannonain boundary, since they are covered with Sarmatian sediments.

The pyroxene andesite of the Magastér and Kis Tokár peaks (No 359, 642) in the vicinity of Telkibánya form the uppermost level of volcanic rocks in this area. The young K/Ar age is in accordance with this stratigraphic position. On the other hand, the dacites near Pálháza and Ujhuta (No 476, 480) represent an older volcanic phase. The too young K/Ar apparent ages disagree with the stratigraphy; still they can be explained by the decomposed character of the dated samples.

The age of the dacite forming the Kopasz-hill at Tokaj is determined on 2 samples (No 543, 715). The mean age of 10.4±0.4 m.y. evi-

TABLE 1

Petrographic data and K/Ar age of volcanic rocks from the Tokaj Mts

No	Locality and rock type	Phenocrysts	Groundmass	Dated fraction (No of analyses)	K (%)	Age (m.y.)
189	Borehole Telkibánya-2. 19.2–20.0 m Potash trachyte (potash metasomatite)	adularia (sericitized) (pseudomorphose after plagioclase), pyrite	microholocrystalline porphyric with pyrite and secondary quartz	w.r. (2)	9.03	12.4±0.8
211	Borehole Telkibánya-2. 173.6 m Hypersthene andesite with amphibole	plagioclase (labradorite) little/chloritized hypersthene amphibole, biotite	microholocrystalline porphyric	w.r. (2)	1.57	13.1±1.2
192	Borehole Telkibánya-2. 182.5–183.0 m Hypersthene andesite (chloroandesite)	plagioclase (labradorite) chloritized and clay mineralized hypersthene, biotite	microholocrystalline porphyric	w.r. (2)	1.87	11.8±1.0
190	Borehole Telkibánya-2. 947.7–949.5 m Potash trachyte (siliceous-pyritic)	low temperature sanidine	porphyric, accompanied by pyritization, adularization and silification	w.r. (3)	5.95	13.8±0.5
191	Borehole Talkibánya-2. 1024.2–1024.6 m Andesitogene propylite	plagioclase (carbonatized, epidotized) hypersthene (chloritized, carbonatized) epidotized, little opacitized amphibole	microholocrystalline porphyric	w.r. (4)	1.92	13.3±0.8
285	Alsókéked, Száraz-hill Pyroxene andesite with amphibole	plagioclase, fresh hypersthene, augite, amphibole	hyalopilitic porphyric	w.r. (1)	1.68	13.3±1.2
283	Telkibánya, Csengő-adit Hypersthene andesite (chloritized)	plagioclase (sericitized) hypersthene (chloritized)	microholocrystalline porphyric with carbonatization and silification	w.r. (1)	1.58	12.2±0.8
360	Telkibánya, between Kányaand Gyepü-hill Pyroxene andesite with amphibole	plagioclase, hypersthene (prismatic elongated) amphibole (opacitized)	microholocrystalline porphyric	w.r. (4)	1.79	10.6±0.8

641	Telkibánya, Baglyas-valley Amphibole andesite (potash metasomatized)	sanidine (sericitized) plagioclase, amphibole (opacitized)	hyalopilitic porphyric with silification	w.r. (1)	2.46	10.9±0.5
640	Telkibánya, Hársas-hill quarry of Cenkely Pyroxene phenoandesite with amphibole (dacite)	plagioclase (fresh), sanidine, hypersthene, augite, amphibole, biotite	hyalopilitic porphyric	w.r. (1)	2.51	11.6±0.7
485	Sátoraljaújhely, Várhegy Pyroxene amphibole dacite	plagioclase, augite, hypersthene, amphibole	hyalopilitic porphyric	w.r. (1)	2.33	11.5±0.5
286	Sátoraljaújhely, Sátor-hill Amphibole andesite with pyroxene	plagioclase fresh, amphibole	hyalopilitic porphyric	w.r. (2)	2.05	12.0±0.5
731	Borehole Sátoraljaújhely-7 130.3–135.3 m rhyolite flood tuff	acid plagioclase, potash-feldspar, biotite	vitroclastic	b. (1)	6.38	11.9±0.7
735	Borehole Sátoraljaújhely-7 241.3–246.3 m rhyodacite flood tuff	quartz, acid plagioclase, biotite	vitro-crystalloclastic	b.(1)	6.80	11.8±0.6
477	Rudabányácska, Nagy Szava pseudotrachyte (potash metasomatite)	plagioclase, adularia	hyalopilitic porphyric	w.r. (2)	5.19	11.8±0.5
713	Erdőbénye Szokolya 605.7 m peak, pyroxene andesite with olivine	plagioclase (bytownite), clinopyroxene, orthopyroxene, olivine	pilotaxitic	w.r. (1)	1.38	10.9±0.5
714* 483	Erdőbénye, Mulató-hill Pyroxene andesite	plagioclase, augite, hypersthene	pilotaxitic	2 w.r. samples	2.78 2.54	11.3±0.5
719	Borehole Erdőbénye-163. 30.7–34.5 m Pyroxene dacite	acid plagioclase (oligoclase-andesine), augite, hypersthene	hyalopilitic porphyric	w.r. (1)	3.29	11.1±0.6
716	Borehole Erdőbénye-163. 43.4–47.1 m Acid pyroxene andesite	plagioclase, augite, hypersthene	pilotaxitic	w.r. (1)	2.07	11.1±0.9
359	Telkibánya, Westward end of Magostér Pyroxene andesite	plagioclase (oligoclase), hypersthene (prismatic elongated), augite	hyalopilitic porphyric with little clay mineralization	w.r. (2)	1.89	11.2±0.5

(Continuation Tab. 1)

No	Locality and rock type	Phenocrysts	Groundmass	Dated fraction (No of analyses)	K (%)	Age (m.y.)
642	Telkibánya, Tokár peak Pyroxene andesite	plagioclase (clay mineralized), hypersthene, augite	hyalopilitic porphyric	w.r. (1)	2.00	10.7±0.6
476	Pálháza Rostalló, Ördög-valley Pyroxene amphibole dacite	plagioclase, amphibole augite, hypersthene	hyalopilitic	w.r. (2)	2.08	10.2±0.4
480	Újhuta, Kecskehát Pyroxene amphibole dacite	plagioclase, augite, hypersthene, amphibole	hyalopilitic	w.r. (1)	4.16	10.2±0.9
543	Farcal, Kopasz-hill quarry No III. Pyroxene dacite (phenoandesite)	plagioclase, hypersthene, augite	hyalopilitic porphyric	w.r. (1)	2.80	10.3±0.5
715	Tokaj, Kopasz-hill Patkó-quarry Pyroxene dacite	acid plagioclase, sanidine, quartz, augite, hypersthene	hyalopilitic-vitroporphyric	w.r. (1)	2.80	10.5±0.5
697	Kopasz-hill, Lebuj inn rhyolite perlitic	quartz, acid plagioclase	fluidal, lithophylitic	w.r. (1)	4.21	11.6±0.6
595	Borehole Kishuta-10. 9.0–9.3 m perlitic rhyolite	acid plagioclase, biotite, quartz	perlitic	b. (2)	6.53	11.7±0.4
718	Between Olaszliszka-Tolcsva fluidal rhyolite	acid plagioclase, quartz, biotite	vitreous	w.r. (1)	5.82	12.2±0.6
544* 725	Legyesbénye, quarry alunite	—	—	2 al. samples	7.96 8.11	10.9±0.4
712	Borehole Sárospatak-10. 140.0–142.6 m olivine basalt	plagioclase, pyroxene, olivine	intersertalic porphyric	w.r. (2)	1.15	10.9±1.0
484	Borehole Sárospatak-10. 91.2–94.9 m basalt	olivine (iddingsitized), plagioclase (labradorite-bytownite), hypersthene, augite, biotite	intersertalic porphyric	w. r. (2)	1.15	9.4±0.5

Isotopic constants; $\lambda_e = 0.581 \times 10^{-10} y^{-1}$; $\lambda = 4.962 \times 10^{-10} y^{-1}$; $^{40}K/K = 1.167 \times 10^{-4}$ mol/mol. Error: 1δ: w.r.: whole rock; b: biotite; al: alunite. * Two samples dated from the same locality.

TABLE 2

Petrographic data and K/Ar age of volcanic rocks from the north-eastern part of the Great Hungarian Plain

No	Locality and rock type	Phenocrysts	Groundmass	Dated fraction (No of analyses)	K (%)	Age (m.y.)
679	Borehole Komoró-1. 1833.7–1833.8 m andesite	quartz, plagioclase	microholocrystalline porphyric	w.r. (2)	2.72	12.1±0.4
680	Borehole Komoró-1. 2395.3–2395.7 m Dacitogene propylite	quartz, plagioclase (sericitized) carbonatized	microholocrystalline porphyric (silificated)	w.r. (1)	1.45	12.1±0.6
678	Borehole Komoró-1. 2438.3–2438.7 m Dacitogene propylite	quartz (sericitized), sanidine (sericitized), plagioclase	microholocrystalline porphyric (silificated, pyritic impregnation)	w.r. (2)	2.81	11.2±0.5
737	Barabás, quarry plagioclase-rhyolite	plagioclase hypersthene, biotite	spherolitic, fluidal, porphyric	w.r. (1)	2.73	11.3±0.6
730	Borehole Barabás-1. 78 m welded rhyolite tuff	quartz, plagioclase hypersthene, biotite	pseudofluidal, in some places pumiceous texture	w.r. (1)	2.44	11.2±0.6
578	Borehole Gelénes-1. 631 m redeposited rhyolite tuff	acid plagioclase, quartz, biotite, amphibole, montmorillonite	vitroclastic	b. (4)	5.65	11.0±0.6
461*	Tarpa, quarry dacite	plagioclase, hypersthene, augite	microholocrystalline porphyric	2 w.r. samples	2.10 1.98	10.5±0.4
462	Tarpa, quarry Dacite	plagioclase, hypersthene, augite	microholocrystalline porphyric	w.r. (2)	2.10	10.4±0.5
170	Borehole Nagyecsed-1. 1109.0–1110.5 m Pyroxene andesite	plagioclase, augite, hypersthene	microholocrystalline porphyric	w.r. (1)	1.76	11.1±0.7
176	Borehole-Nagyecsed-1. 3017.0–3019.0 m Andesitogene propylite	feldspar (sericitized), hypersthene (chloritized)	fluidal with pyritic impregnation	w.r. (1)	1.69	10.2±0.6

508	Borehole Nyiregyháza-5. 1930.0–1932.0 m Pyroxene andesite	plagioclase, hypersthene (clay mineralized, opacitized)	microholocrystalline porphyric	w.r. (2)	2.54	10.0±0.6
676	Borehole Nyirmártonfalva-1. 716.0–721.0 m rhyolite	quartz, plagioclase, biotite	spherolitic	b.(2)	6.98	15.8±0.5
347	Borehole Nyirmártonfalva-1. 932.0–935.0 m Dacite tuff	quartz, plagioclase, biotite, amphibole	lithoclastic	w.r. (2)	6.91	17.1±0.5
348	Borehole Nyirmártonfalva-1. 2183.0–2184.0 m rhyolite	quartz, sanidine, plagioclase, biotite	microholocrystalline porphyric (silificated)	w.r. (2)	5.31	16.0±0.6
736	Borehole Józsa-2. 1633.0–1637.0 m rhyodacite	acid plagioclase, quartz, biotite	vitrophyric	w.r. (2)	6.05	16.5±0.9
346	Borehole Hajdúnánás-1. 1997.0–2000.0 m rhyolite	quartz, sericitized potash-feldspar, biotite	spherolitic	w.r. (2)	3.87	11.4±0.7

(Notations as in table 1).

dences the Pannonian age of volcanism. The possibility of radiogenic argon loss is unlikely, since a definitely older age of 11.6±0.6 m.y. was obtained on a perlitic rhyolite sample (No 697) underlying this dacite. Rhyolite and fluidal rhyolite samples (No 595, 718) represent the same level as sample No 697. The common age of these three rocks is supported by the similarity of their radiometric ages.

In the southwestern part of the Tokaj Mts near the Szerencs town the Sarmatian pyroclastics were studied by Zelenka (1964). He distinguished five tuff levels and fixed the time of the postvolcanic activity at the Sarmatian-Pannonian boundary. Alunite crystals collected from alunite veins in the quarry at the Legyesbénya village (No 544, 725) yielded an average age of 10.9±0.4 m.y. which approximates well the datum of the Sarmatian-Pannonian boundary.

The borehole Sárospatak-10 cut three basalt lava flows separated by pyroclastic layers. The age of the upper basalt flow (No 484) is 9.4±±0.5 m.y.; this is the youngest volcanic rock of the mountains; it was formed unequivocally in the Pannonian stage. The radiometric age of the lower basalt flow (No 712) has a greater analytical error, therefore the time difference can be neither established nor excluded between the formations of the lower and upper basalt flows.

East of the Tokaj Mts the borehole Komoró-I. penetrated andesite and dacitogene propylite (No 679, 680, 678). Their K/Ar ages are in the 11.2—12.1 m.y. interval, the scatter of analytical ages may be due to analytical error or to some radiogenic argon loss from one of the propylitized samples (No 678). These ages agree remarkably well with the radiometric data of the Vihorlat-Popričny Mts published by Ďurica et al. (1978).

Along the Hungarian-Soviet boundary, on the Hungarian territory the Miocene volcanics were studied by Kulcsár (1968). Radiometric ages obtained by Soviet authors on rhyolites were in the 9±2—13.2±±1.9 m.y. interval in accordance with our previous determination (Balogh, Rakovits, 1976). Based on geological consideration Лазаренко et al. (1968) suggested the Lower Sarmatian, Мерлич and Спитковская (1974) the Pannonian age for the rhyolite volcanism, and emphasized that the time of ore mineralization of the area can be established by the exact dating of the rhyolite volcanism. Our data obtained on rhyolite, rhyolite flood tuff and on biotite separated from rhyolite tuff (No 737, 730, 578) are in the 11.0—11.3 m.y. interval, near the datum of the Sarmatian-Pannonian boundary.

The dacite forming Nagyhegy at the Tarpa village (No 461, 462) is regarded by Soviet authors as the surface continuation of the great dacite mass reached by deep drilling in Transcarpathia. Its stratigraphic age is uncertain, different authors advocate the Lower Sarmatian as well as the Pannonian age. The mean value of our determinations — 10.5±0.3 m.y. — strongly supports the assumption of the Pannonian volcanism. Similar ages (10.2—11.1 m.y.) were obtained on the andesite and andesitogene propylite penetrated by the borehole Nagyecsed-1 (No 170, 176).

A pyroxene andesite reached by borehole Nyiregyháza-5 (No 508) yielded an age of 10.0±0.6 m.y.; this shows the regional spreading of the andesite volcanism.

The age of a dacite tuff reached by borehole Nyirmártonfalva-1 (No 341) is 17.1±0.5 m.y. which indicates a Carpathian volcanism. The dating of rhyolites from this borehole (No 676, 348) resulted in ages of 15.8–16.0 m.y., which shows the continuation of volcanic activity until the Lower Badenian. A similar age – 16.5±0.9 m.y. – was obtained on a rhyodacite sample reached by borehole Józsa-2 (No 736). A younger phase of rhyolite volcanism in the inner part of the Pannonian Basin is indicated by sample No 346 from borehole Hajdúnánás-1 which yielded an age of 11.4±0.7 m.y.

REFERENCES

Balogh K., Rakovits Z. (1976) ÉK-Magyarország néhány miocén vulcanitjának K-Ar kora. *Ann. Rep. of the Hung. Geol. Inst. of* 1974, p. 471–476.

— Mórik Gy. (1978) Mágness tömegspektrométer K-Ar kormeghatározáshoz. *ATOMKI Közl.*, 20, p. 215–228.

Durica D., Kaliciak M., Kreuzer H., Müller P., Slávik J., Tözser J., Vass D. (1978) Sequence of volcanic events in eastern Slovakia in the light of recent radiometric age determinations. *Vest. Ust. ust. geol.*, *53*, p. 75–88.

Gyarmati P. (1977) Tokaj-hegység. *Annals of the Hung. Geol. Inst.*, LVIII, Müszaki Kiadó, Budapest.

Kulcsár L. (1968) A magyar-szovjet határmenti vulkánosság a legujabb szovjet és hazai, kutatások tükrében. *Acta Geogr. Debr.*, XIV/VII, p. 143–160.

Molnár J., Paál A. (1980) Mágneses tömegspektrométer mikroproceszszoros kiértékelö rendszere K-Ar földtani kormeghatározáshoz. *Mérés és Automatika*, XXVIII, 12, p. 469–472.

Pantó G. (1966) In "Magyarázó Magyarország 200000-es földtani térképsorozatához, M-34--XXXIV. Sátoraljaújhely". Ed. Balogh, K., MÁFI, Budapest.

Széky-Fux V., Balogh K., Szakál S. (1980) The age and duration of the intermediate and basic volcanism in the Tokaj Mountains, North-East Hungary, with respect to K/Ar datings. *ATOMKI Közl.*, 22, p. 191–201.

— (1970) Telkibánya éreesedése és kárpáti kapcsolatai. Akad. Kiadó, Budapest.

Vass D., Tözsér J., Bagdasarjan G. P., Kaliciak M., Orlicky O., Durica D. (1978) Chronológia vulkanickych udalosti na vychodnom Slovensku vo svetle izotopickych a paleomagnetickych vyskumow. *Geol. Práce*, 71, p. 77–88.

Zelenka T. (1964) A "Szerencsi-öböl" szarmata tufaszintjei és fáciesei. *Földt. Közl.*, 94, p. 33–52.

Славик Я., Балдасарсян Г. П., Каличяк М., Тежер Ю., Орлицки О., Ваш Д. (1976) Min. Slov., 8. p. 319–334.

Лазаренко Е. и др. (1968) Металлогения Закарпатья, Лвов Унив.

Мерлич В., Спитковская С.М. (1974) Глубинные разломи, неогеновый магматизм и орудение Закарпатья. Лвов.

TERTIARY METALLOGENY ZONING IN THE PORPHYRY COPPER DEPOSITS FROM VALEA MORII AND MUSARIU (BRAD AREA, METALIFERI MOUNTAINS)[1]

BY

MIRCEA BORCOŞ[2], ION BERBELEAC[2]

Before the last decade in the Metaliferi Mountains only the Deva Tertiary porphyry copper deposits have been known. Recent studies indicate oth,er porphyry copper deposits, discovered at Roşia Poieni, Bucium-Tarniţa Bucuresci-Rovina, Valea Morii, Musariu, Bolcana and Voia (I o n e s c u et al., 1974, 1975; I a n o v i c i et al., 1977; B o r c o ş et al., 1980; B e r b e l e a c, 1980) (Fig. 1).

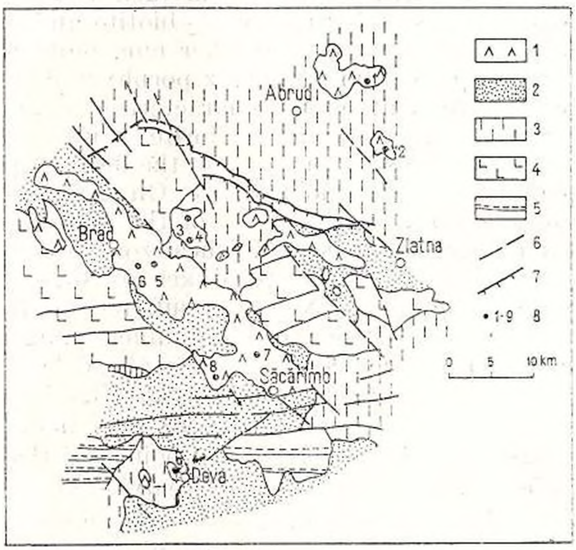


Fig. 1. — Distribution of the Tertiary copper deposits in the Metaliferi Mountains.
1, Tertiary volcanics; 2, Tertiary molasse deposits; 3, Mesozoic sedimentary rocks; 4, Mesozoic island arc volcanics; 5, Paleozoic crystalline schists; 6, fault; 7, overthrust; 8, porphyry copper deposits: 1, Roşia Poieni; 2, Bucium-Tarniţa; 3—4, Bucuresci-Rovina; 5, Valea Morii; 6, Musariu; 7, Voia; 8, Bolcana; 9, Deva.

[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, September 8—13, 1981, Bucharest, Romania.

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At present the investigations reveal that all the Neogene volcanic units in the Metaliferi Mts contain porphyry copper deposits. It is noted that the list of these deposits must remain open. These deposits as well as the genetically associated native gold, tellurides and polymetallic ore, mainly of vein type, show the complexity of the Tertiary metallogeny of the Carpathian range in general, and especially of the Metaliferi Mts.

General Geology and Structure

In the vicinity of the Valea Mare and Musariu deposits, Mesozoic island arc volcanics (basalts, andesites, diorites), Miocene sedimentary and volcano-sedimentary (Badenian) deposits and volcanic and subvolcanic rocks of Sarmatian-Pannonian age are to be found. The Tertiary volcanics consist especially of pyroclastics, flows and intrusions of the Barza andesite type (Ghițulescu, Socolescu, 1941; Ianovici et al., 1969; Borcoș et al., 1980; Berbeleac, 1980). All these rocks belong to the second cycle in the Metaliferi Mts (Sarmatian-Pontian in age, Rădulescu and Borcoș, 1967).

The porphyry copper deposits and the associated gold and base metal mineralizations are located in the two distinct Barza and Musariu composite eruptive structures. Due to a polyphase activity, two polygenous apparata (Teiul and Barza), numerous satellite apparata and two important subvolcanic intrusions are built up (Figs 2, 3, 4). The intrusions were formed as two major events with one minor intrusive event following them. The main intrusion may be divided into two textural units: 1) the oldest one consists of hornblende ± biotite and pyroxene andesite-quartz porphyry andesite and 2) another one, more recent, which comprises dyke-like porphyry diorite-quartz porphyry diorite. The minor intrusive event is not frequent and is characterized by quartz porphyry hornblende ± pyroxene dykes (younger diorite) (Figs 3, 4). These dykes, located in the centre of the structure and in the deep eruptive bodies, show a gradational contact with the older diorite. On the contrary, in the upper levels of the mining works the contacts of the younger diorite, like the older diorite, are marked by breccia and shear zones. Gradually, this aspect disappears to the depth. The younger dykes are slightly pyritic; one of them contains some chalcopyrite. None has suffered the fracturing or the mineralization associated with the major and mineralizing event. The dioritic rocks appear especially in the deepest levels of the Valea Morii and Musariu mines, where they tend to form a unitary dyke (Fig. 3); within the roofs of the subvolcanic structure these intrusions send many irregular dykes into andesite (Figs 3, 4). However, it is to point out that in the upper part of the subvolcanic bodies the older diorite dykes are very rarely found.

The fault measurements from underground exposures revealed three dominant fault systems (Fig. 2): 1) one, the most prominent, strikes 45—70°W and dips 70—75°SW, being especially widespread in the western part of the Barza neck (Musariu area); 2) the other one, proper to the Valea Morii structure, with N—S strike and roughly dipping 60—80°W, and 3) the last one, less important, with W-E strike, lies between the above-

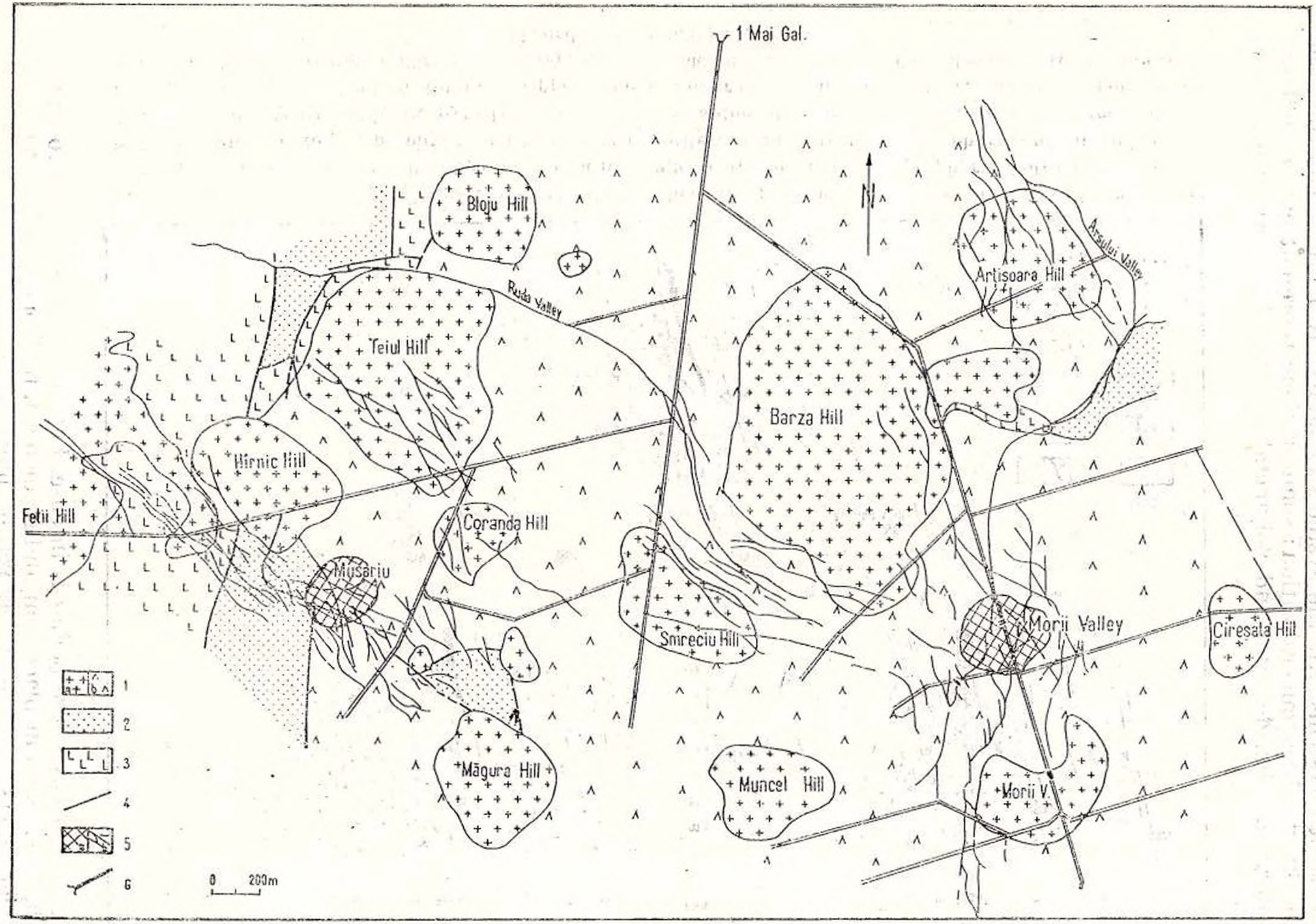

Fig. 2. — Geological sketch map of the Barza-Musariu area.

1, andesite-quartz andesite (Upper Badenian-Pannonian); a, neck and dyke; b, pyroclastics and lavas; 2, sedimentary and volcano-sedimentary deposits (Upper Badenian in age); 3, island arc volcanics (Upper Jurassic-Lower Cretaceous); 4, fault; 5, porphyry copper (a) and native gold and base metal sulphide gold deposits (b); 6, 1 Mai gallery.

mentioned systems. All the systems of faults are more recent than the porphyry copper mineralization; some of them represent native gold and tellurium, tellurides ores and polymetallic — Au, Ag — ore veins. Fault-

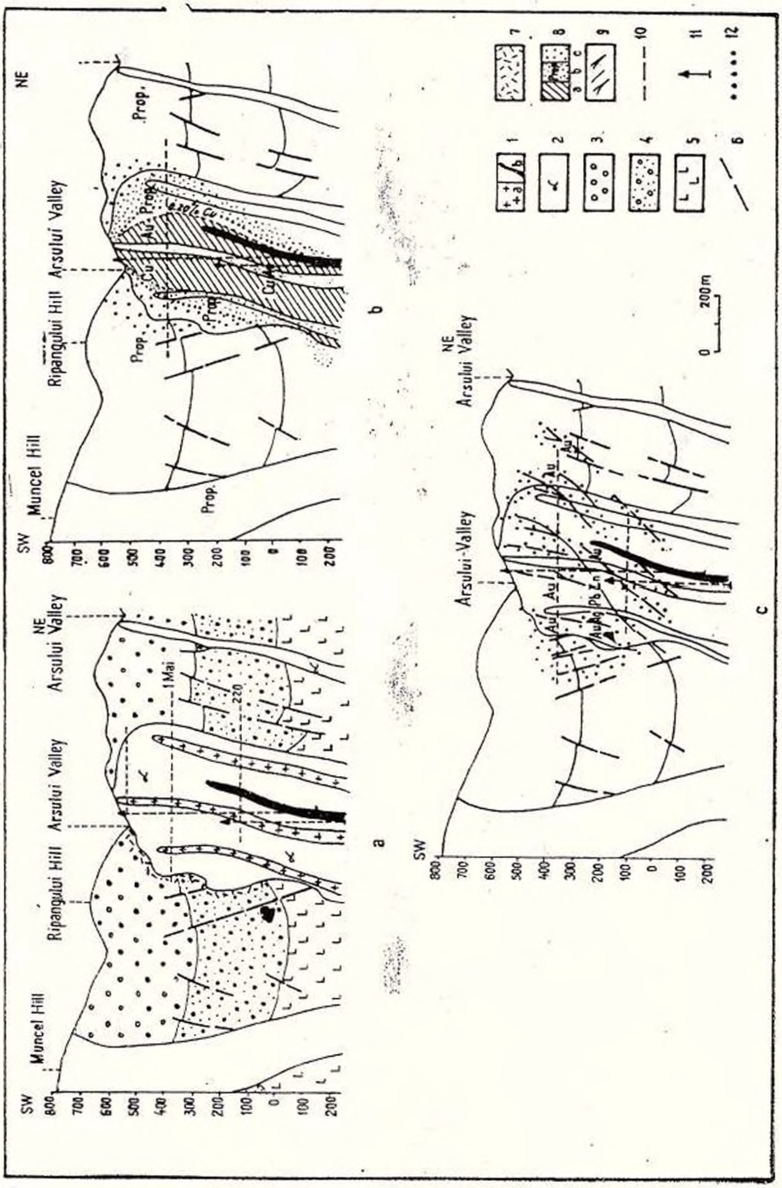

Fig 3.— Genetic model of the Valea Morii deposit. a, schematic geological cross section; b, early alteration-mineralization stage; c, late alteration-mineralization stage. 1 a, diorite-quartz porphyry diorite (older diorite); b, hornblende ±pyroxene porphyry diorite (younger diorite); 2, andesite quartz-porphyry andesite intrusions; 3, andesitic and quartz andesitic pyroclastics and flows; 4, sedimentary and volcano-sedimentary rocks (Upper Badenian in age); 5, island arc volcanics (Upper Jurassic-Lower Cretaceous); 6, fault; 7, breccia; 8, potassic (a), propylitic (b) and sericitic-argillic alteration (c); 9, native gold and base metal sulphide deposits; 10, mining level; 11, drilling; 12, 0.2% Cu isoline.

ing in porphyry ore bodies is intense, but most faults appear to have a relatively small displacement and grade out into "horse tail" fracture zone, which shows post-ore mineralization movements.

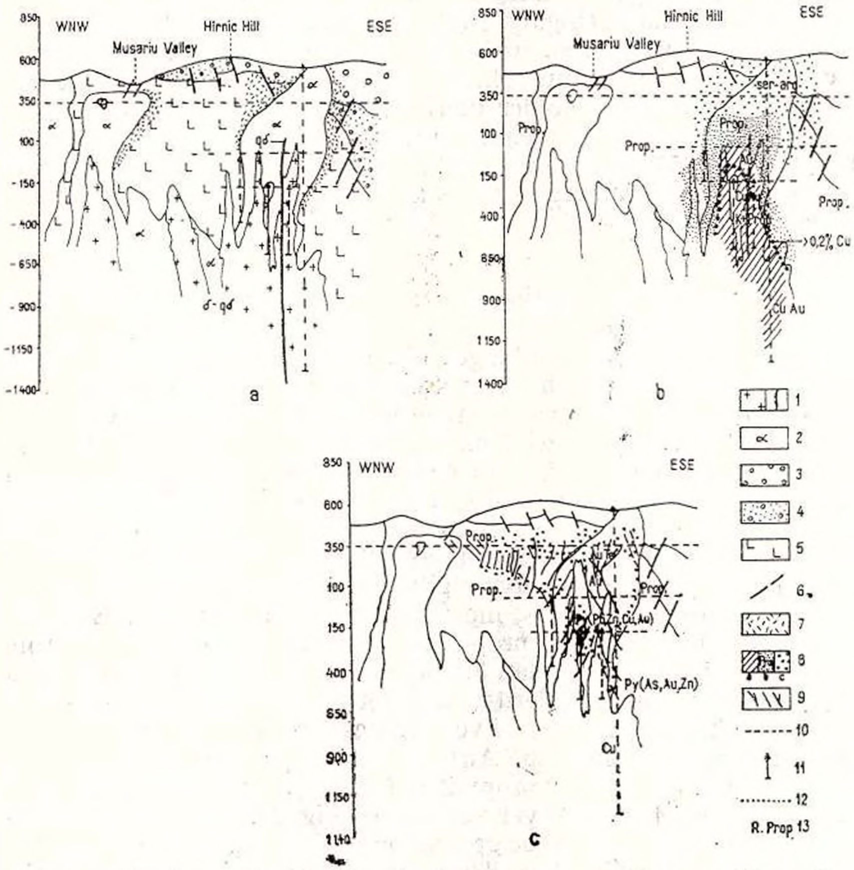

Fig. 4. — Genetic model of the Musariu deposit (for explanation, see Figure 3).

Relative Age Relations

The contacts between the Tertiary intrusions and alteration-mineralization show a multi-stage nature of hydrothermal processes. The last one is clearly demonstrated especially by the truncation of many quartz veinlets from the subvolcanic bodies by younger quartz porphyry diorite and vein deposits. The stockworks of quartz sulphide veinlets from the porphyry copper bodies consist of at least three systems of veinlets. The widest veinlets range from one to less than 10 cm. The veinlets appear to be parallel and subparallel in each system and sometimes show a small displacement. Most of the veinlets, trending NW-SE and E-W, dip steeply southwest, south and north or are nearly horizontal. The majority of the veinlet system seem to be tension veinlets, which, generally, have been opened and filled simultaneously or nearly simultaneously. Yet, some of the veinlets have been filled in different time intervals.

As regards the relative age of the rocks from the intrusion bodies it is to remark : 1) the older diorite-quartz porphyry diorite intruded the andesite-quartz porphyry andesite and are penetrated by the younger quartz porphyry diorite (Fig.3); 2) the andesite and older diorite are intensely hydrothermally altered, while the younger diorites are nearly fresh; 3) the porphyry copper mineralizations are spatially and genetically associated with andesite and older diorite intrusions; 4) the younger quartz porphyry diorites are very weakly mineralized and no quartz veinlets have been observed; 5) the native gold veins, tellurium and telluride veins and polymetallic — Au, Ag — veins are the most recently deposited from the ore-forming solutions.

Alteration-Mineralization Zoning

The background assemblages define well-developed zonal patterns in the polyascendent porphyry system (Figs 3, 4) in which one notices a decrease of the copper values together with a progressive decrease of veining and fracturing away from the central part of the intrusions into wall rocks. The copper veins also show a gradual decrease from depth to the upper part of the deposits. There is also a sulphide zoning in the deposits: 1) the central zone with chalcopyrite-pyrite (Cu, Au), very characteristic in the Valea Morii deposits and 2) the border zone, with disseminated pyrite (Au, Cu) and polymetallic Au-Ag, pyrite -native gold, native tellurium and telluride veins, more rarely breccia deposits. Especially in the Valea Morii deposit, the proportion of chalcopyrite increases from null at the outer edge to more than 30 per cent in the inner zone. Minor amounts of bornite, sphalerite, magnetite and hematite have been found here. The gold and silver values are restricted to the central chalcopyrite-pyrite zone (Cu, Au). Here the potassic (K-feldspar, oligoclase, biotite-quartz) and propylitic (epidote, albite, chlorite, quartz, carbonate) assemblages are very common (Fig. 5). In the potassic assemblage, the phenocrysts and the groundmass feldspar are replaced by orthoclase and probably oligoclase. These feldspars are also present on fissures. The obliterated plagioclase phenocrysts and the common porphyry texture are noted. The basic phenocrysts are also replaced by K-feldspar, oligoclase, biotite and quartz. The grey-smoky coloured quartz veinlets $\pm$ pyrite and chalcopyrite are abundant. These veinlets have a biotite-oligoclase-anhydrite-quartz halo. It is important to underline that the biotitization in the Valea Morii and Musariu porphyry copper deposits is not more characteristic than in other similar deposits in Romania (Roșia Poieni, Ionescu, 1974; Ionescu et al., 1975; Bucium-Tarnița, Deva, Ianovici et al., 1977), Chile (El Salvador, Gustafson and Hunt, 1975), Philippines (Marcopper, Loundon, 1976) and other areas in the world.

In propylitic assemblages the new minerals partly or wholly replaced the feldspar and mafic phenocrysts or microcrysts from the groundmass of the andesite-quartz porphyry andesite and diorite-quartz porphyry diorite (older diorite). Quartz and carbonate usually occur as veinlets. The quartz from the veinlets has a grey colour, banding structure and abundant pyrite and chalcopyrite.

The sericite-argillaceous assemblages from the early substage form an envelope of the deposits and consist mainly of sericite, clay minerals (kaolinite, illite), quartz and pyrite. These minerals are characteristic of border zones — pyrite (Au) zone. Mention should be made of the fact that in the deepest exposures of the Valea Morii and Musariu mines the sericitic-argillaceous alteration occurs in quartz and pyrite veinlet halo. There is a general correlation between the abundance of pyrite and the degree

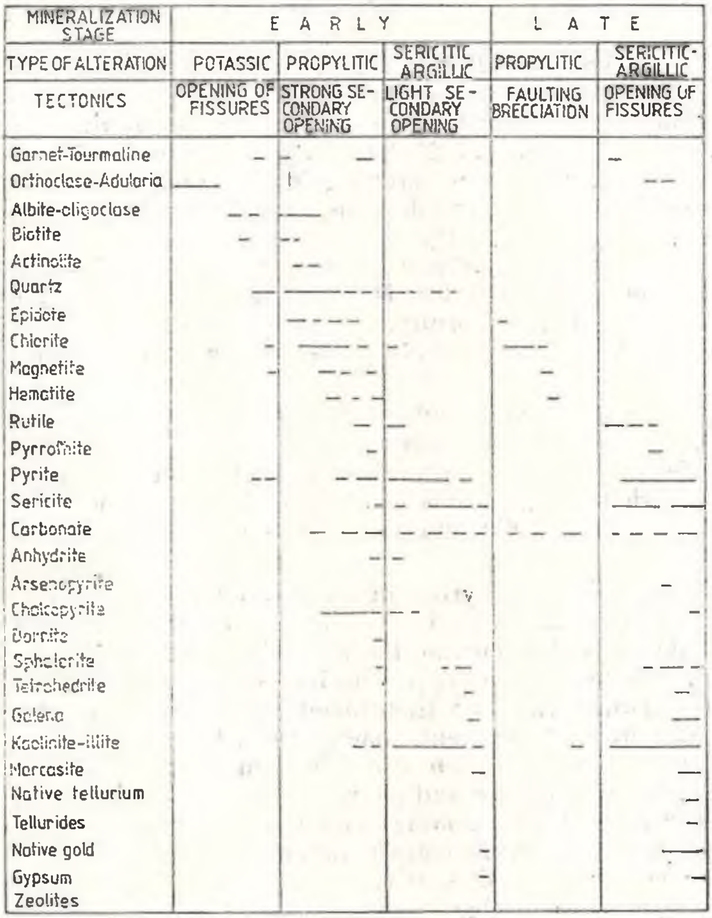

Fig. 5. — Scheme of the sequence of the mineral deposition in the Valea Morii and Musariu porphyry copper deposits.

of sericite-argillaceous alteration. The pyrite appears as isolated fine-grained crystals, usually in cubic forms or as fine aggregates.

In the roofs of the subvolcanic bodies lies the deposit veins from the border zone and the last stage of alteration — mineralization (Figs 3, 4). These types of mineralizations also show a vertical zonality : native gold veins, native tellurium and telluride veins in a large envelope with sericitic-argillaceous alteration in the upper part and polymetallic Au-Ag veins in the very limited argillaceous area, downwards. These veins usually have a sericite-adularia-clay minerals-quartz±tourmaline alteration (Fig. 5).

The strike of the veins (NW-SE, NNW-SSE and WNW-ENE) has been influenced by the regional faults and by the morphology of intrusions (Fig. 2).

Remarks

1. The products of alteration-mineralization show a polyascendant feature and have been released during two stages : 1) early stage with porphyry copper mineralizations and potassic, propylitic and sericitic-argillaceous assemblages and 2) late stage, characterized by veins with polymetallic — Au, Ag — ores, native gold ores, and native tellurium and telluride ores, and sericitic-argillaceous alteration. These two alteration-mineralization stages are clearly associated with the multi-stage evolution of the porphyry intrusion and metallogeneses from the subvolcanic bodies. Thus, the copper mineralization is connected with the major intrusive events — andesite-quartz porphyry andesite and diorite-quartz porphyry diorite, while the ore veins—post-younger quartz diorite have been deposited.

2. A vertical alteration-mineralization in the porphyry copper deposits has been observed : 1) a central zone (chalcopyrite-pyrite (Au) zone) with potassic and propylitic alteration and 2) a border zone (pyrite(Au) zone) with sericitic-argillaceous alteration and polymetallic (Au, Ag) veins, native gold and tellurium and telluride veins (only in the Musariu deposit).

3. Chalcopyrite and pyrite, in an approximate ratio of 2 : 1 in the Valea Morii deposits and 1 : 4 in the Musariu deposits, are the main primary sulphide minerals occurring both as veinlet fillings and as disseminations. Minor amounts of pyrite, pyrrhotite, rutile, sphalerite, tetrahedrite, galena accompanied the two mentioned sulphides. Sometimes the ore sulphides contain 1—2 per cent magnetite with some specularite. Molybdenite can be found accidentally. The minor amounts of Au closely associated with chalcopyrite and pyrite is noted.

As regards the lateral zoning, which is also present, it is to mention the presence of the porphyry copper mineralization in the central part of the complex volcano-intrusive structure and the polymetallic Au-Ag and native gold veins at their edge.

4. The Valea Morii and Musariu copper mineralization deposits show important affinities to the dioritic model of the porphyry copper deposit (e.g. Schmitt, 1968; Creasey, 1968; Lowell and Guilbert, 1970; Rose, 1970; Hollister, 1975; Ianovici et al., 1977) in ore minerals, ore textures, alteration minerals, host rocks, associated igneous rocks and geological setting. The chemical trends and field relations suggest that the alteration-mineralization is genetically related to younger calk-alkali andesite-quartz porphyry andesite and diorite-quartz porphyry diorite subvolcanic bodies from the second phase of the second cycle (Sarmatian-Pannonian in age).

5. The space-time relation between the Au, Ag, Te, Pb, Zn and Cu ore veins and porphyry copper ore from the region have demonstrated the

complexity of the Tertiary metallogeny in the Carpatho-Balkan range or the Tethyan-Eurasian metallogenic belt. For this reason we propose to name this genetic model the Barza type from the Metaliferi Mountains.

REFERENCES

Berbeleac I. (1980) Native tellurium and telluride mineralization from Musariu, Brad Region (Metaliferi Mountains), Romania. *An. Inst. Geol.*, LVI, p . 153—168, București.

Borcoș M., Berbeleac I. (1979) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

— Berbeleac I., Gheorghiță I., Bratosin I., Colios E., Zămîrcă A., Anastase Ș., Verdes G., Stănescu I. (1980) Geochemical remarks on the Valea Morii porphyry copper ore deposits (Metaliferi Mountains). *D. S. Inst. Geol., Geofiz.*, LXIV/2, p. 17—36, București.

Cioflica G., Istrate G., Popescu Gh., Udubașa G. (1966) Contribuții la cunoașterea vîrstei produselor vulcanice din regiunea Hărțăgani-Trestia (Metaliferi Mountains). *Stud. cerc. geol., geofiz., Seria geol.* 11, 1, p. 171—182, București.

Creasey C. S. (1968) Hydrothermal alteration in geology of porphyry copper deposits in southwestern North America. Titley R. S. and Hicks L. C. (Editors), p. 51—74 Tucson, Arizona.

Ghițulescu T. P., Socolescu M. (1941) Étude géologique et minière des Monts Metaliferi. *An. Inst. geol. Rom.*, XXI, București.

Gustafson B. L., Hunt P. J. (1975) The Porphyry Copper Deposit at El Salvador, Chile. *Econ. Geol.*, 70, p. 857—912, Lancaster.

Hollister F. V. (1975) An appraisal of the nature and source of porphyry copper deposits. *Minerals Sci. Eng.*, 7, 7, p. 225—233.

Ianovici V., Giușcă D., Ghițulescu T.P., Borcoș M., Lupu M., Bleahu M., Savu H. (1969) Evoluția geologică a Munților Metaliferi. Ed. Acad. R.S.R., 742 p., București.

— Vlad Ș., Borcoș M., Boștinescu S. (1977) Alpine Porphyry Copper Mineralization of West Romania. *Mineralium Deposits*, 12, p. 307—317, Berlin, Heidelberg, New York.

Ionescu O. (1974) Mineralizația cupriferă de tip diseminat de la Roșia Poieni. *Acad. R.S.R., Stud. cerc. geol., geofiz., ser. geol.*, 19, p. 77—84, București.

— Soare C., Gheorghiu M. (1975) Contribuții la cunoașterea zăcămîntului cuprifer Roșia Poieni. Alterații hipogene. *Stud. cerc. geol., geofiz., ser. geol.*, 20, p. 159—170, București.

Loundon G. A. (1976) Marcopper Porphyry Copper Deposits, Philippines. *Econ. Geol.*, 71, p. 721—732, Lancaster.

Lowell J. D., Guilbert J. M. (1970) Lateral and vertical alteration-mineralization zoning in porphyry copper deposits. *Econ. Geol.*, 65, p. 373—408, Lancaster.

R ă d u l e s c u D., B o r c o ş M. (1967) Spätsubsequenter alpiner Magmatismus in Rumänien. *Acta Geol. Acad. Sci. Hung.*, 11, 1–3, p. 139–152, Budapest.

R o s e A. W. (1970) Zonal relations of wall rock alteration and sulfide distribution in porphyry copper deposits. *Econ. Geol.*, 65, p. 920–936.

S c h m i t t A. H. (1968) The Porphyry Copper Deposits in Their Regional Setting. In : Geology of the Porphyry Copper Deposits in Southwestern North America. Titley R. S. and Hicks L. C. (Editions), p. 17–33, Tucson, Arizona.

S u p e r c e a n u I. C. (1979) The Copper Metallogeny in Romania (abstract). Inter. Symposium "The European Copper Deposits".

SOME FACTS AND OPINIONS CONCERNING THE NEOGENE MAGMATISM AND METALLOGENESIS IN THE METALIFERI MOUNTAINS OF TRANSYLVANIA [1]

BY

TOMA PETRE GHIŢULESCU [2]

Introduction

The Metaliferi Mountains Neogene magmatism actually belong to the great Carpatho-Balkan magmatic province (D i m i t r e s c u, 1959). Still we continue to treat separately the respective problems, since the effort to find out a global solution to corroborate the data concerning the origin and evolution of the different magmatisms, valid for the whole area, in the plate tectonics concept, could not surpass the hypothetical stage (B l e a h u et al., 1973; B o c a l e t t i et al., 1974; R ă d u l e s c u et al., 1973; R ă d u l e s c u, 1953; I a n o v i c i et al., 1976). So, we consider useful to remind some interesting facts and to expound some personal opinions, which, though with reference to the Metaliferi Mountains, could suggest general solutions.

Short Outlook on the Magmatic Processes Evolution in the Metaliferi Mountains

As it is known, the Neogene magmatism is the third manifested in the region, following the initial simatic one, which was a product of the Metaliferi Mountains geosynclinal structure genesis, and the essentially sialic, epirogenic banatitic magmatism, which was not a product of the geosynclinal structure.

The Neogene magmatism, too, could not be considered a product of this eugeosynclinal structogenesis and has in common with the former magmatism the fact that the main structural controlling factor of the petrogenesis seems to be in my opinion the uplift movement of the ground simultaneously with the emission of the respective magma. The evolution of the Neogene magmatism is synthetized in Table.

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[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, September 8–13, 1981, Bucharest, Romania.

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TABLE

Geologic evolution of the Metaliferi Mountains

AGE	PHASE	MOVEMENTS	STAGE	LITHOLOGY EROSION	INTENSITY	NEOGENE ERUPTIVISM SERIES	NEOGENE ERUPTIVISM CHEMISM	ERUPTIVE STRUCT. ROCKS	MINERALIZATION
QUATERNARY PLIOCENE	IV	uplift	Continental alluvial	Volcano-sedimentary alluvia		Rotunda	Gabbroic plagioclasic dioritic	Basalt and andesite volcanoes	
PANNONIAN	III B	uplift	Continental lacustrine	Erosion sands		Cetras	Granodioritic dioritic	Dacoandesite and dacite volcanoes	
SARMATIAN	III A	uplift	Continental lacustrine	Coals, pelite volcano-sedimentary		Barza Săcărîmb	Quartz dioritic Dioritic	apophyses subvolcanoes chimneys andesites volcanoes	
UPPER BADENIAN	II	uplift	Brakish	Volcano-sedimentary, pelites		Cîinel	Dioritic Quartz dioritic Granodioritic	Dacite and daco andesite volcanoes	
MIDDLE BADENIAN		subsidence	Marine	Limestones Sandstones Conglomerates				Mollasse	
LOWER BADENIAN	I		Continental	Conglomerates		Fața Băii Băița	Quartz dioritic Granodioritic Leucogranitic	Mollasse	
LOWER MIOCENE EOCENE		uplift	Continental	Erosion					

It is noticed that the first emission of the sialic magma took place at the end of an uplifting movement that began at the end towards the end of the synkinematic stage (I a n o v i c i et al., 1976), continued in the tardekinematic stage, from Paleocene to the Lower Miocene, before the short interval in the Middle Badenian, when the ground sank and the Tortonian Sea invaded the small intermontane basin. But since the Upper Badenian the uplift movement resumed, the sea water was expelled, a brackish, lagoonal and, at the end, continental stage were installed, existing still today. One should note the significant fact that, during the sinking of the ground, the eruptive process was nearly stopped, but resumed concomitantly with the positive movement.

As it is known the bulk of the Neogene eruptive centres are situated on the NW-SE belt, crosswise on the general direction of the geosynclinal structure, but some of them on axial belt (G h i ţ u l e s c u, B o r c o ş, 1966). The magma chemistry is calc-alkaline, somehow similar to the banatitic one. In the Neogene the first coming lavas, which are in many places the first geological event, have an acid and very acid composition, namely leucogranitic, granodioritic and quartz-dioritic, the last one prevailing during the whole Neogene magmatic evolution, noticing that the few ones with basic compositions appeared in the last volcanic phase.

The manner in which the Neogene magmatism appeared in the Metaliferi Mountains and the matter that gave it life supports the idea that it originates from magmatic chambers older than the Lower Badenian, being in full differentiation process at the very start of the volcanism. These magmatic chambers were situated on volcanic belts that we used to call petrogenetic lines (G h i ţ u l e s c u, B o r c o ş, 1966). Chemical evolution of the magma was the same in all the Neogene subprovinces of the Romanian Carpathians.

The Source of the Neogene Magmatism

The old scholars, as for instance P a l f y (1916) considered the Neogene volcanism linked to the subsidence of the intermontane basins and to their silting, an idea longtime accepted, agreed to by many scholars. M r a z e c (1935) suggested that the Neogene Carpathian and Apuseni Mountains volcanism was produced by the sinking of the Central Transylvanian Basin and the associated intermontane basins, the rising of the magma in volcanoes being a compensation of the sinking sediments. This idea does not match the facts, as before mentioned.

Later, S t i l l e (1953) considered the subpushing palyngenetic process to be the genetic cause of the Neogene Carpathian magmatism, an idea anticipating the subduction theory from "the plate tectonics" concept (I a n o v i c i et al., 1976). As before told, this last theory does not succeed in the Metaliferi Mountains to surpass the hypothesis stage of contradictory hypotheses.

In spite of that situation the author thinks that one may retain from this concept the idea of thermal fluxes, punctiform or linear, plausibly explained in the plate tectonics concept. These caloric fluxes could cause anatexis and palyngenesis phenomena, generating magmatic chambers that fed the explosive volcanism and the intrusive magma in stocks (and subvolcanoes). The magma chambers became the plutons and batholiths, like those described by Emmons (1937). In this concept, one may think that the banatitic and Neogene eruptive belts were created by thermic fluxes, structurally controlled by ruptural and disjunctive tectonics, associated with the uplifting movement of the ground. In these, there formed by anatexis and palyngenesis processes lithomagma chambers, situated at moderate depth, representing plutons with their cupolas and apophysis (Ghițulescu, Socolescu, 1935).

On Volcanism, Subvolcanism and Plutonism

The volcanism in the Metaliferi Mountains was exclusively of the central type (the fissural volcanism is unknown). The essential element of the structure is the volcanic chimney (funnel) sometimes of important dimensions. The volcanic cone high might reach some thousands of meters, as results from theoretical reconstruction (Ianovici et al., 1969). The volcanic funnels are whithin the volcanic belts, but none of them is localized on fractures or settled fracture zones. In the chimney-room, preferably on their borders, volcanic plugs and subvolcanoes are to be found, the best illustrative being the great chimney between Hârnic, Teiul, Coranda and Musariu (Ungureanu, Aldea, Ispas, 1978), with many plugs and subvolcanoes. The volcanic plugs are pillars of magma consolidated in vents or pipes, whose horizontal dimensions are variable at different levels, sometimes with a tendency to diminish at depth, as at Măgura Artișoarei. The mechanical logic supports the idea that the magma chamber that produced it was not too deep.

In the same sense plead the forms and the dimensions of the breccia pipes, for instance the many ones in the Baia de Arieș Mine. Our studies proved that they are in direct connection with the magmatic chamber they came off, though its horizontal sections are not greater than some hundred meters (Ghițulescu et al., 1979).

The subvolcanic bodies are numerous and the greater part of these are built by andesites of the Barza Series. Their dimensions are small near the present surface and many of them do not outcrop. At depth, their dimensions are considerably greater, reaching one million of m^2, as for instance the Musariu Nou subvolcano.

In some favourable conditions we could understand the mechanism of intrusion in the Afiniș subvolcano from Baia de Arieș (Ghițulescu, Sprînceană, 1963). Magma was injected under the form of blade and the movement induced oriented textures (Cochet, 1957) or even fluidal textures in the andesite at the contact (Ghițulescu, 1959). Oriented texture zones separate the blades.

All these data lead to the conclusion that the subvolcanic bodies represent apophyses of deeper plutons. On the basis of certain data we argue for the existence of a Neogene plutonism, that is in direct filiation with the known subvolcanic bodies and the volcanoes. The idea is supported by the presence of the peri-batholithic zonal distribution of the mineralization (Ghițulescu, 1934; Emmons, 1937).

On the Hydrothermal Activity and its Relation with the Mineralizing Process

In the author's conception, the hydrothermal activity occupies a well defined position in the stratigraphic scale, namely after the third eruptive phase, that in certain mining districts was after the eruption of the Barza-Săcărîmb Series and in others after the daco-andesite Cetraş Series (Ianovici et al., 1969).

There are two distinct processes: propylitization and the potassic-argillitic metasomatosis.

The first manifested as a regional process on a relatively great area (Ghiţulescu, Socolescu, 1941; Ianovici et al., 1969), but never in direct relation with the mineralization. The propylitization took place only in certain Neogene eruptive rocks, more frequently in the enrooted bodies, but it is not known in the pre-eruptive basement rocks (crystalline schists, ophiolites, sediments) (Richthofen, 1860; Palfy, 1916; Bürg, 1931; Rădulescu, 1953; Ghiţulescu, Socolescu, 1941). This process is not controlled by the veins fractures or other fracture systems. From the chemical point of view, it is a hydration in the presence of CO_2 and SO_2.

Unlike with the propylitization the potassic-argillitic metasomatosis is intimately associated with the mineralizing hydrothermalism, which preceded it, preparing the necessary conditions for its realization. This alteration process is confined to the neighbouring zone of the circulation ways of the hydrothermal solution, limited to some decimeters or meters width, with the exception of some very permeable rocks, such as Cretaceous and Tertiary sandstones and conglomerates. The metasomatic alteration affected all the silicate minerals in the Neogene eruptive rocks, leaching the Ca^{++}, Mg^{++}, Fe^{++}, $Mn^{++}$and Na^{+} ions, the last being nearly completely eliminated. At their places came K^{+}, B^{+++} and Si^{++++}, the intensity of the process may be evaluated by the value of the K_2O : Na_2O ratio, that diminishes rapidly from the vein to the exterior (Ghiţulescu, 1960). The alteration took place not only in the Neogene eruptive rocks, but also in the crystalline schists, ophiolites and the Cretaceous and Tertiary sediments, till the disintegration. From the chemical point of view, there results a neutralization of the environment as regards the mineralizing solutions.

The hydrothermal metasomatic transformation took place under a geothermic control, namely only in zones whose average temperature was convenient for the chemical activity that formed the neominerals. The medium preheating became possible by the modification of the isothermic surfaces due to the hot magma intruded in the enrooted structures created by the Neogene magmatism (Ghiţulescu, 1934). This way one can explain the possibility of the active hydrothermal solutions to migrate on a relatively great distance from the magma chamber and also the localization of the mineralized area within or around the enrooted bodies.

Another important effect of the hydrothermal solutions was the increasing permeability of the hydrothermalized medium, by the cations leaching and the microfissuration of the rocks due to the formation of neominerals with different molar volumes. In consequence, the hydrothermalized rocks density is by 10–20% lower than that of propylitized or fresh rocks.

About the Hydrothermal Solutions and Their Circulation

In the author's opinion, the hydrothermal solutions are an exclusive product of a perfect magmatic differentiation, able to extract the most soluble magma components. For this achievement it was necessary the intervention of a depressing in the magma chamber by the hood uplifting concomitantly with the magmatic differentiation. In consequence, the author believes that in the Metaliferi Mountains the bulk of the useful metals originate from the magmatic chamber and the eventual addition of metals extracted by dissolving from the medium rocks was unsignificant. This opinion is supported by the idea that the active mineralizing solutions could preserve their useful initial dowry content only when traversing neutral medium, namely environment rocks previously hydrothermalized. In the Metaliferi Mountains there is no valuable mineralization out of the hydrothermalized zones.

A second argument of general order is the fact that in the Metaliferi Mountains all the mineral deposits of a given kind preserve the same characteristics in the whole region independent of the petrographic composition of the environment, in crystalline schists, ophiolitic or sedimentary rocks. All the arguments support the opinion that the main geochemical control of the mineralization was done by the magmatic differentiation in the batholithic cupolas and their apophyses.

As concerns the circulation road and the moving way of the hydrothermal solution we shall mention our interesting observations in the gold district at Baia de Arieş, where we could prove that the hydrothermal solution came out from the magmatic chamber exclusively through the breccia pipes, with very few exceptions (Ghiţulescu et al., 1979). The solution motion was by diffusion, going through the porous (spongy) matrix of the polygenous breccia that filled the pipe, especially on the dorsal part when the pipe was inclined or curved. From the chimney the solution passed by diffusion in the surrounding rocks, causing their hydrothermalization, in a first stage, and their mineralization in the last one. Detailed investigations showed that the hydrothermal solutions penetrated also through the apophysis of some subvolcano bodies (Mălai and Ambru andesite types) that was in advanced hydrothermalized conditions, nearly to the disintegration. The solution moved by diffusion too. We believe that these considerations concerning the role of the diffusion mechanism of hydrothermal solution circulation have a large application in other mineral deposits too. This may be the case of the exceptionally rich gold deposit from Musariu, where the mineralizing solutions could come in on the margin of the huge volcanic chimney mentioned in the other chapter and pass by diffusion in the vein network.

The Formation of the Ore Concentration and the Role of the Trapping Effect

We should point out that the solutions moving by diffusion generally produced non-paying mineral impregnation. This phenomenon is striking in the Baia de Arieș breccia pipes, where the polygenous breccia is intensely hydrothermalized, but there is few concentrated ore in it. These are to be found in the surrounding andesites or crystalline schists, intensely brecciated and loosed, in which the voids had represented some ten percents of the volume. The crystallization starting was caused by the depressing in the moment the hydrothermal fluid penetrated the void spaces. The concentration of the ore was intensified by the stopping of the solution in the stockwork space, namely by the trapping effect that played an important role in the formation of great rich ore concentration (Ghițulescu, Sprinceană, 1981).

In our opinion, the vein fractures may be regarded as traps, where the hydrothermal solution penetrated by diffusion or infiltration and was stopped or hindered in its movement. This may be the case of vein nets that have no lateral or downward communication. The vertical distribution of the ore may be attributed to this phenomenon.

Zonal Distribution of the Mineralization

The author described in 1932 the zonal distribution of the mineralization in some groups of hydrothermal Tertiary ore deposits in the Metaliferi Mountains and Baia Mare region. The distribution of peribatholithic type and the mineral zones are approximately in the order indicated by Emmons (Emmons, 1937; Ghițulescu, 1934). This zonal distribution was confirmed by other research workers, as downward changes in the mineralization and in a horizontal sense for a group of ore deposits in relation with the same pluton. The way this zonal distribution is developed may be considered a mighty argument for the existence of a Neogene plutonism.

Conclusions

The opinions presented in this work are the result of a long gestation process.

The author believes that the Neogene magmatism manifested under a triple aspect : batholithic, subvolcanic and volcanic, at three different petrogenetic and metallogenetic levels.

The plutons are at the deeper level, that, though not too profound, has not been discovered until now in the Metaliferi Mountains.

In the cupolas and apophyses of these plutons took place the petrogenetic and metallogenetic differentiations. From these magma chambers originated the magma that created the intrusive subvolcanoes and fed the volcanism, developed under the structural control of the ruptural and disjunctive tectonics, the most important effect of these, for the evolution of the magmatism being an uplifting movement of the ground, continually resumed.

The Neogene magmatism was associated with a substantial modification of the isothermal surfaces in the respective area, determined to mould the subvolcanic structures and partially the volcanic ones. This new thermic regime and the chemical activity of the hydrothermal fluids reconditioned the geological medium that became favourable for a migration of the metallogenetic level toward the surface and allowed the occurrence of a newer metallogenetic zone than those normally known in the cupolas and the hood of the batholiths (Emmons, 1937). The concentration of the ore was efficiently controlled by structural traps, created by the magmatic tectonics at the subvolcanoes level and at that of the volcanoes infrastructures.

In the author's opinion, the geological and geophysical prospecting works shall investigate all these metallogenetic levels.

REFERENCES

Berbeleac I. (1970) Consideraţii tectonice, magmatice şi metalogenetice în partea centrală şi sudică a Munţilor Metaliferi. *Stud. cerc. geol., geofiz., geogr., Seria geol.*, 15, 2, p. 329–339, Bucureşti.

Bleahu M., Boccaletti M., Manetti P., Peltz S. (1973) The Carpathian arc a continental arc displaying the features of an "Island arc". *J. Geophys. Res.*, 78, 23, p. 5025–5032.

Boccaletti M., Manetti P., Peccerillo A. (1974) Hypothesis on the plate tectonic evolution of the Carpatho-Balcan Arcs. *Earth a. planet sci. lett.*, 23, p. 193–198, Amsterdam.

Bürg G. H. (1931) Charakteristik der grünsteinartigen Andesitfazies, ihre Ursachen und Beziehungen zur Kaolinisierung und Verkjieselung. *Z. prakt. Geol., Halle a., S.*, 39, p. 11–12.

Cioflica G., Savu H., Borcoş M., Ştefan A., Istrate G. (1973) Alpine Volcanism and metallogenesis in the Apuseni Mountains. Symposium volcanism and metallogenesis, *Guidebook Series*, 13, 70, Bucharest.

Cochet R. (1957) Contribuţi geologice asupra zăcămintelor aurifere de la Baia de Arieş. *Rev. minelor*, VIII, 10, p. 467–475.

Dimitrescu R. (1966) Beiträge zur Kenntnis der Magmatisch-Tektonischen Verhältnisse in Karpatisch-Balkanischen Raum. *Acta Geol. Acad. Sci. Hung*, X, p. 357–360, Budapest.

Emmons W. H. (1937) Gold deposits of the world. p. 378–382, New-York.

Ghiţulescu T. P. (1934) Distribution de la minéralisation dans les gisements d'âge tertiaire de Transylvanie. *Bull. Soc. Roum. Géol.*, II, p. 56–97.

— Socolescu M. (1935) Relations entre la tectonique et la métallogenèse dans la quadrilatère aurifère des Monts Apuseni. *Congr. Int. des Mines, Mét. et Géol. appl.*, VII, Paris.

— Socolescu M. (1941) Étude géologique et minière des Monts Métallifères. *An. Inst. geol. Rom.*, XXI, Bucureşti.

— (1959) Report, the archives the I. M. Baia de Arieş.

— Sprinceană M. (1959) Report, the archives of the I. M. Baia de Arieş.

— (1960) Report, the archives of the Brad Gold Trust.

— Borcoş M. (1966) Încadrarea funcţională a magmatismului alpin din Munţii Metaliferi. *Stud. cerc. geol.*, 11, 2, Bucureşti.

— Pitulea G., Ghițulescu I. (1979) The role of the Volcanic Breccia Pipes in Metallogenesis in the Baia de Arieș Mining District. *Rev. Roum. Géol., Géophys., Géogr., Ser Géol., Acad. R.S.R.*, 23, 1, București.

— Romanescu D., Pitulea G., Ghițulescu I., Georgescu A., Neștianu T. (1981) Études pétrogénétiques et pétromagnétiques dans la région de Baia de Arieș. *Rev. Roum. Géol., Géophys., Géogr., Ser. Géophys., Acad. RSR*, 25, București.

— Sprînceană M. (1983) Le rôle de l'effet de piège dans la concentration de la minéralisation dans le gisement de Baia de Arieș, Monts Métallifères de Transylvanie. *Asoc. Géol. Carp.-Balc. XII*, (1981), *An.* LXI, București.

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

Ianovici V., Giușcă D., Ghițulescu T.P., Borcoș M., Lupu M., Bleahu M., Savu H. (1969) Evoluția geologică a Munților Metaliferi. Ed. Acad. RSR, 742 p., București.

— Borcoș M., Bleahu M., Patrulius D., Lupu M., Dimitrescu R., Savu H. (1976) Geologia Munților Apuseni. Ed. Acad. RSR, București.

Mrazec L. (1935) Les mines d'or de Roumanie. Agence économique et financière – 1934.

Palfy M. (1916) Über die Propylitisierung der Eruptivgesteine. *Földt. Közl.*, XLVI, 133, 188, Budapesta.

Peltz S. (1970) Contribuții la cunoașterea vulcanismului neogen din regiunea Groși-Bulza-Fîntoiag (extremitatea sud-vestică a Munților Metaliferi). *D. S. Inst. Geol.*, p. 109–133, București.

— Vasiliu C., Bratosin I. (1972) Petrologia rocilor bazaltice plio-cuaternare din România. *Rev. Roum. Géol., Géophys., Géogr., Ser. Géol.*, 15, 1, p. 77–125.

Rădulescu D. (1953) Contribuții la cunoașterea fenomenului de propilitizare a rocilor vulcanice. *Rev. Univ. Buc. Polit., Seria șt. nat.*, 3, p.220–233, București.

— (1973) Plate tectonics in Tertiary Volcanic Areas. An example : The Carpathians. *Intern. Symp. on volc. a. assoc. metallog. General adress*, 12 p., București.

— Săndulescu M. (1973) The plate-tectonics concept and the geological structure of the Carpathians. *Tectonophysics*, 19, 3, p. 155–161, Amsterdam.

Richthofen F. (1860) Tertiäre Eruptivgesteine von Ungarn und Siebenbürgen. *Jahrb. d.k.k. geol. R. A.*, 153, Viena.

Stille H. (1953) Der tektonische Werdegang der Karpaten. *Beihefte Geol. Jahrb.*, 8, 239, Hannover.

Udubașa G., Istrate G., Anton O., Dafin E., Braun Al. (1976) Mineralizațiile polimetalice de la Bocșa-Leopold (nord Săcărîmb-Munții Metaliferi). *D. S. Inst. Geol. Geofiz.* LXII/1, p. 97–124, București.

Ungureanu B., Aldea I., Ispas G., Chiorpec M. (1971) Report, the archives of the Geological Enterprise of Prospection for Solid Mineral Substances, Bucharest.

LE ROLE DE "L'EFFET DE PIÈGE" DANS LA CONCENTRATION DE LA MINÉRALISATION MÉTALLIFÈRE DANS LE GISEMENT DE BAIA DE ARIEȘ (MONTS METALIFERI DE LA TRANSYLVANIE)[1]

PAR

TOMA PETRE GHIȚULESCU [2], MARIA SPRÎNCEANĂ [3]

Le problème de la concentration de la minéralisation dans les gisements métallifères, afin de constituer le minerai exploitable, présente un intérêt général, théorique et pratique, tout aussi important que dans les gisements de pétrole, mais la littérature de spécialité ne l'abordant que dans de rares études.

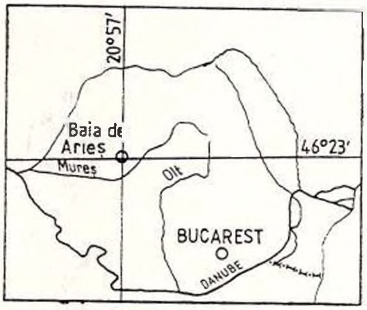

Fig. 1. — Position géographique de Baia de Arieș.

Les recherches effectuées à Baia de Arieș, dans des conditions particulièrement favorables, ont abouti à des résultats intéressants qui peuvent contribuer à l'éclaircissement du cas général.

Vue générale sur la structure géologique et la métallogenèse dans le district

Le gisement est situé dans une structure relativement simple, constituée par des roches cristallines et des magmatites néogènes (C o c h e t, 1957).

[1] Note présentée au 12ème Congrès de l'Association Géologique Carpatho-Balkanique, 8—13 septembre 1981, Bucarest, Roumanie.

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Pendant le Sarmatien, des intrusions d'andésites à quartz de la série de Barza-Săcărîmb (Ghiţulescu, Socolescu, 1941), ont formé de nombreux corps subvolcaniques et un cortège d'apophyses. Un volca-

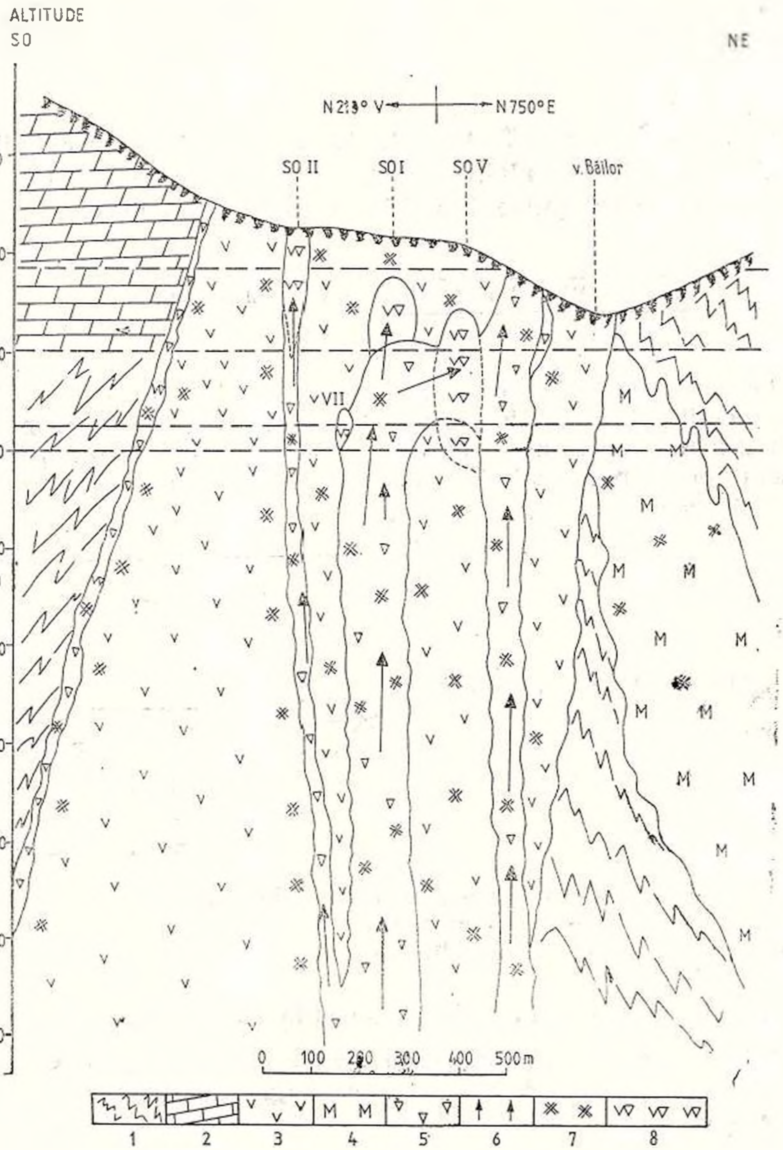

Fig. 2. — Coupe schématique à travers le corps subvolcanique de AFINIS et quelques-uns des stockwerks aurifères, localisés au bord de "breccia pipes".

1, schistes cristallins ; 2, calcaires cristallins ; 3, andésite type Afiniş propylitisée ; 4, andésite type Malai faiblement hydrothermalisée ; 5, breccia pipes ; 6, le flux des solutions hydrothermales ; 7, altération potassique-argilitique ; 8, stockwerks aurifères no. I, II, V.

nisme explosif à gaz a suivi, donnant naissance aux cheminées volcaniques (breccia pipes) qui ont joué un rôle déterminant dans la formation des pièges structurales dans lesquelles se sont localisés les stockwerks et ont assuré la voie d'accès aux solutions hydrothermales émises par les voussoirs (Ghițulescu et al., 1979). Vers la fin de cette étape a eu lieu la principale phase de minéralisation qui a généré d'importants gisements aurifères et de métaux non-ferreux dans toute la sous-province métallogénique des Monts Metaliferi.

Les deux dernières manifestations du magmatisme ont été exclusivement volcaniques, associées à un hydrothermalisme très faible, comme la série pannonienne des dacoandésites de Cetraș ou complètement stérile comme la série pontienne d'andésites et de basaltes de Rotunda.

Le principal gisement aurifère se trouve dans la structure volcanique de AFINIS, étant constitué par quelques stockwerks disposés sur la périphérie des "breccia pipes", surtout à la partie dorsale de ceux-ci (fig. 2) et par de rares et minces filons en relation directe avec les mêmes "pipes". Les formes tubulaires de ceux-ci ont des diamètres variant entre 50 et 250 m et les travaux miniers les ont reconnus sur plus de mille mètres vers la profondeur, où elles se reliaient vraisemblablement à des voussoirs magmatiques. Leur remplissage est constitué par une formation bréchiforme polygène à matrice de cendre volcanique bien poreuse. Autour des "breccia pipes" l'andésite est fissurée et souvent bréchifiée aux endroits de courbure. La roche est hydrothermalisée dans la zone de contact, l'intensité de la transformation diminuant d'une façon graduelle vers l'extérieur. Outre la brèche polygène qui remplie les "breccia pipes" et les zones périphériques des ceux-ci (environ 5—30 m de largeur), l'altération hydrothermale est absente dans la structure AFINIS, ce qui prouve que ces cheminées constituaient l'unique voie par laquelle les solutions hydrothermales montaient lors de la genèse du gisement (fig. 2). La structure AMBRU (située au SE d'AFINIS) a des riches concentrations de mine ai de plomb, zinc et pyrite. Elle s'est formée par des schistes et calcairres cristallins disloqués et pénétrés par des subvolcans andésitiques et des"breccia pipes". Le gisement est constitué d'amas de minerai, générés principalement par substitution métasomatique dans le mur des calcaires cristallins et dans les "breccia pipes".

Les corps de minerai, leur morphologie et minéralisation

La concentration de la minéralisation a été déterminée par un phénomène de captation des solutions hydrothermales dans des pièges structuraux aussi bien que dans le gisement aurifère d'AFINIS que dans celui plombo-zincifère d'AMBRU. Dans cet article nous nous limitons d'illustrer le processus en décrivant les cas où il s'est manifesté de la manière la plus convaincante, par exemple dans les stockwerks no. I et no. II du gisement aurifère.

Le stockwerk no. I est localisé entièrement dans le corps subvolcanique AFINIS, sur la partie dorsale d'un "breccia pipe", á l endroit où ce dernier subit une inflexion (fig. 2). Le stockwerk a la forme d'un paraboloïde de révolution, le diamètre à la base étant d'environ 70 m et la hauteur de 145 m. La partie apicale a la forme d'une voûte d'équilibre. Le stockwerk repose sur la bosse d'une cheminée volcanique. Le stockwerk est remplis par des dalles et des blocs d'andésite, tombés des parois par un phénomène d'autoremblayage. Ces dalles et blocs ne dépassent pas 2—3 m de largeur et 0,5—1 m de puissance, étant disposés d'une

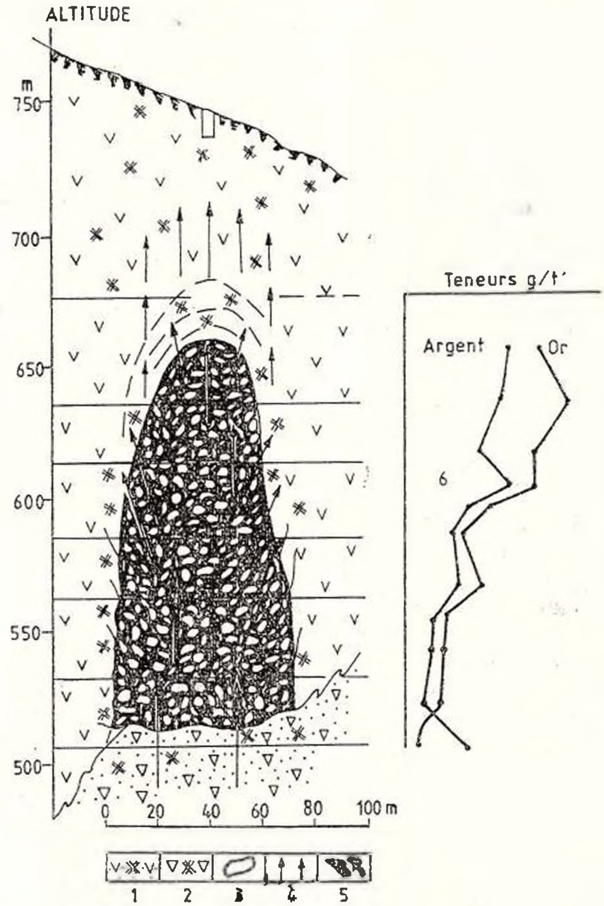

Fig. 3. — Coupe schématique à travers le stockwerk no. I, reposant sur la partie dorsale d'une cheminée volcanique.

1, andésite hydrothermalisée; 2, brèche polygène; 3, dalles et blocs d'andésite minéralisée; 4, flux de la solution hydrothermale; 5, minerai aurifère; 6, variation de la teneur en or et argent à des niveaux divers.

manière chaotique. Les espaces libres d'entre eux représentent environ de 15 à 20% du volume total du stockwerk (Ghiţulescu, Sprînceană, 1959) (fig. 3).

La genèse de ce type de structure, en relation directe avec la formation des "breccia pipes" a été expliquée dans des travaux antérieurs (Ghiţulescu, 1959; Ghiţulescu et al., 1979 a) de même que le rôle que les "breccia pipes" ont joué dans le processus de minéralisation (Ghiţulescu et al., 1979 b). Les solutions minéralisantes, en continuation de celles qui ont produit l'altération hydrothermale potassique-argilitique, se sont élevées en circulant par diffusion à travers le contenu poreux qui remplissait les "breccia pipes", en le minéralisant par imprégnation diffuse de pyrite aurifère. La teneur en or dépasse rarement la limite d'exploitabilité.

En échange, ces solutions ont produit une riche minéralisation aurifère dans les structures à grands espaces libres des pièges structuraux décrits plus haut, en générant des stockwerks aurifères à teneur d'or élevée. Il en a résulté deux types de minéralisation.

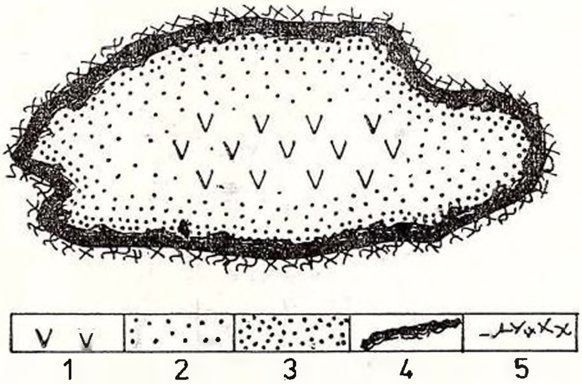

Fig. 4. — Coupe à travers un bloc d'andésite minéralisée. 1, noyau d'andésite propylitisée; 2, zone faiblement imprégnée; 3, zone richement métallisée; 4, croûte noire; 5, dépôt de gangue et de sulfures métallifères.

D'abord, les solutions minéralisantes ont pénétré à l'intérieur des blocs d'andésite, déjà entamés par l'altération hydrothermale, en produisant une imprégnation de sulfures métallifères à teneur d'or. Ce processus s'est déployé sous forme d'une métasomatose, dont l'intensité diminuait de la périphérie vers le noyau du bloc, de façon que ce dernier est resté stérile dans certains cas (fig. 4).

La minéralisation la plus riche a pris naissance par le dépôt des minéraux de gangue et métallifères dans l'espace libre. La composition de cette minéralisation est relativement simple, comprenant environ 6 % de pyrite, 3 % mispickel, 1 % marcassite, 0,2 % fer, oligiste et magnétite, 0,1 % blende, 0,05 % alabandine, 0,05 % chalcopyrite, 0,01 % galène et 0,01 % stibiotellurure d'or et argent (Sb (Ag. Au) Fe_2) (N i ţ u l e s c u et al., 1957). L'or est intimement associé au mispickel; il n'est pas visible même au grossissement de 5000 X (N i ţ u l e s c u et al., 1957).

Le produit le plus caractéristique de ce type de minéralisation est un dépôt de silice microcristalline en concrescence avec des cristaux microscopiques de pyrite, marcassite et mispickel aurifère. Comme formes de cristallisation on remarque des panaches de marcassite filiforme ornés de cristaux rhombiques de mispickel aurifère (fig. 6). Ce dépôt se présente sous forme de croûtes siliceuses compactes de couleur gris foncé (les croûtes noires enveloppant les dalles et les blocs d'andésite). L'espace resté libre entre les blocs est rempli, plus ou moins complétement, par un dépôt de cristaux courts et minces de quartz blanc, associés à des petits cristaux de sulfures métallifères. Dans la gangue on a remarqué aussi la présence de quelques carbonates (calcite, sidérite, rhodochrosite) et de la fluorine. La silice libre est l'élément prédominant dans les minerais, dépassant 60 % dans le tout venant et atteignant 90 % dans le minerai riche. On remarque une certaine proportionnalité entre la teneur en or ou arsenic et la quantité de silice.

L'andésite des parois du stockwerk, sur une largeur de 3 à 4 m, était intensivement fissurée, hydrothermalisée et minéralisée et la teneur en or a justifié son exploitation aux niveaux supérieurs.

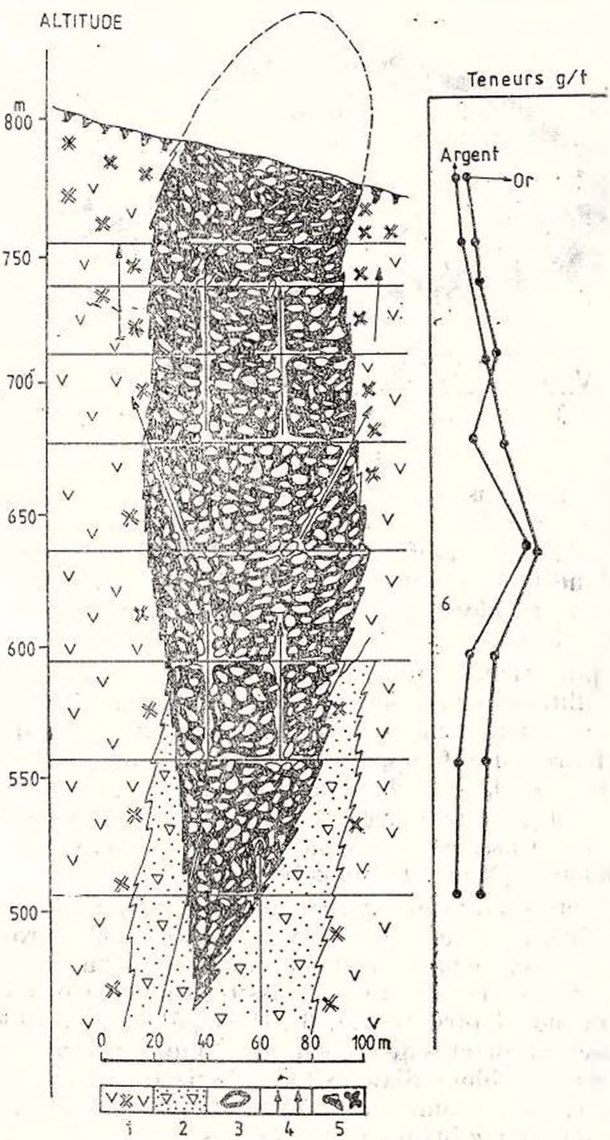

Fig. 5. — Coupe schématique à travers le stockwerk no. II, situé au sommet d'un "breccia pipe". 1, andésite hydrothermalisée; 2, brèche polygène dans le "breccia pipes"; 3, dalles et blocs d'andésite minéralisée; 4, flux de la solution hydrothermale; 5, minerai aurifère; 6, variation de la teneur en or et argent à des niveaux divers.

Il s'ensuit que la partie liquide des solutions hydrothermales quittait le piège en diffusant et filtrant par les parois, surtout dans la partie apicale, où la structure de l'andésite du toit est complètement décomposée.

Le stockwerk no. II, situé à 100 m du no. I est compris entièrement dans l'andésite du même corps volcanique, reposant sur le sommet d'un "breccia pipe".

La forme du stockwerk est presque cylindroïde, à sections elliptiques, évasée vers la surface (fig. 5) où les axes horizontaux atteignent jusqu'à 90 m. La partie apicale a été détruite par l'érosion ; toutefois la hauteur actuelle dépasse 200 m. La morphologie intérieure du stockwerk est similaire à celle décrite pour le stockwerk I, à la différence que les dalles et les blocs d'andésite ont des dimensions plus réduites et l'espace libre d'entre ceux-ci ne représente que 5–15 % du volume total. La minéralisation est pareille à celle décrite pour le stockwerk I, mais les croûtes noires sont plus minces et la teneur moyenne en or est plus réduite.

La distribution de la minéralisation dans les stockwerks aurifères

On dispose d'une information complète concernant la variation de la composition minéralogique et le contenu en or (réserves de minerais pour chaque tranche) en sens horizontal et vertical, vu que le stockwerk no. I a été complètement exploité et que du no. II il n'en reste que quelques tranches vers la base.

Il en résulte de ces données que la teneur moyenne en or a diminué sensiblement du sommet vers la base (voir les diagrammes des fig. 3 et 5) et que cette baisse a été provoquée par un changement dans la composition minéralogique, respectivement par la réduction de la fréquence et des dimensions des croûtes noires, principal élément de la métallisation aurifère. Il paraît, d'après les données de l'exploitation, que la concentration de la métallisation n'a pas atteint son maximum dans le sommet même du stockwerk no. I, mais un peu plus bas, dans les tranches autour du niveau + 130 (par rapport à la base). De même, les tranches les plus riches en or dans le stockwerk no. II ont été trouvées entre les niveaux + 152 m et + 127 m au-dessus du même repère. A ces niveaux, l'abondance des croûtes noires était évidente. Leur puissance atteignait à 3–4 cm, dans le stockwerk I, et environ à 2 cm dans le stockwerk II. Il était frappant de constater la rareté, le rapetissement, voir la disparition de ces croûtes vers la base du stockwerk et leur absence dans la brèche polygène qui remplie les "breccia pipes". Nous connaissons une seule exception à cette remarque d'ordre général, qui a eu lieu dans la cheminée de la base du stockwerk no. II, au niveau — 40 m sous le repère de la galerie principale. On y a exploité une accumulation de minerai riche, à croûtes noires, dans une partie contenant des blocs d'andésite. En sens horizontal, il faut signaler dans le stockwerk no. I une tendance d'enrichissement du contenu aurifère vers les bords.

En guise de conclusion, on apprecie que l'indice de concentration de la métallisation a été d'environ 4,5 fois dans le stockwerk I et de 3,2 fois dans le no. II. Ces indices caractérisent approximativement la différence de teneur et de contenu aurifère entre les parties les plus riches du niveau supérieur par rapport aux parties plus pauvres de la base des stockwerks.

Le rôle de "l'effet de piège" dans la concentration de la métallisation

Nous attribuons les différences de concentration de la métallisation sur la verticale, celles décrites plus haut pour les deux stockwerks, à "l'effet de piège" exercé sur les solutions hydrothermales par des structures ameubliées et à espaces libres. Le processus a pu se dérouler de la façon suivante.

Il est évident que les solutions hydrothermales se sont élevées des voussoirs magmatiques par diffusion à travers la matrice, remplissant les "breccia pipes", dans un état de pression et de température suffisamment élevées pour fournir l'énergie nécessaire à vaincre la contre-pression hydrostatique et surtout la résistance beaucoup plus grande opposée par les forces d'adhésion du liquide aux particules très fines du cendre volcanique. Nous imaginons un mouvement assez lent de la solution composée d'une phase liquide et de nombreuses phases gaseuses correspondant aux divers composants volatils des minéralisateurs.

On constate que dans cet état l'activité de métallisation a eu une intensité très réduite, produisant une imprégnation de pyrite avec très peu d'or.

En débouchant dans le piège, la détente et la baisse correspondante de température ont déterminé le dégagement des phases gaseuses, qui se sont concentrées vers la partie supérieure, avec leur plus grande richesse en arsenic et or. La phase liquide a échappé du piège en filtrant par la paroi, ce qui a permis un renouvellement continuel de la solution minéralisante. Pendant la durée assez longue du processus a pu intervenir un certain changement dans la composition de la solution et un refroidissement avec déposition du gel de SiO_2 et de FeS_2 (déposition du melnicovite). Les déterminations géothermométriques (Borcoş, 1968) ont indiqué des températures de formation dans ces stockwerks aurifères de 190°—200°, mais par endroits celles-ci ont pu être plus basses.

La conception génétique exposée fournie une explication plausible pour tous les faits, concernant la concentration de la minéralisation dans le gisement de Baia de Arieş.

Quelques observations sur d'autres corps de minerais

La minéralisation des autres stockwerks aurifères (no. III, IV, V, VI, VII) ressemble à celle décrite plus haut. Par endroit, on observe certains écarts dans la distribution de la minéralisation en relation avec les différences dans la morphologie des pièges structuraux.

Les filons aurifères, simples fentes dans le massif andésitique, peuvent être considérés également des pièges structuraux, qui ont été alimentés par les mêmes cheminées volcaniques.

Dans le gisement polymétallique d'AMBRU, les solutions hydrothermales, arrivées par de nombreux "breccia pipes" ont diffusé dans la structure extrêmement tectonisée et ont été captées dans les pièges constitués par les écrans de bancs de calcaires cristallins. Il en a résulté les amas extrêmement riches de galène, blende et pyrite.

Conclusions

L'expérience de Baia de Arieș représente un exemple convaincant pour le rôle que les pièges structuraux peuvent jouer dans la concentration des minerais et la distribution de la minéralisation en sens vertical. Dans une solution hydrothermale qui a subit un arrêt et une détente dans un piège structural, les phases volatiles des divers minéralisateurs se dégagent et se distribuent en fonction de la tension de vapeur de la composition, de la température et de la pression ou du système. Les différentes phases deviennent chimiquement actives par affranchissement à des niveaux convenables, ce qui fournit une explication pour la distribution de la minéralisation dans des systèmes métallogéniques plus compliqués que celui de la Baia de Arieș.

BIBLIOGRAPHIE

Borcoș M. (1968) Observații în legătură cu determinarea condițiilor termodinamice de formare a unor filoane și zăcăminte hidrotermale din regiunea Munților Metaliferi. II. Sectorul Brad-Bulza-Baia de Arieș. *Stud. cerc. geol.*, 13, 1.

Cochet J. R. (1957) Contribuții geologice asupra zăcămintelor aurifere de la Baia de Arieș. *Rev. Minelor*, VIII, 10,, p. 467–475.

Ghițulescu T. P., Socolescu M. (1941) Études géologiques et minières des Monts Métallifères. *An. Inst. geol. Rom.*, XXI, București.

— (1959) Rapport , les archives de I. M. Baia de Arieș.

— Sprinceană M. (1959) Rapport. les archives de I. M. Baia de Arieș.

— Pitulea G., Ghițulescu I. (1979 a) The role of the Volcanic Breccia Pipes in Metalogenesis in the Baia de Arieș Mining District. *Rev. Roum. Géol. Géophys., Géogr., Ser. Géol.*, 23, 1, Acad. RSR, București.

— Pitulea G. Ghițulescu I. (1979 b) Petrogenesis of the Volcanic Breccia Pipes at Baia de Arieș (Metalliferous Mountains). *Rev. Géol., Géophys., Géogr., Ser. Géol.*, 23, 2, Acad. RSR, București.

Nițulescu I., Mocanu O., Nicolau V. (1960) Studiul comparativ al stocurilor aurifere de la Baia de Arieș. *Cerc. metalurg. min., Publicații tehnice ASIT*, 2.

STRUCTOGENETIC AND PETROMAGNETIC METHODOLOGICAL STUDIES CONCERNING THE NEOGENE MAGMATITES IN THE METALIFERI MOUNTAINS (TRANSYLVANIA)[1]

BY

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Introduction

This study continues the petromagnetic investigations initiated by Romanescu in 1963 in the Metaliferi Mountains and extended in some other regions of the country. The results were partially published (Romanescu, 1963 a, 1963 b, 1969, 1970, 1972; Ghițulescu et al., 1981).

The investigated area — a classic one for the Neogene magmatism and its associated metallogenesis knowledge in the Romanian Carpathians Mountains — is to be found in between Crișul Alb and Arieș rivers. The main aim of this study is to solve some petrologic and prospecting problems using geologic and petromagnetic data. That's the reason why we have chosen some geologically well-known structures in the Brad and Baia de Arieș areas with a view of correlating some magnetic and petromagmatic data. The influence of the very intensive hydrothermal processes in this zone on the magnetic parameters, has also been taken into account.

A Brief Geological Survey on the Neogene Magmatism in the Studied Area

In a classic geological conception, the Neogene eruptivism represents the third magmatic episode in the Mureș eugeosyncline after the initial simatic magmatism and the banatitic epirogenic sialic one. The Neogene eruptivism has developed in 4 phases, whose ages could be well established

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in comparison with the Neogene molasse sediments. We could separate eruptive series of ejecta as stratigraphic categories (Ghițulescu, Socolescu, 1941; Rădulescu, Borcoș, 1969).

Detailed geological surveys at the surface, in galleries and wells have permitted distinguishing two kinds of structures: infrastructures within the pre-eruptive basement and suprastructures lying above it. The first ones contain the volcanic chimneys enlarged near the contact with the surface, which we named "initial crater". The volcanic chimneys are filled with polygenous breccia composed of rock fragments and blocks in a volcanic ash matrix. Lavas piled up by successive ejecta are sometimes to be found in the "initial crater". The chimney and the crater are penetrated by necks and subvolcanoes which can fill the whole cavity. We use the synthetical denomination of "enrooted structures" for chimneys, initial craters, necks, apophyses and dykes and also for breccia pipes which rove through all mentioned formations.

The suprastructures contain also volcanic tuffs and pyroclastics, as well as lavas and pyroclastic lavas which build up the volcanic cones and the effusive ejecta around them. We call these formations "extracraterial" in comparison with the initial crater. The above-mentioned structural differentiations and classifications play an important part in the prospecting works in the Metaliferi Mountains, because all the economically important deposits are directly bound up with the enrooted magmatic structures. Even Palfy (1912) remarked the preferential positions of the gold veins on the edge of the volcanic necks. Actually, the most favourable positions are the subvolcanic body apexes and their apophyses as well as the boundary of breccia pipes (Ghițulescu et al., 1979 b). The special interest in the structural category identification in the prospecting work is to be developed. One of our main goals is to find out specific petromagnetic characteristics for each kind of structure to be used as a quantitative criterion for the petrologic and prospecting works.

Petromagnetic and Petrologic Data

An important number of magnetic susceptibility (MS) and natural remanent magnetization (NRM) measurements on oriented samples have been performed. The results are indicated in Table (lines 1—24 for the Brad area and 25—60 lines for the Baia de Arieș area). The columns show the age, the eruptive phase, the geographical position, the rock species, the type of alteration (propylitic, hydrothermal, hypergenetic) and the mean values of magnetic parameters for each structure. The fresh state does not exclude modifications taking place during the magmatic stage, of importance in the interpretation of magnetic parameters.

A significant difference between the two regions is to be noticed. In the Brad area, the Neogene magmatism settled up from its very beginning, in the Lower Badenian or even earlier, developing in the first and second phases, which are missing in the Baia de Arieș area. The third phase (Barza Series) was more intensive, diversified and largely developed in the Brad area than in the other areas. On the contrary, the fourth phase is better represented in the Baia de Arieș area. This suggests some differ-

TABLE

Summary of petromagnetic data for the structures in the Metaliferi Mts (Brad and Baia de Arieş areas)

No	Age Phase	Location	Structure species	Rock	Alteration process	MS x.10⁻⁶ CGSu min.	max.	mean	In.10⁻⁶ CGSu NRM min.	max.	mean	Q
0	1	2	3	4	5	6	7	8	9	10	11	12
		Brad Area										
1	Triassic	Vălişoara, Fornădia	inla	pba, gab	fresh	900	2535	1970	180	9485	1320	1.8
2	Jurassic	Draica	pyla	pba	hyd	1450	2185	1780	120	135	130	0.2
3	Cretaceous	Valea Ruda	exla	pan	fresh	30	3680	2700	70	185	120	0.1
4	Lower Badenian phase I	Orminda, Petriceaua	subv	pan	fresh hyp	1230	2810	1680	285	4420	1740	2.2
5	Upper Badenian phase II	Clinclu, Gherghina Valley	neck	dan	(hyd)	2150	2610	2330	500	2285	1410	1.4
6		Măgura Curechiu	inla	an	fresh	1630	3310	2560	95	1105	560	0.5
7		Criscior, Pietrosul Hills	inla	an	fresh	1150	1620	1440	30	610	390	0.6
8		Criscior, N flank Pietrosu	pyla	an	hyp	120	210	170	1790	2280	2090	26.2
9		Criscior S flank Pietrosu	pyla	an	hyp	1890	3220	2640	450	1950	975	0.7
10		Barza, Zdraholţ	inla	an	ppy ; hyd	30	2420	1790	55	200	120	1.0
11	Sarmatian	Muncelu	neck	an	ppy ; hyp	2170	2860	2590	85	27081	3960	2.7
12		Măgura Musariu	neck	an		845	2785	1950	3650	12125	7340	8.2
13	phase IIIA	Dealul Fetii	neck	an	hyd	10	480	170	30	720	200	5.2
14		Muncelu Hill S flank	pyla	an	hyp	240	2090	1020	160	725	325	1.0
15		Tărăţel, Gruieţ Hill	pyla	an	hyp	920	3000	1460	520	555	540	1.0
16		Prihodişte	pyla	an	hyp	980	1090	1035	80	140	110	0.2
17		Arsului Valley, Măgura	inla	an	ppy (hyd)	60	2870	2120	15	540	290	0.3
18		Tărăţel, Criscior Hill	inla	an	fresh	2460	3680	3020	370	1295	750	0.5
19		Criscior, Valea Morii	pyla	an	hyp	2545	2560	2535	295	440	355	0.3
20		Barza Valley	pyla	mdp	ppy	1790	3190	2600	445	1015	715	0.7
21		Brad, Goşa Hill	pyla	an	hyp	2060	2550	2360	925	1845	1245	1.1
22	Pannonian	Hărţăgani, Gurguiata Hill	subv	dan	fresh	2460	2590	2510	1220	1870	1500	1.3
23a	phase	Cetraş Summit	neck	dan	fresh	1570	1980	1710	52170	190510	137740	171.5
23b	IIIB	Under Cetraş	aph	dan	fresh	2290	6620	2450	7290	7955	7685	6.6

24	Pliocene phase IV	Zimbriţa	neck	an	fresh	1260	2850	2190	1295	31965	7100	7.3
		Baia de Arieş Area										
25	Sarmatian phase III A	Afiniş; Vinăt	subv	an	(hyd)	655	1850	1260	865	4670	2005	3.4
26		Prunilor	subv	an	ppy	1720	2000	1860	10865	19420	15000	17.0
27		Afiniş	subv	an	hyd	870	1525	1200	1070	1795	1435	2.5
28		Hărmăneasa	inla	an	ppy	1215	4245	2375	125	765	430	0.4
29	Pannonian phase III B (Colţii Lazărului volcano)	Valea Lacului	inla	an	fresh	1520	2765	1985	155	1500	325	0.3
30		Valea Ambrului	inla	an	fresh	1105	3545	1935	80	4360	810	0.9
31		Valea Ambrului confl. with Valea Obirşiei	inla	an	fresh	975	1810	1385	595	1700	740	1.1
32		Valea Obirşiei	inla	an	fresh ; hyp	95	2015	110	30	2585	875	1.7
33		Catz Brook	inla	an	fresh ; hyp	225	2760	1810	305	2790	1455	1.7
34		Dealul Tuţii	inla	an	fresh	2015	3575	2960	1070	2520	1660	1.2
35		Colţii Lazărului (all lavas)	inla	an	fresh	95	3575	1865	30	4360	670	0.8
36		Valea Lacului	aph	an	fresh							
37		Valea Ambrului confl. with Valea Obirşiei	dyke	an	fresh	1410	2025	2750	3315	14595	10300	1.4
38		Under Baia Roşie	neck	an	fresh	1680	1685	1680	3480	5600	4540	5.7
39		Dealul Catz	neck	an	fresh	1280	2290	1800	1295	26340	13050	15.3
40		Dealul Tuţii	aph	an	fresh	1305	2050	1675	2100	55260	21500	27.0
41		Colţii Lazărului		an	fresh	2345	3185	2890	10500	24550	19465	14.2
		(all enrooted bodies)				1230	3185	1960	1295	55260	12770	14.8
42	Pannonian phase III B	Poieniţa	neck	dan	fresh	690	1700	1340	1295	112295	35155	48.6
43		Poieniţa — Virful Dorului	aph	dan	fresh	340	2390	1295	6100	106890	49370	80.8
44	Pliocene phase IV	Măzăratu Hill	exla	an	fresh	—	—	1365	—	—	1045	1.6
45		Măzăratu Hill	neck	an	fresh	355	2160	1175	57300	138400	85895	153.9
46		Colţii Mizernicului	aph	an	fresh	—	—	775	—	—	5715	15.5
47		Colţii Mizernicului	exla	an	fresh	1200	1400	1280	615	1355	1055	1.7
48		Tirşii Hill	exla	an	fresh	1510	1855	1695	635	1725	1225	1.5
49		Suligata Summit	neck	an	fresh	425	1235	720	5000	11320	7825	22.9
50		Puiul Suligatei	neck	an	fresh	1505	2215	1860	2430	10100	6265	7.1
51		Cărpiniş	pyla	an	fresh	245	1685	1020	200	2525	1120	2.3
52		East Geamăna	inla	an	fresh	260	405	335	1075	2210	1645	10.3
53		East Geamăna	neck	an	fresh	1280	1735	1440	1625	59575	20825	30.5
54		West Geamăna	inla	an	fresh	190	1335	740	910	2720	1735	4.9
55		West Geamăna	neck	an	fresh	2130	4460	3295	4560	236365	120460	77.0
56	Pannonian phase III B	Valea Steuinilor	inla	da	fresh	950	1530	1265	45	270	130	0.2
57		Valea Steuinilor	aph	da	fresh	1560	2115	1840	3775	6640	5210	6.0
58		Teuşeni Valley	inla	da	fresh	25	345	205	40	90	60	0.6
59		Colţii Cioranului	neck	da	fresh	1595	3945	2770	6805	191735	99270	75.5
60		Poieniţa Hill	inla	da	fresh	820	1680	1210	730	1880	1385	2.4

Abbreviations : pba = palaeobasalts ; pan = palaeoandesites ; an = andesites ; dan = daccandesites ; mpd = microporphyrydiorites ; da = dacites ; gab = gabbro ; inla = intracraterial lavas ;ex la = extracraterial lavas ; pyla = pyroclastic lavas ; subv =subvolcano ; aph = apophyses : ppy = propylitized ; hyd = hydrothermalized ; (hyd) = incipient hydro-thermalization ; hyp = hypergenetic alteration.

ences concerning the contemporary rock series formation conditions which could have influenced the magnetic parameter values.

Other magnetic parameters which are not presented in Table are to be discussed in some future studies on the above mentioned structures.

Some Petromagnetic Characteristics of the Investigated Eruptive Structures

Intracraterial lava structures

Many initial crater infrastructures filled with andesitic lavas become visible due to the high erosion degree in the area. Others could have been identified in galleries. In the Baia de Arieş area we studied two structures of this type; one in the Valea Socilor (Hărmăneasa) and the other in the Colţii Lazărului volcano (south of Baia de Arieş). In the Brad area we investigated the Măgura Curechiului, Pietrosul of Criscior, Măgura on the Arsului Valley and Criscior Hill of Tărăţel structures.

The petromagnetic characteristics of the Barza andesite series and the Cetraş dacoandesite series[5] which filled the initial crater, are indicated in Table (lines 6,7,17, 18, 28–32). The lava NRM has usually values lower than 1000.10^{-6} CGSu, but some exceptions occurring in the Colţii Lazărului volcano, where the measured values in the lava like rocks exceed this limit, are to be mentioned. Such situations appear in the subvolcanic apophyses and neck injection area, for example near the Catz and Tuţii necks (Tab. lines 33 and 34) where the mean values are over 1000.10^{-6} CGSu, or in the Valea Ambrului spring zone, the mean value being of 810.10^{-6} CGSu. These cases require a more detailed petrologic investigation in order to delimit the injection areas. The values presented in histograms (Fig. 1 a) point out dominant petromagnetic characteristics of the intracraterial lavas and the mentioned exceptions, too. MS values are normal for the respective type of rocks.

Extracraterial lava structures

These types of lavas are well developed in the Brad area, where the effusive ejecta created by the Barza and Cetraş andesite series were not eliminated by glyptogenesis. Most of them are of pyroclastic type (Ungureanu et al., 1978).

The petromagnetic parameter values are represented in Table, in lines 14–16 and 19 for the normal lavas and in the 8, 9 and 21 lines for the pyroclastic ones. In the pyroclastic lavas the somehow greater NRM values are explained by the presence of older andesite fragments caught in the lava matrix. The NRM measurements showed that the lava matrix has NRM values in the interval $400-500.10^{-6}$ CGSu and the polygenous blocks, in the interval $1900-2950.10^{-6}$ CGSu.

Among the structures created in the Barza Series the structure we discovered on the Barza Valley is to be mentioned. All the geologists have considered these rocks as lava flows, so far. The microscopic studies show that the rock is a microporphyrydiorite made up of a plagioclase crystal paste and femic mineral phenocrysts, completely transformed in different neominerals.

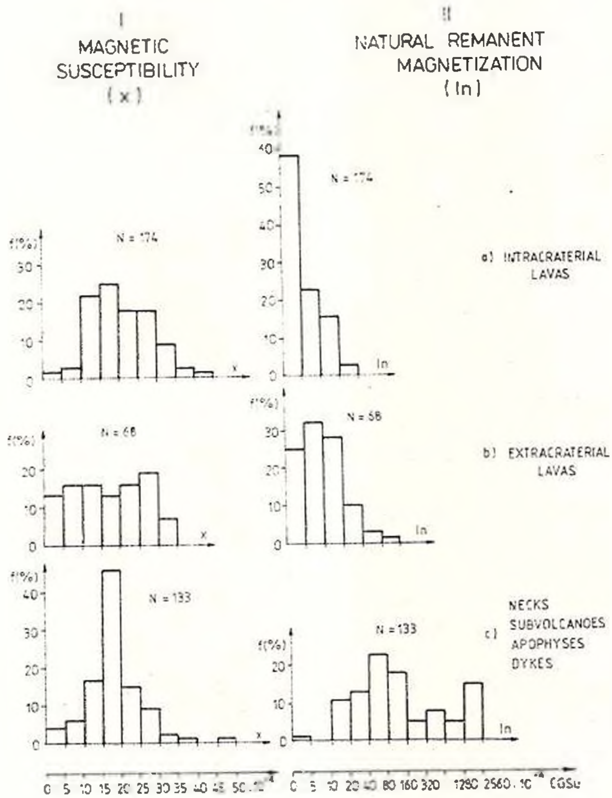


Fig. 1. — Synthetic histograms of the magnetic susceptibility (I) and natural remanent magnetization (II) for the structures in the Metaliferi Mountains (Brad and Baia de Arieş) areas).

The MS values are within normal limits for this type of rock, but the NRM has relatively low values.

As mentioned above, the fourth phase produced andesites and basalts, very well developed in the Baia de Arieş area. The extracraterial lavas of the Rotunda-Cărpiniş-Vîrtop volcano, some of them alternating with pyroclastics and in pyroclastic facies, have a normal mean susceptibility for igneous formations and relatively low NRM values (Tab., line 5).

The extracraterial lava flows in the same phase occur on the Tirşii Hill (Tab., lines 44, 47 and 48) having NRM values lower than for the rocks belonging to the necks of the same phase, for example at Suligata (Tab., lines 49, 50).

Regarding the petromagnetic characteristics of the Neogene lava in the area, we conclude that the intracraterial lavas (Fig. 1 a) and the extracraterial lavas (Fig. 1 b) have low and very low NRM values, the exceptions being petrologically explained.

Necks, subvolcanoes, apophyses, dykes

These kinds of structures appeared in a large number during the second and the third A phases, both in the Brad and in the Baia de Arieș areas, but few of them are in a fresh state, for example, the necks from Dealul Măgura and that from the mine south of Musariu (Brad area). The petromagnetic study of many oriented samples shows that the MS values are similar to those of lavas but the NRM values are much greater (Tab., line 12; Fig. 1 c).

In the Baia de Arieș area the Afiniș subvolcano (R o m a n e s c u, 1963; G h i ț u l e s c u et al., 1981), partially propylitized, partially intensively hydrothermalized, has been investigated. In the first zone the MS varies within normal limits, and NRM has high and very high values (Tab., line 26).

One can see an obvious difference between the lava and the enrooted body magnetic parameters, in the Barza Series rocks, namely, the NRM is low and very low for lavas and high for necks and subvolcanoes.

This difference is more striking for the dry magma of the third B and fourth phases. In the Colții Lazărului volcano, the andesites constituting the necks and subvolcanoes in the Catz, Baia Roșie and Vîrful Tuții hills and the apophyses and dykes in the volcanic area present considerably higher values for the NRM, for example $55260 . 10^{-6}$ CGSu in the Catz neck center (Tab., lines 38—40), the NRM values diminishing towards the edge zone (G h i ț u l e s c u et al., 1981).

The dacoandesites in the same phase constituted a volcanic cluster in the Poienița Mountain, containing lava craters such as the Valea Steuinilor structure (Tab., lines 56 and 57) and necks on the Poienița Peak (Tab., lines 42 and 43). In the latter case we get high and very high NRM values, but normal MS values.

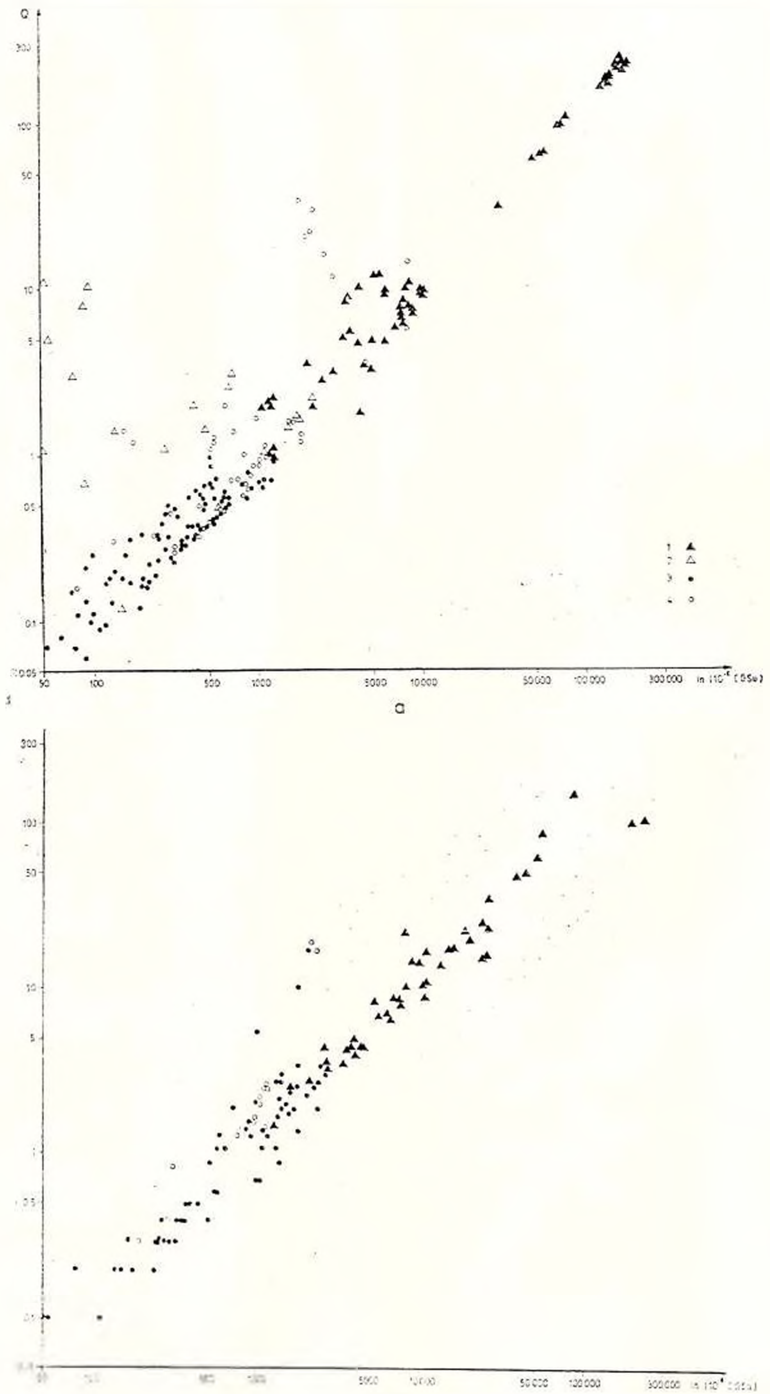
In the Brad area, the dacoandesites and the dacites in the third B phase are developed in the Cetraș-Duba-Bulzu Belt. R o m a n e s c u (1970) carried out some petromagnetic investigations for the Cetraș neck, getting very high values for NRM and for the Königsberger coefficient, too (Tab., line 23). A part of these exceptional high values might be attributed to the isothermal remanent magnetization produced by thunderbolts.

In the Baia de Arieș area, the Măzăratu (Tab., line 45), Suligata (Tab., lines 49, 50) and East Geamăna (Tab., line 55) structures have been studied; all of them are macroporphyric andesite necks formed during the fourth phase (Pliocene). NRM and Königsberger coefficient have high values contrasting with the low or average ones of the surrounding lavas of the same phase (Tab., lines 44, 47, 48).

The only known neck in the Brad area (Zîmbrița), created in the fourth phase has been investigated (Tab., line 24); its petromagnetic characteristics are similar to those of the necks in the Baia de Arieș area.

Pyroclastic deposits and volcanic chimneys

The characteristic formations of these structures are volcanic breccia made up of ash and fine fragment matrix and blocks of different origin



a

b

Fig. 2. — Q— $I_n$diagram for the structures in the Brad (a) and the Baia de Arieș (b) areas.
1, Necks, subvolcanoes, apophyses, fresh rocks; 2, hydrothermalized rocks; 3, intracraterial lavas; 4, extracraterial lavas.

and dimension rocks. Some matrix samples drawn in a borehole penetrating 800 m in the Colţii Lazărului volcanic chimney provided MS values of $210-780.10^{-6}$ CGSu and NRM values of $70-260.10^{-6}$ CGSu.

The petromagnetic investigations could become useful in establishing the origin of the endogenetic rock fragments to be found in the above mentioned formations.

The effect of the hydrothermal alteration process on the magnetic parameters

The papers dealing with this subject (G h i ţ u l e s c u et al., 1981) pointed out that the regional propylitization process might not affect the magnetic parameters while the potassic alteration process reduces both the MS and NRM values. Mention should be made that this diminishing is more rapid for NRM, suggesting the possibility of some practical applications in the prospecting works.

Conclusions

Our experimental results, partially used in this paper, led us to some interesting conclusions with regard to the possibility of getting some characteristics which allow an identification and separation of the Neogene magmatic structures in the Metaliferi Mountains.

The first conclusion is that the MS values of the Neogene igneous rocks in fresh state (Fig. 1, Ia, b, c) are within the same limits for any kind of structure (intracraterial lavas, extracraterial lavas, necks, subvolcanoes, apophyses, dykes) and any petrographic composition of the andesites, dacoandesites and dacites.

The second conclusion is that the NRM values (Fig. 1, II a, b, c; Fig. 2 a, b) are much lower for the intracraterial or extracraterial lavas than for the rocks of the same petrographic species in fresh state constituting necks, subvolcanoes, apophyses and dykes — the NRM values for the last being high or even very high. We noticed that the NRM values are sometimes lower in the edge zones of the enrooted bodies (Fig. 1, II c). Also, we pointed out that the Königsberger coefficient values are much higher for the igneous rocks generated in the final phases (Tab., Fig. 2 a, b) after the main mineralization phase, than for the rocks created in the earlier phases. We think that it might be due to the different magma state.

The potassic hydrothermal alteration process acted differently in diminishing the MS and NRM values. The way the mineralizing hydrothermal phenomena are reflected in the modifications of petromagnetic rock parameters might offer new perspectives in the prospecting works.

The last conclusion is on the usefulness of diversifying the petromagnetic researches. Therefore it is necessary to determine the primary remanent magnetization, the remanent magnetization vector orientations and the magnetic susceptibility anisotropy. Thus, it would be possible to develop a complex geological — petrophysical method able to solve many of the structural and metallogenetic problems in the Neogene magmatism area.

[4] Basalt belonging to the alpine ophyolites.

[5] Series composed of dacites and andesites

REFERENCES

Ghiţulescu T. P., Socolescu M. (1941) Étude géologique et minière des Monts Métallifères. *An. Inst. Geol. Roum.*, XXI, Bucureşti.

— Pitulea G., Ghiţulescu I. (1979 a) The role of the Volcanic Breccia Pipes in Metallogenesis in the Baia de Arieş Mining District. *Rev. Roum. Géol., Géophys., Géogr., Ser. Géol., Acad. RSR*, 23, 1. Bucureşti.

— Pitulea G., Ghiţulescu I. (1979 b) Petrogenesis of the Volcanic Breccia Pipes at Baia de Arieş (Metalliferous Mountains). *Rev. Géol. Géophys., Géogr., Ser. Géol., Acad. RSR*. 23, 2, Bucureşti.

— Romanescu D., Pitulea G., Ghiţulescu I., Georgescu A., Neştianu T. (1981) Études pétrogénétiques et pétromagnétiques dans la région de Baia de Arieş. *Rev. Roum. Géol., Géophys., Géogr., Ser. Géophys., Acad. RSR*, 25, Bucureşti.

Palfy M. (1912) Geologische Verhältnisse und Erzgänge des Siebenburgischen Erzgebirges. *Jahr., d. k. Ung. geol. Anst.*, XVIII, 45, 4, Budapest.

Rădulescu D., Borcoş M. (1967) Spätsubsequenter Alpiner Magmatismus in Rumänien. *Acta geol., Acad. Sci. Hung.* XI, 1—3, p. 139—152, Budapest.

Romanescu D., (1963 a) Cercetări magnetice asupra eruptivului neogen din partea de est a Munţilor Metaliferi (Baia de Arieş, Zlatna). *Asoc. Geol. Carp.-Balc. Lucr. Congr.* V (*1961*) VI, Bucureşti.

— (1963 b) Cazuri particulare în magnetizarea rocilor eruptive din Munţii Metaliferi. *Stud. cerc. geofiz., Acad. RSR*, I/1, Bucureşti.

— (1969) Trăsăturile caracteristice ale magnetizării rocilor din Munţii Apuseni şi din partea de nord a masivului Poiana Ruscă. *Stud. cerc. geol., geofiz., geogr., ser. geofiz., Acad. RSR*, 7, 2, Bucureşti.

— (1970) Iavlenia obratnogo namagnicivania voblastiah rasprostranenia izverjenih parod gor Metaliferi, Ruminii. *Rev. Rom. Géol., Géophys., Géogr. Ser. Géophys., Acad. RSR* 14, 1, Bucureşti.

— (1972) Stabilitatea magnetică a rocilor eruptive din Munţii Metaliferi. *Stud. cerc. geol. geofiz., geogr., ser. geofiz., Acad. RSR*, 10, 1, Bucureşti.

Ungureanu B., Aldea I., Ispas G., Chiorpec M. (1971) Report, the archives of the Geological Enterprise of Prospections for Solid Mineral Substances, Bucharest

URANIUM HYDROTHERMAL MINERALIZATION IN THE BORAČ-RUDNIK AREA (ŠUMADIJA) – POSSIBLE RELATION WITH BURIED STRATIFORM ORE DEPOSITS [1]

BY

DJORDJE KLAJN [2]

Introduction

The world energetic crisis, becoming apparent recently, requires intensive investigations of energetic mineral raw materials. Among them the nuclear raw materials occupy an important place.

The nuclear raw materials, especially uranium, appear as one of the most important energy sources. The need for uranium as source of energy will increase in the next period of time. Thus the attention must be directed to the discovery of new ore deposits. In the last twenty years sedimentary rocks were predominant in uranium production with further tendency to increase.

Geographic location. The studied area occupies some 500 sq. km in south Šumadija, aligned with Zapadna Morava, Ljig, Gruža, Rudnik and Belanovica. This is a hilly area of Rudnik, Borač, Kotlenik, Rajac and other mountains, forming northwest trending ranges with altitudes of about 600 m, with the highest peak Šturac (1132 m).

Previous investigations. Researches of nuclear raw materials started in Šumadija in 1951 and in this area in 1964. According to the information obtained by airborne and hydrogeochemical prospection, geological, geophysical and geochemical investigations have been done, resulting in the discovery of numerous ore occurrences, anomalies of radioactivity and geochemical anomalies, as well as one uranium ore deposit.

Research works were reopened in 1976, after preliminary metallogenetic studies and reconnaissance of the ore controlling factors in the areas promising for the discovery of buried ore deposits.

[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, September 8–13, 1981, Bucharest, Romania.

[2] Geoinstitute, Beograd, Rovinjska 12.

Geological-Structural Setting

The Borač-Rudnik area consists chiefly of Paleozoic schists, Mesozoic molasse — carbonate sediments and a Cainozoic volcanic — sedimentary sequence, forming Middle and Upper structural stages. The Lower structural stage constitutes (1) the sequence of sericite-chlorite and phyllite schists, then schistose sandstones and limestones. The oldest rocks are sericite schists from the lower part of this structural stage, including lenses of graphitic schists without any faunal evidence of age. Comparing with the age of the corresponding schists of the Serbian-Macedonian mass, Eastern Serbia and Macedonia, the writer adopts the Lower Paleozoic, most probable Cambrian-Ordovician age for the sericite-chlorite schists of the studied area. The upper part of this structural stage consists of phyllitic schists, probably Silurian in age, then of Devonian schistose sandstones, as well as of Upper Permian sandstones and limestones.

The Middle structural stage incorporates Lower-Upper Triassic limestones, marls, sandstones and dolomites, unconformably underlain by a metamorphic sequence, then Jurassic serpentinites and serpentinized peridotites, gabbros, limestones, argillites, sandstones, conglomerates and conglomeratic sandstones, and finally Cretaceous sandy marls, limestones, flysch, clastics and sandy-marly limestones.

The Upper structural stage consists of (2) Oligocene dacitic-andesitic pyroclastics, dacites, andesites, quartz-andesites and altered andesites, up to Miocene conglomerates, argillaceous marls, sandstones, tuffs, quartzlatites, quartzlatitic ignimbrites and pyroxene dacites, then of Mio-Pliocene clastics, limestones and carbonaceous clays, Pliocene tuffs, tuffites, labrador andesites and trachyandesites, up to Quaternary deposits.

In the Rudnik Mountain the limited outcrops of Paleocene granitic rocks (quartzdiorite, granodiorite, quartzmonzonite, granitemonzonite, minette and graniteporphyry) are known.

Volcanic-activity is related to the development of Neogene lacustrine basins. Several volcanic phases were inaugurated by alternated tectonic movements starting from Oligomiocene up to Pontian, influencing the form of lake basins.

During the first, pre-Miocene phase daciteandesites and large masses of pyroclastics were largely extruded; the second-Miocene phase produced quartzlatites and dacites; in the third — post-Miocene phase came amphibole-pyroxene andesites, andesite-basalts and basalts followed by large quantities of pyroclastic material.

Tertiary volcanics are restricted to the calcalkalic type, mostly of the granodioritic and opdalitic magma. According to the chemical composition, andesites and dacites are acidic rocks, rich in silica and potassium in all the rock types, indicating the latite-quartzlatite characteristics of magma.

The Borač-Rudnik area is restricted to the geotectonic zone of the Internal Dinarids, exhibiting complex structural features, including internal, medial and external subzones. It is situated in the internal border of the Dinaric orogen, faced with the Serbian-Macedonian zone. The area is accordingly restricted to the Vardar-Šumadija geotectonic zone and the Šumadija ore district. It is bounded to the east by the Central-Šumadija north-northwest trending fault zone (8). It has been reactivated in different epochs, also recently, influencing the facial development of this area as well as the distribution of ore districts including Pb-Zn and other mineralizations.

Beside this fracture system of the first order, numerous large second order northwesterly trending fault structures are present, accompanied by numerous third order fractures of the east-west and northeast trend.

The Paleozoic schist sequence constitutes the Divčina anticline with a northwest trend of the axis, restricted to the Caledonian orogenesis. Schists exhibit minor fold structures, of metric and decimetric size, showing foliation with both northeast and southwest dips.

Alpine post-Cretaceous orogenic movements produced the folding of the Mesozoic sedimentary complex, created the new fault structures as well as the reactivation of ancient structures. Serpentinites overthrust the Triassic and Permian sediments pronouncing imbricated structures.

Finally, in this area, many morphostructural units of different shape and lithofacial development are to be distinguished : depressions with volcanogenetic complexes, depressions with pyroclastics, dislocations, formations of volcano-sedimentary deposits and volcanogenetic zones.

Ore Deposits

Uraniferous area is disposed in the Internal Dinarids as a general geotectonic unit with specific metallogeny and petrogenetic complex. Tectonic style and metallogeny have been developed according to the tonalitic system of deep faults of planetary size, resembling the lineament in the Earth crust, which was active from Archaic up to Quaternary, as a line of the continental drift. This system of deep fractures was reactivated and served for the penetration of the intrusive and extrusive igneous rocks.

The Šumadija ore district was the scene of possible Caledonian or Variscan and Alpine metallogenetic events, followed by pre-Alpine and Neo-Alpine stages. The typomorphism of the uranium mineralization is typical for the Variscan and Alpine orogenic and postorogenic stages, less expressed in geosynclinal stages.

In the Dinarids, the metallogenetic climax was reached in the Caledonian and Variscian epoch and in the Serbian-Macedonian mass it culminated in the Alpine or Neo-Alpine epoch. This late event coincides with the metallogenetic climax in the Šumadija ore district.

Uranium ore occurrences are found in volcanic, volcano-sedimentary and metamorphic rocks. Uranium in volcanics has been related to the shear zones, single veins and also occurs as impregnation. In altered dacites and quartzlatites the thickness of these zones varies from 0.1—0.8 m and uranium grades from 0.01—0.25 %, coinciding with cryptocrystalline apatite. In the altered andesitic breccias the thickness of the uranium bearing zones varies from 0.5—1.25 m, and grades of uranium are 0.0047 — 0.066 %.

Most of uranium ore occurrences are restricted to the Borač calderas, most of them connected to the Trijeska neck, where the uranium has been deposited in fissures with opal veinlets and apatite.

In the Livadak locality uranium mineralization occurs in fine-grained kaolinized volcanic breccias and clayey veins with illite, apatite and sporadic pyrite and chalcopyrite. In a similar way appear the Laz and Strane uranium mineralizations.

The Spasovina uranium ore occurrence in the volcanic-sedimentary rocks is disposed in the altered andesitic breccia, accompanied by clay, opal, limonite and apatite as bearer of uranium and polymetallic accessory minerals. In the Jasike locality, uranium with apatite is associated with pyrite and magnetite. The depth of uranium of this type is limited, about 50 m from the surface.

The origin of these mineralizations is polyphase (6) being different for apatite, originated from solphatara and fumorala activity, and uranium, leached by surface water from the surrounding rocks. It is supposed that chief uranium amounts have been deposited much more in depth connected with volcanic cupolas of the subvolcanic bodies etc, as well as trapped by fault clay and other screen types.

Uranium mineralizations in metamorphic rocks are deposited in graphitic schists, during the hydrothermal stage associated with paleovolcanic centres remarkable in scanograms as ring structures. The Mandre uranium ore occurrence is tracable for 100 m, 0.7—1.7 m in thickness; uranium grades are 0.0050—0.093 %. The vein filling is crushed and argillated graphitic schist, uranium mineral is pitchblende, associated with pyrite, chalcopyrite, tetrahedrite, marcasite, galena, ankerite, dolomite and calcite, as well as with sporadic martite, magnetite and chromspinel, radiobarite and siderite. Pitch blende occurs as isolated grains or coatings.

Out of the uraniferous Borač-Rudnik area, in the northern part of the Šumadija uranium district there occurs the Bukulja uranium ore locality related mostly to granitic rocks, but a considerable amount of uranium is related to the sedimentary and volcanic rocks. Those are hydrothermal-epithermal and infiltration types of mineralizations represented by pitchblende as chief uranium mineral.

Metallogenetic Studies

Metallogenetic studies of the area, realized by formational analysis (7), suggested the postvolcanic-postmagmatic hydrothermal origin for sporadic uranium concentrations, characterized by the following types of uranium ore deposits.

a) Endogenous epithermal ore deposits in permeable rocks of volcanic-tectonic depressions, with sedimentary fillings, and profound geologic significance.

b) Exogenous epigenetic ore deposits in the marginal parts of intermountainous depressions, built up of continental grey bituminous-limestone, carbonaceous or red sedimentary formations, also in limited intermountainous depressions, old river beds and marginal zones of epicontinental basins with alluvial-proluvial variegated rocks.

There is a certain probability that the vein and stratiform epithermal uranium ore deposits, related to paleovolcanic structures, be found in structures of calderas, necks, explosive breccias and dislocations of second and third order, in altered rocks, traced by aureoles of hydrothermal uranium. Thus, the most promising are the marginal parts of the Borač caldera and Mandra, less in the marginal parts of other calderas.

Here is also a probability that the corresponding epithermal uranium ore deposits be discovered in the volcano-sedimentary complex near Mandra. This is evidenced by numerous aureoles of dispersed uranium oxide and mercury, outcrops of altered rocks and other phenomena characteristic for the volcanic complex.

It is possible that the epithermal vein and stratiform uranium ore deposits be discovered in the sericite-chlorite schists, under graphitic schists, than in altered volcanic breccias, in the form of the stratiform diagenetic-sedimentary mineralization near Mandra.

Finally, the buried exogenous infiltration-diagenetic or hydrothermal uranium concentrations may be found in clastic continental sediments with organic material in the gulfs of the Borač and Rudnik volcanic complexes.

Origin of Uranium

The origin of uranium studies has generally been considered to be of profound geologic significance for the discovery of buried uranium ore deposits.

Difficulties arose because of the insufficient reconnaissance of the metamorphic rock sequence, the lack of isotopic age determinations, limited outcrops and the complexity of the structural and geologic features, marked by repeated tectono-magmatic activity. To exceed the conflicting evidence of incomplete information, different studies and considerations have been performed and discussed. These are structural-formational metallogenetic considerations, remodelling of the geologic development of the Šumadija area, familiarizing with the evolution of the world uranium deposits of corresponding type (5) which are largely commented.

As a result of these studies it was concluded that the rocks of the pre-Cambrian or Cambrian central mass, existing beneath the Mandra metamorphic (Silurian?) schists, were the primary source of uranium, dispersed after the destruction of the core and partially redeposited in black shales and other geosynclinal rocks. Later orogenic and magmatic events introduced uranium either of magmatic origin or remobilized from uraniferous sedimentary rocks and black shales.

Borač and Mandra uranium ore occurrences were actually created by the remobilization of uranium from black shales, in the way of leaching by postvolcanic hydrothermal solutions during the Alpine orogenesis.

Primary syngenetic-diagenetic uranium, associated with the carbonaceous material, was redistributed during the metamorphism. Tectono-magmatic reactivations, accompanied by postmagmatic hydrothermal activity in the Hercynian and the Alpine orogenic stage,could restore distribution of uranium, resulting in the creation of buried polygenetic economic uranium mineralizations. Mandra and Borač uranium ore occurrences could be the indicators of such a buried mineralization (9) or concentrations of uranium in underlying black shales.

The sharp decline of uranium contents in the deep parts of the Mandra ore veins speaks about separate and subsequent deposition of uranium. Argillization of country rock prevented the introduction of larger uranium quantities in this deposit by open channels. By this means, the size of the actual ore occurrences is not a good indicator for uranium concentrations in the underlying metamorphic rocks. However, reduced uranium in the deeper part of veins, speaks about the relatively limited depth of uranium-bearing rocks (500—1000 m).

The relation of other uranium ore occurrences of Borač with the uranium from black shales has been confused by three or more successive stages of the volcanic activity.

At the end, it should be defined that the origin of uranium in the aspect of discovery of the buried stratiform uranium concentrations in the anticlinal core of the metamorphic sequence in this area, is of essential importance for further orientations of investigations in Šumadija and Yugoslavia as a whole.

REFERENCES

Geologija Srbije (1975) Prekambrijum i paleozoik. knjiga II-1. Beograd.

Filipović V. (1968) i 1871 Prospekcija urana u oblasti Šumadije. FSD Geoinstituta, Beograd.

Ilavski J. (1974) Conclusions concernant la metalogenie du système plisse alpin de l'Europe Centrale et Sud-orientale en 2 500 000. K-BGA X congress, Bratislava.

Kazanskij V. I. i dr. (1978) Evolucija uranovogo rudnoobrazovanija. Moskva.

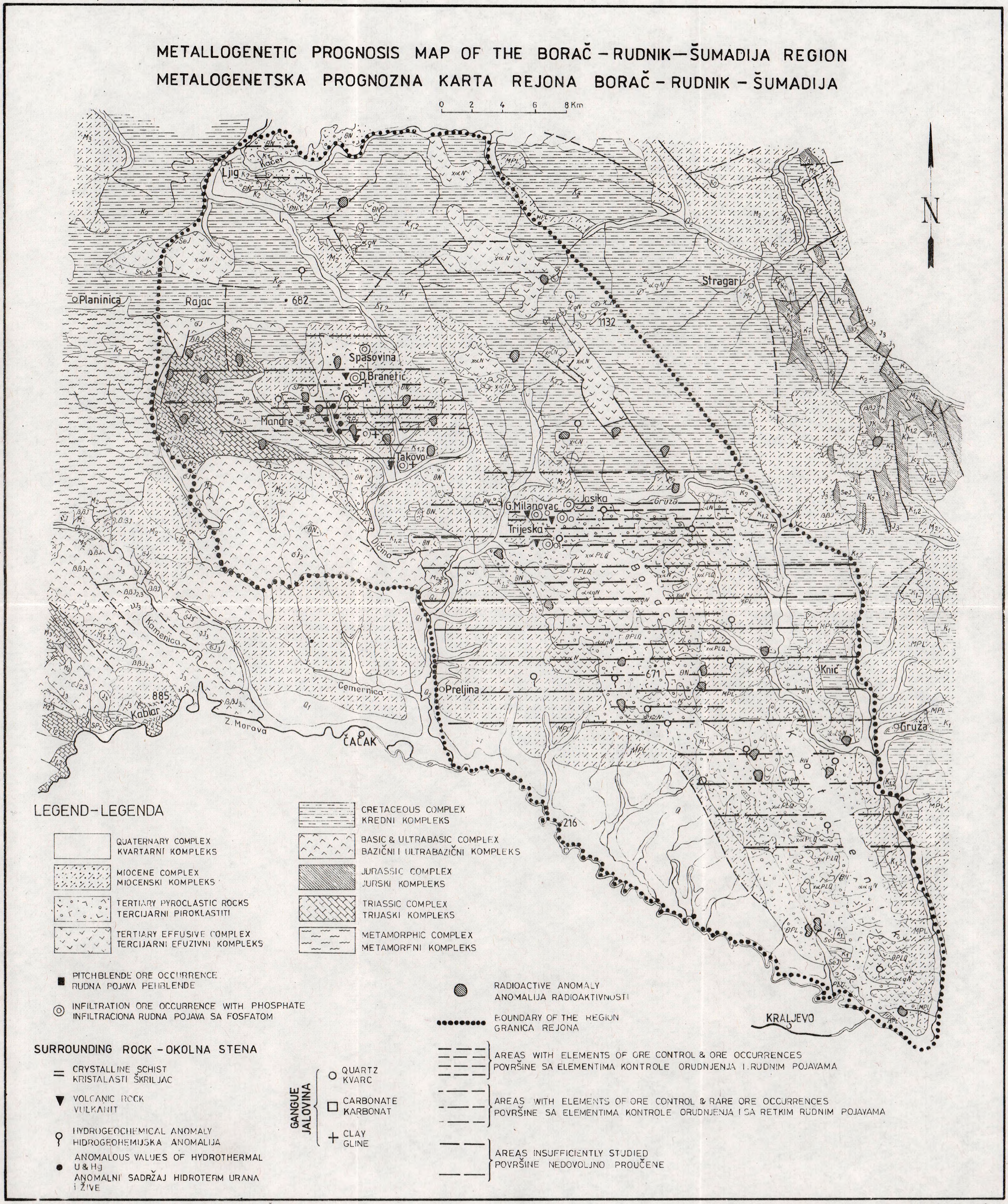
Ristić M. (1964) Genetsko-strukturni tipovi mineralizacija urana Jugoslavije. Doktorska disertacija RGF, Beograd.

— Klajn D. (1976) Metalogenetska izučavanja uranonosnoh rudnog rejona Šumadije. FSD Geoinstituta, Beograd.

Smirnov V. J. (1965) Geologija poleznih iskopaemih. Moskva.

Volfson F. N. (1978) Gidrotermalnije mestoroždenija urana. Moskva.

Vukašinović S. (1963) Aeromagnetska ispitivanja. FSD Geoinstituta, Beograd.



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PETROLOGY OF THE MAGMATITES IN THE ŞAUA LILIANA-VALEA GALBENA REGION (BIHOR MASSIF)[1]

BY

ALEXANDRU Z. MANEA [2]

Introduction. The magmatic rocks of the region between Şaua Liliana and Valea Galbena have not been mentioned or studied so far. They make up a series of dykes and bodies that pierce the sedimentary cover. The above mentioned area is situated in the NW part of the Bihor Massif that was investigated as early as the 19th century by P o s e p n y (1874) and P r i m i c s (1892).

In our century, namely in the 50^{ies}, the geological research of the massif gains ground due to the researches undertaken by A r a b u (1941), K r ä u t n e r (1941), G i u ş c ă (1950), B l e a h u (1957), B l e a h u, D i m i t r e s c u (1957), D i m i t r e s c u (1959, 1969), B l e a h u, M a n t e a (1962), R a f a l e t (1963), I s t r a t e, P r e d a (1970), B o r d e a, B o r d e a (1973), M a n e a (1973), M a n e a et al. (1975), B o r d e a et al. (1975), I s t r a t e (1975), I a n o v i c i et al. (1976), M a n e a, S e r i n i (1980), that bring important petrographic, stratigraphic and structural contributions.

Region's geology. The investigated area is characterized from the point of view of geology by the existence of sedimentary formations that structurally belong to the Bihor Autochthon, being assigned to the Kimmeridgian-Tithonian and Barremian-Aptian, as well as to the Arieşeni Unit, with epicontinental formations assigned to the Permian and formations representing the Codru Nappe system, assigned to the Lower and Middle Trias. Among the mentioned units there are only disjunctive relations. So, between the Bihor Autochthon and the formations belonging to both the Arieşeni Unit and the Codru Nappe system there appears the Cresuia-Galbena-Arieşeni fault, with regional character, trending NNW-SSE; between the Arieşeni and Codru Units there is the Muncelul-Plăiuţ-Păuleasa fault, trending ENE-WSW. Besides the sedimentary

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formations, the magmatic rocks of the banatite group participate in the geologic configuration of the region.

Magmatic rocks petrography. According to their mineralogic composition, on the basis of structural, textural and chemical characters, the studied magmatic rocks have been grouped in : andesites, rhyolites, porphyric granites and lamprophyres.

Microscopic studies have permitted the separation in the andesite group of the following types : basaltic andesites, andesites with pyroxenes $\pm$ hornblende $\pm$ biotite and quartziferous andesites. Generally the andesites are made up of a pilotaxitic mass with microlitic character, in which feldspar phenocrysts, femic minerals and even quartz are included. The feldspars are calc-sodic and appear as phenocrysts or microlites or even porphyric fractions. Function of the petrographic type, the plagioclases appear under the form of long, often zoned with 55—60% An (basaltic andesites) or hypidiomorphous phenocrysts, and sometimes cataclased ones, with complex macles (andesites with pyroxenes $\pm$ pyroxenes $\pm$ biotite). In this case the phenocryst core is more basic and the edges more acid, corresponding to andesine (the extinction angle is 20—26°). The phenocrysts are sometimes replaced by big epidote nests,or are slightly argillized, carbonated or more rarely sericitized, and in the case of quartziferous andesites, together with plagioclase crystals there appear quartz phenocrysts corroded by the groundmass. The phenocrysts are sometimes pierced by veinlets filled with chlorite, epidote and calcite.

Femic minerals are represented by pyroxenes (augite), that appear in short prism crystals, with twins and chloritizations. Crystals with octagonal outline or basal sections of rhombic pyroxenes (hypersthene ?) have also been noticed, but in most cases there are typical basal sections of idiomorphous pyroxenes corresponding to clinopyroxene, and in some sections there appear actinote pseudomorphoses and especially cases in which the pyroxenes are replaced by sheaf-like actinote. The pyroxene phenocrysts are not big and do not differ from those in the groundmass, and in the cases of replacements by actinote, this one is associated with carbonates or with chlorite and epidote. But for pyroxenes, in the andesites with pyroxenes $\pm$ hornblende $\pm$ biotite, the amphibole is represented by hornblende that appears in intensely resorbed phenocrysts with idiomorphous outlines. They exhibit poikilitic structures, marked by small plagioclase and pyroxene inclusions. Often, the phenocrysts have a rather big, brown core, with intense pleochroism and a 5—6° extinction angle. In the quartziferous andesites, hornblende appears in lamellar phenocrysts, with long habitus and outlines irregular in shape or changed by weathering in chlorite, epidote and hydromuscovite.

At all the petrographic andesite types, the groundmass is fresh, with microlitic character, made of plagioclases, partially chloritized augite, rod-like in the case of basaltic andesites-microlites and grains with irregular outlines of feldspars, quartz and femic minerals of slightly modified composition at the other types. The rock is pierced by veinlets filled with epidote or calcite and adularia.

Basaltic andesites build up an almost circular body placed west of Poiana Virciroagele. They marginally pass to the facies of andesites with pyroxenes ± hornblende ± biotite and pierce the Anisian dolomites of the Codru Nappe system.

The andesites with pyroxenes ± hornblende ± biotite together with the lamprophyres make up pairs of veins placed between the Werfenian quartzitic sandstones with Anisian dolomites and the black limestones with Ladinian siliceous accidents that also belong to the Codru Nappe system situated east of Poiana Vîrcioroagele.

Quartziferous andesites represent elliptical bodies inserted on the Juleşti-Plăiuţ-Păuleasa fault, trending ENE-WSW and having a character of depth consolidation.

The ignimbritic rhyolites appear as rocks with lithocrystalvitroclastic structure that include fragments of rocks different in composition and structure, contained in the groundmass which is made up of glass and ash, exhibiting benthonitization phenomena. Often it also appears as a microlitic mass made up of quartz and feldspars, the latter being calc-sodic and including quartz and microperthite phenocrysts and fragments that they corrode. The groundmass also contains quartz and microperthite, feldspar chippy crystallites as well as biotite lamellae. The biotite is chloritized and deferrized being moulded by joined glass fragments. The latter are decomposed in potash feldspar and quartz or in argillaceous mass made up of fine illite and montmorillonite lamellae. Fragments with pseudofluidal structure, marked by the parallel disposition of the biotite lamellae as well as argillaceous mineral aggregates have been noticed. The included fragments are represented by quartzitic sandstones, sericitous quartzitic schists and flattened glass fragments. These rocks build up ellipsoidal bodies that pierce the epicontinental formations belonging to the Arieşeni Unit in Dealul Vîrsecilor and veins trending NNW-SSE.

Porphyric granites appear as porphyric structure rocks and rocks of spherulitic groundmass, with massive texture, orbicular sometimes. Their mineralogic composition is marked by the presence of alkaline and calc-sodic feldspars, of quartz, biotite and of a groundmass with spherulitic character, accompanied by epidote, quartz, potash feldspar. In the porphyric fraction, potash feldspar is represented by orthose that appears in small phenocrysts with characteristic twins and often sericitized. The plagioclase feldspar is albite-oligoclase, very rarely andesine, and biotite appears in almost completely chloritized lamellae.

The groundmass is made up of almost equigranular grains with spherulitic character, with fibroradial structure made up of calcedony and quartzine as well as potash feldspar, accompanied by epidote crystals. It also contains granophyric, radial, aciculaı quartz and feldspar concretions that are often disposed on potash feldspar crystals. These rocks make up a dyke trending NNW-SSE and pierce the Werfenian quartzitic sandstones that crop out west of Poiana Tomnatec.

Lamprophyres (malchites and spessartites) appear as veins with trends characteristic of the banatitic system. Malchites appear as fine-grained grey-green or dark green rocks, with porphyric structure and massive texture. The porphyric separate is made up of calc-sodic feldspars and amphiboles, in various chloritization stages. The groundmass consists of feldspars, actinote, epidote nests, chlorite (engendered at the expense of biotite) and quartz. Plagioclase phenocrysts (32–35% An) with partial epidotizations also appear.

Spessartites appear as microgranular rocks, sometimes with porphyric areas, grey blackish or dark green in colour, with porphyric structure and massive texture. These rocks are made up of green and brown hornblende that participates in the composition of both the groundmass and of the porphyric separate in which diopside also participates under the form of slightly chloritized prismatic crystals. Plagioclases appear as twinned phenocrysts, with 20–25% An. Only seldom appear slight argillization phenomena. The groundmass is made up of amphibole, in which there are feldspars and apatite grains.

Petrochemical studies. In order to establish the petrochemical characters of magmatites, 23 samples have been analyzed that have been taken from various magmatite sampling localities of the region under investigation.

As the diagram of Figure 1 points out, andesites appear in domains, ranging from semifemic to paralfemic, isofalic, semisalic, salic and subfemic, while rhyolites appear only in the salic rock domain. Porphyric granites are distributed in the domain of salic and semisalic rocks, and the femic character of lamprophyres is marked by their distribution in the domain assigned to femic and semifemic rocks.

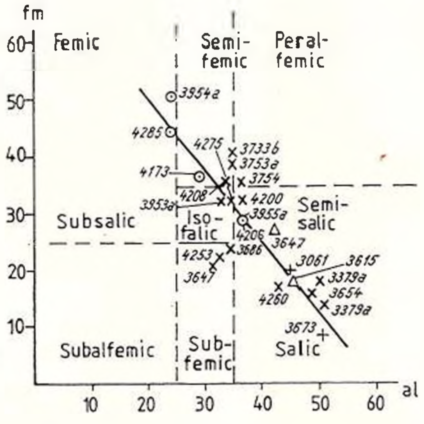

Fig. 1. — *al* : *fm* diagram.

The distribution of samples in diagram *QLM* (Fig. 2) points to the grouping of the most basic rocks in the middle of the diagram, towards the PF line, and of the acid ones towards the upper part, in the area of saturated rocks.

In the diagram of Figure 3, the position of the samples used for tracing the median line reflects an evolution of the magmas similar with the line characterizing the alkaline basaltic series.

In the diagram of Figure 4 the andesites and the lamprophyres are grouped towards the Ca point and there appears a marked dispersion of the other petrographic types.

The linear evolution of the alkalinity index between 1.0–2.5, corresponding to the values of SiO_2 of 50–70% (Fig. 5) marks the fact that all the magmatite types belong to the calc-alkaline series.

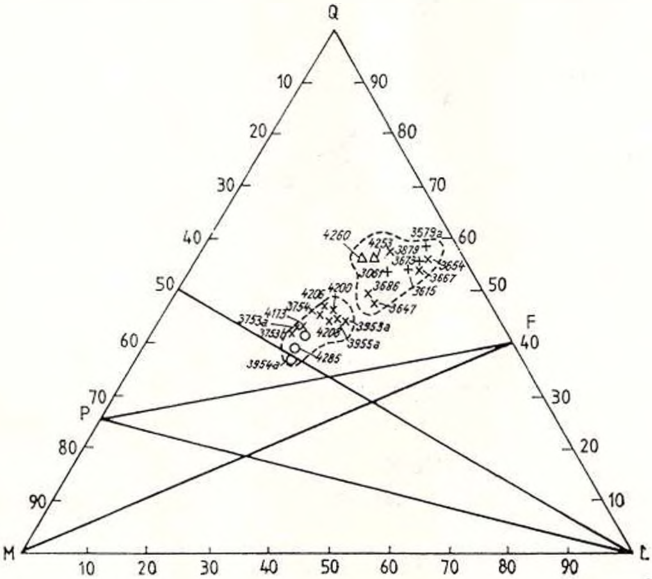

Fig. 2. – *QML* diagram.

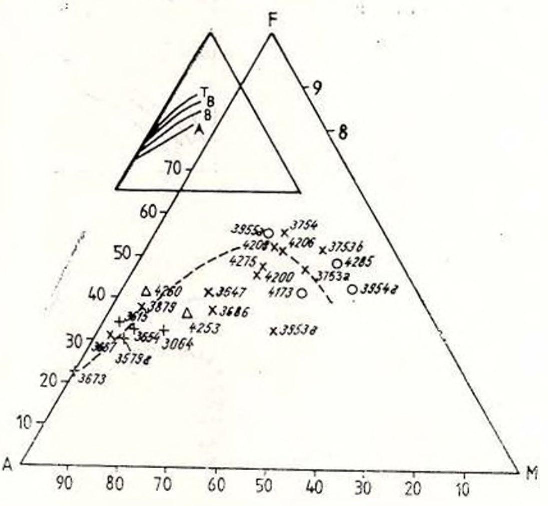

Fig. 3. – FMA diagram.

In the diagram of Figure 6 the andesites and lamprophyres are concentrated between the 10–25 isofrequency curves and the other types are rather dispersed between the 0–10 curves of field B. That proves that the respective rocks originate in sialic magma, that has been generated during orogenic processes.

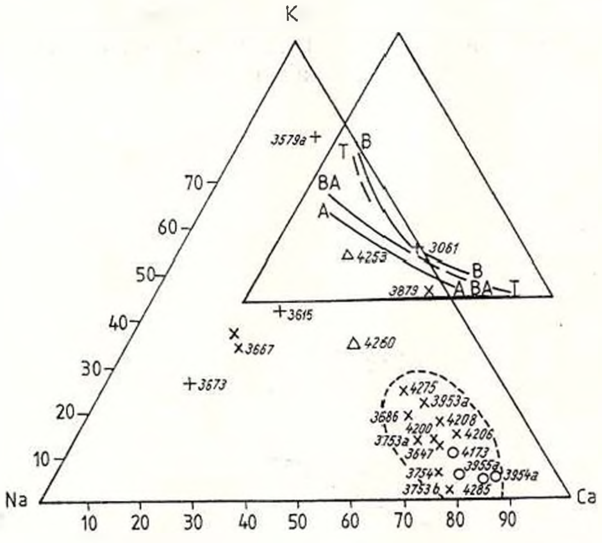

Fig. 4. — KNaCa diagram. A = Crater Lake, B = Californian Batholith, T = = Tholeiitic series, BA = Alkaline basaltic series.

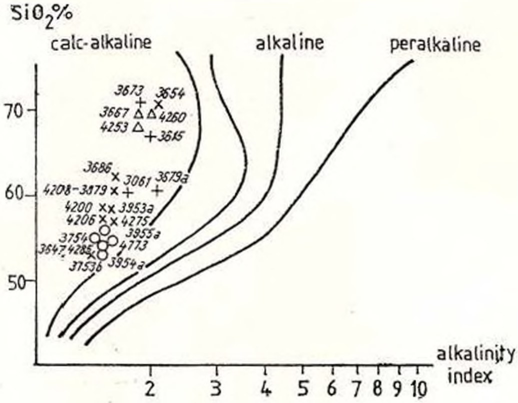

Fig. 5. — Diagram of the variations of the alkalinity ratio.

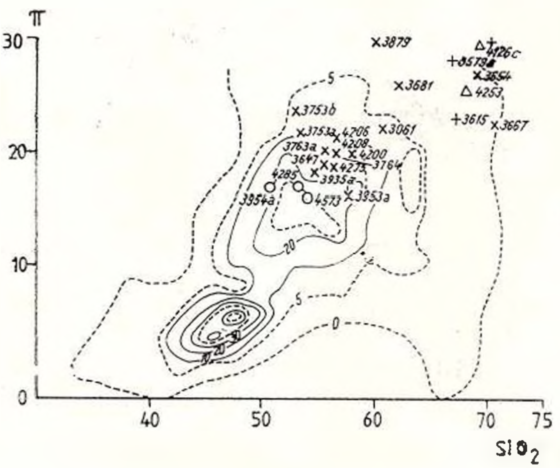

Fig. 6. — π /SiO_2 Gottini diagram.

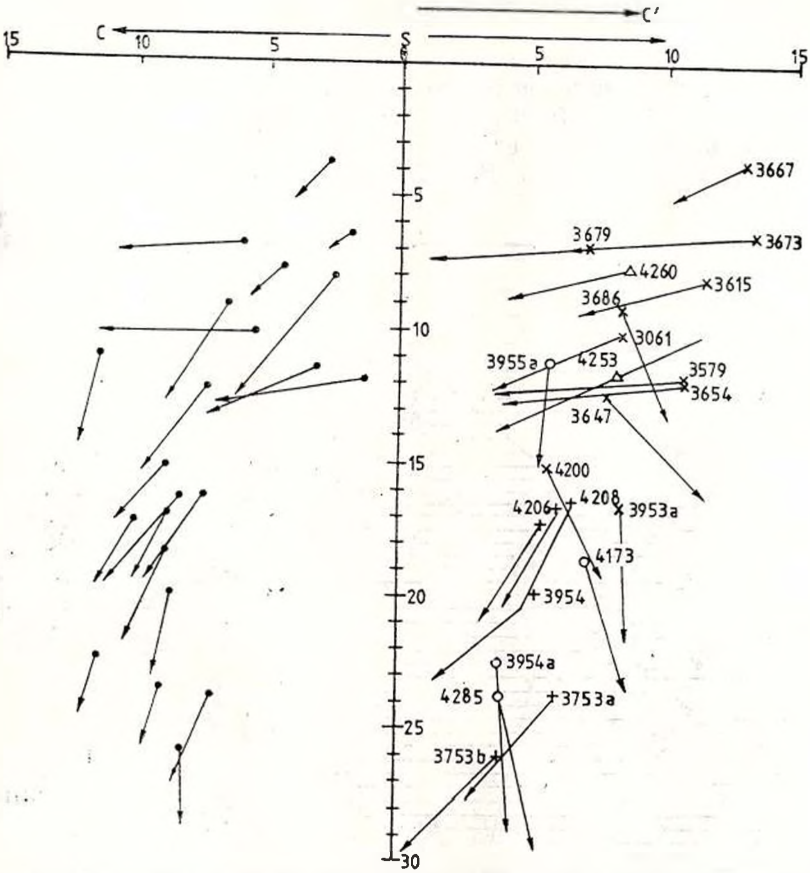

Fig. 7. — Diagram of the magmatic rock chemical composition after Z a v a r i ţ k i..

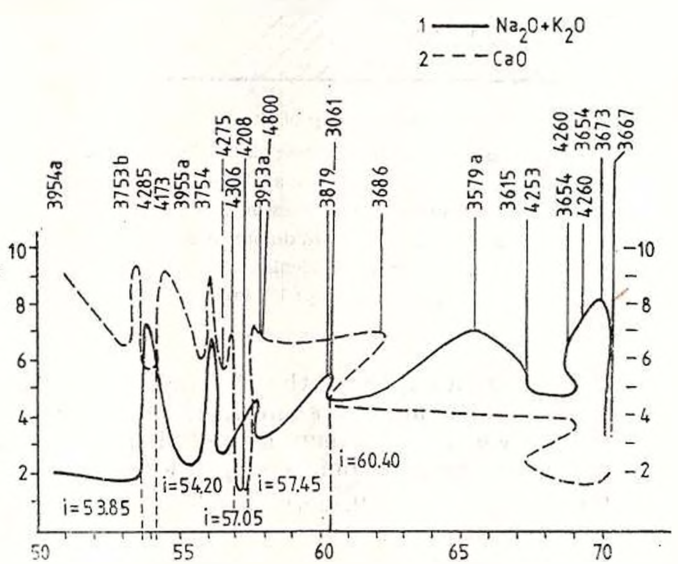

Fig. 8. — Peacock index diagram.

The vector distribution in the diagram of Figure 7 is oriented starting from the acid rock field up to that of basic rocks and their position in the CSB plane marks the existence of plagioclase feldspars in the basic rocks and of potash ones in the acid rocks.

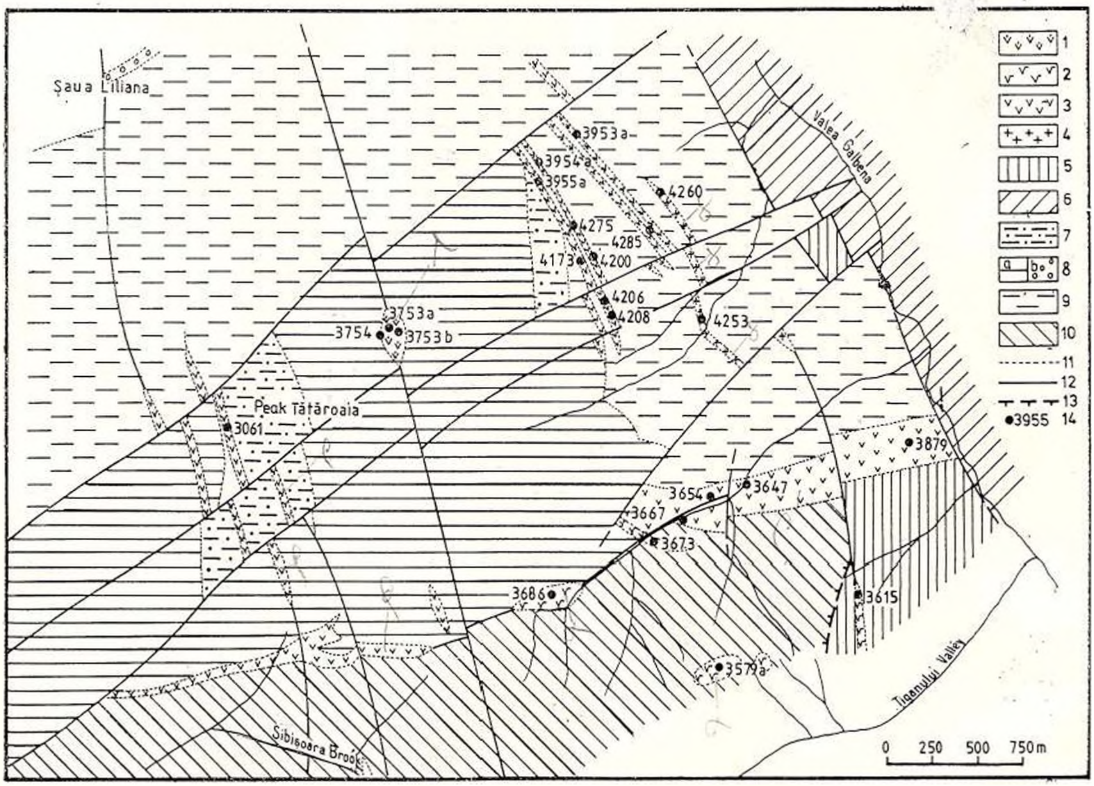

Fig. 9. — Geological map of the Şaua Liliana-Valea Galbena region.

Laramic magmatites : 1, lamprophyres ; 2, andesites ; 3, rhyolites ; 4, granites. Cretaceous-Barremian-Aptian : 5, limestones, black argillaceous shales. Jurassic-Tithonian-Kimmeridgian ; 6, white limestones, dolomites, black limestones, limy shales. Ladinian : 7, black limestones. Anisian-Campillian :8 a, grey dolomites, b, dolomitic limestones with brucite. Seisian : 9, conglomerates, sandstones, shales. Permo-Werfenian : 10, conglomerates, vermicular sandstones, violet shales ; 11, geological boundary ; 12, fault ; 13, thrust line ; 14, sample's location.

Out of the analysis of the Peacock index values (Fig. 8) it results that in the basaltic andesites and in the lamprophyres „i" has 53.85 and 54.20 values, which make them be included in the alkali-calcic rock series, while the other petrographic types are characterized by the values i = = 57.05, 57.45 and 60.40, corresponding to the calc-alkaline rock series.

REFERENCES

Arabu N. (1941) La géologie des environs de Băiţa. *D. S. Inst. Geol.* XXV, p. 47– 94, Bucureşti.

Bleahu M. (1957) Cercetări geologice in regiunea Padiş-Cetăţile Ponorului. *D. S. Inst. Geol.*, XLI, p. 35– 36, Bucureşti.

— Dimitrescu R. (1957) Stratigrafia şi tectonica Munţilor Apuseni. *An. Rom. Sov. Geol.*, 2, Bucureşti.

— Mantea G. (1962) Le Rhetien de Monts Apuseni (Carpates Roumains). Com . des Congr. de Mesozoique, Luxemburg.

Bordea S., Bordea J. (1973) Date noi stratigrafice şi structurale in nord-vestul Munţilor Bihor. *D. S. Inst. Geol., Geofiz.*, LIX/5, p. 58– 59, Bucureşti.

— Bleahu M., Bordea J. (1975) Date noi stratigrafice şi structurale asupra Bihorului de vest. *D. S. Inst. Geol., Geofiz.*, LXI/5, p. 61– 83, Bucureşti.

Dimitrescu R. (1959) Kratkii ocerk gheologhii zapadnih gor. Sov. Gheol., Moskva.

— (1969) Le volcanism permien en Roumanie. *Geologische Rundschau*, 48, Stuttgart.

Giuşcă D. (1950) Le masif de la Vlădeasa. *Ann. Com. Géol.*, XXIII, p. 200– 252, Bucureşti.

Gottini V. (1969) Nuovo metodo di calcolo petrochimico per distinguere i magmi anatettici crostali da quelli provenienti dal mantello superiore. *Boll. d. Sed. Accad. Gienia Sci. Nat.*, IV, p. 609– 618, Catania.

— (1970) Serial character of the volcanic rocks of Pantelleria. *Bull. Volc.*, 33, 3, p. 818– 827, Napoli.

Ianovici V., Giuşcă D., Borcoş M., Bleahu M., Patrulius D., Lupu M. Savu H. (1976) Geologia Munţilor Apuseni. Ed. Acad. R.S.R., Bucureşti.

Istrate G., Preda I. (1970) Prezenţa rocilor spilitice în valea Peşterii-Meziad (Munţii Pădurea Craiului). *Stud. cerc. geol., geofiz., geogr. Ser. geol.*, 15, 1, Bucureşti.

— (1975) Formaţiunea riolitelor ignimbritice din partea vestică a masivului Vlădeasa (Munţii Apuseni). *D. S. Inst. Geol., Geofiz.*, LXI/1, p. 99– 142, Bucureşti.

Kräutner Th. (1941) Etudes géologiques dans la Pădurea Craiului. *C. R. S. Inst. Géol.*, XXVI, p. 145– 156, Bucureşti.

— (1941) Observations géologiques dans les Monts du Bihor. *C.R.S. Inst. Géol.*, XXVI p. 99– 105, Bucureşti.

Mancea Al. Z. (1973) Asupra prezenţei unei formaţiuni cristalofiliene in baza pînzei de Arieşeni (Munţii Bihor). *D. S.Inst. Geol.*, LIX/5, p. 13– 22, Bucureşti.

— Covaci S., Samoilă I. (1975) Mineralizaţiile legate de masivul banatitic din Munţii Bihor. *D. S. Inst. Geol.*, LXI/4, p. 77– 92, Bucureşti.

— Serini V. (1980) Contribuţii noi privind conţinutul micropaleontologic al kimmeridgian-tithonicului din Autohtonul de Bihor (sector Chişcău-Juleşti). *An. Univ. Buc.*, XXIX, p. 86– 98, Bucureşti.

Posepny F. (1874) Geologische Ungr. Sitzgeber. d.k.k. Acad. Ungr., XLIII, p. 40, Wienn.

Primics G. (1892) Skizzenheschtor Beritt über die in nördlichen Theille des Bihargebirges in Jahre 1890. *Jahreb. d.k. Ungr.*, J., p. 44– 62, Budapest.

Rafalet A. (1963) Asupra rocilor din contact din regiunea Pietroasa-Budureasa (Munţii Bihor). *Asoc. Geol. Carp. — Balc. Congr.* V, II, p. 199– 205, Bucureşti.

CONTRIBUTIONS TO THE PETROLOGY OF OPHIOLITIC PERIDOTITES AND RELATED ROCKS OF THE MEHEDINȚI MTS (SOUTH CARPATHIANS)[1]

BY

MARCEL MĂRUNȚIU[2]

Introduction

Ultramafic rocks, mainly serpentinites, have been recognized as characteristic components of the Obîrșia Complex from the Mehedinți Mts represented by clay shales, siltstones, micaceous and carbonatic sandstones, limestones, siliceous depositions, volcanic tuffs and basalts. C o d a r c e a (1940) considered these rocks to be Upper Jurassic-Neocomian in age, the basin in which they are formed being located between the Getic (westward) and the Danubian Domain (eastward). At the present time, many investigators have proposed that the Obîrșia Complex is in a Parautochthonous position (Severin Nappe), supporting the Getic Nappe.

Recently in the Obîrșia Complex have been recognized some sedimentary rocks similar to the Upper Cretaceous rocks of the Wildflysch Formation from the Danubian Domain. In this manner, S t ă n o i u (1980) has suggested that this association represents an olistostromal mixture of blocks of various rock types in an Upper Cretaceous matrix, located at the middle part of the Upper Cretaceous Wildflysch Formation.

R ă d u l e s c u and S ă n d u l e s c u (1973), impressed by the lithological association, have proposed that the ophiolitic rocks represent the ocean lithosphere fragments. This assumption was comfirmed by chemistry of the mafic rocks (M ă r u n ț i u et al., 1977 , C i o f l i c a et al., 1981).

The purpose of this article is to discuss the history of the ultramafic rocks from the Mehedinți Mts, in the light of some textural characters.

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Field Relations

The ultramafic rocks occur as bodies having a wide range in size, incorporated by sedimentary rocks. Usually, the contacts between peridotite and country rocks (sedimentary or igneous mafic ones) are subhorizontal and a lot of the large bodies are like a sheet in shape, (e.g. Cracul Muntelui, Ciolanul, Margina). These contacts are always tectonic, being accompanied by mylonitization and shearing of the serpentinitic rocks. The cleavages of the serpentinites are corresponding in strikes to the schistosity of the adjacent sedimentary rocks, and could be related to the moment of the cold intrusion of the serpentinites at the tectonic level of sedimentary rocks, being a product of shallow deformation without neomineralization. In some outcrops, the second shear-planes can be observed cross-cutting the first ones and corresponding to the axial surface of the microfolds in schistosity of the sedimentary rocks.

A lot of lenses consisting of basalts, mafic tuffs, sedimentary of metamorphic rocks, are found inside the ultramafic bodies.

Ophiolitic Peridotites

Lithology

The allochthonous ophiolite bodies consist of an alpine-type peridotite of the harzburgite subtype of N i c o l a s and J a c k s o n (1972), dominated by a harzburgite-dunite association, which is associated with scattered seams of pyroxenites. Serpentinization is prevalent, usually approaching 80 to 100 percent, but the original mineralogical composition can be recognized.

Harzburgite, the most abundant rock type, has a rather uniform modal composition (68—85 percent olivine, 10—30 percent orthopyroxene, 0.1—2 percent clinopyroxene, 0.5 percent chrome-spinel). In addition, the harzburgite has also a limited variation in bulk rock chemistry. The 100 Mg/Mg + Fe ratio in bulk rock analyses ranges from 82.15 to 85.35, according to the chemical data quoted by C o l e m a n (1977) for typical harzburgites from ophiolite metamorphic peridotites. The modal compositional variations are due to irregular olivine/pyroxene ratio throughout the bodies. A compositional layering, representing alternating bands of pyroxene-rich and pyroxene-poor harzburgite 1 to 5 cm thick, occur sporadically. The foliation marked by the parallelism of orthopyroxene grains and elongated and flattened spinel grains, is more characteristic for harzburgite. The relation between layering and foliation is not clear. In all appearances the foliation cross-cuts the layering at an angle of 10—15°.

The dunite occurs as irregular patches among the harzburgite. This rock has usually a massive texture and consists of the primary minerals olivine, accessory chrome-spinel and sometimes trace of pyroxene. Because of the quite monomineralic nature of the dunite, the definite field observable foliation results from local slight planar concentration of the chrome-spinel grains. Associated with dunite are lenticular iron oxides deposits with cumulus texture.

The pyroxenites are represented by olivine-orthopyroxenite and clinopyroxenite. These are commonly found with a boudinage structure among the harzburgite.

Petrography

Olivine. Olivine from harzburgite is slightly more iron-rich than the olivines reported for other tectonic harzburgites (Loney et al., 1971; Coleman, 1977; Himmelberg and Loney, 1980). The optical study reveals a restricted range in the forsterite content from 85 to 88 percent.

Although the serpentinization is a widespread process, the morphologies and textural relations of the olivine can partially be seen. The olivine shape is characteristic of plastic deformation, and some different textural types were identified. Commonly, the olivine in harzburgite is anhedral with mutually interfering grain boundaries (xenoblastic granular texture — Himmelberg and Loney, 1980, protogranular texture — Mercier and Nicolas, 1975). In foliated harzburgite it is quite possible to distinguish a kink-banded coarse grained olivine (3—4 mm in size) and a fine recrystallized granular one (much less than 0.5 mm in size) with straight grain boundaries meeting in triple points (porphyroclastic texture — Mercier and Nicolas, 1975). Sometimes elongated olivine grains parallel to compositional layering or foliation were also remarked.

The olivine microfabric of the harzburgite is characterized by the presence of X = [010] maxima oriented approximately normal to the foliation plane. The generally interlocking texture and equant grain shape of olivine with quoted fabric is the result of the syntectonic recrystallization under high-temperature conditions (Avé Lallemant and Carter, 1970; Loney et al., 1971; Nicolas et al., 1973, 1980).

The (100) kink bands of olivine porphyroclasts with related {okl} [100] glide system are also likely to develop at high-temperature.

Usually, the result of the olivine serpentinization process is a lizardite- chrysotile assemblage with mesh structure. The banded structure with elongated olivine relics is a noticeable pattern occurring likely by serpentinization of the olivine porphyroclasts. It appears that this structural type of serpentinization is the result of the olivine internal discontinuity related to the glide planes inferred from kink bands.

Orthopyroxene. Orthopyroxene occurs in harzburgite as grains with 3 to 10—15 mm in size, more distinct in weathered surfaces. The enstatite content of the orthopyroxene is quite variable depending on the structure and mineralogical composition of the rocks. In zones with visible layering and high-modal amounts of clinopyroxene, the enstatite content of orthopyroxene ranges from 85 to 87 percent. In foliated and massive harzburgites with no or very poor clinopyroxene, the orthopyroxene is slightly more magnesium-rich (En_{90-92}). The clinopyroxene exolution lamellae are rare and significant for deformed orthopyroxene from banded harzburgite.

The orthopyroxene grains have different morphologies. Frequently there occur anhedral, lobate grains with interlocking grain boundaries in the olivine matrix, but usually there are elongate and flattened orthopyroxene prismatic grains with the "c" axis marking the foliation plane. Foliated and banded harzburgite showed a good example of large and elongated strained orthopyroxene grains (3—5 mm in size) (porphyroclasts) and small polygonal strain-free ones of about 0.5 mm. The same porphyroclastic texture is met in the olivine-orthopyroxenite with the strongest development of typical textural characters.

C l i n o p y r o x e n e . Clinopyroxene occurs as an accessory mineral in foliated harzburgite. It is characteristically smaller (about 0.02 mm) than the coexisting olivine and orthopyroxene. Clinopyroxene exhibits anhedral cuspate, embayed grains, interstitial or intergranular to other minerals. These small grains tend to cluster and the clinopyroxene pockets are parallel to the foliation. Adjacent to orthopyroxene grains or poikilitically enclosed by this, the clinopyroxene is commonly abundant.

In the banded harzburgite, clinopyroxene may be as small anhedral strain-free grains, usually clustered in the olivine bands and as large (2—3 mm in size) anhedral strained grains which occur in the pyroxenitic bands and contain thin exolution lamellae.

Two types of the clinopyroxene textural relations were met in the clinopyroxenites. In some occurrences there are strong-stained anhedral clinopyroxene grains with small $2V_\gamma$ (about 35°) and abundant spinel exolutions. In addition, along the kink bands, this clinopyroxene recrystallized as polygonal strain-free grains with large $2V_\gamma$ (about 55°) and no spinel exolutions. In other clinopyroxenites, the clinopyroxene has no signs of deformation and shows smooth curved grain boundaries in an equigranular texture.

S p i n e l . Chrome-spinel in foliated harzburgite from the Mehedinți Mts, is dark red-brown to brown in thin section like the spinel quoted for the mantle peridotite (N i c o l a s et al., 1980). It is characterized by anhedral grains and occurs in some textural relations. Large anhedral, partially vermicular grains (2—3 mm in size) are associated with orthopyroxene pockets, sometimes enclosed by the latter.

Small anhedral grains occur in elongate clusters among the small clinopyroxene grains and flattened in the plane of foliation. Spinel grains within the olivine groundmass were also observed, displaying a textural pattern similar to holly-leaf shape and parallel to the orthopyroxene foliation. It is a proof of the increasing strain penecontemporaneous with the development of the porphyroclastic texture.

Comments on the textures

Textures of the peridotites point out a complex evolution of these rocks, in the subcrustal environment. Features of the textures are those that generally originate differently during deep-seated plastic deformation and recrystallization. The different textures are present in varying degrees throughout the tectonic harzburgite, one obscuring another.

A xenoblastic granular texture of the olivine and orthopyroxene as anhedral lobate grains with interlocking boundaries has been interpreted by D i c k (1977) to be a result of the dissolution-reprecipitation creep "if melt was present or forming during deformation". Some clinopyroxene and spinel data presented above can be used to point out that partial melting occurred. The clusters of the clinopyroxene and vermicular spinel grains adjacent to orthopyroxene, suggest the incongruent melting of the orthopyroxene. The X = [010] olivine microfabric is in agreement with a hypersolidus temperature for the plastic flow with large to not appreciable strain (N i c o l a s et al., 1980), accompanied by melting of the fusible components and extraction of melt, leaving a depleted residue. In this manner, the scarcity of clinopyroxene in this harzburgite is explained by its removal during partial melting. The small interstitial, intergranular grains and clusters of the clinopyroxene and spinel may be a result of the magmatic crystallization of the trapped melt (M e n z i e s and A l l e n, 1974). I suggest that the banded (layered) harzburgite occurring as isolated outcrops throughout the massive or foliated harzburgite, represent a relatively undepleted peridotite containing large anhedral strained clinopyroxene grains with exolution lamellae ($2V_\gamma = 48-50°$) and magmatic pockets of the small clinopyroxene grains ($2V_\gamma = 55-60°$).

The occurrence of the clinopyroxenite must be ascribed to crystallization from the mafic melt resulted by partial melting. Because two clinopyroxenite structural-groups exist, I am stressed to assert that they represent different periods of emplacement with slightly different composition (two periods of partial fusion) separated by an intense deformation. Quite similar a situation is reported by E l a i n e H a r k i n s et al. (1980) for the magmatic dunite bodies from the Vorinous Ophiolite which occur as multiple intrusive events both before and after the deformation.

The porphyroclastic texture has been interpreted as a consequence of the plastic flow, the operating mechanism being intracrystalline gliding (M e r c i e r and N i c o l a s, 1975). In our situation, this process has been operated at temperatures which emerge from the development conditions of the olivine kink-bands. Experimentally, the {okl} [100] glide system was estimated to develop at temperatures from 900 to 1.200° (A v é L a l l e m a n t et al., 1968). These temperatures must be lower than those which occurred during partial melting, because the kink-banding and recrystallization of the clinopyroxene from the strained clinopyroxenite group, is accompanied by the exolution of spinel. According to V a r n e (1977) this textural relation of spinel may be the result of the sub-solidus reaction and re-equilibration in response to falling temperature.

In the layered harzburgite, mylonitic texture is well represented along shear bands parallel to the layering and appears to be subsequent to the porphyroclastic texture. It is characterized by small grains (0.01 – 0.05 mm in size) of olivine and pyroxene, completely recrystallized. This type of texture is reported for the lower contact of the peridotite sequence of ophiolites as the result of the dynamic metamorphism during overthrusting of peridotite on the oceanic side of an active subduction zone (P a r r o t and W h i t e c h u r c h, 1978; N i c o l a s et al., 1980).

An important point is that of the dunite genesis. Their texture was obscured by serpentinization and relations with harzburgite are unclear. But, the occurrence of the cumulate iron oxides (magnetite euhedral grains with inclusion zonation patterns) and interlayered sulfide-bearing clinopyroxenite and dunite strongly militates against the residual origin of the dunite, and favors a magmatic one. This evidence suggests that the dunite is an initial cumulus deposit infolded into the residual harzburgite, similar to the mechanism invoked by George (1978) to explain the dunite-harzburgite relation for the Troodos Ophiolite.

Related Rocks

Extrusive rocks

The extrusives are represented by a sequence of submarine flows of the mafic rocks with distinct pillow-structure among which tuffaceous and sedimentary rocks are found. Subophitic and intersertal textures are common, the sparsely porphyritic ones bearing phenocrysts of subcalcic clinopyroxene or plagioclase. A lot of secondary minerals (chlorite, epidote, zoisite, uralitic amphibole, calcite) are found in the mafic rocks. The field relations between extrusives and ultramafics are always tectonic. Throughout the ultramafic bodies consisting of massive or sheared serpentine, there are isolated masses of basalt. These brecciated inclusions have undergone a low-grade metamorphism responsible for the growth of some new minerals such as hydrogarnet, diopside, prehnite, zoisite, chlorite. This rodingite metasomatism (Coleman, 1967, 1977) has preserved the original texture of mafic rocks.

Metamorphic rocks

Metamorphic rocks occur as tectonic inclusions in serpentinized ultramafic rocks. They are very brecciated or mylonitized and consist of plagiogneisses, micaschists, amphibolites, the parageneses being as follows:

— quartz-plagioclase-biotite (chlorite)-hematite;
— quartz-biotite (chlorite)-muscovite-apatite;
— hornblende-plagioclase An_{30}-apatite.

The intense deformations of these rocks are accompanied by a substantial change in the mineralogical composition. Near the contact with the serpentinite, the plagioclase is completely replaced by prehnite and hydrogarnet; the hornblende is replaced by diopside. In addition, variable amounts of chlorite and zoisite occur. Numerous monomineralic veins of prehnite cut the metamorphic rocks. The occurrence of this low-temperature metasomatism, closely related to serpentinization, proves that the metamorphic rocks have been tectonically incorporated from the crust, by a rising ultramafic mass undergoing a contemporaneous serpentinization.

Conclusions

The restricted mineralogy and textures of the harzburgite give evidence of the residual character of this, and of its complex tectonic and magmatic mantle history. In fact, the peridotite from the Mehedinţi Mts, is similar to the residual tectonic peridotite described at the lower part of the ophiolite massifs (Coleman, 1971, 1977; Menzies and

Allen, 1974; Sinton, 1977; George, 1978) on which cumulate sequences of olivine ± clinopyroxene ultramafic rocks, more or less deformed, were deposited.

After the deep-seated evolution, the ultramafic rocks were incorporated at the shallow crustal level. The upward diapiric movement, in favourable tectonic conditions was facilitated by the decrease of the density connected with the transformation of the peridotite to serpentinite. In this manner, any ultramafic bodies can move, if it has critical size depending on the degree of serpentinization, density of the country rocks and the cohesion forces between serpentinite and crossed rocks (Şeclăman and Mărunţiu, 1981). The upward plastic movement also determined the inclusion of various crustal blocks by the serpentinitic mass and must be the most important reason for the dismembering of the original ophiolitic sequence. The successive deformations during the serpentinite movement are well illustrated by the local recrystallization of the serpentine minerals, the mylonitization of the inclusions and finally by the different shear-stages.

The serpentinite-bearing Obîrşia Complex is similar to the ophiolitic mélange formation revived by Gansser (1974). Some quoted features of the ophiolitic mélange were recognized for the occurrence from the Mehedinţi Mts (Mărunţiu et al., 1977):

— mixture of ophiolitic rocks (serpentinite and basalt) together with non-ophiolitic rocks, especially metamorphic ones;

— the matrix is tectonically sheared serpentinites and sedimentary rocks;

— there are tectonic relations between components, locally accompanied by metasomatic reaction zones;

— the mixture was tectonically carried out with the probably quite important participation of an olistostromal sedimentation.

The origin of ophiolitic rocks from the Mehedinţi Mts, was in an Upper Jurassic-Lower Cretaceous spreading zone, more probably situated between the Danubian Domain and the Moesian Platform. The closing of this trough and the westward subduction of the oceanic crust during the Upper Cretaceous led to the formation of the ophiolitic "mélange" and its thrusting over the Danubian Domain.

REFERENCES

Avé Lallemant H. G., Schiffmann P. M., Carter N. L. (1968) Ductile flow of peridotite and quartzite. *Trans. Am. geophys. Un.* 49, 313–14 (abstract).

— Carter N. L. (1970) Syntectonic recrystallization of olivine and modes of flow in the upper mantle. *Geol. Soc. America Bull.*, 81, p. 2203–2220.

Cioflica G., Vlad Ş., Nicolae I., Vlad Catrinel, Bratosin Irina (in press) (1981) Copper metallogenesis related to mesozoic ophiolites from Romania.

Codarcea Al. (1940) Vues nouvelles sur la tectonique du Banat méridional et du Plateau de Mehedinţi. *An. Inst. Géol. Roum.*, XX, p. 1–74.

Coleman R. G. (1967) Low-temperature reaction zones and alpine ultramafic rocks of California, Oregon and Washington. *U. S. Geol. Surv. Bull.*, 1247, p. 1—49.

— (1971) Plate tectonic emplacement of upper mantle peridotites along continental edges. *J. Geophys. Res.*, 76, p. 1212—1222.

— (1977) Ophiolites, ancient oceanic lithosphere? Springer-Verlag, 229 p., Berlin.

Dick H. J. B. (1977) Partial melting in the Josephine Peridotite I, The effect on mineral composition and its consequence for geobarometry and geothermometry. *Am. Jour. Sci.*, 277, p. 801—832.

Gansser A. (1974) The ophiolitic mèlange, a world-wide problem on Tethyan examples. *Eclogae Geol. Helv.*, 67, 3, p. 479—507.

George R. P. Jr. (1978) Structural petrology of the Olympus ultramafic complex in the Troods ophiolite, Cyprus. *Geol. Soc. America Bull.*, 89, p. 845—865.

Harkins Elaine M., Green H. W., Moores E. H. (1980) Multiple intrusive events documentated from the Vourinos Ophiolite Complex, Northern Greece. *Am. J. Sci.*, 280 A, p. 284—295.

Himmelberg G. R., Loney R. A. (1980) Petrology of ultramafic and gabbroic rocks of the Canyon Mountain Ophiolite, Oregon. *Am. J. Sci.*, 280 A, p. 232—268.

Loney R. A., Himmelberg G. R., Coleman R. G. (1971) Structure and petrology of the alpine-type peridotite at Burro Mountain, California. *U.S.A. Jour. Petrologyy* 12, p. 245—309.

Mărunțiu M. et al. (1977) Unpublished data, the archives of the Institute of Geology and Geophysics. Bucharest.

Menzies M., Allen C. (1974) Plagioclase lherzolite-residual mantle relationships within two eastern Mediterranean ophiolites. *Contr. Mineralogy Petrology*, 45, p. 197—213.

Mercier J. C. C., Nicolas A. (1975) Textures and fabrics of upper mantle peridotites as illustrated by xenoliths from basalts. *J. Petrology*, 16, p. 454—487.

Nicolas A., Jackson E. D. (1972) Répartition en deux provinces des péridotites des chaines alpines longeant la Méditerranée : implications géotectoniques. *Bull. Suis. Min. Petr.*, 52, p. 479—495.

— Boudier E., Boullier A. M. (1973) Mechanisms of flow in naturally and experimentally deformed peridotites. *Am. J. Sci.*, 273, p. 853—876.

— Boudier F., Bouchez J. L. (1980) Interpretation of peridotite structures from ophiolitic and oceanic environments. *Am. J. Sci.*, 280 A, p. 192—210.

Rădulescu D., Săndulescu M. (1973) The plate tectonic concept and the geological structure of the Carpathians. *Tectonophis.* 16.

Sinton J. M. (1977) Equilibration history of the basal alpine-type peridotite, Red Mountain, New Zealand. *Jour. Petrology*, 18, p. 216—246.

Stănoiu I. (1982) Orizontarea formațiunii neocretacice de tip olistostromă din partea nord-vestică a Podișului Mehedinți. *D. S. Inst. geol. geofiz.*, LXVII/5, p. 155—168, București.

Șeclăman M., Mărunțiu M. (1981) Emplacement mechanism of some serpentinic rocks the South Carphatians (in press).

THE NEW STRUCTURAL ASPECT OF BANATITIC BODIES IN THE BIHOR MOUNTAINS[1]

BY

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Geological investigation carried out in the Bihor Mountains (geological mapping, structural drilling, main tunnels) showed that at the depth of 1000–1500 m all intrusive, banatitic massifs join in a unique batholitic body; the area of this batholitic body is over 500 km^2. The orientation of the batholite is approximately north-south; its dimensions on the two axes are 45 and respectively 4–20 km (see the attached (Fig. 1) structural sketch). The actual relief suggests a distorted aspect of the batholite in which erosion opened up a few large stocks here and there, but most of it remained concealed. The shape of the batholite is very complicated, having many protuberances, apophyses and domes and slopes occasionally sliding gently (especially in the central Băița-Luncșoara area, bearing many structural drillings) and having mostly abrupt marginal walls. The most important differences between the structural aspects of the central part of the batholite and its limits are preserved, regarding its petrographic composition: its central area is mainly granitic, its northern area is predominantly of granodioritic character with quartz-diorite and diorite segregations (Budureasa, Seacă) Valley and its southern area contains granodiorite-granite with subordinated diorites and even salic gabbros (Luncșoara, Găina Mountain respectively).

The hypabyssal phenomena associated with the banatitic intrusion are well represented in the area by a whole series of vein-rocks of various petrographic composition; for the purpose of a more systematic display these rocks were divided in 4 main groups. Several diagrams after Niggli, Harker, Rittmann, Peacock and Streckeisen were drawn, pointing out the main petrochemical characteristics of the Bihor magmas.

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H a r k e r's diagram shows the variation of the main oxides in accordance with the SiO_2 quantity. The diagrams presenting the tendency of differentiation of those magmas which generated the Laramian batholite

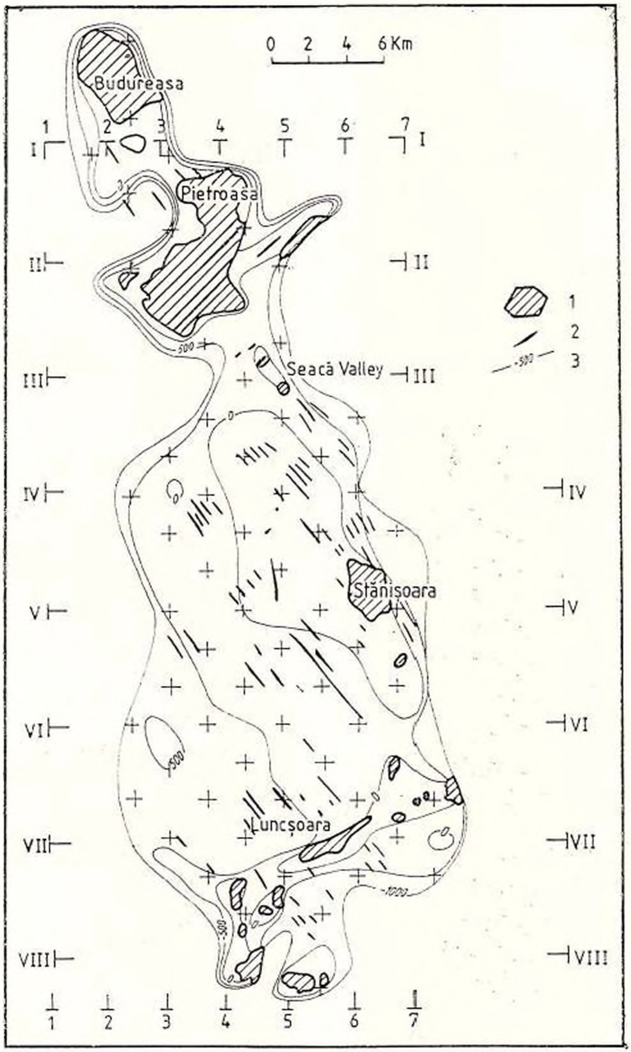

Fig. 1. — Structural sketch of the banatitic batholite in the Bihor Mountains.
1, intrusive bodies outcrops; 2, dyke outcrops; 3, isobath.

and the associated veins in the Bihor Mountains (Figs 2, 3) generally indicate the normal aspect of a subalkaline differentiation : the variation diagrams are slightly sloped and parallel two by two. They overlap at certain Si values which are characteristic for the whole banatitic region in our country, pointing out their consanguinity (G i u ş c ă et al., 1966). But the magmatic differentiation for the plutonic rocks ranges from Si = = 160 to Si = 480, that is from the quartz-dioritic magmas towards

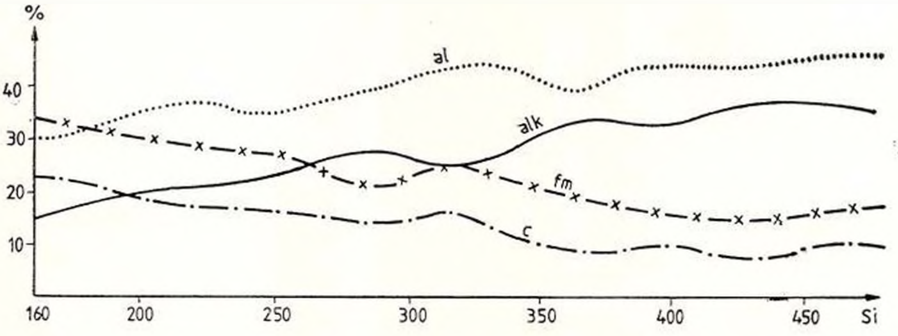

Fig. 2. — Niggli diagram of banatitic magmas, differentiation in the Bihor Mountains (batholite)

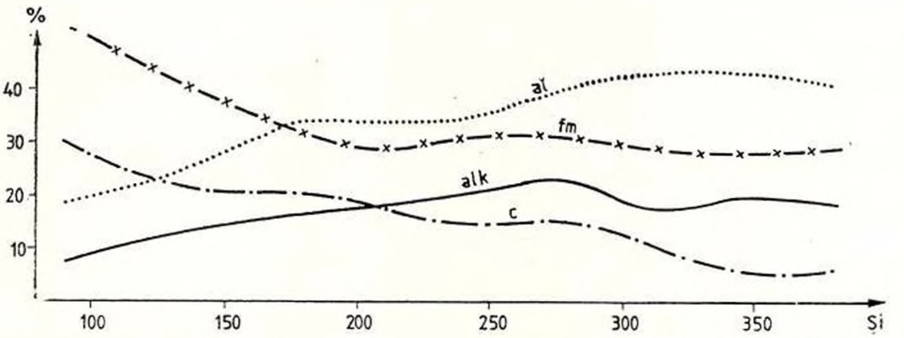

Fig. 3. — Niggli diagram of banatitic magmas, differentiation in the Bihor Mountains (dykes).

leucogranitic magmas, while the variation rank for the hypabyssal rocks ranges between Si 95—390.

mg—k diagram also points out the subalkaline character of the plutonic rocks (almost entirely listed between the diagonals 0.5—0.9), at the same time revealing the alkaline tendencies of most hypabyssal rocks. Diagram 4 (*al-fm* diagram) presents the isofalic character of the quartz-dioritic magmas and the enlistment of the granodioritic, granitic and leucogranitic magmas in the isofalic-salic field. Unlike plutonic rocks, most of the hypabyssal rocks present an isofalic-half-femic character. The Rittmann diagram $K_2O + Na_2O/SiO_2$ (Fig. 5) illustrates that plutonic rocks belong to the magmas of the calc-alkaline series and some vein-rocks have an affinity towards alkaline magmas. *al-alk* diagram (Fig. 6) manages to convey a certain tendency of the granitic and leucogranitic rocks from central Bihor, towards the alkaline field. The calc-alkali index value is 58 and it fits into the established limits for the whole banatitic region of 55—60. These limits were established in 1966 by G i u ş c ă and others. Diagram *QLM* (Fig. 7) illustrates that most plutonic rocks concentrate the saturated rocks field, above the PF line : only two rocks (a Seacă Valley diorite and a Găina Mountain salic gabbro) range below the MF line. The hypabyssal rocks though, are included in the PFM triangle in the field of

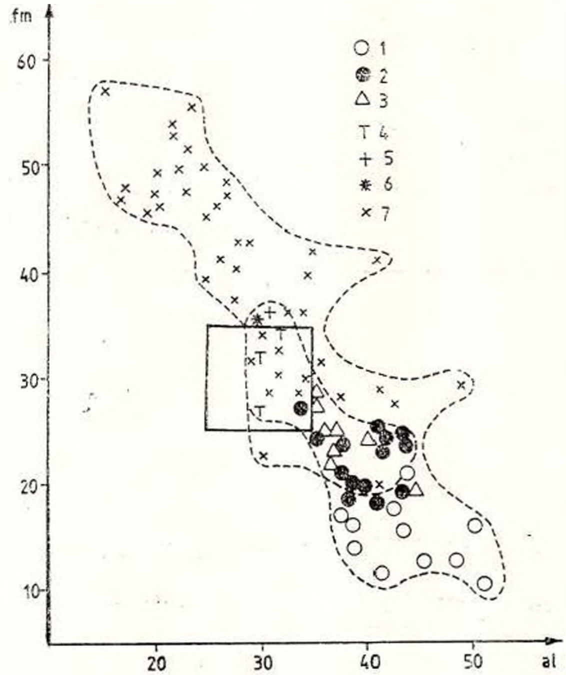

Fig. 4. — *al-fm* diagram of the Bihor banatites.
1, leucogranites; 2, granites: 3, granodiorites; 4, quartz-diorites; 5, diorites; 6, leucogabbroids; 7, hypabyssal rocks.

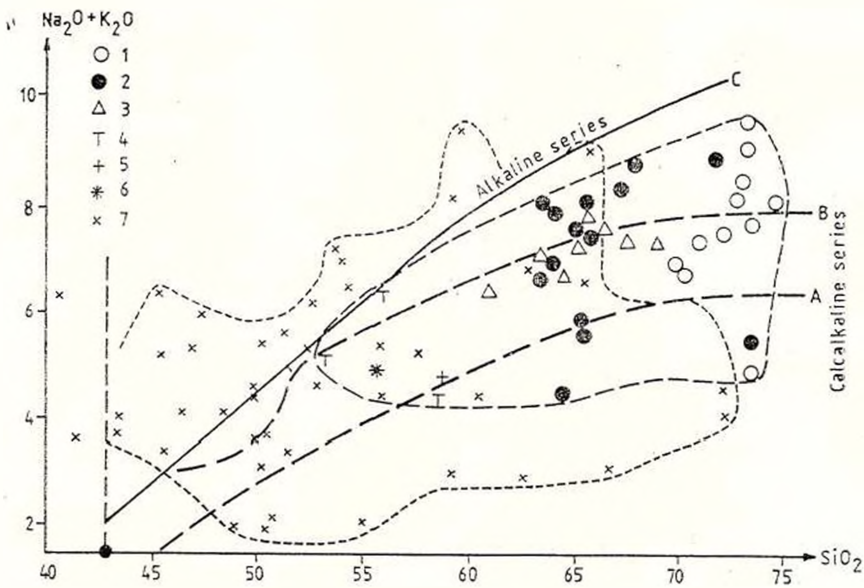

Fig. 5. — $Na_2O + K_2O/SiO_2$ diagram. Rittman serial index.
1, leucogranites; 2, granites; 3, granodiorites; 4, quartz-diorites; 5, diorites; 6, leucogabbroids; 7, hypabyssal rocks.

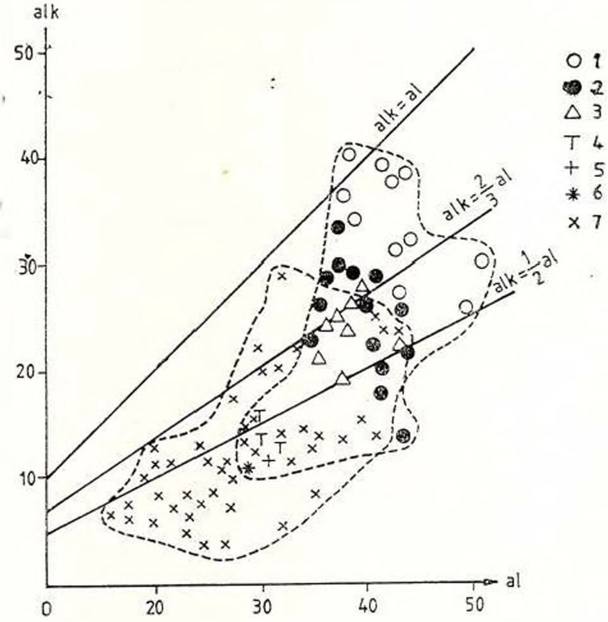

Fig. 6. -- *al-alk* diagram of the Bihor banatites.
1, leucogranites; 2, granites; 3, granodiorites; 4, quartz-diorites; 5, diorites; 6, leucogabbroids; 7, hypabyssal rocks.

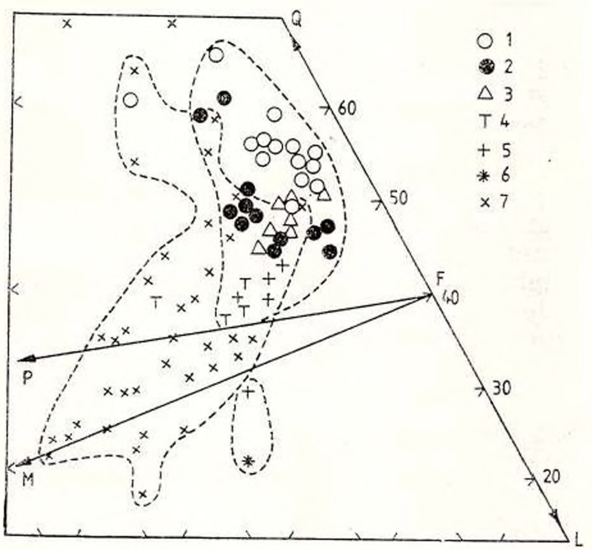

Fig. 7. — *QLM* diagram (Niggli) of the Bihor banatites. 1, leucogranites; 2, granites; 3, granodiorites; 4, quartz-diorites; 5, diorites; 6, leucogabbroids; 7, hypabyssal rocks.

saturated rocks, as well as above and below this triangle, indicating as such their petrographic variety. The calculation of the CIPW norm located the Bihor Laramian rocks in the *QAP* triangle (Streckeisen diagram, Fig. 8) for the purpose of descriptive mineralogical classification. One can notice that plutonic rocks preferentially enlist rather towards the granitic rocks (precisely monzogranitic) then towards monzodioritic and monzo-

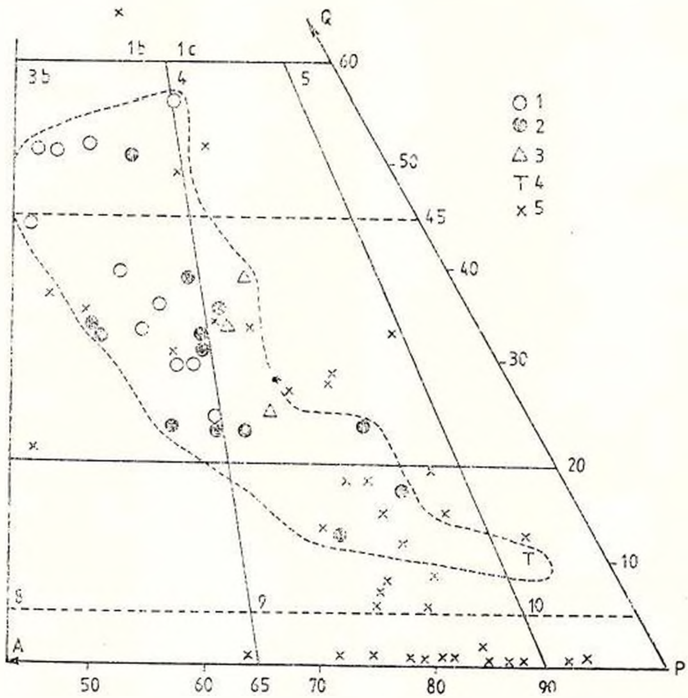

Fig. 8. — Position of the Bihor banatitic rocks on the *QAP* diagram (Streckeisen). 1, leucogranites; 2, granites; 3, granodiorites; 4, quartz-diorites; 5, hypabyssal rocks.

gabbroic rocks. On the contrary, hypabyssal rocks are mostly monzodioritic and monzogabbroic; the granodioritic and granitic varieties are not so significant. *Or-Ab-An* diagram (Fig. 9) also proves the acid character of plutonic rocks; proof is brought by the great quantity of albite accompanied by orthoclase and by a less significant quantity of anorthite. As a distinctive feature vein rocks present an increase of the anorthite quantity and an equivalent decrease of the acid plagioclase and especially the potash-feldspar quantity.

The consanguinity of intrusive and hypabyssal rocks and mainly the granitic feature of the batholite being well established, we have to discuss the problem of the petrographic variety among the limiting zones: the northern and southern limits as well as the dome area of the concealed batholite of central Bihor present intermediary and even basic rocks. Beside the two ways natural differentiation of some magmas of granitic composition (towards paracidic and intermediary terms) established in 1966 by Giușcă, one can say that the presence of the rocks with a more basic character in the granitic batholite could be due to recurrent intrusions of magmas with varied characteristics; these magmas originate in various chambers located at different depths of the same magmatic basin. Among the magma recurrent intrusions one could point out two significant,

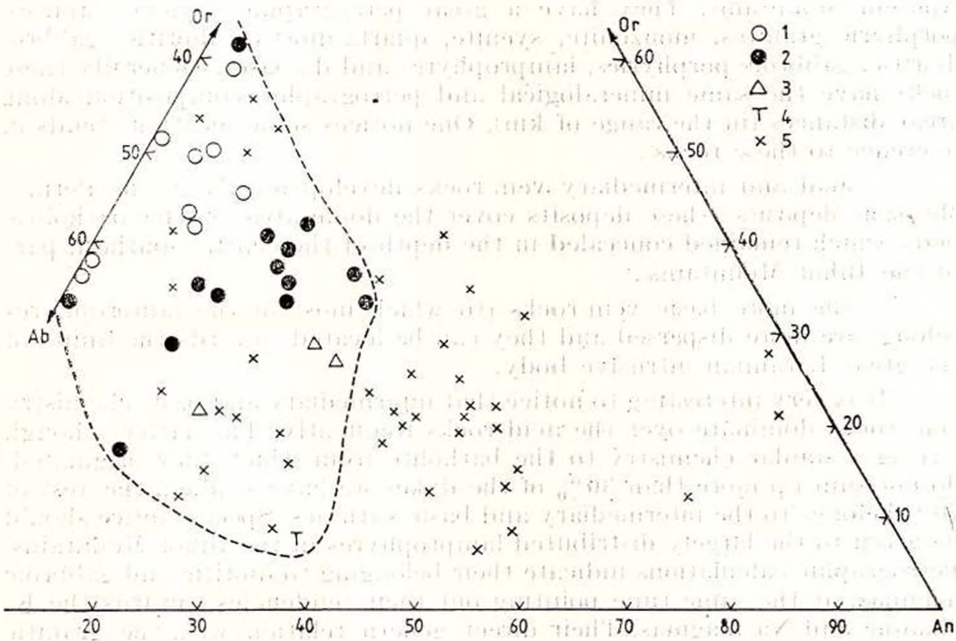

Fig. 9. — *Or-Ab-An* diagram (CIPW norm) of the Bihor banatites. 1, leucogranites; 2, granites; 3, granodiorites; 4, quartz-diorites; 5, hypabyssal rocks.

time — spaced stages; the primary stage when the granodiorites, the quartz-diorites and the diorites originated from the granodioritic magma (this stage is more pregnant in the Budureasa, Găina Mountain and Pietroasa regions); the secondary stage, which is more significant in time, when the granites and leucogranites making up most of the batholitic body originated from the recurrences of granitic magmas. More or less, the secondary stage rocks replaced or partially enclosed the primary stage rocks; the latter can be often found in the granites body as enclaves. Many researchers pointed out this ascertainment.

The influence of the enclosing rocks on the chemistry of the ascending Laramian magmas is only of local interest and not very significant, both for the Permian cherty-argillaceous and the Mesozoic limestone deposits. But one can notice though that the batholite had a significant influence on the enclosing rocks; this influence is indicated by an important thermal contact halo. Two stages can also be distinguished in the emplacement of vein-rocks. The primary stage took place shortly after the consolidation of the batholite and it is connected to the operation of some magmatic basins placed at great depths; aplites and the vein-granites originated during this stage. The secondary stage is held responsible for the origin of intermediary and basic dykes: it took place late in the magmatic activity.

The banatitic vein rocks of the Bihor Mountains are oriented predominantly towards NW-SE, in terms of the Laramian fault system of the

Apuseni Mountains. They have a great petrographic variety: aplites, porphyric granites, monzonite, syenite, quartz-dioritic, dioritic, gabbro-dioritic, gabbroic porphyries, lamprophyres and diabases. Generally these rocks have the same mineralogical and petrographic composition along great distances (in the range of km). One notices some location trends in reference to these rocks:

— acid and intermediary vein-rocks develop mostly in the Permo-Mesozoic deposits; these deposits cover the dome area of the batholitic body which remained concealed in the depth of the central-southern part of the Bihor Mountains;

— the more basic vein-rocks (to which most of the lamprophyres belong) are more dispersed and they can be located towards the limits of the great Laramian intrusive body.

It is very interesting to notice that intermediary and basic chemistry vein rocks dominate over the acid rocks frequently. The latter, though having a similar chemistry to the batholite from which they originated, do not sum up more than 30% of the dykes we have studied, the rest of 70% belongs to the intermediary and basic varieties. Special notice should be given to the largely distributed lamprophyres in the Bihor Mountains. Petrographic calculations indicate their belonging to dioritic and gabbroic magmas, at the same time pointing out their tendencies towards the K, alkaline and Na magmas. Their direct genetic relation with the granitic batholite is very obvious: they represent the last differentiated products of the granitic magma; they formed by injecting towards some residual differentiated products deeply located in the magmatic basin. Around the banatitic veins in the Bihor Mountains there are many post-magmatic changes such as the greisen and skarn processes, or various hydrothermal changes which affected both the so-called magmatic rock and the enclosing rocks. Greisen and skarn processes are present especially in the dykes with a more acid composition, at their contact point with the carbonatic rocks (limestones and Mesozoic dolomites) in the region of the Crişul Negru Springs. Here there took place bimetasomatic processes which formed skarn type deposits. Subsequent circulation of hydrothermal solution on the same contact areas formed partially overlapping hydrothermal mineralizations. In those parts where the same dykes intruded the Permian rocks or the crystalline schists, the hydrothermal phenomena are either absent or not very pregnant. Dykes of intermediary and basic composition act differently; they indicate hydrothermal changes of various intensities, even in the Permian rocks which are intruded by them. At the contact area of 12 dykes in Bihor (8 lamprophyres and 4 diabases) one can notice the following types of hydrothermal metasomatosis: Na (forming albite, paragonite, riebeckite); K (forming sericite and biotite and rarely adularia and orthoclase); carbonatic (forming carbonates and epidote and rarely apatite and sphene); quartzoze-silification; ferric (forming ferric oxides and ferric sulphide, chlorite, argillaceous minerals and siderite); B — tourmaline enrichments; P — resulting apatite and S (with depositions of metallic sulphides). Mostly the different types of hydrothermal metasomatosis overlap — a single type occurring quite rarely.

REFERENCES

Abdulaev H. M. (1957) Daiki i orudenenei. Ed. Kartfabrika Gosgeoltehnizdaka, Moskva.

Burri K. (1959) Petrochemicals calculations. Basel-Stuttgart.

Giuşcă D., Cioflica G., Savu H. (1966) Caracterizarea petrografică a provinciei banatitice. *An. Com. Geol.* XXXV, p. 13–40, Bucureşti.

Goga I. (1973) Teză de doctorat, Univ. Bucureşti.

Moorhouse W. W. (1959) The study of rocks in thin section. New York

Niţu Gh. (1977) Teză de doctorat, Univ. Bucureşti.

Raguin E. (1975) Les roches plutoniques dans leur rapports avec les gites mineraux. Paris.

— (1976) Géologie du granite. Paris.

Stoicovici E., Selegean I. (1970) Contribuţii la cunoaşterea magmatismului din Munţii Bihorului. *Studia Univ. Babeş-Bolyai, seria geol.-mineral.*, p. 2–15, Cluj.

IRON MINERALIZATION ASSOCIATED WITH THE NEOGENE VOLCANISM IN THE SOUTH-WEST OF THE HARGHITA MOUNTAINS (EAST CARPATHIANS) [1]

BY

SERGIU PELTZ [2], MARGARETA PELTZ [2], IRINA BRATOSIN [2], ROSETTE IANC [3]

Introduction

In the south-west of the Harghita Mts, in the Băile Homorod-Vlăhița-Chirui Neogene andesitic area, the most important iron mineralization — siderite and limonite — on the territory of the Neogene volcanic zone in the Romanian East Carpathians is located. In this region there occurs the Lueta ore deposit, exploited since the 18th century, as well as other accumulations or indications of mineralizations. Up to now the Lueta Mine has represented the only exploitation of the iron ore associated with the Neogene volcanism in Romania. Within the East Carpathian metallogenetic province most of the iron mineralizations occur, however, at the southern and eastern periphery of the Harghita Mts or in the neighbouring zones, in the Ciuc and Baraolt Depressions (Fig. 1). At the present stage of knowledge the presence of the most significant complex and gold mineralizations in the north of the subprovince (Gutîi Mts) and of the most important iron mineralizations in the south of the subprovince (Harghita Mts) is considerable.

In the last years (1975—1980), the authors have studied the territory of the Lueta metallogenetic field and the neighbourring areas, the main aim being the prospect of the iron mineralizations in the extension of the Lueta ore deposit. On this occasion there have been obtained scientific results, which substantiate the geoeconomic prospect as well as the geologic activity for its control. Some of these results will be presented further on.

The geological researches related to the south-west of the Harghita Mts were less numerous before 1900. In the period 1900—1946, the contributions of the researches carried out by Atanasiu (1939), Codarcea and Petrulian (1941), Ghika-Budești (1949) to the

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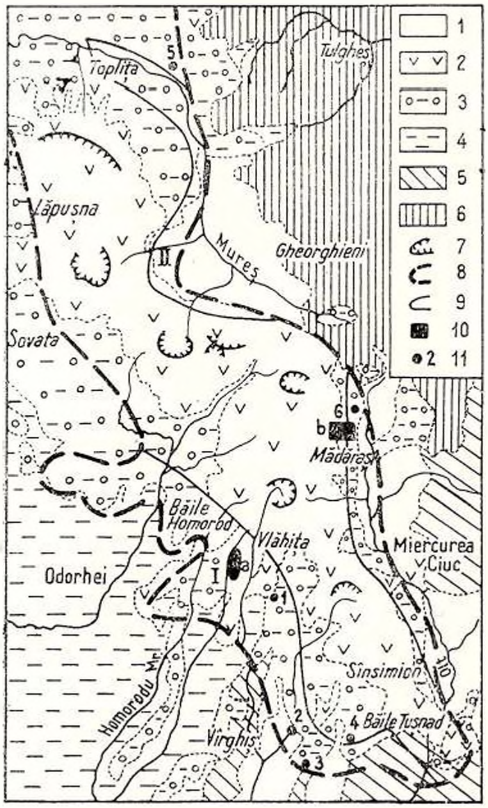

Fig. 1. — Distribution of the iron mineralization in the Corund-Lueta-Herculian and Toplița-Sinsimion metallogenetic districts (geological and metallogenetic data according to the geological and metallogenetic maps of Romania, scale 1 : 1,000,000, completed by the authors)

1, Quaternary deposits; 2. Neogene volcanic formations of the upper compartment; 3, deposits of the volcano-sedimentary formation (lower compartment); 4, Neogene molasse formations; 5, Cretaceous flysch formations; 6, Crystalline schists; 7, crater and caldera; 8, limit of the East Carpathian metallogenetic province; 9, limit of the ferri-ferrous metallogenetic districts: I. Corund-Lueta-Herculian; II. Toplița-Sinsimion; 10, metallogenetic field with siderose and limonite: a, Lueta-Vlăhița; b, Mădăraș; 11, mineralization location at the exterior of the metallogenetic fields: 1, Chirui; 2, Filia; 3, Biborțeni; 4, Herculian; 5, Toplița; 6, Cirța.

knowledge of the Lueta ore deposit and the geological structure of the region are to be mentioned. An intense activity of geologic and gravimetric prospections, drilling exploration and assessment of the geological reservoirs in the Vlăhița region was carried out in the periods 1959—1965 and 1974—1980.

The study region is constituted of volcanic and sedimentary formations. Over 90% of the territory is represented by andesitic volcanics belonging to the lower structural (volcano-sedimentary) compartment of the Harghita Mts. The Miocene and Pannonian (zones C+D and E) molasse deposits of the eastern border of the Transylvanian Depression occupy the easternmost part of the territory. At the same time some of these deposits have been met with in the bore-holes and the Lueta ore deposit.

In the study region andesites are products of the first stage of manifestation of the subduction volcanism, which in the Călimani-Gurghiu-Harghita area took place between the Pannonian (zone E) and the Pontian. The totality of volcanics in this region, as a matter of fact in the whole area of the lower compartment of the Harghita Mts, constitute a volcano-sedimentary formation (Fig. 2).

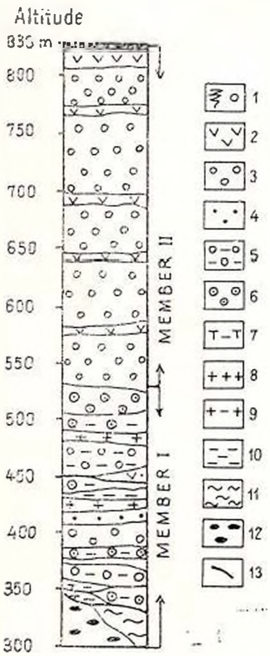

Fig. 2. — Synthetic lithological column of the volcano-sedimentary formation in the south-west of the Harghita Mts.

1, soil with blocks ; 2, andesitic lava ; 3, pyroclastic breccia ; 4, pyroclastic microbreccia ; 5, hybrid pyroclastic breccia ; 6, conglomerate, hybrid andesitic microconglomerate ; 7, andesitic sandstone ; 8, tuff ; 9, tuffite ; 10, clay ; 11, marl ; 12, conglomerate ; 13, limit of the Vlăhiţa iron formation.

Characterization of the Iron Types

The siderite and limonite mineralizations are known in two different geological situations, as follows :

1. At the contact of the andesitic volcanoclastics, belonging to the lower member of the volcano-sedimentary formation, with the Miocene and/or Pannonian sedimentary deposits. The mineralization forms concretions and lenses with sizes and contents of Fe % (> 20 %), which justify their exploration. Such accumulations represent the only ore in the Lueta ore deposit which can be turned to account. Peltz, Peltz (in Peltz et al., 1981) use for this mineralization the denomination "Lueta type mineralization" as it characterized in the course of time and at present the Lueta exploitation.

2. At different levels in the volcano-sedimentary formation (usually in the lower member) .The mineralization consists in the local sideritization of the binder of the volcanoclastics or of the tuffs, constituting either concretions or, more rarely, lenses with irregular distribution in the pyroepiclastic horizons. Peltz, Peltz (in Peltz et al., 1981) use for this mineralization the denomination "Vlăhiţa type mineralization", as is well-known from numerous boreholes, which in the area of the locality of Vlăhiţa investigated the mineralization from the volcano-sedimentary formation. In the Chirui region, in a geological situation similar to that which characterizes the Vlăhiţa type mineralization, there is a mineralization of limonitized siderite and limonite in association with opalites. For this mineralization, which has a genesis different from the Lueta and

Vlăhița types, P e l t z , P e l t z (in P e l t z et al., 1981) use the denomination of "Chirui type mineralization".

The classification of the types of iron mineralization was made taking into account the geologic environment, the factors which control the metallogenesis, the ore types, and the associated rocks (Tab. 1).

TABLE 1

Mineralization types and correlation criteria (South-West, Harghita Mts)

Mineralization type	Lueta	Vlăhița	Chirui
volcanism		e x t r u s i v e	
sedimentation environment	small basin lake, swamp		—
tectonic activity	—	fault, tension fracture	
volcano-sedimentary lithogenesis	—	+	—
circulation	fumarolian and hydrothermal emanations		
associated volcanic centres	+	±	±
associated eruptive bodies	—	±	±
associated rocks :			
volcanic	+	+	±
sedimentary	±	±	—
mineralization :			
siderite(s), limonite(l)	sl	sl	ls
mineralization phase	I	II	II+III?

The Lueta Type. The mineralization of this type crops out in isolated sectors at Minereni and Chirui. It also constitutes the main objective of the exploitation of the Lueta ore deposit. The ore bodies occur as : lenticular bed, lens, concretion. The mineralized rocks are represented by compact siderite, cavernous siderite, sideritized sandstone, sideritized microconglomerate.

The data on the position of the ore bodies in the Lueta ore deposit point out the special role played by the paleogeographic and lithologic control factors in metallogenesis. The mineralization is usually located in the terminal parts of the Lueta-Vlăhița Basin, which probably represents ramifications of an older water course (paleo-Homorod?) within the basin. In connection with the paleogeographic situation, the mineralization is encompassed by the Minereni conglomerates and pebbles, which represent the old alluvia and deluvia.

The Vlăhița Type. The drillings carried out in the area of the Lueta ore deposits and in its extension pointed out the presence of iron mineralizations at different levels of the volcano-sedimentary formation. The siderite mineralizations located in the lower member are much more numerous. The ore bodies consist of concretions of siderite ± sideritized binder of the coarse volcanoclastics± portions of the tuff levels (usually coaly). The spatial distribution of the mineralization is irregular ; however one can observe its tendency to be ordered on certain depth intervals. The mineralized rocks are represented by compact siderite, coaly siderite, sideritized tuff.

The Chirui Type. In the Chirui region (Linii Brook), in a volcanic area with numerous fractures, affecting the volcano-sedimentary and sedimentary deposits, as well as with numerous small-sized eruptive bodies, a surface with limonitized siderite and limonite accumulations associated with opalites is located. However, at Chirui there are several sectors with opalites, the only ones with such a development in the Călimani-Gurghiu-Harghita volcanic area. The ore bodies appear in the volcano-sedimentary formation at different depths. They consist of concretions and lenses of limonitized siderite, limonite, limonite+opalizations.

The role of the control factor, played by fractures in the Chirui mineralization, has been pointed out by Peltz, Peltz (in Peltz et al., 1978).

Chemism of the Mineralization

Major Elements. On the basis of 45 complete chemical analyses, the distribution of the major constituents has been examined in the types of mineralized rocks (Fig. 3). The SiO_2 values decrease continuously from the original rocks to siderites. In limonites, the SiO_2 values are higher due to the genetical particularities. The Fe_2O_3 values are significant for limonite and high in the limonitized siderites. The FeO contents increase directly proportional with the sideritization ; FeO is lacking in limonite. The significant increase of CO_2 in correlation with the high values of FeO can be observed ; CO_2 contents are low in limonite. In comparison with the original rocks, in the mineralized rocks the CO_2 values decrease, the MnO values increase (in all types), and the TiO_2 and P_2O_5 values are higher in certain types (Fig. 3).

The examination of the variation of the main major components from the original rocks to the mineralized ones indicates : increase of the FeO, Fe_2O_3, CO_2 and MnO values ; decrease of the SiO_2, Na_2O and K_2O values.

The comparison of the major components in the mineralizations of the Vlăhița Formation and of the Algoma type sideritic ones (Gross, 1980) indicates certain similarities.

Trace Elements. Pb, Cu, Ni, Co, Cr, V, Sc, Y, Be, Sr, Ba, Li, B have been determined by emission spectroscopy. The results are presented in Table 2. It is considered that the average values of the Fe_{tot}/Si ratio increases from the original rocks (0.08—0.09) to siderites (7.79) and limonitized siderites (11.71). In comparison with the increase of the Fe_{tot}/Si ratio (Fig. 4) the following behaviour of the trace elements can be noticed : Ni and Cr show a tendency of decrease ; Cu, V, Sc generally show a tendency of decrease, except the low values in sandstones ; Co appears in low contents, with quite close variation domains and average values in all types of rocks ; Sr and Ba generally show a tendency of decrease, except for the sideritized pyroclastics with values close to the original rocks ; Pb, Be, B, Li have similar variation domains and average contents in all types of rocks taken into account. On the basis of the behaviour of the trace elements studied it is proved that the mineralizing solutions were poor or deprived of Ni, Co, Cr, V, Sc, Ba, being a question of postvolcanic solutions

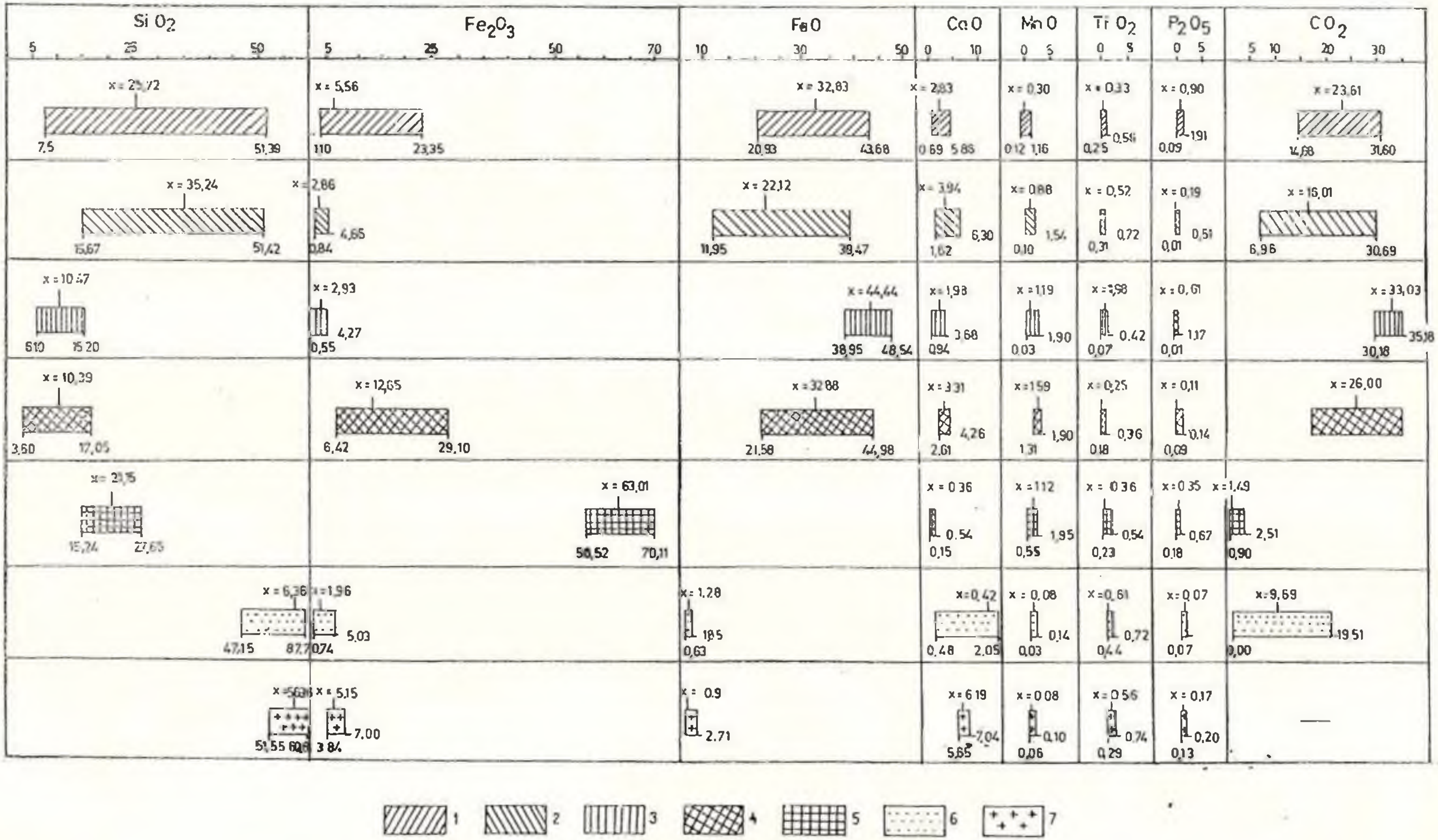

Fig. 3. — Variation domains and average values of oxides, on types of mineralized rocks.
1, sideritized sedimentary (n = 10); 2, sideritized pyroclastics (n = 13); 3, siderite (n = 12); 4, limonitized siderite (n = 6); 5, limonite (n = 4); 6, sandstone; 7, tuff.

TABLE 2

Variation domains and average values of trace elements in siderite limonite and sideritized rocks

Type of rock	Values avera-ge	Pb	Cu	Ni	Co	Cr	V	Sc	Y	Be	Sr	Bc	Li	B	n	Fe tot / Si
Sideritized sedimentary	x min	3	5,5	5	<3	12	15	2	11	<1	22	51	11	14		0.70
	x max	10	18	18	10	48	115	31	54	3	146	146	24	56	11	10.51
	$\bar{x}$	6	12	12.5	10	27.5	49	10.5	36	1.6	68	93	21	37		2.82
Sideritized Pyroclastics	x min	3	4	2.5	9	5	5	6.5	16	<2	34	54	17	30		0.33
	x max	22	30	18	28	61	133	35	60	2	350	420	30	75	15	4.34
	$\bar{x}$	11	13	10	16	30.5	88	17	38	1.4	200	212	23	58		1.61
Siderite	x min	<3	4	2.5	3	6	11	3	26	<1	16	25	8	18		4.83
	x max	10	14	15	18	30	140	39	96	2	152	170	45	95	20	14.02
	$\bar{x}$	5	8	9	10.5	16	45	11.5	51	1.4	42	69	21	41		7.79
Limonitized siderite	x min	<3	3.5	3.5	4.5	2.5	27	4	27	<1	19	36	7	27		4.32
	x max	9.5	13	15	12	34	86	12	35	1.4	117	78	10	43	6	23.52
	$\bar{x}$	/	7	9	8	11.5	55	8	31	/	54	56	13	30.5		11.71
Limonite	x min	<3	4	17	8.5	4	35	10	29	1.9	17	63	<4	<10		3.12
	x max	49	80	120	80	25	440	25	61	3	95	330	11	25	3	6.19
Sandstone	x	5	9	14	3	33.5	19	3	10.5		185	185	10		2	0.09
Tuff	$\bar{x}$	7	54	13	11	35	118	19	15	/	240	300	36	/	2	0.08

and solutions not connected directly with the volcanic activity, bearing and/or enriched in iron. As known in the world, in case of the carbonatic

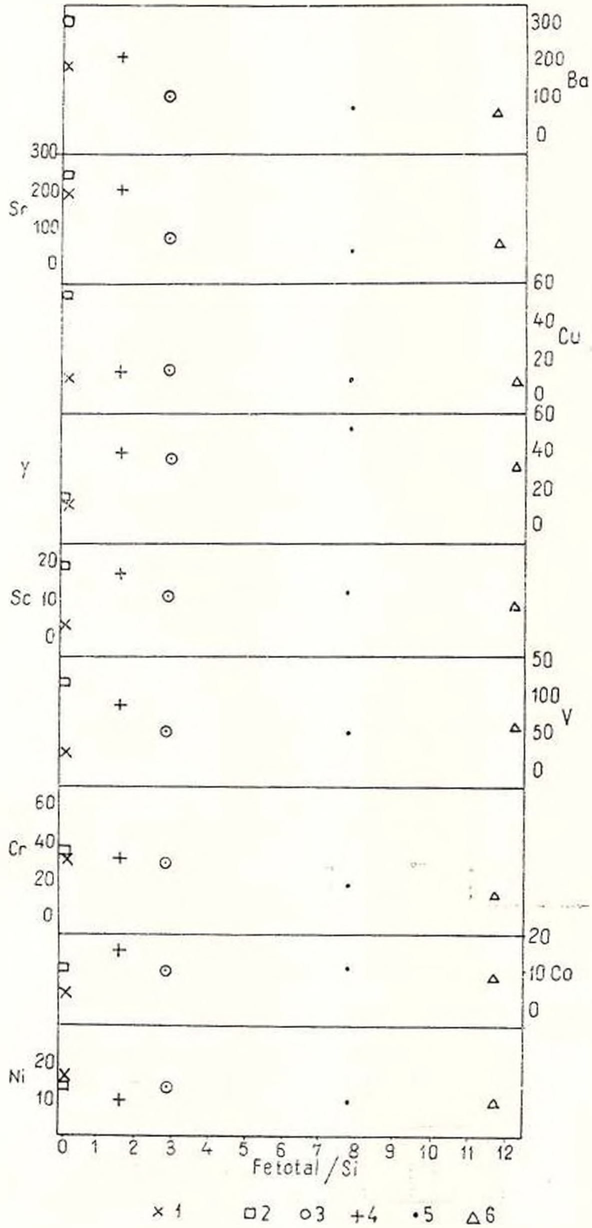

Fig. 4. — Fe_{tot}/Si-Ni, Co, Cr, V, Sc, Y, Cu, Sr, Ba diagram (average values). 1, sandstone; 2, tuff; 3, sideritized sedimentary; 4, sideritized pyroclastics; 5, siderite; 6, limonitized siderite.

ore deposits with a genesis partly similar to the mineralizations in the Vlăhiţa region (G r o s s, M c L e o d, 1980) a negative correlation between Fe^{+3} and V, Ni, Co, Cr, is emphasized.

Genesis of the Mineralization

The solving of the problem on the genesis of the iron mineralization in the south-west of the Harghita Mts is of great importance. It is implied in the drawing up of the programmes which aim at the possible identification of some new zones with iron mineralizations, which can be turned into account.

In the world, certain aspects on the genesis of the iron mineralizations are still unsolved. Numerous researchers expressed their opinions, with general or special character, the genetic interpretation being determined by the specification of the source of the solutions and of the ways of transport, deposition and accumulation of iron. As concerns the genesis of the sideritic mineralizations, which form the Lueta ore deposit, up to now several points of view have been presented; deposition from hot springs (Jekelius, 1938; Codarcea, Petrulian, 1941); deposition from subsurface waters (Atanasiu, 1939; Ghika-Budești, 1949); metasomatic genesis (Petrulian, 1951; Grigore, 1961; Treiber, 1965); hydrothermal genesis (Banyai, 1957; Grigore, 1961); volcano-sedimentary genesis (Kosareva, Catană, 1965).

The available data indicate once more the complexity of the factors which have to be considered with a view to the clearing up of the mineralization genesis. It is due to the complexity of the parameters which control the genesis that the mineralizations are not of a single type, as formerly considered, but of three types, as shown at point 2. Taking into account all the genetic factors related to: a) iron source; b) transport in thermal solutions, subsurface waters, ascendant hydrotherms; c) deposition and conditions which rule it (Eh, pH, T°, organic matter): d) factors of paleogeographic, lithologic, tectonic control, which check the iron accumulation, we consider that in the study region the mineralizations are the result of a complex genesis, volcano-sedimentary+hydrothermal — metasomatic + hydrothermal. If we study the genesis of each type of mineralization it is considered: the volcano-sedimentary genesis for the Lueta type mineralization; volcano-sedimentary and hydrothermal — metasomatic genesis for the Vlăhița type mineralization; the hydrothermal and hydrothermal-metasomatic genesis for the Chirui type mineralization.

The Lueta type mineralization accumulated in the Minereni conglomerates and pebbles (lithologic control), probably also above this deposit in a partly barred basin (paleogeographic control). Postvolcanic solutions, generated by apparata situated in the north of the Vlăhița region, preferentially circulating on the slopes of the volcanoes, have determined the alteration and leaching of the volcanics. The result was an important enrichment of the iron solutions. The existence of gases (CO_2, H_2S) led to the alteration of the mostly volcanoclastic rocks, as well as to the iron. The solutions enriched in iron penetrated into the lacustrine basin (Fig. 5), became diluted, more basic, at the values pH > 5 the iron participation being achieved. The mineralization accumulated in the first period of volcanic calm, but with a rich post-volcanic activity, which started in this region at the beginning of the Pontian.

The Vlăhiţa type mineralization accumulated during several periods of volcanic calm, subsequent to the Lueta metallogenetic phase. The Vlăhiţa metallogenetic phase was probably the longest one; it covers the

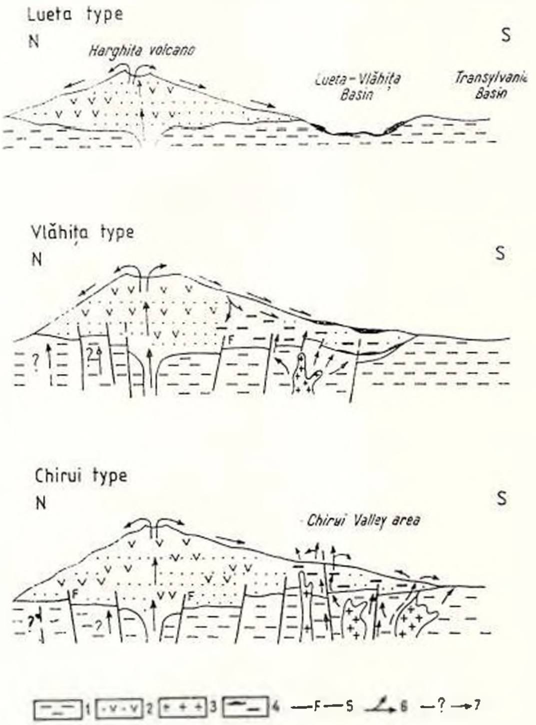

Fig. 5. — Geological environment of deposition of the Vlăhiţa iron formation (no scale).

1, pre-volcanic basement, Miocene molasse of the Transylvanian Depression; 2, volcano-sedimentary formation; 3, eruptive body; 4, mineralization; 5, fault; 6, hydrothermal flux; 7, supposed hydrothermal flux, north of the volcanic apparata.

entire time period during which the lower member of the volcano-sedimentary formation accumulated. The cyclicity of the calm moments controlled the spatial position of the mineralization. The iron source as well as the transport modalities were similar to those considered for the Lueta type mineralization. The paleogeographic and lithologic control factors were different; a less important role was played by the tectonic control. The formation of small depression areas at the surface of the volcanoclastics during the period of volcanic calm favoured the formation of pools and swamps with their lithogenetic and biogenetic processes. In these sedimentation environments the iron accumulations from post-volcanic solutions occurred under favourable conditions of Eh and pH. Concomitantly, the post-volcanic solutions enriched in iron, which infiltrated in the volcanoclastic beds, preferentially sideritized their porous binder; in this case the metasomatic action of the solutions is obvious. At the same mineralization moment, ascendant hydrotherms, usually mobilized on systems of fractures in the aureole of eruptive bodies, metasomatically enriched the binder mineralization. The spatial position of the volcano-sedimentary siderite, constantly associated with coaly tuffs, sedimentary rocks, epiclastics, is obviously controlled by the accumulation in pools and swamps.

The spatial position of the hydrothermal metasomatic siderite is controlled by the solution thermodynamics, as well as by the porosity + composition of the associated volcanoclastics.

The Chirui type mineralization accumulated in the volcano-sedimentary formation and is characterized by the presence of opalites in association with limonite and siderite. The opalization was partly contemporaneous with the volcanics accumulation, and partly subsequent, being connected with solutions usually accompanying the volcanic activity. Iron comes from the alteration of the volcanics under the influence of the products of the post-volcanic activity. Silica results from the decomposition of the volcanoclastics "*in situ*", as well as from hydrotherms. The solutions which vehicled iron and silica circulated on fractures of the aureole of the eruptive bodies. In the case of the genesis of this type of mineralization the deposition conditions particularly interfere. Iron precipitates as hydroxide or carbonate, depending on the pH of the solutions. Silica is less dependant on pH (K r a u s k o p f, 1971), being connected firstly with the ionic changes and the decrease of temperature. The precipitation of the gel results from the suprasaturated solutions, and the siliceous gel itself is flocculated by gels of iron hydroxide (H a r d e r, F l e h m i n g, 1970).

Comparative and Final Considerations

The presence in the Băile Homorod-Vlăhiţa-Chirui region, in a large area, of the siderite and limonite mineralizations at the base of the volcano-sedimentary formation as well as in the lower member, give the character of "iron formation" to this lithologic unit. We call it "the Vlăhiţa iron formation". This formation includes, besides the Lueta ore deposit, numerous areas with iron rock accumulations (10–15% Fe) belonging to the Vlăhiţa type, as well as accumulations with poor mineralizations (15–25% Fe) of the Vlăhiţa and Chirui type. Considering the geological and tectonic environment, the associated rocks and the lithofacies, the Vlăhiţa type formation has some similarities with the Algoma type iron formation (G r o s s, 1980). In the case of the iron metallogenetic districts associated with the andesitic volcanism in the East Carpathians, similar mineralizations are known, as follows: 1) Mădăraş area (Topliţa-Sînsimion district). The Vlăhiţa and Lueta type mineralizations are associated with a special type (T ă n ă s e s c u, 1969), represented by the siderite accumulated in metamorphic rocks (schists, limestones, dolomites), hydrothermally and hydrothermal-metasomatically. On the basis of the features pointing to the mineralogy of the mineralization, genesis, geological environment, associated rocks, we consider the presence of the Mădăraş iron formation, similar to the Vlăhiţa formation, in the Mădăraş metallogenetic field. 2) Topliţa region in the northernmost part of the Topliţa-Sînsimion district encompasses sideritic mineralizations of the Vlăhiţa and Mădăraş type. At the present stage of knowledge the presence of an iron formation in this region cannot be specified. 3) Filia-Biborţeni-Herculian region (south of the Corund-Herculian district) includes isolated sideritic and limonitic accumulations with a volcano-sedimentary genesis, of the Vlăhiţa type. The poor mineralizations and the ferri-ferrous rocks predominate. Locally, limonite occurs associated with diatoms. As in the Topliţa region, no iron formation can be pointed out here.

REFERENCES

Atanasiu I. (1939) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Codarcea A., Petrulian N. (1941) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Ghika-Budeşti Ş. (1949) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Grigore I. (1961) Contribuţii la dezvoltarea bazei materiale a minereurilor de fier. *Rev. Minelor*, XII/7, Bucureşti.

Gross A. (1980) A classification of iron formations based on depositional environments. *Canadian Mineralogist*, 18/2, p. 215—222, Toronto.

— Mc Leod C. R. (1980) A preliminary assessment of the chemical composition of iron formations in Canada. *Canadian Mineralogist*, 18/2, p. 223—229.

Harder H., Flehming W. (1970) Quartzsynthesen bei Tiefen Temperaturen. *Geochim. Cosmochim. Acta*, 34, p. 295—305, Oxford.

Jekelius E. (1938) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Kosareva T., Catană N. (1965) Citeva consideraţii privind geneza minereului de fier de la Lueta (masivul Harghita). *Rev. Minelor*, 3, p. 112—116.

Krauskopf B. (1971) The source of ore metals. *Geochim. Cosmochim. Acta*, 35, 7.

Peltz S., Peltz M., Bratosin I., Olteanu R. (1978) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

— Peltz M., Bratosin I., Ianc R. (1981) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

— Peltz M., Bratosin I., Ianc R. (1982) Contribuţii la cunoaşterea mineralizaţiei de fier din regiunea Lueta-Vlahiţa-Chirui (judeţul Harghita). *D. S. Inst. Geol. Geofiz.*, LXVI/2 (1978—1979), Bucureşti.

Petrulian N. (1951) Zăcăminte de minerale utile. In Manual Ing., Mine, I, Ed. tehnică, Bucureşti.

Tănăsescu L. (1969) Date noi asupra formaţiunilor geologice de adincime din Bazinul Mădăraş şi asupra mineralizaţiei de siderită. *St. tehn. econ.*, seria A/7, Bucureşti.

Treiber I. (1965) Contribuţii la geologia şi petrografia Harghitei de Nord cu privire specială la geneza zăcămintelor de fier din regiunea Lueta-Vlahiţa. *Stud. Univ. Babeş-Bolyai*, Cluj.

LE MAGMATISME D'ÂGE MÉSOZOÏQUE DANS LES CARPATHES ORIENTALES[1]

PAR

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Les plus importants produits de l'activité magmatique d'âge mésozoïque des Carpathes Orientales affleurent dans la zone cristallino-mésozoïque et dans les plus internes parties de la zone du flysch. C'est surtout de ces formations-là que nous nous occupons dans cette note.

Du point de vue tectonique les magmatites mésozoïques se trouvent dans les nappes transylvaines, les nappes centrales est-carpathiques et les nappes daciques externes. Les nappes transylvaines ont la position la plus haute — et donc l'origine la plus interne — de l'échafaudage structural des Carpathes Orientales internes. Elles sont constituées d'un complexe ophiolitique et de roches sédimentaires d'âge triasique, jurassique et éocrétacé. Il n'y a pas de formations cristallophylliennes dans les nappes transylvaines; ce sont des nappes d'obduction provenant d'une suture ophiolitique représentée par les Transylvanides (Săndulescu, 1975, 1980). Les nappes centrales est-carpathiques sont des nappes de socle constituées chacune de formations métamorphiques préalpines et de dépôts sédimentaires (Trias, Jurassique, Eocrétacé) qui surmontent normalement les premières. De haut en bas les nappes centrales est-carpathiques sont : la nappe bucovinienne, la nappe sub-bucovinienne et les nappes infrabucoviniennes. Des nappes daciques externes c'est surtout la nappe du "Flysch Noir" qui offre la possibilité d'une analyse plus approfondie du complexe basique situé à la base de celle-ci. A part les roches magmatiques, dans ces nappes affleurent également des formations de type flysch et des niveaux calcaires.

La tectogenèse principale des nappes transylvaines et des nappes centrales est-carpathiques et mésocrétacées. Les nappes daciques externes ont subi une tectogenèse paroximale à la fin du Crétacé supérieur.

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Complexe ophiolitique des nappes transylvaines

Le complexe ophiolitique transylvaine est constitué de péridotites plus ou moins serpentinisées, de dolérites, de basaltes (pillow-lavas et coulées massives) et aussi de plagigranites. Dans l'une de ces nappes transylvaines (nappe de l'Olt) sont associées également des roches alcalines (trachytes et syénites quartzifères). Des jaspes (de différents âges ?) sont connus aussi dans ce complexe ophiolitique.

Les péridotites sont représentées par des hartzburgites et lherzolizes, interstratifiées à l'échelle microscopique avec de minces lamines de websterites et d'orthopyroxénites. Certains aspects microscopiques rappellent les produits cumulatifs, suggérant leur origine dans le niveau des péridotites métamorphiques ou dans les cumulats situés au-dessus de celles-ci.

Les roches basiques sont représentées par des basaltes à structures variées, certaines étant spilitisées, des dolérites, des gabbros et aussi des tufs de différents types. Elles montrent une tendance de différenciation magmatique de type tholéïtique (Russo-Săndulescu et al., 1982). Il faut aussi remarquer la présence, dans les basaltes, du prehnite, du pumpellite et des zéolites, fait qui peut suggérer l'existence — dans le domaine transylvain — d'un métamorphisme de haute pression, qui aurait suivi un métamorphisme hydrothermal soutenu par les phénomènes de spilitisation. Dans des klippes sédimentaires inclues dans la Formation du Wildflysch bucovinien et provenant des séries transylvaines ont été identifiées des schallsteines et des ophicalcites.

Les roches alcalines associées aux nappes transylvaines affleurent seulement dans les Monts Perşani (dans la nappe de l'Olt) (Patrulius et al., 1976). Elles y sont représentées par des trachytes quartzifères, des syénites quartzifères, des bostonites et des rhyolites alcali-feldspathiques (Cioflica et al., 1965 ; Săndulescu et al., 1981). Les roches alcalines traversent des basaltes, englobant aussi des morceaux de celle-ci.

Les magmatites du complexe ophiolitique des nappes transylvaines ont des âges différents. Les plus anciennes sont d'âge triasique (moyen), les plus jeunes jurassique supérieur et, peut-être même néocomien (Săndulescu, Russo-Săndulescu, 1981). Ce fait montre qu'elles proviennent de différentes parties de la zone de spreading qui a été entrainées dans les processus d'obduction générant les nappes transylvaines.

Complexe basique de la nappe du "Flysch Noir"

Affleurant seulement dans les Monts du Maramureş, le complexe basique de la nappe du "Flysch Noir" est constitué de coulées massives (par endroits des pillow) de roches basiques, de dolérites et de roches d'origine mixte (stromatites et tuffites) ; rarement y ont été trouvées des roches à faible tendance alcaline. Bien que de constitution minéralogique homogène on y remarque l'abondance de l'augite titanifère par rapport aux basaltes des nappes transylvaines. Les basaltes de la nappe du "Flysch Noir" montrent une tendance de différenciation de type tholéïtique alcalin. Les processus de spilitisation sont relativement rares, mais la présence généralisée du chlorite et de l'épidote soutient, quand même, l'existence d'un métamorphisme hydrothermal,sous-marin. Un second moment de métamorphisme est marqué par le développement général du pumpellite. De ce moment est lié également la genèse du prehnite (identifié par endroits) et des noyaux d'amphiboles bleus (dans des dolérites — vallée de la Rica). De ce fait nous admettons que ce deuxième moment de métamorphisme est de type hP/bT. Il serait contemporain à l'apparition de la schistosité qui est développée en liaison avec les écaillages qui ont précédé le charriage de la nappe.

L'âge de l'ensemble basique de la nappe du "Flysch Noir" est au moins jurassique supérieur. Il n'est pas encore certain mais il n'est pas exclu qu'il a débuté même avant le Callovien.

Magmatites ensialiques des nappes centrales est-carpathiques

Le magmatisme d'âge mésozoïque lié aux nappes centrales est-carpathiques — ou au moins à leurs équivalents dans la partie sud de la courbure interne des Carpathes — est le mieux exprimé dans la région de Holbav et Poiana Mărului-Sinca. Dans la nappe gétique (Holbav et Brașov) — qui correspond aux nappes infrabucoviniennes — affleurent autant des roches effusives que des roches filoniennes, tandis que dans les nappes supragétiques (Monts Făgăraș) il n'y a que des roches filoniennes.

Les roches effusives prennent part à la constitution d'une formation volcano-sédimentaire d'âge liasique (Holbav). Elle comprend des pyroclastites de type ignimbritique à chimisme alcalin (rhyolites et tufs rhyolitiques alcalins) alternant avec des dépôts sédimentaires et des tuffites ; des sills et des coulées minces s'y intercalent (rhyolites alcalines à arfvedsonite et trachytes). Cette formation volcano-sédimentaire est traversée par des petits dykes de basaltes olivino-pyroxéniques. Les roches filoniennes sont largement développées dans l'unité de Brașov (partie interne du domaine gétique dans la courbure des Carpathes) et dans la nappe de Sinca (une des nappes supragétiques correspondant à la nappe sub-bucovinienne), donc dans des unités plus internes que celle de Holbav (à effusions). Il s'agit surtout de camptonites, de syénites quartzifères et de bostonites, ces dernières largement développées dans la nappe de Sinca.

Certainement que, parmi les magmatites qui traversent les formations cristallines des Carpathes Orientales, il faut ranger aussi le massif alcalin de Ditrău. Etant un problème en soi, nous nous bornerons de souligner son âge jurassique (S t r e c k e i s e n, H u n z i k e r, 1974), et son caractère lié à une période de distension et pas de compression. Du point de vue tectonique il se trouve dans la nappe alpine la plus haute — donc la plus interne — du système central-est-carpathique, celle bucovinienne (S ă n d u l e s c u, 1972, 1975 a).

Considérations géochimiques

Nous allons analyser les roches magmatiques d'âge mésozoïque des Carpathes Orientales d'une manière globale sans découpages séparées par unités tectoniques.

Roches ultramafiques. Les éléments mineurs caractérisant les roches ultramafiques se trouvent à des valeurs élevées par rapport aux moyennes citées dans la littérature pour ce type (G o l e s in C o l l e m a n, 1977). Ces valeurs du Ni varient entre 2100 et 2900 ppm dépassant même celle des péridotites métamorphiques (2250) ; le Cr varie entre 2600 et 3000 ppm, se situe au-dessus de la moyenne caractérisant les roches ultramafiques, mais plus faible que celles de péridotites métamorphiques. Ces valeurs élevées du Ni et du Cr sont liées à la présence des sulfures de Ni et des cromites. Les autres éléments mineurs analysés — Co, V, Sc — présentent des valeurs rapprochées de celles caractérisant les roches ultrabasiques.

Roches basiques. L'étude géochimique des roches basiques fournit des conclusions intéressantes sur les conditions géotectoniques dans lesquelles ont été engendrées les roches magmatiques mésozoïques. Suivant le diagramme Perace et Cann (1973) on a distingué (R u s s o - S ă n d u l e s c u et al., 1982 ; S ă n d u l e s c u et al., 1980) les suivantes catégories : (1) basaltes de type fond océanique, (2) basaltes de type intraplaque océanique (hot spots) (1 et 2 dans les nappes transylvaines) et (3) basaltes de type intraplaque continentale (rift continental) passant à un rift de type Mer Rouge. Les roches basiques associées aux complexes alcalins ensialiques (Holbav, Sinca) ne trouvent pas leur place dans l'ensemble des basaltes cités ci-dessus à cause de leur caractère très alcalin : teneurs élevées en Zr et Nb.

Pour comparer les roches basiques situées dans des unités différentes nous avons utilisé les teneurs en Zr, Y, Yb et Nb, qui sont, de la même manière que le Ti et le P, peu affectés par les processus de métamorphisme et métasomatisme hydrothermal.

Le rapport Y/Nb sépare distinctement, dans les nappes transylvaines (fig. 1) les basaltes de fond océanique de celles de hot-spot ; les premières contenant moins de 1o ppm pour le Nb et 10 — 90 ppm pour l'Y, les secondes étant caractérisées par des contenus de 14 — 32 ppm pour l'Y et de 15 — 85 ppm pour le Nb. Les basaltes de type intraplaque continentale (nappe du "Flysch Noir") se superposent partiellement aux deux précédents. Les basaltes alcalins ensialiques (Holbav, Sinca) montrent des valeurs semblables avec celles des basaltes du même type des nappes transylvaines.

L'Yb varie d'une manière semblable avec l'Y, largement dans les roches de type fond océanique (jusqu'à 9 ppm) et plus serré dans celles de type hot-spot (2—3 ppm).

Le La se trouve sous la limite de détection (30 ppm) dans les roches de type fond océanique, mais monte jusqu'à 70 ppm dans celles de type hot-spot. Les basaltes de la nappe du "Flysch Noir" sont de ce point de vue dans une position intermédiaire : 75 % au-dessous de la limite de détection, le reste variant entre 30 — 52 ppm.

La différence entre les roches de type fond océanique et de type hot-spot sont soulignées aussi par les teneurs en Ba (fig. 2), rapportés au FeO^{tot} /MgO. Les premières présentent des valeurs au-dessous de 100 ppm, les autres entre 140 — 1000 ppm. La nappe du "Flysch Noir" a encore une fois une position intermédiaire.

Le Co, V et Sc ont des valeurs plus élevées dans les roches de type fond océanique. Pour les types fond océanique $\overline{X}$Co = 51 ppm, $\overline{X}$V = = 275 ppm, $\overline{X}$Sc = 39 ppm ; pour ce qui est de type hot spot les valeurs sont plus basses ($\overline{X}$Co = 29 ppm, $\overline{X}$V = 160 ppm, $\overline{X}$Sc = 16 ppm). Les

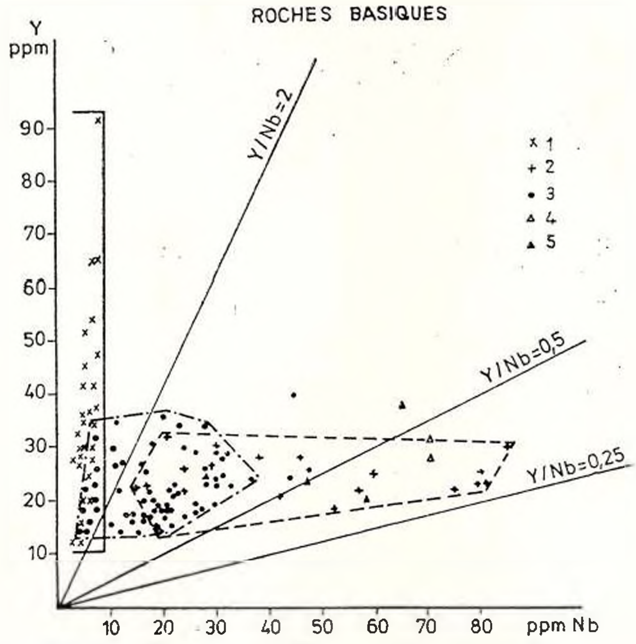


Fig. 1. — Diagramme Y-Nb pour les roches basiques.

1, basaltes de type fond océanique-Rarău, Hăghimaş, Perşani ; 2, basaltes de type intraplaque océanique (Hot spot)-Rarău, Hăghimaş, Perşani ; 3, basaltes de type intraplaque continental (rift continental)-Maramureş ; 4, basaltes alcalins de Holbav-Braşov (nappe gétique) ; 5, basaltes alcalins-Monts Făgăraş (nappe supragétique).

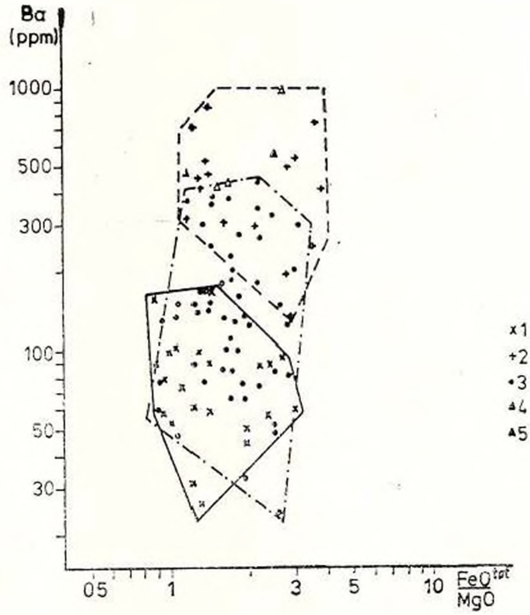


Fig. 2. — Diagramme Ba/FeO^{tot} /MgO (légende voir la fig. 1).

basaltes de la nappe du "Flysch Noir" sont toujours en position intermédiaire ($\overline{X}Co = 43$ ppm, $\overline{X}V = 200$ pp, $\overline{X}Sc =$ ppm).

Les roches alcalines associées aux basaltes analysés ci-dessus sont caractérisées par des valeurs élevées de Zr, Nb, Y, Yb et La (valeurs maximales : Zr = 1900 ppm, Nb = 300 ppm, Y = 125 ppm, Yb = 10 ppm,

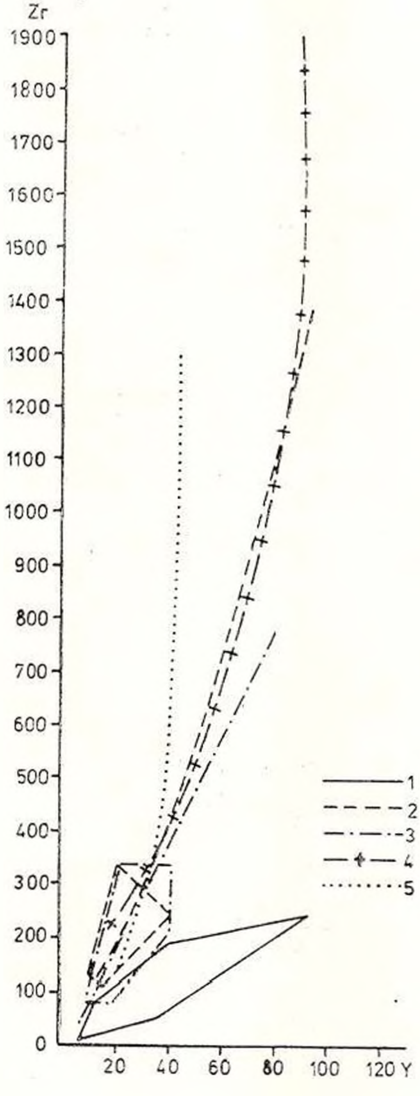


Fig. 3. — Diagramme Zr/Y (légende voir la fig. 1).

La = 280 ppm, reconnues dans l'unité de Holbav). On observe que les roches alcalines subissent des tendances de différenciation spécifiques suivant les catégories correspondantes de basaltes (fig. 3).

Utilisant les rapports P_2O_5/Zr ppm (fig. 4) on arrive à la conclusion que les magmas qui ont généré les roches de type fond océanique des nappes transylvaines sont d'origine nettement tholéïtique (les valeurs les

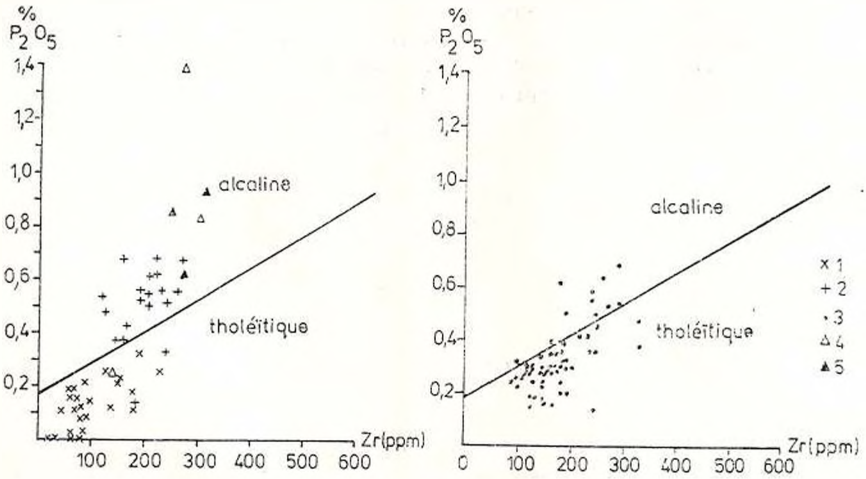


Fig. 4. — Diagramme P_2O_5 /Zr (légende voir la fig. 1).

plus faibles en Zr et P), tandis que ceux qui ont généré les basaltes intraplaques continentaux (nappe du "Flysch Noir") bien que tholéïtiques ont des valeurs plus élevées. Les roches basiques de type hot-spot correspondent à des magmas alcalins, ainsi qu'une faible partie des basaltes de la appe du "Flysch Noir".

Evolution et position géotectonique

Les trois circonstances structurales dans lesquelles affleurent les roches magmatiques d'âge mésozoïque dans les Carpathes Orientales répètent les trois ambiances géotectoniques qui ont gouverné leur genèse. En effet les nappes transylvaines — provenant d'une suture ophiolitique bien marquée (les Transylvanides, Săndulescu, 1980) — présentent les caractères les plus nets et les plus typiques d'une croûte océanique. Les traces du spreading prolongé (Trias moyen-Jurassique supérieur voir même Néocomien) ne fait que souligner cette conclusion. Le complexe basique de la nappe du "Flysch Noir" avec ces caractères ambigus trouve bien sa place dans un contexte de rifting continental passant aux caractères de croûte amincie. D'ailleurs l'apparition — le long de la chaîne — dans le même sillon de la croûte océanique (nappe de Severin) (Săndulescu, 1980) marque la différence par rapport à la suture des Transylvanides : l'ouverture parallèle du sillon dacique externe est plus hésitante, "imitant" seulement le spreading transylvain.

Enfin le magmatisme ensialique qui s'est manifesté dans une aire située entre les deux zones de distension est dominé par les produits alcalins. Bien que cette situation n'ait pas trouvé — encore — une explication satisfaisante, il est à remarquer que cette zone a gagné ces caractères spécifiques par rapport à deux autres. La présence — réduite — des magmatites alcalines dans les nappes transylvaines (seulement au niveau du Trias moyen) suggère l'influence du rifting primaire qui a précédé (et en même temps a déclenché) le spreading.

Du point de vue de l'évolution temporelle on constate que : (1) le spreading transylvain (son début) a été suivi par (2) l'apparition du magmatisme alcalin plus à l'extérieur, qui à son tour a précédé (3) le rifting encore plus externe qui a découpé la marge continentale.

[4] Représente les moyennes des teneurs.

BIBLIOGRAPHIE

Cioflica G., Patrulius D., Ionescu J., Udubașa G. (1965) Ofiolitele alohtone triasice din Munții Perșani. *Stud. cerc. geol., ser. geol.*, 10, 1, p. 161–182, București.

Colleman R. G. (1977) Ophiolites. 229 p., 72 fig., Springer Verlag Berlin-Heidelberg-New York.

Patrulius D., Popescu I., Mirăuță E., Gheorghian D., Antonescu E. (1976) Rapport, les archives de l'Institut de géologie et géophysique, Bucarest.

Russo-Săndulescu D., Udrescu C., Medeșan A. (1982) Petrochemical Characteristics of the Mesozoic Ophiolites of the Rarău-Hăghimaș Marginal Syncline. *D. S. Inst. geol., geofiz.*, LXVI/1, p. 153–185, Bucharest.

Săndulescu M. (1972) Considerații asupra posibilităților de corelare a structurii Carpaților Orientali și Occidentali. *D. S. Inst. geol. geofiz.*, LVIII/5, p. 125–150, București.

— (1975) Essai de synthèse structurale des Carpathes. *Bull. Soc. géol. Fr.*, (7) XVII, 3, p. 299–358, Paris.

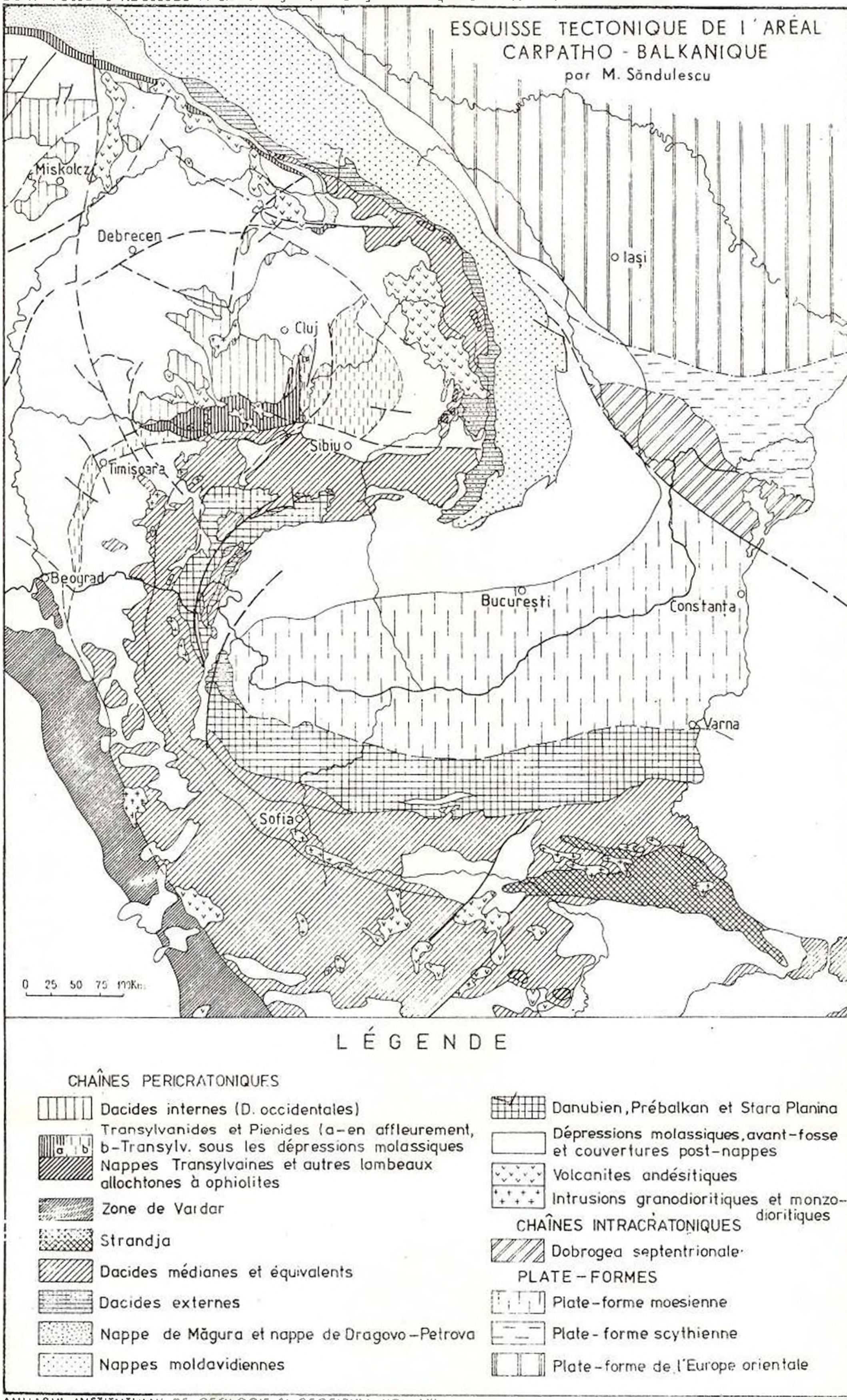
— Russo-Săndulescu D., Bratosin I., Medeșan A. (1980) Rapport, les archives de l'Institut de géologie et géophysique, Bucarest.

— Russo-Săndulescu D., Udrescu C., Medeșan A. (1981) Rapport, les archives de l'Institut de géologie et géophysique, Bucarest.

— (1980) Analyse géotectonique des chaînes alpines situées autour de la Mer Noire occidentale. *An. Inst. geol. geofiz.*, LVI, p. 5–54, București.

— Russo-Săndulescu D. (1981) The Ophiolites of the Rarău and Hăghimaș Synclines-Their Structural Position, Age and Geotectonic Evolution. *D. S. Inst. geol. geofiz.* LXVI/5, p. 103–114, București.

Streckeisen A., Hunziker J. C. (1974) On the Origin and Age of the Nepheline Syenite Massif of Ditro (Transylvania, Romania). *Sweiz. Min. Ptr. Mitt.*, 54, 1, p. 59–77.



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GEOTECTONIC AND MAGMATIC EVOLUTION OF THE MUREȘ ZONE (APUSENI MOUNTAINS) — ROMANIA [1]

BY

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The Mureș zone, known also as the Southern Apuseni, the Mureș Mountains (Macovei, Atanasiu, 1933) or the Metaliferi Mountains (Ianovici et al., 1967), formed in the initial stage of the Alpine cycle. Unlike the other two local names, that of the Mureș zone has a genetic sense, being confounded with the suture of the Mureș Ocean ; it is therefore more comprehensive and includes also the ophiolites outside the Southern Apuseni. It crops out along a distance of 200 km, between Turda and Pătirș, but continues in Yugoslavia towards the Shumadia zone.

As long as it was considered to be the result of the evolution of a geosyncline (Stille, 1953) or of a classic oceanic zone (Rădulescu, Săndulescu, 1973; Herz, Savu, 1974; Bleahu, 1974) several aspects relating to its development could not be explained. First, not all the magmatic rocks supposed to have formed in the oceanic zone could be defined as ophiolitic rocks (Savu, 1962 b). Some other aspects were not clear and the situation was even more complicated when all the magmatic rocks of the Mureș zone were considered to have formed in an island arc zone (Cioflica et al., 1980).

Our concept, according to which the Mureș zone functioned primarily as an ocean floor zone, becoming finally a zone of island arcs (Savu, 1980), can solve the still existing problems. This model has been gradually achieved. It started from the observation that the closing of the oceanic zone is due to a compression process accompanied by a bilateral subduction one (Savu, Udrescu, 1973; Savu, Nicolae, 1975) and from the discovery that the basic rocks formed prior to subduction correspond to ocean floor basalts, while those generated by the subduction process (Savu et al., 1978) to island arc volcanics. According to this model we shall present the evolution of the Mureș zone in two stages : (1) the spreading stage and (2) the closing stage.

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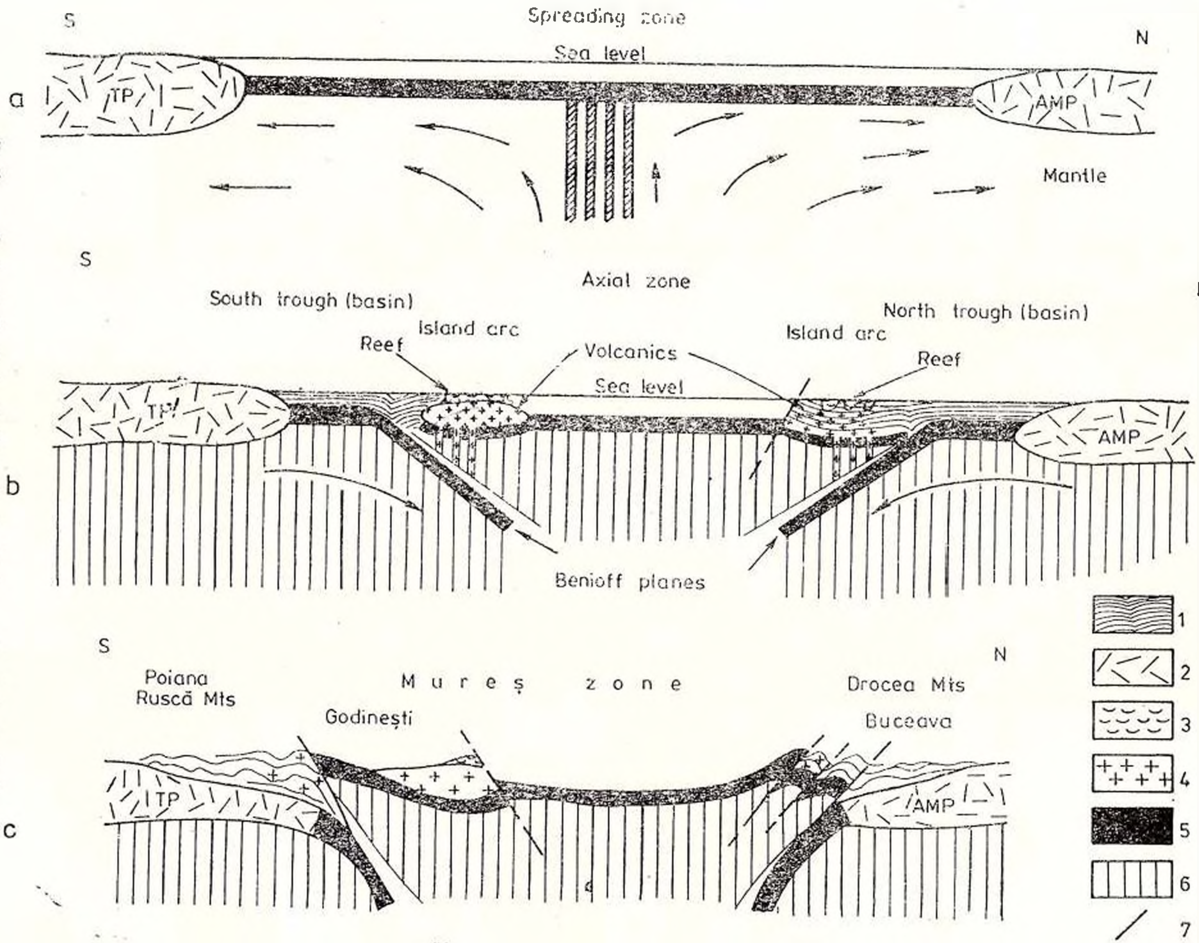

Fig. 1. — Geotectonic evolution of the Mureș zone (a model).

a, the spreading stage of the oceanic zone (Jurassic-Pre-Oxfordian); b, the closing stage marked by bilateral subduction (Upper Jurassic-Lower Cretaceous); c, folding of the Mureș zone during the Austrian and Laramian movements; TP, Transylvanian microplate; AMP, Apuseni Mountains microplate.

1, flysch deposits; 2, continental crust; 3, calcareous reefs; 4, island arc volcanics; 5, ocean crust (ophiolites); 6, upper mantle; 7, major fractures.

Geotectonic Evolution of the Mureş Zone

The formation of the Mureş zone, which appears as an asymmetric orogen with bilateral overthrusting (M a c o v e i, A t a n a s i u, 1933) started in the Lower Jurassic by the splitting of the sialic crust on the south-western margin of the East European continent into two microcontinents, represented by the Apuseni Mountains Microplate and the Transylvanian Microplate (S a v u, 1980). Thus a rift forms, which is marked by a spreading phenomenon (Fig. 1 a)[3].

The spreading lasts to the new Kimmerian movements which start manifesting in the Upper Callovian or Oxfordian. This period corresponds to the first evolution stage of the Mureş zone, when the oceanic crust formed [4], consisting of ophiolitic rocks as defined by C o l e m a n (1977); its age is of 180 m.y.

Dacă spreadu Call-o. ! → numai dece 180 m.y.? de unde — fără citare

With the new Kimmerian movements starts the closing of the oceanic zone, an event previously named "the geosyncline inversion", with which the bilateral subduction process taking place during the second evolution stage of the Mureş zone is associated. This process lasts to the Austrian movements that determined the final closing of the oceanic zone.

According to the proposed model (Fig. 1 b), at the beginning of the second stage the oceanic crust, subjected to compression, is broken in its marginal parts. The external segments joining the northern and southern microplates are subducted under the Mureş zone, thus generating two marginal troughs or basins (S a v u, 1962 b). The magmas forming on the Benioff planes at a depth of about 250 km give rise to two pre-orogenic or submarine island arcs in which typical volcanic rocks erupt (S a v u, 1980). Owing to the Austrian movements, the Mureş zone undergoes a folding and consolidation process, ending with the Laramian movements (P a p i u, 1953). In this way the folded and overthrust nappes structure of the Mureş zone forms (Fig. 1 c).

Geological data are very well correlated with the results of the geophysical (gravimetric and aeromagnetic) researches obtained by A n d r e i et al. (1975) which A n d r e i figures also on the map drawn by B o r c o ş et al. (1979).

Reinterpreting the geophysical data in the light of the plate tectonics concept, we can infer that during the subduction process the two plates, subducted in the north and south, advanced on the Benioff planes, reaching a considerable depth, so that the melting of the subducted oceanic crust gave rise to magmas from which the volcanics of the two island arcs proceed (Fig. 1 b). Owing to this process and to the subsequent tectonic movements, the plates sink to a still greater depth, so that their sialic part comes under the mass of ophiolites, the boundary of which is pointed out by geophysical measurements (Pl. and Fig. 1).

The asymmetric subduction varies in intensity since on the northern border of the Mureş zone the island arc and the marginal trough or basin occupy the same space which seems to have an ophiolitic basement, while on the southern border the marginal basin remains on the border of the oceanic zone and partly settles on the sialic plate in the south; the island arc lies farther inside the Mureş zone, on the ophiolitic basement. The southern island arc is marked by numerous volcanic structures covered by a barrier reef extending between the Căpîlnaş-Zam-Vorţa-Almaşu Mare localities (S a v u et al., 1981); it is pointed out very well by a geophysical negative anomaly, situated in its axial zone (Pl.). The northern volcanic island arc, situated between Pătîrş and Şoimuş-Buceava, was not marked by a geophysical anomaly probably for several reasons: firstly, because it covered the area of the northern trough, the sediments of which obscured the magnetometric effects of the volcanic rocks and, secondly, because the sialic plate of the Apuseni Mountains sinks under the Mureş zone, exceeding in depth the island arc alignment (Pl.) ; this probably effaced the gravimetric effects of the masses of volcanic rocks.

In the eastern extremity of the Mureş zone (Trascău Mountains) which is separated from the main part by the Valea Verde-Inuri Fault, probably a resumed transform fault, and a cordillera (A n d r e i et al., 1975) the depth structure is difficult to know, as only island arc volcanics and Upper Jurassic and Cretaceous sedimentary deposits crop out (Pl.). These volcanics resulted from an island arc, extending towards NNE to Turda and somewhat farther under the Transylvanian Basin.

Owing to the more intense activity of the marine basin in the eastern extremity of the Mureş zone, which continues also after the Upper Jurassic, spilites (G i u ş c ă et al., 1963) and keratophyres (N i c o l a e, B r a t o s i n, 1980) formed here during the Lower Cretaceous, making up a spilite-keratophyric association.

Because during the terminal Cretaceous the Transylvanian microplate breaks again in the Transylvanian Basin zone and this part was subducted under the Apuseni Mountains together with the north-eastern extremity of the Mureş zone, the latter was consumed to a greater extent. It will give rise to the overthrust nappe structure from the east towards the west of the Trascău Mountains and will generate new island arc magmatites as the Laramian and possibly the Neogene ones.

Ophiolitic Rocks

Ophiolitic rocks (S a v u, 1962 a), that formed during the ocean spreading stage, are better represented in the western part of the Mureş zone (Pl.), where they make up a series of basic (tholeiitic) rocks formed on the ocean floor. As this ophiolitic series was affected by tectonic movements, like other geological formations, its two complexes crop out over well-known areas (Fig. 2).

a) The upper Căzăneşti-Roşia Nouă-Toc complex (O_1). This complex shows a basaltic character ; it comprises the ophiolitic rocks situated between the Vărădia-Obîrşia diagonal line and the boundary between ophiolites and the island arc volcanics (Pl.) . From this boundary towards the east the ophiolitic rocks crop out also above the island arc volcanics only in the inliers of Visca, Luncoi and Buneşti, which are marked by geophysical positive anomalies pointed out by A n d r e i et al. (1975), and in the Glodghileşti-Furcşoara tectonic rise. Rocks from the basaltic complex (O_1) crop out also on the north-western border of the Mureş zone, at Troaş, Laleşinţ and Pătîrş.

The basaltic complex generally consists of submarine flows of aphanitic rocks represented by anamesites, basalts, hyalobasalts and variolites, seldom amygdaloidal basalts which are generally in pillow lava facies. These lavas are intercalated with rare tuff and agglomerate levels with which tachylites are associated; the latter rocks are very characteristic of the

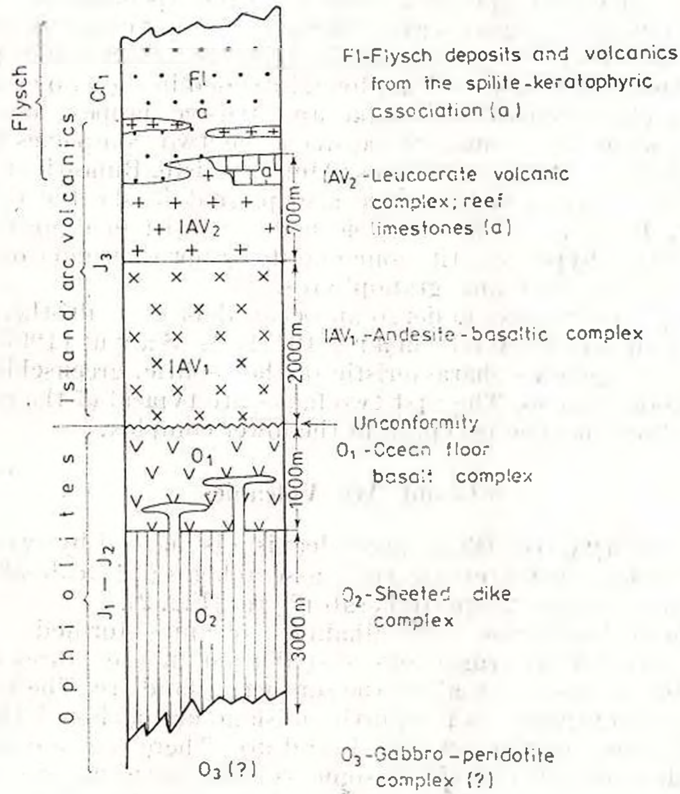

Fig. 2. — Stratigraphic column of ophiolitic rocks and island arc volcanics of the Mureș zone.

submarine tholeiitic eruptions. Some jasper intercalations are found at Dumbrăvița, while recrystallized Jurassic limestone lenses are present among the basalt flows in the Căpîlnaș-Zam-Dumești region (Savu, Nicolae, 1975). The rocks belonging to this complex have characteristic intersertal or intergranular and radial textures. In addition to the plagioclase laths, augite and magnetite, they contain always glass.

b) The lower Troaș-Bata complex (O_2). The second complex shows a doleritic character and is less widespread, covering the western part of the Mureș zone between Troaș, Julița and Zăbalț. It consists mainly of dolerites, spilites and gabbros, seldom granophyres, forming a complex of sheeted dykes trending approximately E-W and NE-SW. The rocks show ophitic or grained textures consisting of plagioclase, clinopyroxene and magnetite ± pyrite.

The doleritic dykes reach rarely also the upper complex, giving rise to basalt flows, dolerite and spilite sills or gabbro bodies, as it results from the maps 1 : 50,000 Roșia Nouă and Săvîrșin (1979).

Unlike the classic ophiolitic zones from the Apennines and Dinarids, the third complex of gabbro-peridotitic rocks does not crop out in the Mureș zone; but such rocks form here small bodies of gabbros and peridotites (S a v u, 1962 a, 1980), 0.3—4 km long, intruded in the two complexes of basic rocks. The gabbro bodies make up intrusive nappes, some of them being situated at the boundary between the two complexes (Căzănești-Ciungani, Julița) and layered dykes (Almaș, Săliște, Bunești) or composite bodies (Almaș, Cuiaș). Gabbros are also present in sheeted dykes of the Dumbrăvița-Baia zone. The gabbroic bodies consist of diopside gabbros, olivine gabbros, hyperites, titanomagnetite gabbros, pegmatoid gabbros, dolerites, microgabbros and granophyres.

The ophiolitic rocks undergo an ocean floor or hydrothermal metamorphism as meant by C o l e m a n (1977). As S a v u (1967) reported, there form parageneses characteristic of the zeolitic, greenschist and epidote amphibolite facies. The first two facies are typical of the upper complex, while the third one is typical of the lower complex.

Island Arc Volcanics

The activity of the island arc volcanism is marked by two successive phases involving characteristic rock assemblages: (1) calc-alkaline and alkaline volcanics and (2) spilite-keratophyres (Fig. 2).

1. The calc-alkaline and alkaline volcanics formed by bilateral subduction show a heterogeneous distribution in the Mureș zone. They form a complete succession along the southern island arc. The volcanics do not form rock complexes in the northern island arc, although they include all the rock types of the southern island arc. There the calc-alkaline and alkaline volcanics concentrate in some volcanic structurs of central type, buried (Pl.) under the formation with jaspers, red argillites and Upper Jurassic limestones (S a v u, 1962 b).

They are absent in the axial part of the Mureș zone, west of the Vărădia-Obîrșia line, being present only in a few outliers (Roșia Nouă, Gomile Valley etc) between the above line and the Basarabeasa-Zam boundary. East of the sinuous Basarabeasa-Zam boundary they cover wholly the Mureș zone — except for the ophiolite inliers — up to the Trascău Mountains. The calc-alkaline and alkaline volcanics make up two principal complexes (S a v u et al., 1981): (a) a lower andesite-basaltic complex and (b) an upper quartz-andesite-rhyolitic complex.

a) The lower complex (IAV_1). This complex consists mainly of pyroclastics — especially agglomerates—of porphyritic amygdaloidal basalts, basaltic andesites and, more rarely pyroxene and hornblende andesites, ankaramites, sometimes intercalated with lava flows in pillow lava facies. The agglomerates are sometimes associated with tuffs, radiolarites and red argillites. We point out the existence of the "polygenous agglomerates", a kind of "volcanic melange", which include elements broken by volcanic explosions from the ophiolitic rocks in the basement or from the reef limestones present in the volcanic structures. Within the volcanics between Cărmăzineşti and Săliştioara there occur exotic blocks of strongly deformed crystalline schists, which might represent olistoliths broken from the Transylvanian Microplate in the southern subduction zone and included within the pyroclastics of the island arc volcanism — by an obduction process. They lend to this formation the character of melange with tuffogenous cement ("coloured melange"), which does not constitute an ophiolitic melange, but a pyroclastic olistostrome. There are also olistoliths of gabbros, peridotites and limestones (Godineşti-Vălişoara).

Levels of amphibole andesites or red dacites showing a perlitic texture are also found : they appear even at the base of the complex (Visca), which indicates the recurrent character of the island arc volcanism and the discontinuity between the ophiolitic series of the basement and the island arc volcanics (S a v u et al., 1981). The thickness of this complex exceeds 2 km in the Visca-Dealu Mare region.

Between Almaş-Sălişte, Almăşel and Roşia Nouă a system of andesite and basalt dykes with a NW-SE trend is known, which crosses the ophiolitic series.

b) The upper complex (IAV_2). In the axial zone of the southern island arc the lower complex is followed by a complex consisting of leucocrate rock pyroclastics, such as amphibole quartz andesites, dacites and rhyolites, some of them showing a perlitic texture. It finishes the island arc volcanism in most of the Mureş zone at the end of the Upper Jurassic, possibly the beginning of the Neocomian (?). Such pyroclastics also occur in the upper part of the volcanic structures of the northern island arc, buried at the base of the Upper Jurassic-Lower Cretaceous flysch of the northern trough (Pl.). The acid volcanics underlie or include Stramberg Limestones, pointing to the age of these eruptions. The limestones overlie directly the lower complex in the Păroasa-Măgureaua Vaţei zone. The acid rocks are present also as veins or dykes crossing the lower complex of the island arc volcanics or the ophiolites cropping out west of the Basarabeasa-Zam boundary, up to Juliţa. At the level of the upper complex, a formation consisting of tuffs, red and grey argillites and seldom radiolarites develops in the Vorţa synclinorium (S a v u, N i c o l a e, 1975).

The calc-alkaline rocks are associated with rocks showing an alkaline character, represented by limburgites, oligophyres, trachyandesites, orthophyres (paleotrachytes) occurring as veins and more rarely as pyroclastics or lava flows (S a v u, 1962 b).

The island arc volcanics were emplaced by an intense intermittent and recurrent submarine volcanic activity of some volcanoes of central type, aligned in the two island arcs or isolated, and on the cones of which there form reefs (S a v u, 1962 b). The rocks show a porphyritic or glomeroporphyritic, frequently amygdaloidal texture, with pilotaxitic or fluidal groundmass, typical of the volcanic rocks.

Since theyformed in a marine environment, the island arc volcanics also underwent a weak submarine regional metamorphism in the zeolitic facies, being sometimes (Vorța, Almaș Săliște) overlain by a local hydrometasomatic metamorphism.

2. The spilite-keratophyric rocks are found as flows and veins in the Albian sedimentary deposits of the Trascău Mountains. They are represented by spilites and quartz keratophyres interbedded with rhyolites and paleotrachytes. Irrespective of the process that led to their formation — magmatic differentiation or metasomatosis—their petrochemical and geochemical characteristics differ from those of the above-mentioned rocks.

Conclusions

The following conclusions can be drawn from this paper :

The Mureș zone functioned initially as an ocean floor zone, becoming subsequently an island arc zone as a result of a bilateral subduction process.

The first spreading stage is marked by the formation of ophiolites, while in the second stage, island arc volcanics formed.

The two rock series formed in these stages differ from each other in every respect.

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[3] The dynamics of this model has been given by a system of two convergent cells rather similar to those imaginated by Ciocârdel and Socolescu (*Rev. Roum. Géol., Géophys., Géograf., Série Géophysique*, 16, 2, 1972, p. 179–192, București) in the depth of this region.

[4] If we consider that the spreading of the Mureș Ocean progressed with only 2 cm every year, we arrive to the conclusion that the width of this ocean zone was at least of 400 km.

REFERENCES

Andrei J., Calotă C., Scurtu F. (1975) Considérations structurales sur le sillon ophiolitique des Monts Métallifères à l'aide des données géophysiques et de certains éléments quantitatifs obtenus grace au modelage. *Rev. Roum. Géol., Géophys., Géogr., Géophysique*, 19, p. 101–111, București.

Bleahu M. (1974) Zone de subducție in Carpații românești. *D. S. Inst. Geol.*, LX/5, p. 6–25, București.

Borcoș M., Andrei J., Lupu M., Berbeleac I., Cristescu T. (1979) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Cioflica G., Lupu M., Nicolae I., Vlad Ș. (1980) Alpine Ophiolites of Romania. Tectonic setting, magmatism and metallogenesis. *An. Inst. Geol. Geofiz.*, LVI, p. 79–96, București.

Coleman R. G. (1977) Ophiolites — Ancient oceanic lithosphere? Springer-Verlag, Berlin, 229 p.

Giușcă D., Cioflica G., Savu H. (1963) Vulcanismul mesozoic din masivul Drocea (Munții Apuseni). *Asoc. Geol. Carp.-Balc., (Congr. V, 1961)*, II, p. 31–44, București.

Herz N., Savu H. (1974) Plate tectonics history of Romania. *G. S. A. Bul.*, 85, p. 1329–1440, Boulder, Colo.

Ianovici V., Borcoș M., Bleahu M., Patrulius D., Lupu M., Dimitrescu R., Savu H. (1976) Geologia Munților Apuseni. Ed. Acad. R. S. România, 631 p., București.

Macovei G., Atanasiu I. (1933) L'évolution géologique de la Roumanie — Crétacé *An. Inst. Géol.*, XVI, București.

Nicolae I., Bratosin I. (1980) Petrochemical investigations of the Mesozoic spilite and keratophyre rocks from the Trascău and north-eastern Metaliferi Mountains (Apuseni Mountains). *Rev. roum. Géol., Géophys., Géogr., Géologie*, 24, p. 99—114, București.

Papiu C. V. (1953) Cercetări geologice în masivul Drocea (Munții Apuseni). *Acad. R.P.R., Bul. științif.*, V, 1, p. 107—213, București.

Rădulescu D. P., Săndulescu M. (1973) The Plate-Tectonics concept and the geological structure of the Carpathians. *Tectonophysics*, 16, p. 155 — 161, Amsterdam.

Savu H. (1962 a) Corpul gabbroic de la Almășel și contribuții la cunoașterea chimismului și petrogenezei ofiolitelor din masivul Drocea. *An. Com. Geol.*, XXXII, p. 211 – 248, București.

— (1962 b) Chimismul vulcanitelor jurasice superioare-cretacice inferioare din Munții Drocea. *D. S. Inst. Geol.*, XLVII (1959 – 1960), p. 199—220, București.

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

— (1980) Genesis of the Alpine cycle ophiolites from Romania and their associated calc-alkaline and alkaline volcanics. *An. Inst. Géol., Géophys.*, LVI, p. 55—77, București.

— Udrescu C. (1973) Geotectonic evolution of the Mureș Zone (Apuseni Mountains-Romania) and the distribution of trace elements in its ophiolitic rocks. *Intern. Symp. on Volc. and Assoc. Metallog. IAVCEI*, Abstracts, p. 182, Bucharest.

— Nicolae I. (1975) Evolution of ophiolitic volcanism in the Vorța Region and its position in the Mureș zone tectogenesis (Apuseni Mountains). *D. S. Inst. Geol., Geofiz.*, LXI/5, (1973—1974), p. 179—196, București.

— Berbeleac I., Călinescu E., Florescu R., Zămircă A. (1978) Structure and Origin of Bunești Gabbroic Body (Metaliferi Mountains). *D. S. Inst. Geol., Geofiz.*, LXIV/1, (1976—1977), p. 173—191, București.

— Berbeleac I., Udrescu C., Neacșu V., Nacu D. (in press) Caracteristicile petrologice și geochimice ale vulcanitelor de arc insular jurasic — superioare din regiunea Almaș Săliște-Godinești-Zam (zona Mureș). *D. S. Inst. Geol., Geofiz.*, LXVIII/1, București.

Stille H. (1953) Der geotektonische Werdegang des Karpaten. *Beih. für Geol. Jb.* 8, 239 p. Hanover.

QUESTIONS

V. Széky-Fux: 1. What sort of ultramafic rocks did you find?

2. What is the chemical composition of these rocks?

Answer: 1. Within the complex of ocean floor basalts there are several small bodies of layered ultramafic rocks in the base of which there are peridotites overlain by troctolites and olivine gabbros.

2. The chemical composition of the ultramafic rocks *s. str.* corresponds to a peridotitic magma in the acceptation given by Niggli. The mineral paragenesis is as follows: olivine-clinopyroxene-brown hornblende-biotite (phlogopite) ± chromite-magnetite.

D. Hovorka: 1. According to the previously published papers of Dr. H. Savu and other authors the products of volcanism of Mesozoic age have been divided into three stages.

What is the correlation between the three mentioned stages of the volcanic activity and the newly defined ophiolitic complex and island arc volcanics?

2. The microscopic study of the serpentinite bodies does not give us sufficient data to distinguish among the serpentine-group minerals. Are there any new results on DTA, roentgenographic and electron microscopic study of the serpentine-group minerals of the ultramafic bodies in the Drocea Mountains?

3. The strongly expressed Ti tendency of gabbroic rocks and ultramafics is typical of the whole Carpathian Arc (Szarvasko, Apuseni Mountains). Are there any new correlation study data?

Answer: 1. The ocean floor rocks make up an ophiolitic series corresponding to the magmatites of the first evolution stage from the old classification (J_1— J_2). Concerning the formation time, the island arc volcanics correspond to the stages 2 and 3 (J_3— Cr_1) of the same classification.

2. There are no additional data for determining the serpentine character, which sometimes replaces olivine from peridotites and troctolites. These rocks are usually quite fresh.

3. There are no new correlation data. This element shows a high concentration in the magnetite gabbros, together with Fe and V, this being a tendency characteristic of the differentiation of the tholeiitic magmas.

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LE VOLCANISME MIOCÈNE AFFLEURANT ET RECOUVERT DU NORD-EST DE LA HONGRIE [1]

PAR

VILMA SZÉKY-FUX [2], P. GYARMATI [2], K. BALOGH [3], Z. PÉCSKAY [3]

Introduction

Le territoire étudié est compris entre la frontière est de la Hongrie, la Tisza jusqu'à la ville Csongrád et la Körös sebes jusqu'à la frontière roumaine. Ce sont les Monts de Tokaj qui se situent au-delà de la ligne de la Tisza ; là le volcanisme de profondeur est bien développé. Mais au point de vue géologique, à la frontière nord-est, la dacite de Tarpa,

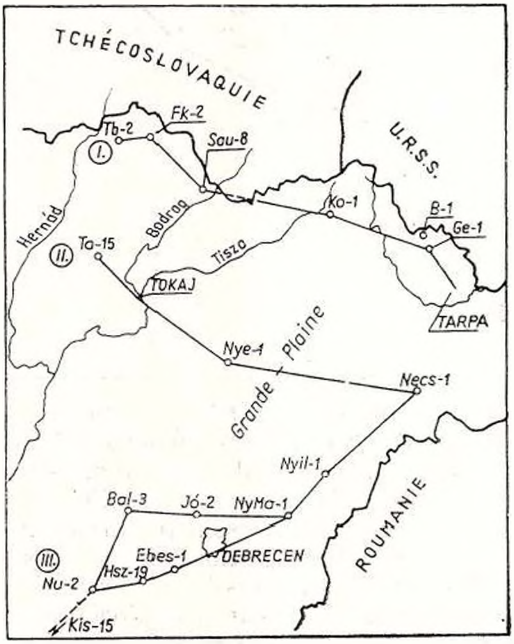

Fig. 1. — La place des sondages et des profils (I – III) dans la partie nord-est de la Hongrie.

[1] Note présentée au 12ème Congrès de l'Association Géologique Carpatho-Balkanique, 8—13 septembre 1981, Bucarest, Roumanie.

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le tuf rhyolitique de Barabás de l'autre côté de la frontière, les Monts de Vihorlát, aussi bien que les territoires de Beregova (Beregszász) et de Vinogradov (Nagyszöllös) sont aussi en relation étroite avec les niveaux profonds de la région transtibiscinne (fig. 1).

A partir de la carte géologique non couverte et des études détaillées faites par Szepesházy (1973) devient évidente une ligne structurale

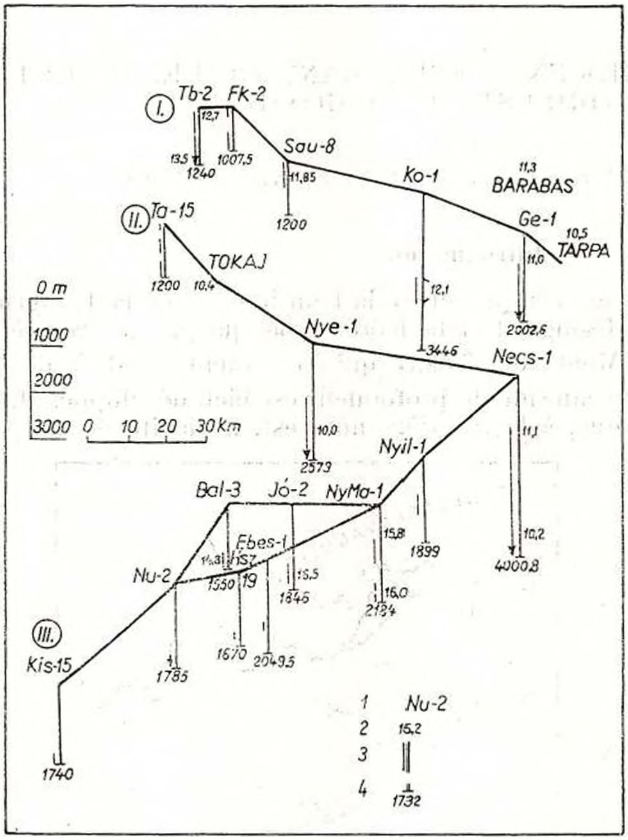

Fig. 2. — Les volcanites miocènes tranchées par les sondages dans la partie nord-est de la Hongrie.

1, le signe et le nombre des sondages ; 2, l'âge en millions années ; 3, les volcanites tertiaires tranchées par les sondages ; 4, la profondeur du sondage.

nette — à orientation SO-NE — qui sépare (suivant les caractères du substratum des formations néogènes) la région transtibiscinne en deux grandes parties. Au sud de cette ligne structurale, les formations pannoniennes reposent directement, par l'intermédiaire d'un mince complexe miocène, sur les anciens schistes cristallins. Au nord de la même ligne, et presque sans interruption les schistes cristallins sont recouverts de sédiments triasiques, jurassiques, crétacés et paléogènes (Flysch interne) et de sédiments et volcanites miocènes souvent à une profondeur bien considérable. Les assises pannoniennes ne surmontent que ceux-ci.

Parmi les nombreux sondages — dont les successions stratigraphiques ont été étudiées par nous — nous avons choisi ceux qui ont traversé les volcanites en quantité considérable et qui sont en relation avec les Monts de Tokaj et les territoires volcaniques situés au-delà de la frontière du pays.

Nous avons indiqué dans la figure 1 les lignes directrices des profils des sondages présentés dans la figure 2.

Nous avons représenté sur la figure 2 les sondages en bloc-diagrammes, en maintenant la position géographique exacte et en indiquant en même temps les échelles. Pour chaque sondage — à ligne épaisse — on a désigné l'extension des volcanites tertiaires en profondeur et leur âge radiogénique à K/Ar. Les chiffres indiquent les cotes finales des sondages.

Les faciès du substratum — intercepté par les sondages — du soi-disant territoire du Flysch sont extrêmement variables, mais il est constitué le plus souvent par des formations éocènes ou, assez souvent, crétacées (par exemple Flysch ou diabase crétacée dans le sondage d'Ebes 1) ou bien jurassiques et triasiques sur le territoire de Hajdúszoboszló. Le sondage de Komoró a été arrêté à 3500 m de profondeur,dans le schiste séricíteux paléozoïque.

Il y a de très grandes différences entre les sondages eu égard aux faciès volcaniques. De nombreux sondages — surtout sur le territoire du Flysch — les laves et les tufs volcaniques miocènes ont une épaisseur de plusieurs centaines de mètres (Nyirmártonfalva, Nyirlugos). La base du complexe volcanique n'a pas été traversée dans bien des sondages, même après avoir atteint 1000 m (bordure ouest des Monts de Tokaj) ou davantage. Le sondage de Gelénes 1 a été arrêté à 2002 m dans les formations rhyolitiques, perlitiques, à 2579 m à Nyiregyháza dans les formations rhyolitiques et à 4000, 8 m dans l'andésite miocène à Nagyecsed.

Les volcanites miocènes au NE de la Hongrie

Profil I. (Monts de Tokaj). Nous en avons choisi les sondages de Telkibánya 2 (Tb 2), Füzérkajata 2 (Fk 2), Sátoraljaújhely 7 (Sau 7) et Tállya (Ta 15). Dans le bloc-diagramme, la colline Kopasz-hegy de Tokaj, la colline Tarpai-hegy et la carrière de Barabás montrent les formations en affleurement.

Tout comme le sondage de Tállya 15 (Ta 15) le sondage profond de Telkibánya 2 n'a pas atteint le substratum et il a été arrêté à 1240 m de profondeur dans la propylite andésitogénique badénienne (tortonienne) de 13,3 $\pm$ 0,8 M.A. L'épaisseur du sédiment dans le sondage est de presque 40 m, alors que celle des formations volcaniques de 1200 m, représentant surtout l'andésite, la propylite andésitogénique, la trachyte pseudo-potassique et en quantité réduite la propylite dacitogénique et tufogénique rhyolitique.

Il est connu que dans le substratum des volcanites miocènes des Monts de Tokaj est développée, sur le versant ouest de ces monts (d'après les informations géologiques-sismiques) une fosse volcano-tectonique de 3—4 km de profondeur. C'est pour cela que ni le sondage de Telkibánya 2 (Tb 2) ni celui de Tállya 15 (Ta 15) n'ont traversé entièrement les volcanites badéniennes ("tortoniennes"). Vers le nord et l'est, le substratum monte progressivement.

Après avoir traversé le complexe rhyolitique, puis andésitique, tufitique et dacitique-badénien à sarmatien, le sondage de Füzérkajata 2 (Fk 2) a été arrêté à 1007,6 m dans le porphyroïde du type des Monts Métallifères de Szepes-Gömör, qui appartient au socle cristallin (P a n t ó, 1966).

Le sondage de Sátoraljaújhely 7 (Sau 7) est arrêté à 1200 m dans l'assise permo-carbonifère, après avoir traversé les volcanites miocènes et les dépôts mésozoïques.

La présence dans le sondage de Sárospatak 10 (Sp 10) du basalte (phase terminale du volcanisme des Monts de Tokaj) dont l'âge isotopique K/Ar est de 9,4±0,5 à 10,9±1,0 M.A. est une information importante. Vu le caractère pétrochimique, il s'agit d'un basalte à chaux alcaline typique, tandis que suivant son âge isotopique il s'est formé durant le Pannonien. Il ne représente pas le volcanisme final alpin, mais le terme final du volcanisme subséquent.

A l'est de la bordure des Monts de Tokaj, c'est-à-dire à l'est de Sátoraljáujhely, le substratum est de nouveau affaissé dans des niveaux plus profonds et dans la sois-disante "fosse de Záhony" — dans le sondage de Komoró — il n'a été atteint qu'à une profondeur de 2859 m. D'ailleurs, le sondage de Komoró a fait une grande sensation, parce que, jusqu'à présent, aucun sondage — profond au N de Debrecen — n'a pu traverser le complexe volcanique miocène à grande épaisseur. C'est pourquoi le sondage de Hajduböszörmény 1 a été arrêté à 1514 m dans les tufs et les agglomérats gris volcaniques et le sondage de Nyiregyháza 1 (Nye 1) à 2579 m dans la rhyolite miocène.

Par contre, après avoir traversé le complexe volcanique pannonien et miocène (par endroits à intercalations sédimentaires) et les formations triasiques, le sondage de Komoró 1 (Ko 1) s'est arrêté dans le schiste cristallin.

Le complexe miocène est épais et variable. Il est constitué surtout d'andésite, propylite andésitogénique, dacite, propylite dacitogénique et tuf dacitique sarmatiens, ensuite d'alternance de couches gris foncé, siliceuses tufacées aleuritique et tuffitique. L'âge isotopique K/Ar de l'andésite situé entre 1833,7 et 1833,8 m de profondeur est de 12,1±0,4 M.A. La dacite située à 2395 m de profondeur est contemporaine à l'andésite susmentionnée.

Au-dessous des couches triasiques (qui représentent le substratum du Miocène) le sondage a traversé des schistes quartzeux à graphite et séricite paléozoïque, respectivement des gneiss, entre 3288,6 m et 3288,5 m de profondeur.

Suivant les ressemblances pétrologiques et les âges isotopiques, on pourrait corréler les roches métamorphiques avec celles du massif cristallin de Cerna Hora, alors que l'andésite du sondage pourrait être rattachée au volcanisme andésitique des Monts de Vihorlát (Ďurica et al., 1978).

Profil II. Dans le profil II nous avons indiqué les formations en affleurement et les sondages profonds suivants.

Selon les mesurages chronologiques à K/Ar exécutés sur plusieurs échantillons de surface, on a établi l'âge de la dacite à pyroxène de la colline de Kopasz-hegy de Tokaj à 10,5±0,5 M.A. (Lengyel, 1924; Gyarmati, 1977). Elle est la plus récente, représentant le volcanisme intermédiaire des Monts de Tokaj, s'étant consolidée pendant le Pannonien.

Comme nous l'avons déjà mentionné, ni le sondage de recherche d'hydrocarbures de Nyiregyháza 1 (Nye 1) n'a atteint la base du complexe volcanique miocène.

Le complexe miocène est en partie marin (de 980 m à 1150 m), en partie continental (de 1150 m à 2579 m). Les formations marines sont représentées d'une manière prépondérante par des marnes argileuses et par des grès tuffitiques. Les formations continentales sont constituées pour la plupart des volcanites. Les formations rhyolitiques y prédominent. L'âge isotopique K/Ar est de 10,8±0,6 M.A., ces formations étant les plus récents niveaux de tuf rhyolitique supérieur.

L'andésite à pyroxènes (propylite andésitogénique) n'est apparue que dans le sondage de Nyiregyháza 5. Son âge est récent, à savoir de 10,0±0,6 M.A., donc correspondant à celui de la colline de Kopasz-hegy de Tokaj.

Nagyecsed 1 (Necs 1) est l'un des plus importants sondages. Les volcanites miocènes à minces intercalations sédimentaires (surtout à marnes calcaires) développent une épaisseur considérable. Entre 1074 m et 4000,8 m le sondage a rencontré peu de tuf rhyolitique et de l'andésite en grande quantité. Au-dessous de 2000 m, la propylitisation de l'andésite est caractéristique.

L'âge radiogénique K/Ar de l'andésite fraîche est de 11,1±0,7 M.A., et comme d'habitude celui de l'andésite propylitisée est plus réduit, toutes les deux étant pannoniennes. Le complexe surtout andésitique (épais de presque 3000 m) appartient aux zones volcaniques de Csop (Csap) et de Vinogradov (Nagyszöllös), si l'on prend en considération le caractère pétrographique et l'âge isotopique.

Profil III. Les sondages du profil III ont été approfondis au soi-disant territoire du Flysch crétacé et paléogène. A partir de Kisújszállás (à travers Debrecen) le profil arrive jusqu'à Nyirlugos, respectivement à Nagyecsed (fig. 1, 2). Vis-à-vis des sondages précédents la quantité des volcanites diminue.

Après avoir traversé les assises pannoniennes inférieures, le sondage de Kisújszállás 15 (Kis 15) a atteint à la profondeur de 1612 m les volcanites miocènes. Il a aussi traversé le tuf rhyolitique, puis le tuf andésitique jusqu'à la cote finale de 1767 m.

Le sondage de Nádudvar 2 (Nu 2) a atteint le complexe miocène de 1707 m. Il a traversé des sédiments réduits de tufs rhyolitiques et puis entre 1781 et 1785 m — jusqu'à la cote finale — la chloro-andésite.

Le sondage de Balmazújváros 3 (Bal 3) a traversé entre 1216 et 1219 m les formations dont l'âge radiogénique est de 14,8 ±1,0 M.A.

Nous ne connaissons pas exactement la succession stratigraphique du sondage de Hajdúszoboszló 19 (Hsz 19). Entre 1390 m et 1470 m le sondage a traversé un tuf rhyolitique cristallisé sous influence thermique.

Le sondage d'Ebes 1 (Eb 1) a traversé le calcaire tufeux sarmatien (entre 1427 m et 1506 m), puis le tuf rhyolitique badénien ("tortonien"). A partir de 1506 m jusqu'à la cote finale de 1830 m le forage passe dans la diabase crétacée.

Le sondage de Józsa 2 (Jó 2) a atteint le Miocène à 1238 m. Le calcaire tufacé et les tuffites caractérisent le Sarmatien (1238 m et 1284 m). Au-dessous des couches sarmatiennes entre 1284 m et 1720 m le calcaire tufacé admet des intercalations de tuf rhyolitique et des rhyo-dacites. L'âge radiogénique K/Ar (d'après la biotite) est de 16,5±0,9 M.A. correspondant à l'étage karpatien (Helvétien). Au-dessous du Miocène, le sondage passe dans les couches paléogènes (Flysch).

L'âge du tuf dacitique (entre 932 m et 935 m) dans le sondage de Nyirmártonfalva 1 (Ná 1) est de 17,1±0,5 M.A., ce qui correspond à l'étage karpatien . Jusqu'à la cote finale de 2184 m, le sondage a traversé le Paléogène qui montre une épaisseur de plus de 1000 m. Nous y avons distingué des faciès variables, avec des filons de rhyolites à la partie inférieure.

Le Sarmatien du sondage de Nyirlugos 1 (Nyil 1) d'épaisseur réduite est caractérisé par une alternance de calcaires fossilifères, tufs rhyolitiques et tuffites (846 m à 869 m) ; le Badénien (869 m à 1194 m) y présente une alternance de tuf rhyolitique, tuf rhyolitique bentonitisé et rhyolite. Jusqu'à la cote finale de 1847 m, le Paléogène renferme des intercalations de sables et grès. Le complexe rhyolitique amincit, progressivement vers Nyirlugos. Selon les données bibliographiques le sondage approfondi à Carei (Nagykároly) — au-delà de la frontière — ne renferme pas des volcanites.

Dans le territoire entre Barabás et Tarpa, situé le long de la frontière soviétique, K u l c s á r (1968) a décrit le volcanisme et ensuite a synthétisé en 1976, vu les informations sur l'Ukraine subcarpathique fournies par L a z a r e n k o (1963, 1968) et par M e r l i c s et S z p i t k o v s z k a j a (1974).

Si on considère l'âge radiogénique à K/Ar (10,5±0,3 M.A.) et la position stratigraphique, la dacite de Tarpa est également récente. En raison de son âge, elle est engendrée en même temps que la colline de Kopasz-hegy de Tokaj à laquelle elle ressemble autant par sa position géologique que sa composition pétrochimique.

La petite partie sud de la colline de Barabás — présente des volcanites rhyolitiques en affleurement, explorées dans une carrière. On peut l'attribuer au Sarmatien supérieur, d'après sa position géologique et son âge isotopique K/Ar (11,3±0,6 M.A.). Le tuf rhyolitique écoulé du sondage de recherche de Barabás 1 (K u l c s á r, 1976) — profond de 100 m — est du même âge (11,2±0,6 M.A.). Sur toute l'épaisseur du sondage, le tuf rhyolitique écoulé est caractérisé par la riche distribution de pyrite et marcasite aussi bien que par une forte altération en montmorillonite.

Le sondage de recherche de Beregdaróc 3 (Bd 3) a exploré une formation d'un même âge que le précédant : le tuf rhyolitique et les couches tuffitiques et tufacées à sédiments fossilifères. A la cote finale de 500 m, le forage passe dans le tuf rhyolitique à grains variés, à pierre ponce, souvent richement pyriteux.

Le sondage de Gelénes 1 (Ge 1) a exploré le tuf rhyolitique à caractère semblable. D'après la détermination à K/Ar (à partir de la biotite) son âge est de 11±0,6 M. A., ce qui le situe à la limite entre le Sarmatien et le Pannonien.

Conclusions

Dans la fosse volcano-tectonique longeant le flanc ouest des Motns de Tokaj (Telkibánya 2, Tállya 15), au-dessous des complexes volcaniques sarmatiens — 1000 m — et badéniens tortoniens — 1000 m — le substratum se situe à une profondeur de 2000 m, même plus grande. Vers le nord-est, respectivement sud-est, l'épaisseur du complexe volcanique miocène diminue, alors que le substratum arrive en position de plus en plus élevée (Füzérkajata 2, Sátoraljaújhely 7, Sárospatak 10).

Mais on a constaté que vers la région de Nyirség, le substratum descend progressivement (Nyiregyháza 1) vers l'est, vers Komoró et aussi vers la région de Bodrogköz, respectivement de la soi-disante fosse de Záhony.

Sur la carte à isohypses de la base des formations miocènes est bien illustrée la structure caractéristique du substratum qui comprend des zones à direction NE-SO. L'étude des volcanites affine considérablement cette tectonique. Les volcanites badéniennes (tortoniennes) se trouvent à la surface à Aknaszlatina. Le sondage de Gelénes (situé près de la frontière) a été arrêté au-dessous de 2000 m dans les volcanites rhyolitiques et perlitiques badéniennes ("tortoniennes") (P a n t ó, 1976). Dans le sondage de Nagyecsed apparaissent les volcanites sarmatiennes même dans la profondeur de 4000 m. On a relevé que les failles en escalier, presque parallèles à la frontière soviétique (ligne de Szamos = Samech) montrent une direction tectonique carpathique, orientée NO-SE.

Outre la détermination des conditions tectoniques de la Grande Plaine de Hongrie, respectivement au NE de la région transtibiscine nous avons aussi pu distinguer 4 types de faciès :

1. Le caractère pétrographique et l'âge isotopique (13,8 à 11,1 M.A.) de la soi-disante andésite à pyroxène acide rattache l'alignement principal des Monts de Tokaj à la chaîne volcanique andésitique de la zone carpathique intérieure.

Les informations sur la composition et l'âge de l'andésite à pyroxènes concordent avec les informations des chercheurs slovaques sur les Monts de Vihorlát (12,3 à 11,6 M.A.) et sur les andésites (12,3 à 11,6 M.A.) traversées par le sondage de Komoró 1.

La phase intermédiaire la plus récente des Monts de Tokaj (dacite, andésite lamelleuse à pyroxène de 11 à 10,3 M.A.) s'est déroulée pendant le Pannonien. Cette phase se rattache à la zone volcanique la plus récente qui tout comme la colline de Fekete-hegy de Vinogradov (Nagyszöllös) est l'annexe du lignement volcanique couvert de Csap et Naghyszöllös (l'âge de la dacite de Tarpa est de 10,5 et celui de Fekete-hegy de 10,2 M.A.). On peut y attribuer aussi le complexe andésitique du sondage de Nagyecsed, épais de 3000 m environ et l'âge de 11,1 à 10,2 M.A.

2. Les volcanites acides affleurant ou recouvertes du territoire de Barabás — tufs dacitiques et rhyolitiques — présentent les annexes de la chaîne volcanique la plus interne qui alternent avec les complexes de sédiments ; ils ont été attribués aux étages badénien et sarmatien. Leur grandes effusions dénotent des caractères des tufs-laves et des tufs écoulés.

L'épais complexe rhyolitique — exploré dans le sondage de Nyiregyháza 1 (Nye 1) — y appartient aussi. Le faciès et l'âge le rattachent étroitement aux rhyolites de Beregova (Beregszász) et de Bégány. On trouve de tels tufs sarmatiens dans le bassin de Kosice (Kassa) et dans l'île de Zemplin.

3. Les volcanites formées sur le territoire du Flysch interne mettent en évidence des faciès spéciaux. Les tufs sarmatiens n'atteignent que

quelques dizaines de mètres d'épaisseur, toutefois les tufs et les laves acides badéniens — alternant avec des sédiments — sont considérablement épais. Les cotes finales des sondages se trouvent fréquemment dans les volcanites miocènes et le Flysch, plus rarement dans la diabase crétacée.

4. Au sud du territoire du Flysch, l'épaisseur des volcanites miocènes diminue de plus en plus ; elles sont surtout représentées par des tufs et des tuffites alternant avec des sédiments, et par des conglomérats à éléments de schistes cristallins à leur base (par exemple les sondages de Biharnagybajom). Au sud de la rivière Körös, les couches pannoniennes reposent immédiatement sur le socle cristallin.

Les dimensions du volcanisme miocène des niveaux profonds se manifestent bien, si nous les comparons à celles de nos monts volcaniques à la surface. Les plus hauts monts, ceux de Mátra, n'atteignent qu'une altitude de 1000 m. Les sondages de Gelénes et de Komoró ont traversé les volcanites miocènes sur une épaisseur de 2000 m et celui de Nagyecsed de 3000 m. C'est ainsi qu'on peut supposer que la plupart des volcanites se trouvent en profondeur.

Suivant le volcanisme en affleurement, la durée du volcanisme recouvert est de 17,1 M.A. (sondage de Nyirmártonfalva 1 tuf dacitique de $9{,}4 \pm 0{,}5$ M.A.) (sondage de Sárospatak 10 basalte).

L'étude des volcanites est aussi importante du point de vue des hydrocarbures. Dans le bassin de la Slovaquie orientale, les géologues slovaques ont trouvé des hydrocarbures dans le tuf rhyolitique poreux. Les indications de minerais sont particulièrement intéressantes. Dans le sondage de Barabás 1 (100 m) la série du tuf rhyolitique est sur toute son épaisseur pyriteuse et marcasiteuse. Nous avons trouvé des disséminations de pyrrhotine et pyrite dans le sondage de Beregdaróc 3 (500 m) ; entre 400 et 500 m le tuf rhyolitique sarmatien est aussi richement pyriteux. Le tuf rhyolitique du sondage de Hajdúböszörmény 1 est jusqu'à la fin fortement pyriteux. Le complexe rhyolitique du sondage de Nyiregyháza 1 (Nye 1) est caractérisé par altération en andulaire, séricitisation, carbonatisation et dissémination de pyrite.

La propylitisation de l'andésite est fréquente dans de nombreux sondages. Au-dessous de 900 m, le sondage de Telkibánya 2 a traversé un filon de minerai sulfureux à épaisseur considérable.

Il en résulte que ce volcanisme situé en profondeur présente un grand intérêt tant au point de vue de réserves de hydrocarbures que de l'évolution des ressources métallogéniques du pays.

BIBLIOGRAPHIE

Ďurica D. et al. (1978) Sequence of volcanic events in eastern Slovakia in the light of recent radiometric age determinations. Vest. Úst. ust. geol. 53, p. 75–88.

Gyarmati P. (1977) A Tokaji-hegység intermedier vulkanizmusa. Műszaki Könyvkiadó, Budapest.

Kulcsár L. (1968) A magyar-szovjet határmenti vulkánosság a legujabb szovjet és hazai kutatások tükrében. *Acta Geogr. Debr.*, XIV/VII, p. 143–160.

Pantó G. (1966) In "Magyarázó Magyarország 200 000-es földtani térképsorozatához, M-34-XXXIV. Sátoraljaújhely " MÁFI, Budapest.

Széky-Fux V. (1970) Telkibánya érceesedése és kárpáti kapcsolatai. Akadémiai Kiadó, Budapest.

— Kulcsár L. (1977) A Beregdaróc /B-3/ furás komplex földtani feldolgozása. *Kutatási jelentés.* MÁFI Adattár.

— Balogh K., Szakál S. (1980) The age and duration of the intermediate and basic volcanism in the Tokaj Mountains, North-East Hungary, with respect to K/Ar datings. *ATOMKI Közl.*, 22, p. 191—201.

Szepesházy K. (1973) A Tiszántúl északnyugati részének felsőkréta és paleogén korú képződményei. Akadémai Kiadó, Budapest.

Лазаренко Е. и др. (1968) Металлогения Закарпатья. Лвов Унив.

Мерлич В. Б., Спитковская С. М. (1974) Глубинные разломи, неогеновый магматизм и оруденение Закарпатья. Лвов.

QUESTIONS

T. P. Ghiţulescu : 1. Ma première question concerne la succession dans le magmatisme néogène : si vous avez remarqué une succession définie, en partant du faciès le plus acide vers le plus basique?

2. La deuxième question concerne la structure des magmatites, spécialement : si vous avez remarqué de structure enracinée, par exemple cheminée volcanique, sous-volcans ou bien enclos?

3. Troisième question : quelle est votre opinion concernant le processus de propylitisation ; il y a un processus régional ou local en liaison avec la hydrothermalisation et la minéralisation?

Réponse : 1. Le magmatisme néogène de la Hongrie commence en même temps avec l'activité des volcanites acides, alterne, mais indépendamment de l'activité des volcanites andésitiques. Pour ce qui est des andésites, il y a une différenciation en direction basique. A la fin du volcanisme andésitique, apparaissent des andésites basiques et quelquefois des basaltes.

2. Les grandes effusions des volcanites acides montrent des caractères des tufs-laves, des tufs écoulés. Les volcanites intermédiaires et les andésites ont les caractères des sous-volcans et des effusions sans pyroclastites.

3. Selon mon opinion la propylitisation est un processus régional sans être en connexion avec la minéralisation.

T. Wieser : Est-il possible de faire une corrélation exacte des niveaux des tufs à grand développement, suivant les données minéralogiques (minéraux accessoires) et radiométriques dans les exemples hongrois?

Réponse : Oui, il est possible de faire une corrélation entre les tufs rhyolitiques d'après les données minéralogiques, pétrographiques, stratigraphiques et radiométriques, dans ce territoire de la région transtibiscinne de Hongrie.

I. Mărdărescu : L'âge absolu déterminé sur les tufs rhyolitiques a été obtenu sur des roches fraiches?

Les phénomènes d'altération hydrothermale, notamment l'adulairisation, sont liés aux plans de circulation? Connaissez-vous les sources de ces solutions?

Réponse: L'altération en adulaire est en corrélation avec des lignes structurales dans les Monts de Tokaj, mais dans les sondages nous avons trouvé l'adulairisation sur des épaisseurs considérables; les sources de ces solutions hydrothermales nous sont moins connues.

LEUCITE-BEARING ROCKS OF SERBIA AND MACEDONIA (YUGOSLAVIA)[1]

BY

MILESA TERZIĆ [2]

During the Upper Cretaceous and the Tertiary there occurred in Serbia and Macedonia effusions of large masses of calc-alkaline volcanics of different composition and facies : andesites, dacites, quartzlatites, latites with a small amount of trachytes and rhyolites. Typical basalts are absent. In large volcanogenic complexes of these rocks there also occur alkalic volcanic rocks (Sorensen's definition, 1974), sometimes followed by alkaline basalts and shoshonitic basalts.

Alkaline volcanic rocks are not widely spread if compared with other Tertiary volcanics.

The main zones of appearance of the alkaline volcanic rocks are controlled by large tectonic structures ; the Vardar zone, the Inner Dinarides, the Serbian-Macedonian massif and the Carpathian-Balkan Arc.

The alkalic volcanics in Yugoslavia are : leucite-bearing rocks and nepheline rocks. The former are far more widely spread than the latter.

Some of the leucitic rocks discussed in this paper were first described by Žujović (1920), Kojić (1926), Lacroix (1926), Tomić (1929) and Tućan (1931). New petrological and petrochemical data were established by Ristić (1959, 1961, 1963) and the present author (1961—76), while the geological ones were established by geologists who have drawn the Basic Geological Map of Yugoslavia.

Petrology of Leucite-Bearing Rocks

Leucite rocks are the most frequent in Macedonia-Vardar zone, and the least frequent in the Carpathian-Balkan Arc. Localities of Yugoslavia where leucite rocks occur are listed in Table, and are presented per geotectonic units (Fig. 1).

Leucite rocks are the products of separate phases of the volcanic activity in the Tertiary and not everywhere the youngest volcanics, as was previously assumed. Alkaline lava, rich in potassium, partially gene-

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TABLE

Average SiO_2, K_2O and Na_2O contents and SiO_2/K_2O and K_2O/Na_2O ratios of leucite-bearing and related rocks from various localities in Yugoslavia

Locality	Rock type	SiO_2	K_2O	SiO_2/K_2O	Na_2O	K_2O/Na_2O	N	Sources
		Inner Dinarides						
Rudnik Mts	Leucite-	52.72	6.11	8.62	1.70	4.29	3	Terzić and
	trachyte	S.d. = 2.39	0.05	0.35	0.84	2.29		Ristić, 1961
Rudnik Mts	Olivine-	50.27	4.87	10.32	2.28	2.13	1	Pavlović, 1970
	leucitite	S.d. = —	—	—	—	—		
Rudnik Mts	Lc-trachyte	50.25	4.99	10.21	2.73	1.84	3	Terzić and Milojković, 1967
	Lc-lamproite	S.d. = 0.96	0.69	1.60	0.19	0.38		Ristić P. 1959
Rudnik Mts	Alali-	47.31	1.75	28.32	3.69	0.52	5	Terzić and Milojković, 1967
	basalt	S.d. = 2.51	0.43	6.53	1.00	0.24		Terzić M. 1976
Sjenica-	Olivine-	42.69	4.04	10.84	2.27	1.78	3	Terzić and
Nova Varoš	leucitite	S.d. = 0.62	0.79	2.20	—	—		Popević, 1972.
Sjenica-	Lc.-Oliv.	44.36	1.37	34.62	3.15	0.47	5	Terzić and
Nova Varoš	tephrite	S.d. = 1.00	0.41	9.28	0.59	0.26		Popević, 1972.
Gnjilane	Leucite-	50.77	5.07	10.41	3.03	1.83	3	Ristić P, 1963.
	trachyte	S.d. = 0.51	1.27	2.02	0.75	1.02		Terzić (manuscript)
Gnjilane	Phonolite	52.89	7.51	7.04	3.62	2.07	1	Ristić. 1963. %
Gnjilane	Leucitite	47.31	7.95	5.97	2.32	4.00	6	Ristić. 1963
		S.d. = 2.91	0.73	0.30	0.96	1.82		
Gnjilane	Orendite	50.99	3.98	12.96	3.11	1.32	2	Ristić, 1963.
		S.d. = 0.93	0.56	2.07	1.02	0.25		Terzić (manuscript)
		Serbo-Macedonian massif						
Vranje	Leucite-	46.00	3.63	12.79	3.57	1.01	2	Milojković R. 1962
	tephrite	S.d. = 3.25	0.63	1.34	0.38	0.07		
Vranje	Micro-	45.06	4.91	9.47	1.87	2.65	3	Kostić, 1971.
	shonkinite	S.d. = 2.65	1.08	2.20	0.29	0.66		
Preševo	Leucite-	49.52	4.08	12.54	4.34	1.14	6	Terzić (manuscript)
	tephrite	S.d. = 1.00	0.69	2.20	1.69	0.52		Kostić, 1971.

			Vardar zone					
Nagoričano	Olivine-	49.81	5.30	9.60	2.16	2.21	6	Terzić (manuscript)
	orendite	S.d. = 1.26	0.89	1.91	1.10	0.57		Tomić J. 1929.
Kumanovo	Olivine-	47.30	6.96	6.84	2.22	3.15		Filipović B. 1976.
	leucitite	S.d. = 1.06	1.41	0.73	0.27	0.02	2	Terzić (manuscript)
Štip	Olivine-	47.71	5.42	8.80	2.50	2.16		Tomić, 1929.
	orendite	S.d. = 0.14	0.17	0.30	0.07	0.13	2	Terzić (manuscript)
Sv. Nikola-	Olivine-	51.66	5.16	10.39	2.01	2.54		
Kumanovo	orendite	S.d. = 3.09	1.33	3.22	0.38	0.37	2	Terzić (manuscript)
Kumanovo	Basalt shosho-	53.77	5.15	10.59	3.09	1.74		Tomić, 1929.
	nititic	S.d. = 1.98	0.66	1.88	0.98	0.35	3	Terzić (manuscript)
Demir Kapija	Olivine-	50.82	4.69	11.16	1.97	2.42		Tomić, 1929; Filipović, 1976.
	orendite	S.d. = 2.99	0.96	1.91	0.39	0.50	7	Terzić (manuscript)
Negotino	Olivine-	44.72	3.76	11.89	1.85	2.03		Tučan F. 1931.
	orendite						1	Filipović B. 1976.
Pčinja-	Olivine-	53.26	5.14	10.36	3.82	1.35	2	Filipović B. 1976.
Ovče polje	latite	S.d. = 1.44	—	—	0.46	0.13		
			Carpathian-Balkan Arc					
Bogovina	Jumillite	45.86	3.53	13.21	3.23	2.61	2	Terzić, 1974.
		S.d. = 2.21	0.67	1.93	0.73	1.60		Glišić M. 1959.

S.d. = Standard deviation, N = Number of analyses.

rated from the Upper Mantle, penetrated along renewed old fractures and new ones, transversal to them, during the late Alpine tectonic events from Eocene towards Pleistocene.

The leucite-bearing rocks occur as: a) massive or vesicular lava flows, sometimes containing bombs with chilled margins, b) pyroclastics or tuffs, frequently cut by dykes, and c) veins and small "intrusive" bodies.

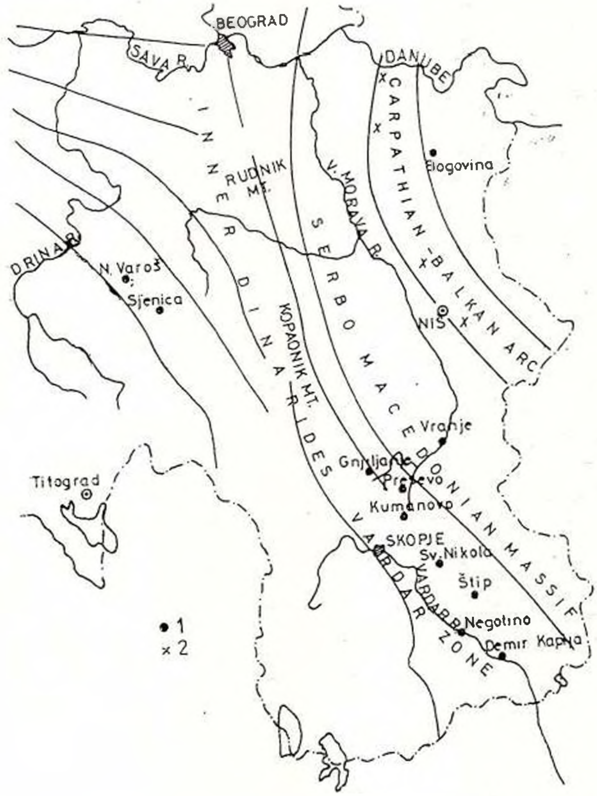


Fig. 1. — Occurrences of leucite bearing rocks (1), nepheline rocks (2), situated in the major tectonic units in Yugoslavia.

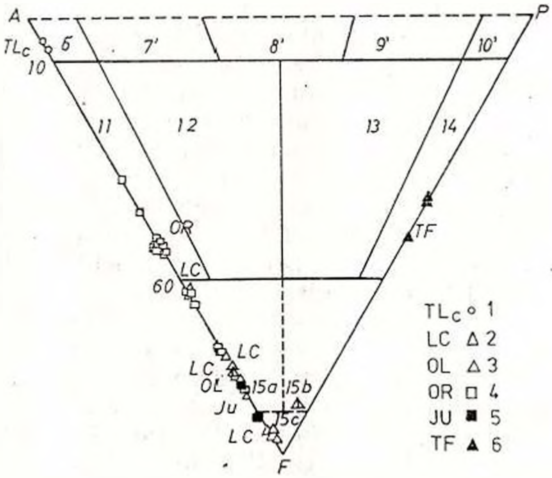
Leucite rocks range from felsic to mafic types. Generally speaking, from the Oligocene to the Pleistocene, they tend to become more mafic. The following types are distinguished: leucite-trachytes, leucitites, olivine-leucitites, leucite-tephrites, leucite-olivine-tephrites, olivine-orendites, jumillite, microshonkinite, as well as their transitional types and types with analcime. In some areas, for example Gnjilane, types without olivine predominate, whereas in others- Macedonia and Sjenica-Nova Varoš- only those with olivine.

For several types of leucite rocks modal mineral content is presented on the A-P-F diagram, after Streckeisen (1978), see Figure 2. They cover the fields of the foid-bearing alkali trachyte, leucite phonolite, sanidine leucitite, tephritic leucitite, leucitite and tephritic phonolite.

In the large complex of calc-alkaline volcanics of Rudnik Mts, the Inner Dinarides, leucite rocks are interbedded with a fresh water sedimentary series of the Lower to Middle Miocene age. The same rocks occur also as fragments in quartz-latite pyroclastics. Flows of quartz-latites, which contain xenoliths of leucite rocks, overlie sediments of Middle Miocene (Pavlović, 1970).

In the area of the Rudnik-Kotlenik Mountains there are: leucite-trachytes, leucitites, olivine-leucitites and metaleucitites. The presence

Fig. 2. — Plot of the modal contents for leucite-bearing rocks from Yugoslavia. 1, leucite trachyte; 2, leucitite; 3, olivine leucitite; 4, orendite; 5, jumillite; 6, tephrite.



of sanidine is characteristic for leucitites, so that these rocks grade into phonolites. A small amount of olivine occurs in the leucite-trachyte. The main pyroxene is titanoaugite. Other minerals of these rocks are: phlogopite, titanomagnetite, zeolite, serpentine minerals, calcite and iron oxide.

Texture is holocrystalline porphyritic to micropoikilitic and trachytic, by exception hypocrystalline porphyritic.

In the Sjenica area the leucite rocks are post-Tortonian in age. In this part of the Inner Dinarides leucite-bearing rocks are separated from the calc-alkaline rocks of the Golija Mountain. Flows of mafic lava were formed along a large dislocation trending WSW-ENE, renewed in the final phase of existence of the fresh water Neogene lake Trijebine. The lava flow lies over the narrow zone of the diabase-chert formation and limestones of Upper Triassic age, and a small part of this covers the Neogene. In Nova Varoš the pyroclastics of these are mixed with Neogene sediments.

Leucite rocks of these regions: olivine-leucitite, olivine-tephrite, analcime-tephrite, are very rich in coloured minerals CI = 65—80. The main constituents are: leucite, analcime, labradorite, augite, olivine, phlogopite, hornblende; then sanidine, magnetite, iddingsite, serpentine minerals, pyrite, hematite, goethite and calcite. The presence of nepheline was established by electronic microprobe analysis. The texture is holocrystalline porphyritic, seldom hyaline.

The greatest masses of leucite rocks in Yugoslavia, are situated in apart of the Morava (Binačka) Depression, in the environment of Gnjilane. Together with their tuffs and breccias they are part of a heterogeneous sedimentary-volcanic complex of Middle Oligocene age (P a v i ć, 1969). The ground of the above-mentioned rocks is formed by Paleozoic slates and Upper Cretaceous flysch. Leucitic rocks cover an area of about 40 km².

The leucitic lava effused in several phases, along cross faults, in relation to the same Dinaric trends. The following high-K unsaturated, mainly felsic rocks, were formed : leucitites leucite-phonolites, leucite-trachytes. Exceptionally there occurs orendite, which is probably the youngest leucite rock of this area. In the wider area of Gnjilane the flow of leucitic lava was preceded by the formation of alkali trachytes and latites. The "paramagma" of leucite rocks, only if one may say that it ever existed as such, had a more alkalic character than the "andesite" one.

The minerals of leucitic rocks are : leucite, analcime, sanidine, augite-aegirineaugite, biotite, melanite, and so on. The uneven size of leucite crystals from 0,2 to 5,0 mm, indicates a long period of their crystallization in the leucite lava of this province. Sanidine corresponds to a high temperature potassium feldspar with different contents of sodium, which is reflected in the optical property. A particularly high content of biotite was established in orendites.

Leucite trachyte dykes have xenoliths of different older rocks: gabbro, sandstones ; as well as alkalic pyroxenite, composed of monoclinic pyroxene, biotite, apatite, magnetite, sphene and some alkali feldspar. The above-mentioned rock represents a mineral association crystallized in a previous period of crystallization of alkali magma, when P-T conditions prevented the crystallization of leucite.

East of Gnjilane, near Vranje, there occur mafic leucitic rocks. In the Klinovačka Reka leucite-tephrites occur as short flows in a facies of sandstones. It is considered that volcanism existed in the upper part of the Upper Eocene, and that the volcanic activity lasted for a very short time.

Near Preševo, the leucite-tephrites, as well as the transitory rocks to trachybasalts, occur as flows and "dykes" through slates of the "Veleš Series", partly also through Neogene sediments, along the dislocation trending NW-SE, at the borders of the Kumanovo Depression.

The alkalic volcanism of the Vardar Zone, in Macedonia, is related to a vertical faulting, with main NW-SE trends.

According to the opinion of Macedonian geologists, leucite rocks are of Pleistocene age. There are also opinions that these rocks, from the Kumanovo area, effused in the Oligocene and Miocene (T o m i ć, 1929 ; K. P e t k o v i ć, 1958 ; P a v i ć, 1969). Pliocene sands below the flow of leucite lava, near the St. Nikola — Nagoričane monastery, do not contain fragments of leucite rocks, nor their minerals. Thermocontact changes were observed, but they are of very slight intensity. Changes of this type were also observed on Kurešnička Krasta near Demir Kapija.

Mafic leucite rocks occur in Macedonia as short flows (up to 2 km near Nagoričane) ; as neck and dykes on the Kurel hill. The lava flow of Nagoričane, later broken and partly curved, with its characteristic feed channel of lava, is particularly outstanding in the relief.

Contrary to high-potassium feldspathoid rocks of the Gnjilane area the leucite lava of Macedonia has no pyroclastics or these are only found in small amounts.

The mafic rocks of Macedonia belong to: olivine-orendite, olivine-leucitite or their transitory types. The previous determination of these rocks as "kajanite" (Lacroix, 1926; Kojić, 1926; Tomić, 1929 and Tućan, 1931) cannot be accepted, since these rocks contain alkali feldspar. The original rock from the Borneo Island, Oele Kajan locality, which H. A. Brouwer (1910) determined as "mica leucite basalt", does not contain any feldspar.

The mineral composition of these leucite rocks is as follows: leucite, sanidine, anorthoclase, augite, phlogopite, olivine, neobiotite, magnetite, apatite, priderite (?), analcime, pseudoleucite, iron-titanium oxides, and exceptionally amphibole and melitite (?).

The content of coloured minerals is high, CI = 62–65, and is independent of the rock type.

The leucites are untwinned and isotropic with varied amounts of inclusion (Štip). In other localities leucites are pseudomorphosed by turbid material-analcime, and pseudoleucite. X-ray analyses indicate that K (Na) feldspars of these rocks approach the high sanidine modification or lie in the field of crystallization for high sanidine. The optical angle ($-2V = 23-71°$) of these feldspars is very variable. Augite has an unusually elongated habit and it is very difficult to determine its optical properties ($2V = 53-60°$). Olivine occurs as small scattered red-rimmed phenocrysts and there is no petrographic evidence of its instability (St. Nikola), or completely altered to serpentine minerals or bowlingite. The CaO content of olivine shows that this olivine was produced under volcanic condition, ($-2V = 87-90°$).

Phlogopite ($2V = 31-42°$) is one of the earliest phases precipitated in the leucite rocks of the Vardar zone in Macedonia. In the olivine leucitite of Macedonia, as well as in all the other localities in Yugoslavia, there occurs neobiotite as accessory mineral. Priderite (?), may be present as inclusions in phlogopite, as well as iron-titanium oxides.

The texture of these rocks is holocrystalline, rarely hypocrystalline.

A unique outcrop of jumillite in the Carpathian-Balkan Arc is found in Bogovina, near Bor. In the Bogovina mine leucite-bearing rocks cut pyroclastic rocks of the second volcanic phase, Upper Cretaceous in age, which are overlain by transgressive conglomerates of Oligocene age.

Chemistry and Genesis

Chemical data for leucite rocks of the Serbian-Macedonian alkaline province in Yugoslavia, are presented in Table. By their chemical characteristics these belong to potassium-rich unsaturated volcanic rocks. The SiO_2/K_2O ratio is below 15. The K_2O/Na_2O ratio is 1–4.29, which is somewhat lower than the one for the same mafic leucite rocks in the world.

Some of the leucite rocks of this province show subalkaline properties as a result of the change of leucite to analcime, and oxidation of the olivine. Only unaltered leucitites, olivine leucitites, olivine orendites and jumillites contain leucite in the CIPW-norm. The type with analcime has hypersthene in the norm.

Plot MgO against SiO_2, CaO against SiO_2, and FeO (total iron) against MgO, weight %, of the K-rich mafic rocks from Macedonia are the same as the orenditic lamproite of the Leucite Hills and the Jumilla District (Sahama Th. G., in Sörensen, 1974).

Relationship between chemistry of the unsaturated nepheline rocks, jumillite and other calc-alkaline volcanics of the Timok area is represented

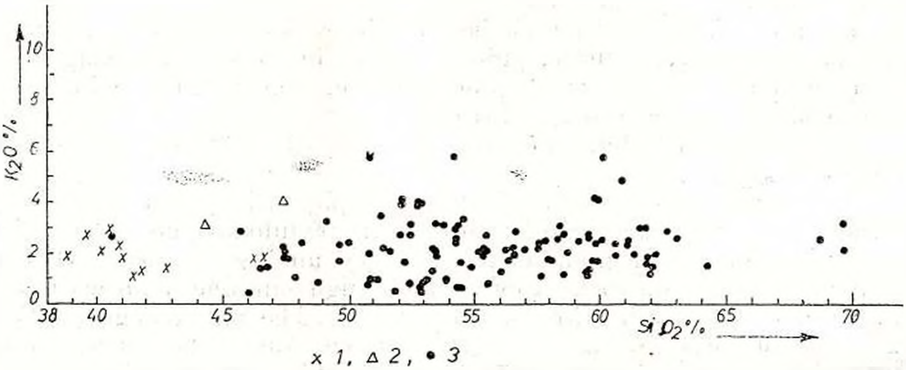


Fig. 3. — Plot of K_2O vs. SiO_2 for the nepheline rocks (1), jumillite (2) and calc-alkaline volcanics (3) of the Carpathian-Balkan Arc.

on Figure 3. This mafic, high K-alkalic rock shows the K_2O/Na_2O ratio of 2.61 (Tab.).

Figure 4 shows the relations between SiO_2 and K_2O of the leucite rocks and of the calc-alkaline volcanics of the same regions (Tertiary

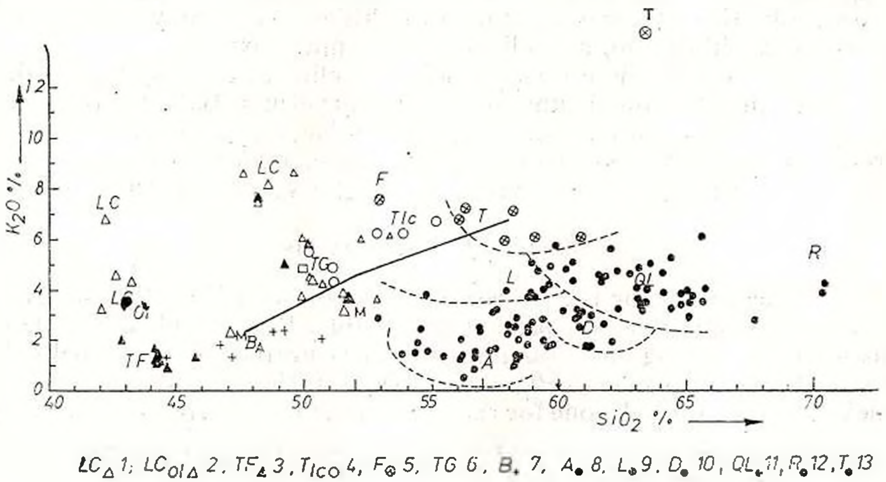


Fig. 4. — Plot of K_2O vs. SiO_2 weight percent for the leucite-bearing and calc-alkaline rocks of the Inner Dinarides in Yugoslavia.

1, leucitites; 2, olivine leucitites; 3, tephrites and analcime tephrites; 4, leucite trachytes; 5, leucite phonolites; 6, transition rocks; 7, alkali basalts; 8, andesites; 9, latites; 10, dacites; 11, quartz latites; 12, rhyolites; 13, trachytes.

age), in the Inner Dinarides. Most of the leucite rocks from the Gnjilane area fall in the high K-alkaline field and those from Rudnik-Kotlenik Mts lie in the transitory among "shoshonitic rocks" and the field of high alkaline rocks. The rocks from the Sjenica-Nova Varoš area have all ultrabasic compositions and fall out of bordered fields.

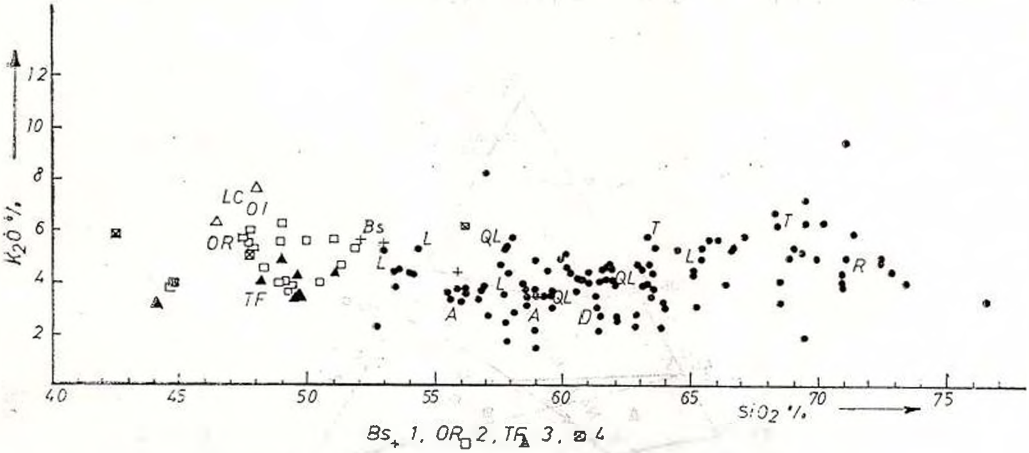


Fig. 5. — Plot of K_2O versus SiO_2 for leucite-bearing rocks and calc-alkaline volcanic rocks of the Vardar Zone and the Serbian-Macedonian Massif.

1, shoshonitic basalt; 2. olivine orendite; 3, olivine tephrite; 4, microshonkinite.

In the diagram SiO_2 vs. K_2O the volcanics of the Vardar zone (Fig. 5), have not a sharp separation between high-K calc-alkaline series (latites and basalts shoshonitic. Olivine orendites and with them associated olivine leucitites lie in the field of high K-alkalic basic rocks.

Data on the course of crystallization of systems similar to the studied rocks are very rare. Neglecting the contents of Fe-Mg and Ca-Mg-Fe of silicate components, the nearest is the studied system SiO_2-$NaAlSiO_4$-$KAlSiO_4$ (S c h a i r e r, 1950, cited by G u p t a, 1980).

All the studied rocks fall in corresponding fields, except for leucite-tephrite which falls in the field of crystallization Na-K feldspar, between the phases Ab, Lc and Ne. This is the consequence of a considerable amount of basic plagioclases which caused a movement of the rock field in the diagram. The content of olivine in the modal composition of the rock, agrees with the distinct unsaturation of this rock, in the diagram. The other leucite rocks, such as orendite, leucite-trachyte, leucite-phonolite, olivine-leucitite and leucitites are in the field of leucite crystallization which agrees with the presence of leucite phenocrystals (with phenocrystals of coloured minerals). Apart from leucite phenocrystals, phenocrystals of sanidine occur only in leucite-trachyte. This indicates that abrupt crystallization began after reaching the cotectic line between leucite and K-Na feldspar (sanidine) during which the ground was formed.

Interrelationship between the various rock types in a specific area is important for the petrogenesis of the potassic volcanics. The leucite rocks which occur in Macedonia are characterized by a more mafic mineral assemblage, whereas those from Gnjilane have felsic minerals. In the

Nova Varoš-Sjenica region, field evidence indicates that more felsic types of the olivine leucitite crystallized at, or near, the surface of the flow, whereas the crystallization of the leucite tephrite took place in its lower part. This differentiation was effected by the enrichment of the alkali on the top through gaseous transfer and by the sinking of the olivine phenocryst (T e r z i ć, P o p e v i ć, 1972).

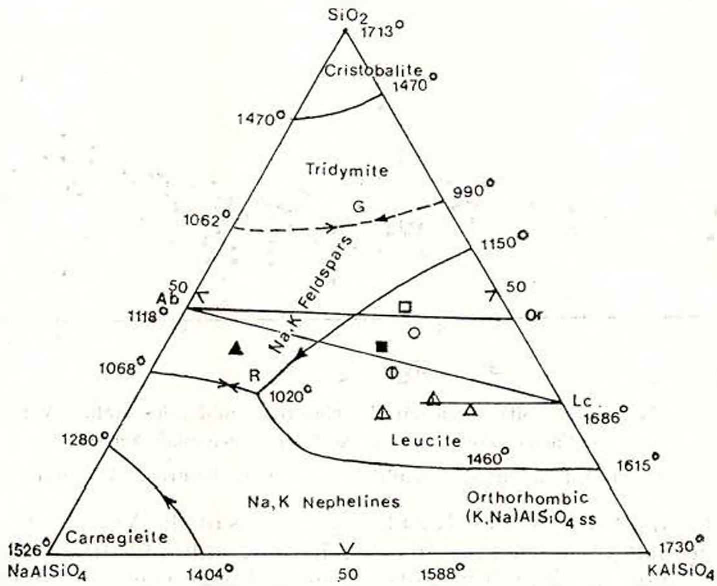


Fig. 6. — Plot of the chemical composition of the selected type of the leucite bearing rocks of Macedonia and Serbia in the nepheline-kalsilite — SiO_2 system (After S c h a i r e r)

The area of the potassic volcanics in Macedonia (Kumanovo, Nagoričane) is not so far from the same region of volcanics of Gnjilane. It could be supposed that this potassium and magnesium deep seated parent magma, was subjected to chemical differentiation during its ascent from the depth to the surface. The felsic portion of this magma erupted in an earlier phase, e.g., in the Oligocene, as pyroclastics of leucitites which alternate with leucite phonolites and leucite trachytes, and the mafic types of leucite rocks were crystallized later, in the Pleistocene, as calm lava flows without pyroclastic rocks.

REFERENCES

C a r m i c h a e l I. S. E. (1967) Cont. Min. Petrology, 15, p. 425—456.
C v i j i ć J. (1906) Osnovi za geografiju i geologiju Makedonije i Stare Srbije.
G l i s i ć M. (1959) Vesnik Zav. za geol. i geof. istraž. knj. XVII, str. 115—119, Beograd.

Gupta A. K. et al. (1980) Petrology and Genesis of Leucite-Bearing Rocks. Spring. Verl. Berlin.

Kojić M. (1926) Comptes rendus, Séances du 8 mars, p. 624–645, Paris.

Kostić A. et al. (1961) Glasnik prir. muz., ser. A, knj. 14–15, str. 59–78, Beograd.

— (1970) Glasnik prir. muz., knj. 25, ser. A, str. 109–127, Beograd.

Lacroix A. (1926) Comptes rendus, Seances du 8 mars, p. 597–601, Paris.

Marić L. (1934) Vesnik geol. Inst. Kr. Jugoslavije za 1934., knj. IV, 1, str. 44–51, Beograd.

Pavić A. (1969) Vesnik Zav. za geol. i geof. istraž., knj., XXVII, A, str. 73–82, Beograd.

Pavlović Z. (1970) VII Kongres geologa SFRJ, Zagreb, 28.IX.4.X, II, str. 297–301, Zagreb.

Protić M., Terzić M. (1976) Congress on thermal waters, geothermal energy and volcanism of the Mediterranean area, Athens, October 5–10, p. 195–210, Athens.

Ristić P. et al. (1959) Geol. anali Balk. pol., knj. XXVI, str. 219–227, Beograd.

— (1963), Arhiv za tehnologiju, 3, str. 1–74, Tuzla.

Sorensen H. (1974) The Alkaline Rocks. London.

Terzic M. et al. (1961) Geol. anali Balk. pol., knj. XXVIII, str. 295–301, Beograd.

— (1967) Geol. anali Balk. pol., knj. XXXVIII, str. 309–315, Beograd.

— (1972) Geol. anali Balk. pol., knj. XXXVIII, str. 179–198, Beograd.

— et al. (1974) VIIII Kongres geologa Jugoslavije, Bled 1–5, X, str. 273–290 Ljubljana 1976.

— (1976) Geol. anali Balk. pol., knj. XL, str. 219–232, Beograd.

— (1976) Geol. anali Balk. pol., knj. XL, str. 253–262, Beograd.

Tomić J. (1929) Glas SKA, knj. CXXXVII, str. 3–42, Beograd.

Tućan F. (1931) Glasnik Skopskog naučnog društva, Knj. IX, Od. pr. nauka, sv. 3, Skoplje.

Tumač za izradu Osnovne geoloske jkarte. List : Uroševac, Kumanovo, Štip, Savezni geol. Zavod., Beograd (1 : 100.000).

Žujović J. (1920) Geologija Srbije. Drugi deo. Beograd.

THE ȚIBLEȘ NEOGENE IGNEOUS COMPLEX OF NORTH ROMANIA: SOME PETROLOGIC AND METALLOGENETIC ASPECTS[1]

BY

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Introduction

The Țibleș igneous complex belongs to the so-called subvolcanic zone within the Neogene volcanic chain of the Eastern Carpathians. It shows the most complicated petrographic relationships as compared to the other units of the subvolcanic zone, i.e. Toroiaga, Rodna and Bîrgău. Peltz et al. (1972) have already reported some general petrographic features of the whole zone. The uniqueness of the Țibleș complex consists in its peculiar rock types and metallogenetic products. Special papers are devoted to each major topic i.e. geologic setting and structure of the Țibleș massif and its surroundings (Edelstein et al., 1981), petrology (Pop et al., 1984), metallogenesis (Udubașa et al., 1984) as well as the peculiar low temperature Sb-rich mineralization occurring in the western part of the massif (Pop et al., 1984). Accounts of previous geological work in the area may be found in these papers too. It is, however, to be noted the first petrographic overview on the Țibleș-Hudin zone given by Pavelescu (1960) and Maier (1962). The present paper summarizes the most important and relevant features of the Țibleș igneous complex.

Geologic Setting

The Țibleș igneous complex is situated south of the important E-W striking Bogdan Vodă-Fault and near the southernmost appearance of the so-called Pienides (Săndulescu, 1980). The basement of the area is plunging eastwards and the Earth's crust is about 30 km thick (Socolescu et al., 1975). The igneous rocks penetrate sedimentary rocks belonging to the Paleogene and Miocene; the age of the igneous complex is thus confined to the post-Lower Miocene time (Edelstein et al., 1981).

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The sedimentary rocks build up three tectonic units with overthrusting relationships: the Autochthon, the Lăpuş or Wildflysch Nappe and the Central Tectonic Unit. The largest part of the igneous rocks appears within the Central Tectonic Unit. The igneous rocks cover an area of about 23 square km and form three main eruptive units: Hudin in the north, Hudieş-Stegioara in the middle, and the biggest, Tomnatec-Ţibleş-Măgura Neagră, in the southern part. The geophysical data (Andrei et al., 1981) indicate for the southern unit a unique eruptive body at a depth of about 1,000 m; the extension of such a unique intrusion largely corresponds to the thermal contact aureole (Fig. 1).

The Structure of the Igneous Complex

The most striking feature of the Ţibleş igneous complex is the lack of any volcanic event. All the rocks occurring there have properties (structures and opaque mineral assemblages) typical of the hypabyssal crystallization.

There are two igneous phases: 1) an earlier, more acidic phase including microgranodiorites and dacites and 2) an intermediate and slightly basic rock sequence consisting of quartz monzodiorites (main type) and quartz monzogabbros, diorites ± quartz, quartz microdiorites, granodiorites, monzogranites, andesites (hornblende and pyroxene-quartz-bearing varieties), tonalites etc. The last rock types have a limited distribution and it is thought that they represent *in situ* differentiation products of the consolidating andesitic magma.

The rocks of the first phase form some unitary developed eruptive bodies, whereas those of the second phase display more complicated settings (Fig. 1). The main intrusion of the southern unit consists of porphyry quartz monzodiorites and granodiorites with scarce diorites, monzogranites, tonalites etc. Very obvious is the development of a ring around the main intrusion, which in places exhibits properties of chilled margins. This ring consists of finer grained rocks of quartz dioritic composition (quartz microdiorites, andesites). Around this ring structure there are tens of small bodies of varying composition ranging from andesites and quartz-diorites to (micro) granodiorites.

The rocks of the second phase develop also toward the north building up the middle eruptive unit; various kinds of rocks have been herefrom reported (Edelstein et al., 1981; Pop et al., 1984). In this area the relationships between the two main igneous phases could be observed: near the Tomnatec Peak a small body of quartz monzodiorites penetrate the dacites.

Mineralogy of Igneous Rocks

The most significant rock-forming minerals of the Ţibleş igneous rocks are the plagioclases, the alkali-feldspars and the clinopyroxenes. All of them occur in at least three generations successively formed during the stadial crystallization of the rocks. The orthopyroxenes (hypersthene), amphiboles (hornblende) and micas (biotite) have been locally observed.

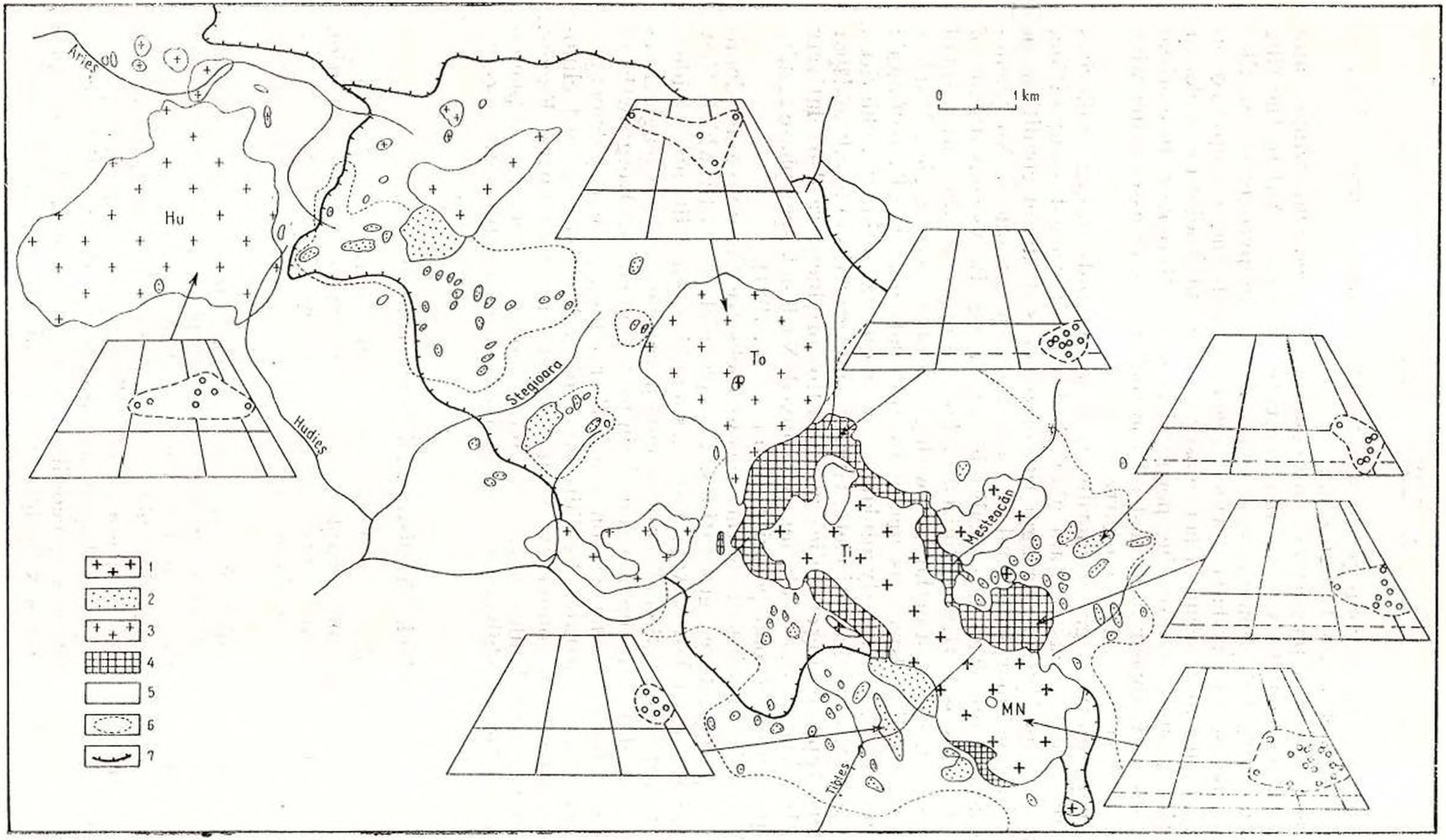


Fig. 1. — Ţibleş igneous complex : distribution of main rock types with *Q A P* plots.

1, quartz monzodioritic rocks ; 2, diorites and gabbros, quartz-diorites, porphyry granodiorites, quartz andesites, dacites, pyroxene and hornblende bearing andesites ; 3, microgranodiorites and dacites ; 4, pyroxene andesites, quartz microdiorites and andesitoid rocks ; 5, Oligo-Miocene sedimentary rocks ; 6, contact aureole ; 7, tectonic contacts.

In the rocks of the second phase abnormal amounts of magnetite and apatite may in places be recognized.

The plagioclase phenocrysts are commonly zoned and twinned and their anorthite content varies from the core (60–85% An) to the ring rocks (30–50% An). Sometimes they include clinopyroxenes. Smaller sized plagioclase grains have formed together with the clinopyroxenes and amphiboles and occur either mostly in the microdioritic rocks or within the groundmass of the porphyritic ones. Moreover, interstitial plagioclases may appear in almost all the rock types in association with the alkali-feldspars and the quartz.

Except diorites, gabbrodiorites and some kinds of quartz-diorites all the remaining rocks contain alkali-feldspars of varied composition. They have formed either as magmatic components or under conditions of the contact metamorphism and as postmagmatic products. No phenocrysts could be observed, the alkali-feldspars being always constituents of the groundmass as partly triclinized orthoclase or as sanidine with small 2 V angle with a ring of orthoclase with greater (48–65°) angles. Veinlets or nests of alkali-feldspars as magmatic residuals may locally be found; they occur together — mostly adularia — with quartz, actinolite, chlorites, magnetite, pyrite, chalcopyrite etc. Graphic intergrowths of alkali-feldspar with quartz representing eutectic crystallization products are also relatively frequently recognized. Exsolution of a mixed Na-K-feldspar phase results in the appearance of some perthites (P o p et al., 1984).

The mafic constituents of the igneous rocks are clinopyroxene dominated. They form three generations, like plagioclases with which they may be genetically related. Augite, titanium augite and salite are the identified mineral species. Sometimes the clinopyroxenes include plagioclases or ilmenite and in some instances contain uralitized cores. Overgrowths of an younger clinopyroxene (II) on the earlier clinopyroxene phenocrysts (I) are sometimes separated by thin bands of uralite too. The last clinopyroxene generation commonly occurs within the groundmass. Except some primary hornblendes (both common and brown) occurring as phenocrysts, the other amphiboles (uralite, actinolite, antophyllite) are of secondary origin.

Rock Chemistry and Nomenclature

More than 90 full chemical analyses are up to date available from the igneous rocks of the Ţibleş complex. The average values and the variation intervals for the main rock types are given in Figure 2. The SiO_2 — and TiO_2 — contents are clearly bimodal, separating the two igneous phases (Fig. 3).

Due to the considerable variations, both structural and chemical, within the individual eruptive bodies the *QAP* plots in the Streckeisen diagram (Fig. 1) occupy more than one field. The greatest spread is to be noted for the rocks of the first phase; it is, however, to emphasize that these rocks rarely occur completely fresh. Similar rocks (e.g. microgranodiorites) of the second phase display a more restricted point distribution. Typical heteromorphic rocks (according to R i t t m a n n, 1973), represented in the Ţibleş complex by quartz monzodiorites, cover six fields on the Streckeisen diagram, suggesting their heterogeneity and the difficulties arose in the microscopic identification of the petrographic type.

The chemical discontinuity between the two main igneous phases is a general feature of all the diagrams constructed by P o p et al. (1984). The calc-alkaline character of the Ţibleş rocks — with a slightly calcic

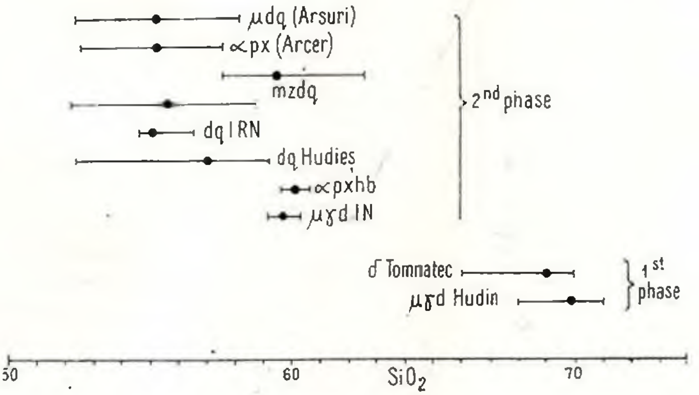


Fig. 2. — SiO_2 variation intervals and average values (points) of the main rock types in the Ţibleş igneous complex.

trend derived from the Peacock index by 62.5 percent SiO_2 — is coupled with a tholeiitic trend (more correct : plotting onto the transitional zone between the calc-alkaline and the tholeiitic rock series), very apparent in the Na_2O+K_2O/SiO_2 diagram (after G i r o d et al., 1978). Most of the

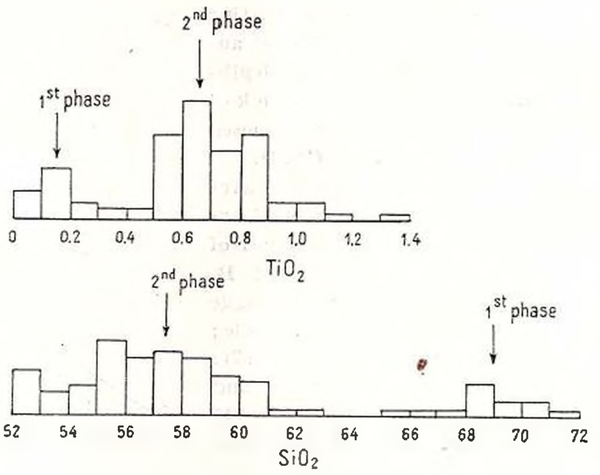


Fig. 3. — TiO_2 and SiO_2 frequency diagram of the Ţibleş rocks.

Ţibleş rocks belong, according to the $MgO/FeO+Fe_2O_3$ diagram (Y o d e r jr., 1969), to the calc-alkaline series developed in the island arcs. All these and further diagrams are largely commented by P o p et al. (1984).

P e l t z et al. (1972) accepted a common differentiation trend for the igneous complexes belonging to the subvolcanic zone. It is, however, interesting to note that the average values for the rocks of the first and second phases in the Ţibleş complex plot onto the Ca-Na-K diagram at a

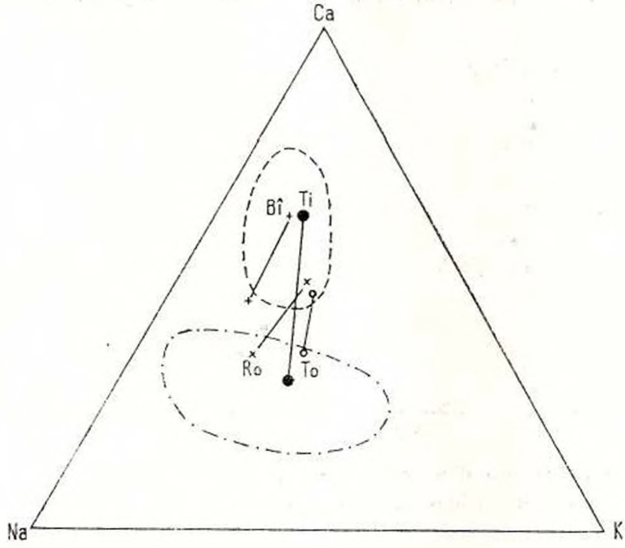
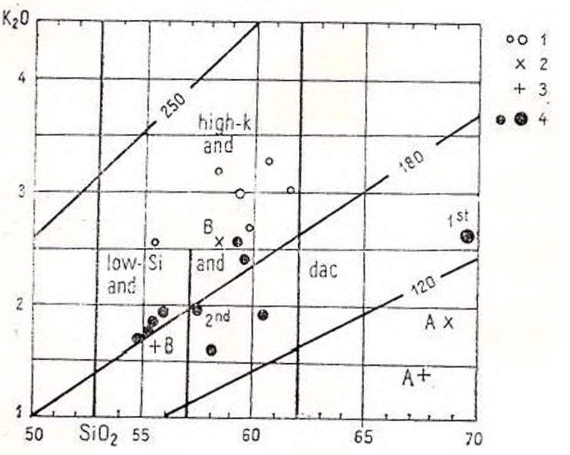


Fig. 4. — Ca—Na—K diagram; abbreviations are: Ti, Ţibleş; Bi, Bîrgău; Ro, Rodna; To, Toroiaga (chemical data for Toroiaga from the paper of B e r z a et al. (1981).

distance twice larger than those for the corresponding rocks in other complexes (Fig. 4). This fact points to different sites of generating magmas for the two Ţibleş igneous phases, rather than to a unique differentiated

Fig. 5. — A combined diagram relating T a y l o r ' s (1969) classification of andesites and the diagram of the depths to the magmatic chamber of N i n k o v i c h and H a y e s (1971), as proposed by D i m i t r i j e v i ć (1974).
1, Toroiaga rocks; bigger circle – average value (n = 52; data from B e r z a et al., 1981); 2, average values of the Rodna rocks (A, acidic n = 2; B, intermediate types n = 20); 3, average values of the Bîrgău rocks (A, acidic; n = 2; B, intermediate types: n = 17); 4. Ţibleş rocks; main rock types and average values for the first (A, n = 11) and the second (B, n = 54) phases.



magma. The same conclusion is reached if the combined diagram of T a y l o r (1969) and N i n k o v i c h & H a y e s (1971) is applied to the subvolcanic rocks (Fig. 5).

Minor elements connected with the Nockolds-Allen index show normal features : Ba and Zr increase with the increment of this index, Cr, Ni, Co, Sc and V decrease, and Y, Yb and Ga remain practically unchanged. Taking into account the basic and intermediate rocks of the Țibleș complex it is to emphasize their higher copper content as compared with the average

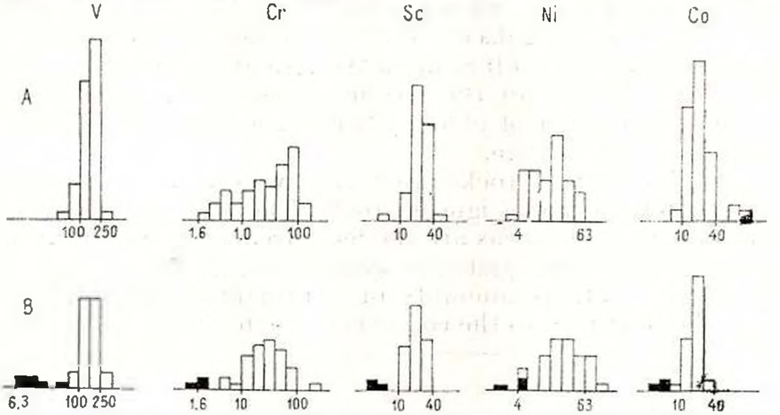


Fig. 6. — Frequency diagrams of the distribution of some minor elements in andesites (A, acc. to T a y l o r, 1969) and in the Țibleș rocks (B ; black, first phase rocks).

values presented by K r a f t & S c h i n d l e r for similar rocks. The frequency diagrams of the distribution of some minor elements in the Țibleș rocks (Fig. 6) are astonishingly similar to the diagrams shown by T a y l o r (1969) to be typical for andesites.

Constitution of the Igneous Contacts

The contact aureole around the Țibleș igneous complex has an average width of about 1 km (Fig. 1) and is developed only in connexion with the rocks of the second phase. Various kinds of hornfelses have been herefrom described (E d e l s t e i n et al., 1981). Strongly silicified rocks, sometimes tourmaline bearing, form locally developed roofs on the igneous rocks. The innermost zone within the contact aureole consists of biotite + + hornblende ± andalusite and clinopyroxene, whereas the outermost one displays features typical of the Knotenschiefer. An overprint with obvious substance exchange has been locally recognized. The contact metamorphic effect may sometimes be observed in the igneous rock itself, mainly at the contacts of the second phase rocks with dacites ; the latter become very fine-grained and bear small grains of red garnet. Generally, the contact aureole of the Țibleș complex contains low-pressure assemblages (indicating pressures lower than 4 kb) formed under conditions of a generalized disequilibrium. The lack of the glass as hornfelse constituent proves the shallow depth of formation (lower than 1 km). Worth noticing is the presence of the stable assemblage pyrrhotite + rutile at the igneous contacts of the Țibleș complex, both within the sedimentary and eruptive rocks. Such a feature reveals a particular evolution of certain intrusions (U d u b a ș a, 1981).

Late Magmatic and Postmagmatic Events Accompanying the Intrusion

Magmatic magnesian skarns, e.g. with spinel, forsterite, phlogopite etc, occurring in the 4 Arcer gallery have been already described by U d u b a ş a et al. (1982). This occurrence represents the first reported magnesian skarn formation related to the Neogene magmatites in Romania.

Breccias were recognized either in relation to the dacites (west of the Tomnatec Peak) as contact intrusive formation or connected with the more complex evolution of the second phase rocks. Such breccias may contain either tourmaline or phlogopite or both and locally show a relative enrichment in chalcopyrite.

Tourmaline bearing rocks are rather widespread and become very characteristic for the Ţibleş igneous rocks. The tourmaline bearing quartzites (hornfelses) and breccias are the most frequent. Slightly transformed igneous rocks (andesites, quartz monzodiorites, dacites, quartz microdiorites) show also various amounts of tourmaline. This mineral seems, however, to be restricted to the contacts of the main intrusion (Figs 7 , 8).

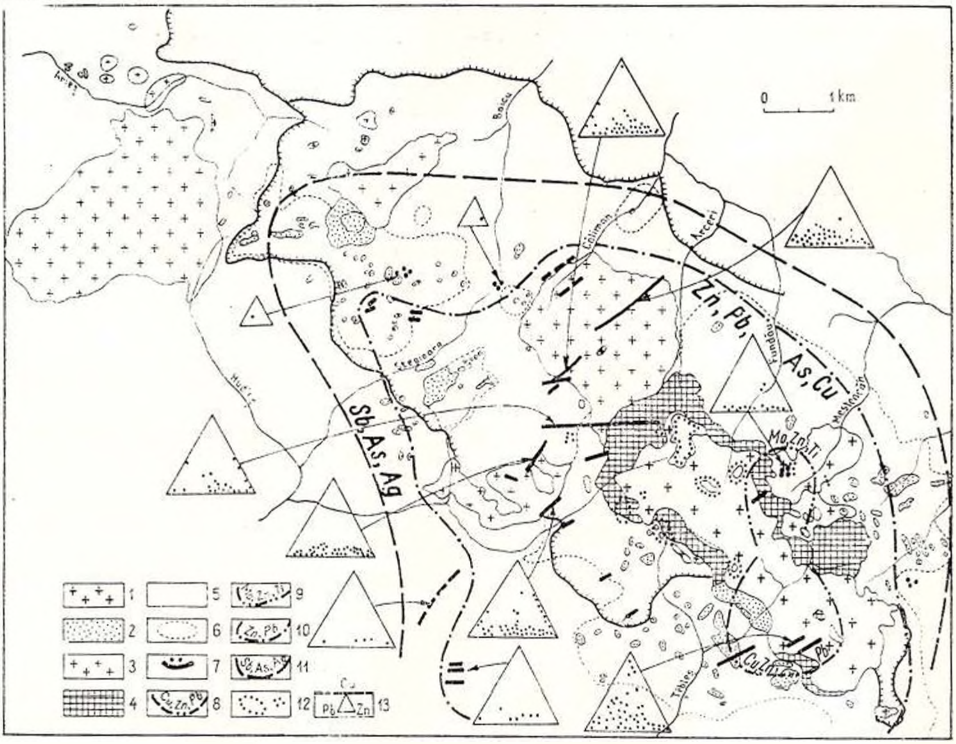


Fig. 7. — Metallogenetic map of the Ţibleş Igneous Complex.
1 - 6, see Fig.1 ; 7, veins and disseminations ; 8, copper enriched ores and the presumed porphyry system ; 9, Mo-Cu-Zn-Ti-B disseminations ; 10, inner zone of high temperature vein assemblages ; 11, external belt of lower temperature vein assemblages ; 12, tourmaline occurrences ; 13, ternary diagrams with analytical data of the primary ores (larger triangles) and limonites (smaller triangles).

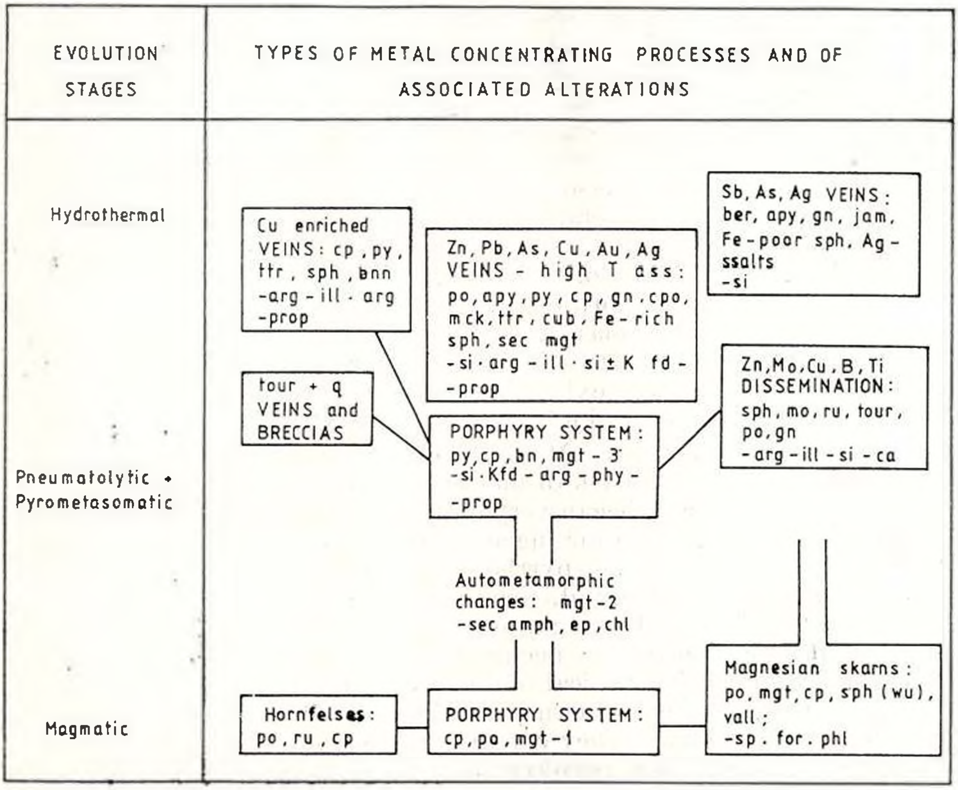


Fig. 8. — Sketch of the metal concentrating processes and of the associated alterations in the Țibleș Igneous Complex.

Special mention should be made for the tourmaline occurrence within the disseminated sphalerite-molybdenite-rutile mineralization on the Mesteacăn Valley (Fig. 7).

Metallogenesis

Types of mineralizations

1. *Veins*. The central north-western vein group (Fig. 7) is the most important type of mineralization occurring in the Țibleș massif. The ores contain high temperature mineral assemblages (Fig. 8). Copper enriched veins develop in the southern part, in which the sphalerite is iron-poor, chalcopyrite occurs not only as exsolution blebs in sphalerite (as in the ores of the main veins) but also in the form of monomineralic aggregates; bournonite has been known for a long time (B o r d e a, 1960). An external belt of lower temperature mineral assemblages (veins and impregnations) envelops discontinuously the central mining area (Fig. 7). Berthierite and arsenopyrite in association with some silver sulphosalts are here the dominant mineral species (U d u b a ș a et al., 1984; P o p et al., 1984).

2. *Impregnations and Disseminated Ores.* In the southern part of the igneous complex abnormal amounts of magnetite, locally associated with chalcopyrite, may in places be observed. This association forms either veinlets or impregnations in fairly transformed, locally brecciated, monzodioritic and granodioritic rocks. There is a special type of magnetite (mgt-3, in Fig. 8) appeared as a result of a certain degree of magma oxidation. Such a non-accessory magnetite is thought to be characteristic for the porphyry type mineralization for it has been observed as yet only in porphyry environments (Butte, Montana, USA; Medet, Bulgaria; Reczk, Hungary; Deva, Romania etc).

Within the monzodioritic small intrusion on the Mesteacăn Valley, eastern part of the Ţibleş massif, a disseminated mineralization with sphalerite-rutile-molybdenite-tourmaline occurs in relation to strongly leached rocks. Very small grains exhibiting properties of topaz and cassiterite have been reported too (Udubaşa et al., 1984).

Succession of Metallogenetic Events. It is presumed that the activation of metals in and from the magmatic system at Ţibleş (second phase) began very early due to the oxidation state of the magma. Its rather high sulfur fugacity and copper concentration is shown by the appearance of the pyrrhotite-chalcopyrite $\pm$ magnetite assemblage at igneous contacts and, together with sphalerite, within the magnesian skarns. The oxidation state of the magma allowed the formation of a special type of magnetite occurring only in the rocks of the intrusions displaying an evolution typical of the porphyry systems. The copper enriched ore veins in the southern part of the massif have probably opened the porphyry system and they reveal — as they do in some other cases — the presence of a hidden porphyry system. The relationships among the metal concentrating events and the associated alterations are given in Figure 8. The Ţibleş igneous complex exhibits thus a polyascendent type of ore deposition resulting in the development of a well, rather contrasting, regional zoning (Fig. 7). A similar zoning (without the central disseminated mineralization types) shows the Kutna Hora mining district, Czechoslovakia (Bernard et al., 1969).

REFERENCES

Andrei J., Ciucur E., Nedelcu I. (1981) Report, the archives of the Institute of Geology and Geophysics, Bucharest.

Bernard J. H. a kolectiv (1969) Mineralogie Ceskoslovensko. Academia Praha, 396 p.

Berza T., Borcoş M., Ianc R., Bratosin I. (1981) Succesiunea intruziunilor neogene din regiunea Toroiaga-Ţiganul (Munţii Maramureşului). *D. S. Inst. geol. geofiz.*, LXVII/1, p. 11—24. Bucureşti.

Bordea R. (1960) Report, the archives of the IPEG Maramureş, Baia Mare.

Dimitrijević M. D. (1974) The Dinarides: a model based on the new global tectonics. In: S. Janković (ed.) Metallogeny and concepts of the geotectonic development of Yugoslavia. *Fac. Mining. Geol.*, Belgrade.

Edelstein O., Istvan D., Pop N., Răduţ M., Kovacs M., Stan D., Bernad A., Andrei J., Pop V., Götz A., Bordea R., Roman L. Alcătuirea geologică a munţilor Ţibleş. (Comunicată în 1981), Bucureşti.

Girod M., Bailey D. K., Baker P. E., Fischer R. V., Maury R., Rocci M., Schmincke H., Upton B. G. (1978) Les roches volcaniques : pétrologie et cadre structural. Doin edit., Paris, 235 p.

Maier O. (1962) Geologia şi petrografia masivului Hudin (Munţii Ţibleş). *D. S. Com. Geol.* XLV, Bucureşti.

Ninkovich D., Hayes J. D. (1971) Tectonic setting of Mediterranean volcanoes. *Acta Int. Sci. Congr. Volcano Thera (Greece)*, 1, 111.

Pavelescu L. (1960) Etude géologique et pétrographique du massif de Ţibleş. *An. Com. Géol.* XXVI – XXVIII, p. 69 – 82, Bucureşti.

Peltz S., Vasiliu C., Udrescu C. (1972) Petrologia magmatitelor zonei subvulcanice neogene din Carpaţii Orientali. *An. Inst. geol.*, p. 177 – 256, Bucureşti.

Pop N., Pop V., Edelstein O., Răduţ M., Istvan D., Götz A., Kovacs M., Stan D., Bernad A., Roman L. (1984) Consideraţii petrologice asupra magmatitelor din Munţii Ţibleş. *D. S. Inst. geol., geofiz.*, LXVIII/1, Bucureşti.

— Răduţ M., Pop V., Edelstein O., Roman L., Coroiu G. (1984) Mineralizaţiile epitermale din sectorul văii Zimbrului (munţii Ţibleş). *D. S. Inst. geol. geofiz.*, LXVIII/2, Bucureşti.

Rittmann A. (1973) Stable mineral assemblages of igneous rocks. Springer, Heidelberg.

Săndulescu M. (1980) Sur certains problèmes de la corrélation des Carpathes Orientales Roumaines avec les Carpathes Ucrainiennes. *D. S. Inst. geol., geofiz.*, LXV/5, p. 163 – 180, Bucureşti.

Socolescu M., Airinei Ş., Ciocirdel R. (1975) Fizica şi structura scoarţei terestre din România. Ed. tehnică, Bucureşti.

Taylor S. R. (1969) Trace element chemistry of andesites and associated calc-alkaline rocks. Proc. Andesite Conf. Oregon, p. 43 – 63.

Udubaşa G. (1981) Rutile of postmagmatic mineral formation. In : Ore genesis 1980 (volume dedicated to Prof. P. Ramdohr) Springer, Heidelberg.

— Edelstein O., Pop N., Istrate G., Kovacs M., Istvan D., Bogancsik V., Roman L. (1982) Magnesian skarns at Ţibleş : preliminary data. *D. S. Inst. geol. geofiz.*, LXVII/2, Bucureşti.

— Răduţ M., Edelstein O., Pop N., Istvan D., Roman L., Pop V., Stan D., Kovacs M., Bernad A. (1984) Metalogeneza complexului eruptiv Ţibleş, Carpaţii Orientali. *D. S. Inst. geol. geofiz.*, LXVIII/2, Bucureşti.

Yoder jr. H. S. (1969) Calcalkalic andesites : experimental data bearing on the origin of their assumed characteristics. Proc. Andesite Conf. Oregon, p. 77 – 90.

CONTRIBUTION À LA CONNAISSANCE DE LA ZONE MAGMATIQUE DE TIMOK-SREDNA GORA EN YOUGOSLAVIE, D'APRÈS LES DONNÉES AÉROMAGNÉTIQUES [1]

PAR

SLOBODAN VUKAŠINOVIĆ [2], BORIS SIKOŠEK [2]

Toute la région occupée, en Serbie, par les Carpatho-balkanides a été étudiée par des investigations systématiques aéromagnétiques. Cette note présente l'interprétation géostructurale des données aéromagnétiques de la région où s'étend la zone magmatique de Timok et son prolongement vers le sud-est. La zone de Sredna gora en Bulgarie représente la continuation directe de cette zone. Etant donné que les magmatites de Timok et de Sredna gora représentent une zone unique, on a dénommé cette zone "la zone magmatique de Timok-Sredna gora" [3]. Les données aéromagnétiques ont permis de tracer les limites de cette zone en Yougoslavie et d'établir son extension. On a également relevé en continuité des magmatites dans les régions qui se trouvent hors de la zone éruptive de Timok (ZET), vers le nord et le sud de celle-ci où affleurent des sédiments mésozoïques et tertiaires ou des roches cristallines. A partir des données aéromagmatiques on a établi la présence des éléments appartenant à cette zone sous la couverture sédimentaire ("les magmatites cachées"). Le prolongement des magmatites en dehors de la ZET a été seulement supposé auparavant, ces limites n'étant pas jusqu'à présent précisées. A la suite du fait que les affleurements des magmatites accessibles à la recherche géologique ne sont pas nombreux, ils ne suffissent pas pour préciser leur extension aréale ou même leur développement en profondeur. Il s'ensuit qu'il est difficile de délimiter les marges de la zone magmatique de Timok-Sredna gora.

Les résultats des mesurages aéromagmatiques de la région entre le Danube et la frontière yougoslave-bulgare ont été illustrés graphiquement dans l'annexe où la zone considérée apparaît comme une zone des anomalies magmatiques généralisées.

Vu la configuration du champ magmatique, on peut le diviser, suivant ses caractères spécifiques, en trois parties : l'aréal de la ZET et les parties au sud et au nord de celle-ci.

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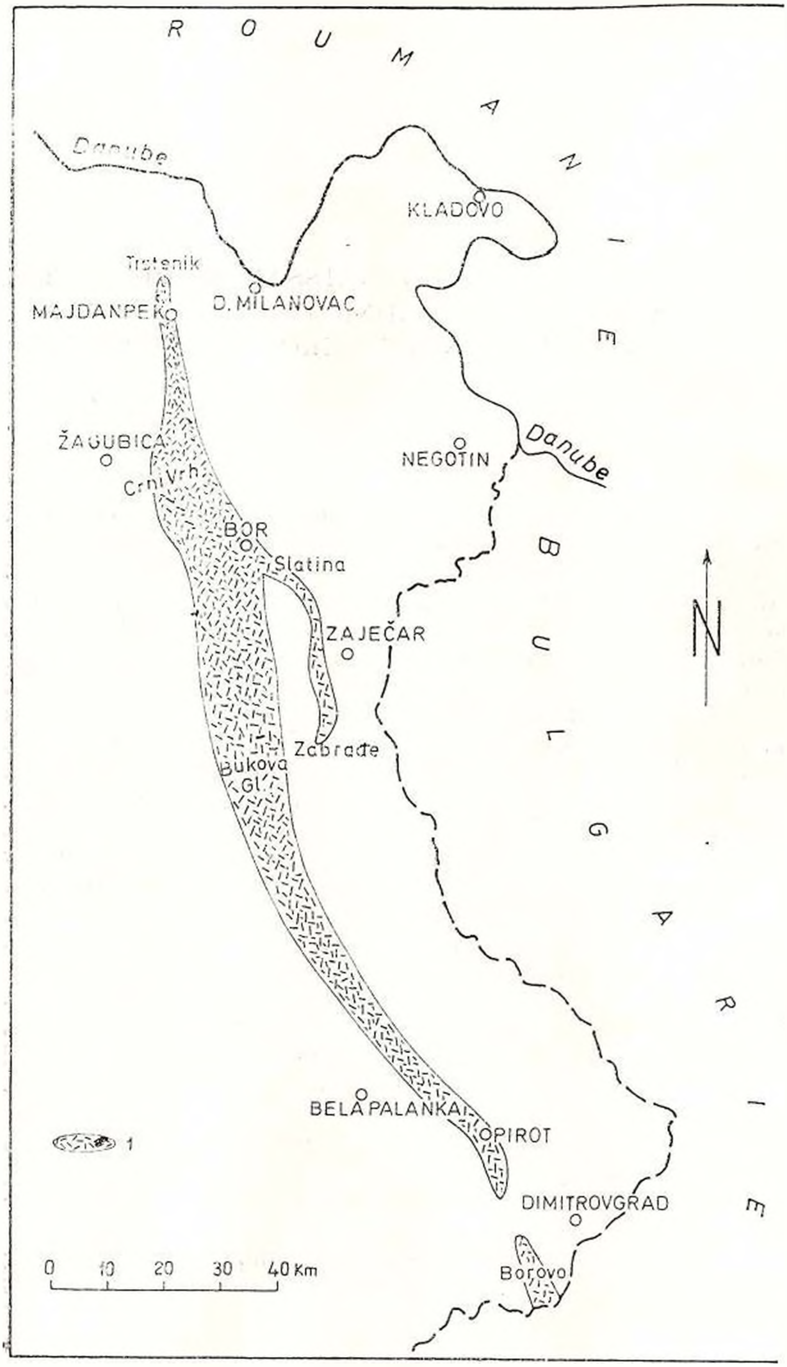
[1] Note présentée au 12 ème Congrès de l'Association Géologique Carpatho-Balkanique, 8—13 septembre 1981, Bucarest, Roumanie.

[2] Yougoslavie.



La zone des magmatites de Timok-Sredna gora en Yougoslavie.
1, La zone généralisée des anomalies magnétiques — des magmatites.

Dans le cadre de la ZET, à partir de Trstenik (Majdanpek) au nord jusqu'à Bukova Glava au sud, on a localisé une zone d'anomalies positives continues ayant des intensités de plusieurs centaines de nT. En réalité, elle est constituée de plusieurs sous-zones de maximums et de minimums Δ T (seulement quelques-unes ayant des valeurs négatives). Le nombre de ces sous-zones augmente vers le sud. Dans la région de Bor-Kirvelj une branche se sépare vers le sud suivant la direction Slatina-Zabradje. Les valeurs extrêmes Δ T dépassant une intensité de 2000 nT ont été enregistrées à Zlotska reka, mais une anomalie tout aussi remarquable est celle de Crni Vrh d'une intensité de 1300 nT.

Vers le sud de la ZET, entre Bukova Glava et Borovo (la frontière yougoslavo-bulgare), le champ anomal s'est manifesté dans une zone d'une linéarité bien exprimée, avec une intensité de 100 à 200 nT. Elle est en fait continuelle, hors une discontinuité mince au sud de Pirot. On y a constaté plusieurs maximums ΔT partiellement individualisés.

Vers le nord, entre Trstenik et le Danube, la zone anomale n'a pas disparue. Cependant il serait très utile de savoir si elle ne s'ouvre de nouveau sur le territoire de la Roumanie.

La nature lithologique des éléments générant les anomalies est bien connue pour la ZET, cette zone est liée directement aux magmatites en affleurement. Elle est constituée notamment par des volcanites du Crétacé supérieur — où prédominent les andésites et les roches granitoïdes, le plus souvent présentant des anomalies positives, mais montrant aussi un champ réduit ou avec des anomalies intensivement négatives. Il est évident qu'il s'agit d'une magnétisation hétérogène, déterminée par la position de la masse des volcanites susmentionnées, par rapport aux necks, par les processus hydrothermaux ou par l'existence du volcanisme polyphasique ayant au moins une phase dans la position inverse des pôles magnétiques. Les volcanites fraîches, le plus souvent en relation directe avec les necks, sont exprimées par des anomalies d'intensité Δ T maximale. Les plaques effusives et spécialement les volcanites hydrothermalement altérées s'expriment par un champ plus bas, tandis que les magmatites de la phase d' inversion magnétique présentent des anomalies intensément négatives. Les granitoïdes laramiques ont provoqué seulement une anomalie vaste dans la région de Crni Vrh, où elle se ferme. Pourtant, il est tout à fait naturel de supposer que quelques-unes des anomalies de cette zone sont provoquées par les granitoïdes laramiques (par exemple dans le district de Zlotska reka), qui se trouvent en profondeur et ont une extension moins grande que celles de la zone de Crni Vrh (connue dans la littérature sous le nom de „Valja Strž").

On ne connaît pas encore suffisamment la nature des éléments perturbateurs de la zone située au sud de la ZET, où affleurent presque exclusivement des roches sédimentaires mésozoïques et tertiaires. Cependant il est évident que ces éléments perturbateurs sont de nature magmatique — des volcanites surtout, mais aussi des roches granitoïdes laramiques — étant donné la continuation de la zone anomale de la ZET et le fait que les rares affleurements des volcanites sont situés à l'intérieur de cette zone. D'après les calculs, la profondeur où se trouvent ces magmatites-volcanites est de 200 à 800 m, au-dessous de la couverture sédimentaire. Il faut admettre que les corps magmatiques relativement réduits peuvent être situés tout près de la surface, là où leurs parties apicales peuvent affleurer.

Donc, on peut affirmer, en considérant le caractère continu de la zone anomale et la nature lithologique de sa source, que la zone magmatique de Timok-Sredna gora s'étend en Yougoslavie à partir de Trstenik (Magdanpek), au nord, jusqu'à Borovo et la frontière yougoslave-bulgare, au sud. Elle se continue aussi bien en Bulgarie. Les magmatites de cette zone ont pratiquement une distribution continue. Les magmatites ne se prolongent pas au nord de Trstenik, au-dessous des schistes cristallins, au moins jusqu'au Danube, d'où nous avons les données magnétiques.

Il est évident que les magmatites, leurs zones profondes (necks, intrusions subvolcaniques hypo-abyssales, granitoïdes) sont contrôlées par une zone de failles profondes (faisceau des dislocations), représentant la partie resserrée de la structure dénommée dans la littérature "Graben synclinorium de Timok", "Zone de Timok-Sredna gora", "Zone de Timok", etc. Dans les limites de la ZET, la zone magnétique de Timok-Sredna gora correspond aux roches magmatiques affleurantes et à celles situées tout près de la surface. La zone avec la plus intense valeur du ΔT marque la zone des magmatites au sens plus resserré — la racine volcanique basique, qui a une forme de dyke de grandes dimensions. Elle se trouve au milieu de la ZET. Les autres parties de la ZET sont constituées de pyroclastites et d'écoulements volcaniques avec des necks secondaires, qui sont les chenaux d'effusion, marqués par une structure positive linéaire de second ordre. On peut dire que parmi les necks apparaissent également des plaques volcaniques avec altération hydrothermale. Vers le sud de la ZET, la zone magnétique est plus simple et plus étroite, située à une profondeur de 800 m au-dessous de la couverture sédimentaire.

On a pu relever les gisements minéraux cachés par la détermination des limites de l'extension de la zone magnétique, la corrélation des anomalies magnétiques avec les aréales de minéralisation connues, les régions et les espaces où se sont déroulés les processus métallogénétiques les plus intenses.

De nombreux gisements minéraux ont été découverts en Yougoslavie et en Bulgarie dans l'espace de la zone magnétique de Timok-Sredna gora. En Yougoslavie sont connus les gisements minéraux de cuivre de la région de Bor-Majdanpek. Dans la ZET sont localisés les plus importants gisements de cuivre connus jusqu'à présent. Cette zone est de première importance économique en ce qui concerne les gisements de cuivre et pas seulement pour la Serbie Orientale mais aussi pour toute la Yougoslavie. Le problème à résoudre est de sélecter les parties de la ZET où les nouveaux gisements minéraux peuvent être trouvés, notamment, les gisements "cachés" sans manifestations notables en affleurement. Puisque les régions, où les roches magnétiques affleurent, ont été intensément étudiées jusqu'à présent il faut tester les perspectives des gisements "cachés", surtout là où les magmatites se trouvent au-dessous de la couverture. Le problème qui reste à résoudre pour les recherches aéromagnétiques est de préciser les endroits où les magmatites se trouvent sous la couverture sédimentaire ou sous les pyroclastites, cette tâche étant de premier ordre pour la recherche du cuivre et des matières premières accompagnantes. C'est le cas surtout de la zone située au sud de la ZET, où la zone magmatique n'est pas large

Dans la partie médiane de la ZET, la zone relativement riche en minéraux est plus large. C'est pourquoi il faut faire une sélection plus approfondie en partant des données aéromagmatiques, avant d'entreprendre des recherches plus détaillées. Vu les positions des gisements minéraux de cuivre connus vis-à-vis de la configuration du champ anomal, il résulte que les terrains productifs sont localisés à côté des anomalies positives qui correspondent aux chenaux effusifs (necks volcaniques) — ou dans le champ anomal diminué, dû aux processus hydrothermaux intensifs, leurs effets ne pouvant pas être observés en affleurement. Les régions où la probabilité d'existence des gisements minéraux est plus grande et qui doivent être investiguées en détail, sont clairement illustrées sur les cartes magnétiques.

[3] On ne comprend sous ce nom que la zone fracturée avec des magmatites (racines volcaniques, corps intrusifs subvolcaniques et hypo-abyssaux, roches granitiques) qui ne peut pas être comparée à la structure connue dans la littérature sous le nom de : "Graben synclinorium de Timok", "Zone de Timok-Sredna gora", "Zone de Timok" etc.

О ГЕНЕЗИСЕ ЗАПАДНОРОДОПСКОГО ПОРФИРОКЛАСТИЧЕСКОГО РИОЛИТОВОГО КОМПЛЕКСА [1]

Д. БАХНЕВА [2]

В группе кислых вулканитов, из-за специфики петрографических особенностей, механизма образования и геологического положения, выделяются особые породы, известные как игнимбриты, туфолавы, невадиты игниспумиты, автомагматические брекчии, флюидпорфиры, флудтуфы и др. Кроме концепций о туфовом или лавововом характере этих пород, по вопросу их генезиса является новый взгляд Масуренкова (1961), что при формировании игнимбритообразующих магм первостепенная роль отводилась процессам плавления пород кислого состава. По существу, с дальнейшим изложением материала предлагается на обсуждение идея о формировании Западнородопского комплекса (представителя игнимбритоподобных ассоциаций) в результате неполного плавления пород кристаллического фундамента.

Комплекс сложен риолитами и риодацитами, в меньшей степени лавокластитами, игнимбритами лавового происхождения и пирокластитами. Вулканические продукты формировались в трех последовательных фазах. Фациальные виды не отличаются по составу и главным структурно-текстурным признакам. Комплекс характеризуется наличием особых литологических и текстурных разновидностей, описанных в дальнейшем в качестве переходных: спекшиеся, сваренные туфы, туфолавы, игнимбриты *(s.s.)*, черные гиалориолиты, автомагматические брекчии, ксено- и кластолавы. Несмотря на небольшие объемы они имеют повсеместное распространение, локализуясь в основании разреза или на периферии вулканических массивов. Вулканиты комплекса отличаются типоморфными петрографическими чертами — большое количество фенокристаллов — от 43 до 68%, чаще всего 50—52% (в автомагматических брекчиях — до 80%) и кристаллокластическим характером вкрапленников (кварц, андезин An_{34-52}, санидин, биотит, амфибол). Степень кристалличности является почти предельной, если согласиться с мнением (Масуренков, 1979), что 60% фенокристаллов можно принять рубежом вулканических пород. Литокласты и минералы метаморфических пород, как и гломеропорфировые сростки плагиоклаза встречаются

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редко в типичных вулканитах, но они характерны для переходных видов придавая им туфовый облик. Типичны также „фьями" стекла.

Если идея о происхождении порфирокластических риолитов в результате неполного плавления субстрата правильна, то реликтовые формы следует ожидать в наиболее ранних вулканических продуктах и в породах, в которых процессы постликвидусного преобразования способствовали сохранности промежуточных состояний. Особо информативными оказались черные гиалориолиты. Как продукты быстрой консолидации и переохлаждения, в них законсервированы отдельные звенья длительного процесса фазовых превращений в системе „кристаллический субстрат — расплав — вулканит". В петрографическом аспекте в работе раздельно рассматриваются особенности твердой фазы (на уровне фенокристаллов и включений) и жидкой фазы (основная масса $\pm$ микролиты).

Вкрапленники в породах с максимальной сохранностью „первичных" признаков обладают рядом особенностей. Кварц по периферии зерен гранулированный. Большие кристаллы (до 7—8 мм) обычно полуавтоморфные, с разрушенной половиной индивидов. Часто встречается „растрескивание" крупных зерен по извилистым концентрическим и радиальным трещинам, вдоль которых выделяется полоса новообразованного (?) кварца без включений (рис. 1). Чаще всего в центре плагиоклаза сохранены корродированные кристаллы более основного состава (An_{37-39}) и псевдоизометрической формы, характерной (К о с т о в, 1978) для метаморфических пород. Более кислый, свежий плагиоклаз (An_{31-34}) обрастающих зон достраивает ксенокристаллы до таблитчатых индивидов

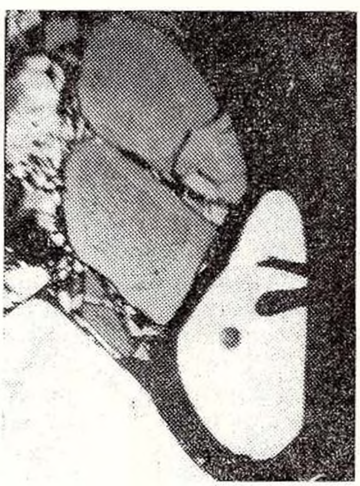

Рис. 1. — Порфирокластический кварц с новообразованной (?) каймой без включений. N+, увел. 15.

и обусловливает шахальбитовую структуру. Нередки случаи когда ядро переполнено каплями стекла или полностью витрофицировано, что определяет ситовидный и губчатый вид минерала. Санидин в переходных разновидностях отсутствует или встречается в незначительном количестве, но зато в основной массе встречаются прямоугольные пятна светлого стекла с ортогональной отдельностью (рис. 2). По исключению наблюдается зональное строение минерала — переплавленное ядро и полуавтоморфная периферия, отличающиеся интерференционной окраской, но с общей, почти одноосной фигурой. Среди цветных минералов амфибол по количеству почти не уступает биотиту. Биотит присутствует в двух генера-

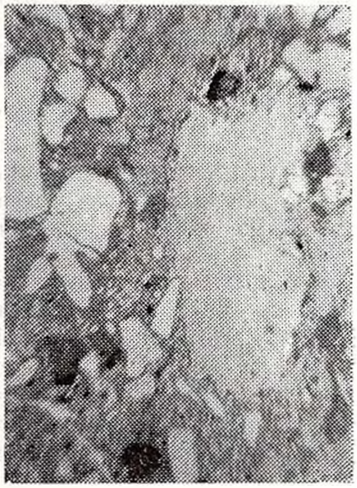

Рис. 2. — Светлое стекло на месте калишпата (?), N||, увел. 50.

циях. В индивидах первой генерации часто сохранены реликты амфибола или спаянность под углом 125°. Сам минерал на ранней стадии преобразования становится коричневато-бурым, а на поздней мутнеет, распадается,

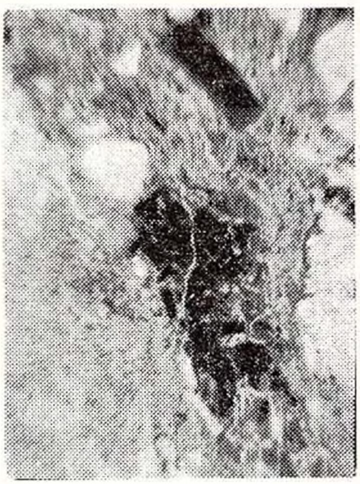

Рис. 3. - Разложенный ксеногенный биотит (I ген.) и магматогенный биотит (II ген.), N||, увел. 50.

образуя агрегат разноориентированных чешуек с общим опацитовым контуром. Вместо биотита, после разложения, в стекле остаются коричневые пятна, содержащие пылевидные обособления Fe. Биотит второй генерации зеленовато-коричневый, с резкими контурами (рис. 3). Амфибол обычно встречается в ксеноморфных сплавленных зернах в переменных размерах и количествах. По трещинам спаянности наблюдаются новообразования биотита. Акцессории относительно крупных размеров (титанит, рудный и апатит) по морфологии очень сходны с акцессорными минералами включений и вмещающих метаморфических пород. По мере увеличения степени преобразования количество и размеры зерен резко сокращаются.

В сильно преобразованных разновидностях, более близких к вулканитам, кварц и плагиоклаз обычно настолько „съедены" стеклом, что часто являются с цементной структурой. Обломки вкрапленников сдвинуты и смещены, сохраняя общую оптическую ориентировку. В ядрах минералов реликтовые признаки практически не встречаются. Биотит и амфибол являются чаще всего в автоморфных индивидах, с чистыми контурами, без явлений плавления и опацитизации.

Включения. В понимании механизма плавления особую роль, по-видимому, следует отвести включениям, как это подчеркивается в последние годы (Масуренков, 1961, 1979; Наседкин, 1975; Кутыев, Шарапов, 1979). При рассмотрении включений в исследованных вулканитах исходилось из того, что постоянное присутствие ксеногенного материала в игнимбритовых комплексах не может иметь случайный характер и является генетически обусловленным. При изучении Западно-родопского комплекса, согласно классификации включений (Масуренков, 1979), к ксенолитам отнесены литокласты песчаников, туфопесчаников палеогена, реже гнейсов, кристаллокласты серицитизирован-

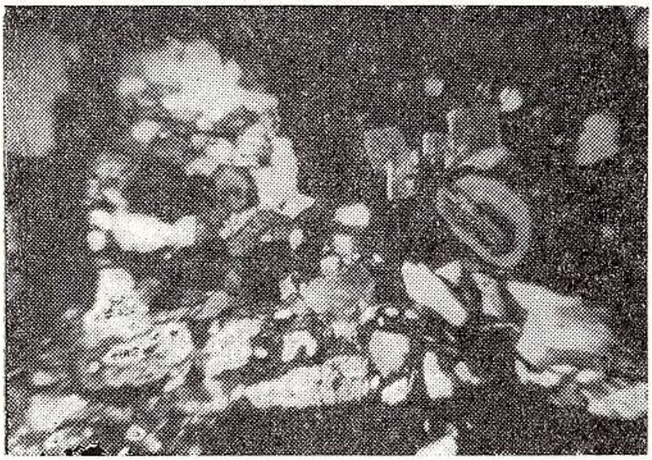

Рис. 4. — Плагиоклаз зонального строения в амфиболитовом реликтолите, N+, увел. 50.

ного плагиоклаза, микроклина, пироксена, граната, часть рудного акцессория, апатита и титанита. При более интенсивных преобразованиях, в связи с наложенным плавлением, ксенолиты теряют резкие контуры, дезинтегрируются частично и превращаются в реликтолиты. По исходному составу разграничены реликтолиты амфибол-биотитовых, двуслюдяных, аплитоидных гнейсов, амфиболитов, редко гранитоидов и мраморов. Новообразованное стекло распространяется вдоль сланцеватости пород, раньше всего между зернами кварца и полевого шпата. Плагиоклаз и кварц, хотя и богаты включениями как во вмещающих метаморфитах, очищаются по периферии, а местами гранулируются. Иногда плагиоклаз амфиболитовых реликтолитов приобретает зональное строение (рис. 4). Наблюдаются случаи нарастания свежего полуавтоморфного плагиоклаза на реликтовом ядре более основного состава в пределах амфиболитовых включений (рис. 5). Мафиты в реликтолитах обычно слабо

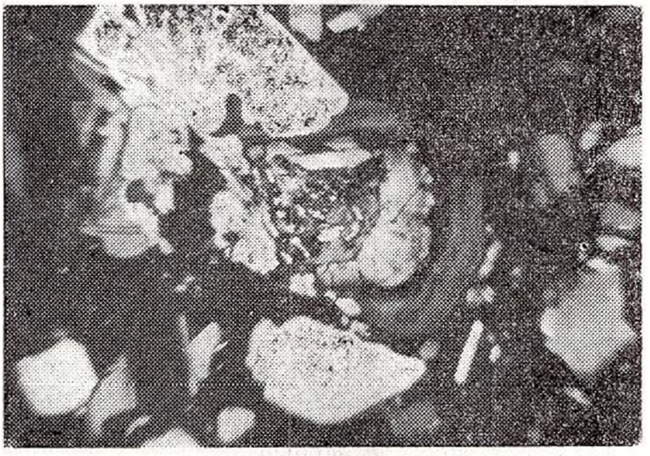

Рис. 5. — Гетерогенный фенокристалл плагиоклаза реликтово-кристаллизационного происхождения, N+, увел. 50.

опацитизированы и содержат капли стекла. Роговая обманка часто замещается микрочешуйчатым биотитом. Наблюдается случай, когда полоски, обогащенные роговой обманкой из включения, сменяются во вмещающем риолите вулканическим стеклом с реликтами амфибола (рис. 6). Кроме полиминеральных реликтолитов, породы заполнены единичными ксенопорфирокластами кварца с мозаичным, волновидным или простым угасанием, осколочной , округленной или полуэвгедральной формы, часто с корозионными контурами. Значительно реже встречается микроклин или калишпат с неоднородным угасанием, двузональный плагиоклаз с ксеногенным ядром и частично опацитизированные мафиты. К автолитам, по-видимому, следует отнести единичные включения с гомеогенным обликом и аплитовым составом, характеризующиеся прозрачными полевыми шпатами. Такой же генезис имеют, по всей вероятности, скопления автоморфных субпорфиров плагиоклаза, встречающиеся в виде пятен в отдельных шлифах, а также гломеропорфировые группы плагиоклазового или кварц-полевошпатового состава.

Хотя в одном и том же образце можно встретить включения разной степени переработки, ксенолиты характерны для спекшихся туфов и туфолав; для черных гиалориолитов и игнимбритов типичны реликтолиты. По мере увеличения степени структурно-минеральных преобразований, в разных типах пород уменьшается количество и степень ксеноморфности порфирокластов, а также содержание ксеногенного материала.

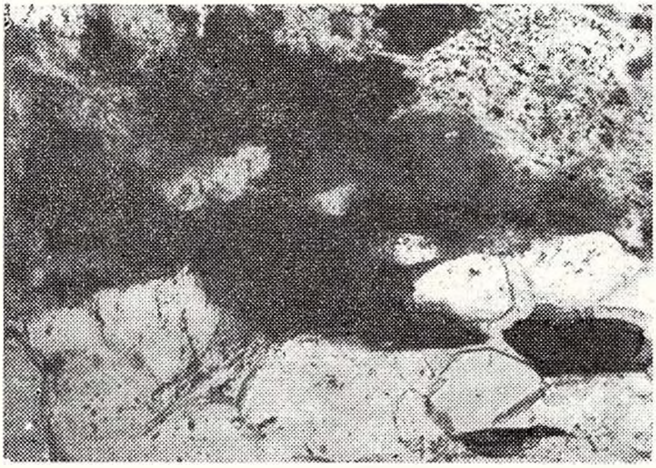

Рис. 6. — Контакт амфиболитового включения с риолитом, включающим реликты роговой обманки, N||, увел. 100.

Основная масса. В вулканитах ранних извержений, в относительно слабо переработанных видах и вблизи реликтолитов, сохранены элементы лепидо-, нематобластовой, свилеватой, очковой текстур, напоминающие полосчатые и порфиробластические разновидности метаморфитов фундамента. В таких участках псевдофлюидальность обусловлена чередованием темноокрашенных, более высокопреломляющих, с бесцветными ($N_{1.490}$) микрополосками. Различие в составе особенно четко вырисовывается по продуктам разложения и раскристаллизации стекла. Контуры двух стекол резкие, часто пламевидные и молниевидные. Само стекло — пузырчатое, микропористое, иногда с перлитовой отдельностью. Реликтовая полосчатость поддерживается и на уровне порфирокластов. В светлых полосках встречаются остатки разрозненных, сдвинутых кварцевых, реже плагиоклазовых обломков с ,,островной'' структурой. Кроме перечисленных минералов, темноокрашенные полоски содержат реликты биотита и амфибола в ассоциации с рудным акцессорием (до 1,5 мм) и титанитом (до 0,5 мм), а также витрофицированные реликты плагиоклаза. Разделение не только по составу, но часто и по крупности, дополняет впечатление об унаследованной порфиробластической текстуре. Почти мономинеральные чечевицы с крупными (до 7—8 мм) ,,глазками'' полуавтоморфного кварца чередуются с мелкообломочными ,,слойками'', содержащими кроме саличных и мафичные минералы.

В более интенсивно преобразованных породах преобладают эмульсионные структуры типа ориентированной, пепловидной, игниспумитовой, микрофлюктуационной. Интерес представляет, установленное рентгеном, по всей вероятности в основной массе, значительное количество, кристобалита (рис. 7).

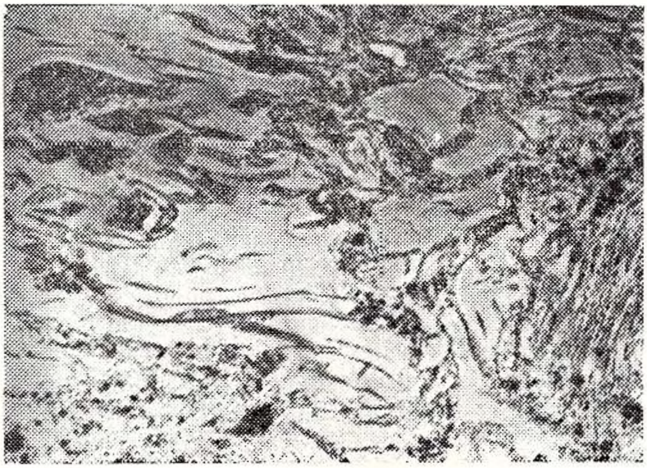

Рис. 7. — Эмульсионная пепловидная структура, N||, увел. 50.

Обсуждение результатов

Прослеживая разностепенные превращения в исследованных породах можно предположить следующий порядок устойчивости минералов при плавлении и реакционном замещении: калишпат-плагиоклаз-амфибол-биотит-кварц. В таком свете часть вкрапленников и акцессориев кварц, пироксен(?), крупный титанит, апатит и рудный минерал можно рассматривать как субликвидусные реликты, а плагиоклаз, амфибол, биотит и санидин — как гибридные. Такой порядок мобилизации вещества является схематическим, имея в виду совместное нахождение в шлифах амфибола, биотита и пироксена, микроклина и санидина, разложенного и свежего биотита, реликтового и новообразованного плагиоклаза. Эти ассоциации обычны для слабо переработанных разновидностей и свидетельствуют о сильно неравновесной системе в условиях резкого охлаждения и переохлаждения. Сходная с вышеуказанной последовательностью известна из экспериментального плавления гнейсов (Steuhle, 1962; цит. по Винклеру, 1979) и гранитов (Kranck, Oja, 1960; Luth, Tuttle, 1969; Кутыев, Шарапов, 1979). Предположительный ряд подвижности элементов в случаях, описываемых нами, следующий: K, Na — Ca — Mg — Fe — Al — Si. Эта последовательность почти полностью совпадает с модельной мобилизацией и выносом флюидов при 2000 атм: щелочей — при T = 650 — 700°, Al — при T 750° и Si — при 800° (Летников и др., 1975).

Установленные нами петрографические особенности хорошо согласуются с изменениями и их последовательностью, наблюдаемыми в частично плавленных пирометаморфических продуктах (Mc Birney, 1979). Отмечается, что в контактном ореоле внедрившихся магматических тел минералы вмещающих пород меняют оптические свойства: калишпат отступает частично, а с приближением к контакту — полностью ортоклазу (Менерт, 1971) или санидину (Al-Rawiy, Charmichael, 1967). Саличные минералы становятся трещиноватыми, нередко гранулированными, а мафиты меняют свой цвет, опацитизируются и разлагаясь выделяют окислы железа, титана, местами ортопироксена и плагиоклаза. Первичные плагиоклазы характеризуются губчатыми и ситовидными структурами. Первые следы стекла появляются на границе кварца с полевым шпатом, причем плагиоклазы исчезают позже или одновременно с калишпатом (Butler, 1961; Al-Rawiy, Charmichael, 1967). Новообразованное стекло в пирометаморфических продуктах черное или коричневое, богатое пузырьками и часто с перлитовым строением.

Сходные явления наблюдались и при экспериментальном плавлении гранитов. При атмосферном давлении (Кутыев, Шаранов 1980) в интервале 800—850° начинается плавление по границам калишпата и кварца, внутри зерен плагиоклаза и калишпата; биотит разлагается на бурое стекло и магнетит, иногда с кристалликами гиперстена; амфибол распадается на рудную пыль и плагиоклаз. Калишпат исчезает полностью при 1000° и вместо него появляется светлое пузырьчатое стекло. В опытах Винклера (1979) при P_{H_2O}=5 kbar и T=680°C при анатексисе мусковит-плагиоклаз-кварцевого гнейса без калишпата получены силиманит и калишпат. Сходные соотношения наблюдались нами когда по серицитизированному плагиоклазу образовался калиевый полевой шпат.

Исходя из экспериментальных результатов, можно выбрать вариант, наиболее близкий к природной системе, предполагаемой в качестве исходной для Западнородопского комплекса. Он включает пятикомпонентную комбинацию Q — Ab — An — Or — H_2O — при невысоких давлениях (2—5 kbar) и температурах (640—750°), при ограниченном количестве воды. При давлении 4—5 kbar, которое, примерно, соответствует предполагаемой глубине магматической камеры (Бахнева и др., 1978), для появления расплава-минимума необходимы 10—6% H_2O и соответственно T°=650—700° (Tutlle, Bowen, 1958; Менерт, 1971). Вряд ли петрологически возможно существование таких количеств воды в начале плавления. Кстати, в исследованных нами образцах содержание ,,обсидановой'' (?) воды ниже этих значений. В таком случае следует ожидать, что расплавится только часть субстрата, так как недостаток воды мешает полному плавлению. Экспериментами показано (Наседкин и др., 1975), что при P=5 kbar в интервале 570—600° и H_2O 7,5=% получено лишь 10—15% расплава. Для 40% расплава (что, примерно, соответствует количеству основной массы в исследованных породах) необходимы H_2O = 5% и T=670°. Так как давление мало влияет на плавление при недостатке воды, то температура является главным фактором в процессе мобилизации. Известно, что игнимбриты характерны для орогенических областей, в которых величина геотермического градиента достаточно высокая — от 50—65 до 150—180°/km, для достижения температуры плавления на глубине нескольких километров (Менерт, 1971). Расплавы, продуцированные анатексисом пород амфиболит-гранулитовой фации, нормально должны быть недонасыщены водой (W i l l i e, 1979). Это тоже указывает на возможность генерирования кислых магм вышеуказанным способом.

Вопрос о генезисе порфирокластических комплексов типа Западнородонского является весьма сложным, чтобы принять высказанные соображения как единственно возможные. Вполне вероятно, что процесс игнимбрито-магмообразования результат многофакторный, в котором не малую роль играет конвергенция, например, черных гиалориолитов, интрузивных и эксплозивных игнимбритов. Хорошо увязываются с этими представлениями геологические и петрографические особенности исследованного комплекса. Трифазовое формирование, например, Западнородонского риолитового ареала интерпретируется как трехкратная мобилизация субстрата с достижением до состояния расплава в объеме почти предельном для вулканических пород (к 50%). Влияние вмещающей среды двоякое. Отрицательный эффект выражен в получении ,,гранитной магмы'', независимо от состава исходных пород, причем только реститы темноокрашенного компонента свидетельствуют о первичном составе. Логическое объяснение этой особенности можно найти в экспериментах плавления глин, граувакков, гнайсов, гнайсов в присутствии воды (В и н к л е р, 1979). Положительное влияние среды демонстрируется парагенетическим рядом ,,включения — фенокристаллы — вулканиты''. В зависимости от состава магмогенерирующей среды включения в вулканитах ранних импульсов только метаморфогенные, а в продуктах второй и третьей фазы — и риолитовые. По-видимому, ассимиляция является существенной причиной для разнообразия не только Са-щелочных, ассоциирующих с андезитами и щелочных пород (M c B i r n e y, 1979), но и для самостоятельно проявляющихся кислых вулканитов.

Выводы

1. Предполагается, что Западнородонский игнимбритоподобный комплекс имеет гетерогенную природу-магматическую для основной массы и отчасти реликтовую для вкрапленников. В образовании порфирокластов предусматривается участие пород амфиболитовой фации регионального метаморфизма.

2. В петрогенезисе вулканитов выделяются два этапа: в первом, ликвидусном (мобилизация субстрата), подобно анатектитам, плавлению подвергаются преимущественно светлоокрашенные, а замещению — темноцветные минералы; во втором, солидусном этапе (консолидация вулканитов) обособляются полигенные фенокристаллы, реликтово-реакционно-кристаллизационного происхождения.

3. Полевые, петрологические и экспериментальные данные свидетельствуют о возможности корового игнимбрито-магмообразования в условиях низких T° и P при недостатке воды. Процесс частичного плавления, подобно мигматизации и в отличие от кристаллизационной дифференциации, протекал при увеличении T°, а также летучих, в особенности щелочей.

4. Можно предположить, что игнимбритоподобные комплексы типа Западнородонского, отражают в значительной мере фосилизированное

ОФИОЛИТОВЫЕ И ОСТРОВОДУЖНЫЕ КОМПЛЕКСЫ УКРАИНСКИХ КАРПАТ [1]

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Идеи новой глобальной тектоники, оформившиеся в концепцию образования верхних оболочек Земли, заложены в 60-е годы исследованиями Г. Хесса, Р. Дитца, Ф. Вайна, Д. Меттьюза, Дж. Т. Вильсона, У. Моргана, Ле Пишона, Б. Изакса, Дж. Оливера, Л. Сайкса и др. Эти идеи широко развиваются в нашей стране и опробируются в разных структурных элементах земной коры с целью выявить геодинамические условия прошлого и историю формирования верхних оболочек Земли. Палеотектонические реконструкции выполнены и для альпийского Средиземноморского пояса, в том числе и для Карпат (Данилович, 1975; Доленко, Данилович, 1975; Доленко и др., 1978, 1980). Большое значение в проведении этих работ имело использование показательных формаций, и в первую очередь магматических, которые формируются в строго определенных геодинамических обстановках. Развитие Карпат с точки зрения концепции тектоники литосферных плит можно представить как образование складчатых областей по типу столкновения континент-островная дуга-микроконтинент.

Петрологические исследования показали наличие на территории Украинских Карпат как пород офиолитовой ассоциации — показателей спрединга, так и известково-щелочных вулканических серий островных дуг.

Породы офиолитовой ассоциации установлены в трех структурно-фациальных зонах: в Пьенинской и на продолжении ее в фундаменте Закарпатского прогиба ; в Мармарошской утесовой зоне и в зоне надвига флиша Раховско-Буркутской группы зон на Силезско-Черногорскую (рис.). В виде фрагментов толщ они входят в состав меланжа, маркирующего зоны субдукции, слагают останцы покровов и олистолиты. Известны триасовые, юрские и нижнемеловые комплексы офиолитов. Мощность фрагментов свыше 400—450 м. Они интенсивно разлинзованы, тектонически нарушены, аллохтонны. Магматические породы ассоциируют с осадками разных глубин седиментации (кремнистые известняки, темноцветные кремнистые аргиллиты, черты, радиоляриты и др.). Среди них главное место занимают диабазы, спилиты, вариолиты, базальты, реже

[1] Работа была представлена на XII-ом Конгрессе Карпато-Балканской Геологической Ассоциации, 8—13 сентября 1981 г., Бухарест, Румыния.

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габбро. Это производные базальтовых магм толеитового типа, менее развиты щелочные разновидности. По химизму они близки к толеитовой серии современных океанов, образующейся в зонах спрединга: в них низкое содержание калия (0,03—0,18%) На диаграмме AFM они расположились вдоль тренда дифференциации гавайских толеитов. Породы с содержанием калия 0,3% и выше имеют щелочной характер, отражающий химизм родоначальной магмы. Петрологические исследования показали

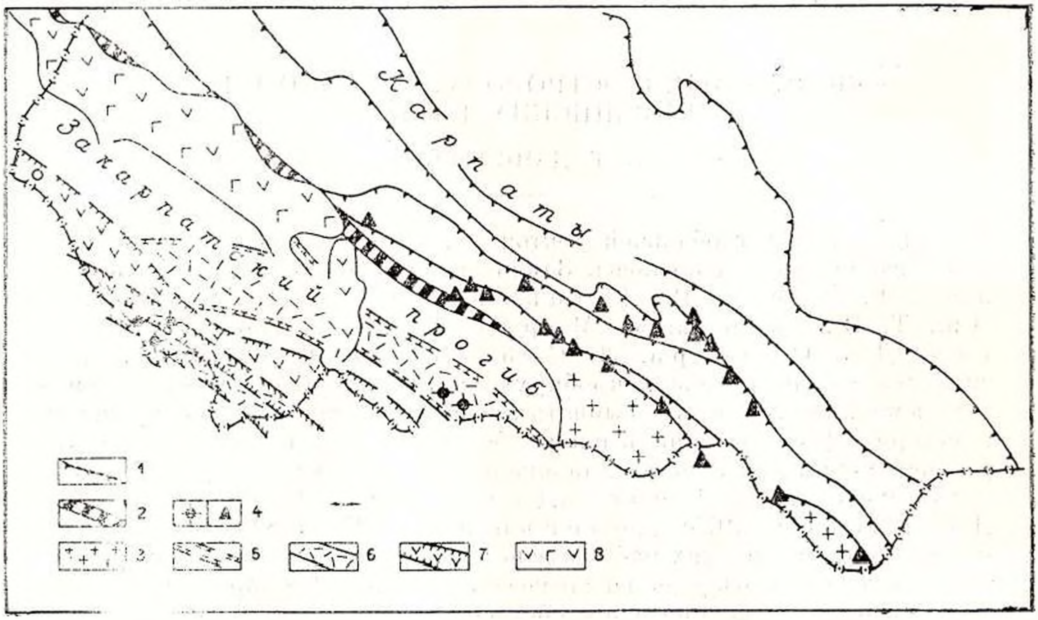

Рис. — Развитие офиолитовых и островодужных комплексов в Украинских Карпатах и Закарпатском прогибе.

1, границы структурно-фациальных зон во Флишевых Карпатах; 2, Пьенинская зона; 3, Мармарошская зона; 4, фрагменты офиолитовой ассоциации (триас — ранний мел): а, в скважинах, б, в зонах надвигов. Известково-щелочные островодужные серии: 5, проявления кислого вулканизма (игнимбриты, новоселицкие туфы, 18—16 млн. лет), 6, проявления кислого и 7, андезитового вулканизма (сармат, 15—9 млн. лет); 8, проявления андезито-базальтового вулканизма Выгорлат-Гутинской гряды (плиоцен, 12—8 млн. лет).

принадлежность офиолитовой ассоциации к серии океанических толеитов с нормативным оливином 7—15% и гиперстеном — 11-16%, менее к серии высокоглиноземистых оливиновых базальтов и щелочных оливиновых базальтов (без нефелина, с низким содержанием нормативного гиперстена). Они имеют примитивную геохимическую характеристику. Отношение Rb/Sr составляет 0,01, K/Rb — 1000 — 3000. Изучение изотопного состава стронция (Д а н и л о в и ч, 1977) показало низкие значения отношения Sr^{87}/Sr^{86} (0,702—0,704).

Гипербазиты (лерцолиты) установлены только в Мармарошской зоне утесов, где находятся вместе с другими фрагментами офиолитовой ассоциации и глаукофановыми сланцами (бассейн р. Малая Уголька и ручей Квасный). По составу соответствуют типичным альпинотипным разновидностям. Характерно преобладание оливина над пироксенами, среди пироксенов преобладает ромбический. Это шпинелевые разности с устойчивым парагенезисом: оливин-клинопироксен-ортопироксен-шпинель, что сближает гипербазиты Карпат с гипербазитами срединно-океанических хребтов.

В целом можно говорить о типичной офиолитовой ассоциации (лерцолит-толеитовый тип), связанной с зонами субдукции, и рассматривать фрагменты толщ как реликты океанской коры Тетиса и отдельных морских бассейнов, разделявших микроплиты.

Проявление кислого, андезитового и андезито-базальтового вулканизма в Закарпатском прогибе характеризует зрелую стадию островных дуг (неоген). Вулканизм протекал в разных обособленных зонах (рис.). В составе кислых вулканических образований (мощность >2 км) преобладают дацитовые игнимбритовые образования (серии поточных туфов, туфолавы), воздушноосажденные туфы, дациты, липариты (Данилович, 1976). Андезитовые комплексы состоят из вулканических (лавы и туфы) и субвулканических фаций (диорит-порфириты, гранодиорит-порфиры). В разрезе Выгорлат-Гутинской гряды развиты стратовулканические комплексы андезито-базальтового состава (серии базальт-липарит). Геохимические исследования и изучение изотопного состава стронция свидетельствуют о глубинном источнике магм во всех зонах вулканизма (Данилович, 1977).

Миграция процессов вулканизма во всех зонах происходила с запада на юго-восток. Вулканические образования принадлежат к известково-щелочным сериям островных дуг. С использованием бинарных диаграмм ,,окись калия — кремнекислота'' были определены расстояния к возможным очагам магм (100—200 км) и тем самым установлено направление наклона погружающейся океанской плиты. Смещение поверхностных проекций магматических очагов объясняется сложной динамикой горизонтальных перемещений всей системы сочленения ,,континент-микроконтинент''. Таким образом, в неогеновое время напряжения горизонтального сжатия, вызванные сближением Африканской и Евразиатской литосферных плит, привели к закрытию межплитных глубоководных альпийских бассейнов и их превращению в складчатые сооружения, они спаялись в единую литосферную плиту. Совмещенность вулканических дуг и меланжа субдукции на рассматриваемой части карпатского орогена свидетельствует о формировании последних по типу столкновения, о большой силе сжимающих усилий. Наличие офиолитовых и островодужных комплексов, меланжа, маркирующего зоны субдукции, показательных формаций разного типа позволили наметить эволюцию Украинских Карпат в альпийский этап. Выделяются стадии:

1. Стадия заложения (или рифтовая), характеризуется деструкцией коры континентального типа (Доальпийская), формированием в триасовую эпоху рифта, в котором накапливались известняки, кремнистые известняки, доломиты, щелочные базальты, конгломераты, песчаники и т.д. Они выявлены бурением в фундаменте Закарпатского прогиба (рис., скв. 2-Залуж, 952-Берегово).

2. Стадия раскрытия океана (или собственно океаническая). Во времени она соответствует юре и отражает формирование коры океанского типа с характерной для нее формацией пород офиолитовой ассоци-

ации (офиолиты, кремнистые известняки, радиоляриты, яшмы, кремнистые аргиллиты). В прилегавшей с внешней стороны области развивавшейся на континентальной коре, шло накопление терригенно-карбонатной формации (известняки, доломиты, мергели, песчаники, конгломераты и т.д.).

3. Стадия сокращения океана охватывает мел-палеоген и характеризуется формированием флиша, что свидетельствует о смене условий растяжения — условиями сжатия.

Островодужный вулканизм протекал южнее. В фундаменте зон вулканизма установлен меланж — продукты тектонического скучивания.

4. Стадия замыкания океана (или орогенная) завершает развитие в конце неогена Карпатского орогена. Она характеризуется образованием молассовой формации в краевых прогибах, проявлением кислого, андезитового и андезито-базальтового вулканизма и последующей инверсией геотектонического режима.

В заключение отметим, что данная модель, построенная с преломлением идей тектоники плит, пока схематична, но она отражает качественно отличные периоды истории развития Карпат и объясняет такие явления, как наличие мощного флиша, его покровное строение, одностороннюю вергенцию в сторону платформы, особенности строения земной коры Закарпатского прогиба (близка к островному типу), типы магматизма и другие и намечает прогрессивный подход в решении ряда проблем геологии, магматизма, тектоники и металлогении Карпат.

ЛИТЕРАТУРА

Данилович Л. Г. (1975) Магматизм Карпат и вопросы геодинамики. — Геология и геохимия горючих ископаемых, 42, с. 3—13.

— (1976) Кислый вулканизм Карпат, Киев, Наук. думка, 148 с.

— (1977) Петрогенезис магматических образований Карпат по данным изучения изотопного состава стронция. *Геол. журнал*, 1977, 37, 4, с. 50—62.

Danilovich L. I. (1978) Geochemistry of trace elements in acid volcanic rocks of the Ukrainian Carpathians. *Geol. Zb. Geologica. Carpathica*, 29/2, p. 241—252.

Доленко Г. Н., Бойчевская Л. Т., Данилович Л. Г. и др. (1980) Глубинное строение, развитие и нефтегазоносность Украинских Карпат. Киев, Наук. думка, 147 с.

ВОПРОСЫ

Д. Ховорка: Существуют ли опубликованные данные касающиеся синих сланцев Советских Карпат?

Ответ: Мне известны только некоторые синие сланцы в долине реки Малая Уголька. До сих пор не были опубликованы результаты петрологических исследований.

Т. Визер: Нашли ли Вы тешениты в Украинских Карпатах? Какое отношение между ними и диабазами и спилитами?

Ответ: Пород типа тешенитов на нашей территории не обнаружено.

СТРУКТУРНЫЕ УСЛОВИЯ ЛОКАЛИЗАЦИИ РУДНЫХ МЕСТОРОЖДЕНИЙ В ОБЛАСТИ ИСКЫРСКОГО УЩЕЛЬЯ (ЗАПАДНАЯ СТАРА ПЛАНИНА)[1]

СИМЕОН КАЛАЙДЖИЕВ[2]

Регион расположен к северу от г. Софии по долине р. Искыр. Здесь сосредоточены наиболее крупные в стране стратиформные месторождения полиметаллических, медных и железных руд. Значительно распространены и различные по составу жильные минерализации.

Исследуемая область сложена преимущественно осадочными породами. В Болгарии фанерозойские осадочные геокомплексы уже выделены (Н а ч е в и Я н е в, 1980). Возле Искырского ущелья разграничиваются следующие осадочные геокомплексы: Берковско-Шипкинский алевролито-аргиллито-сланцево-филлитовый (кембрий, возможно частично и ордовик), Искырский кварцито-аргиллитовый (ордовик-силур-девон), Кучай-Черногорский граувакко-алевролитовый флиш (верхний девон, возможно и нижний карбон), лимнический конгломерато-песчаниковый (намюр-вестфаль, верхний стефан — нижняя пермь), песчаниково-брекчеконгломератовый (нижняя-верхняя пермь), базальный песчаниковый (нижний триас), доломито-известняковый (кампил-карн), регрессивный брекчеконгломератоаргиллитовый (верхний триас), трансгрессивный мергельно-известняковый (геттанг-синемюр), аргиллитово-алевролитовый (аален-средний байос), регрессивный песчаниково-известняковый (верхний байос и бат), известняковый (верхняя юра), Ниш-Троянский граувакко-алевролитовый флиш (титон и берриас), Пирот-Тырновский мергельно-известняково — глинисто-песчаниковый (берриас-апт), Среднегорский туфо-тефроидно-карбонатный (коньяс-кампан) и галечно-песчанистый (плиоцен) (рис. 1).

Установлен и ряд магменных тел, преимущественно группы „старопланинских плутонитов" доверхнекарбонового возраста (Д и м и т р о в, 1927), включающихся в габбродиорит-гранодиоритовую формацию (Д и м и т р о в а и др., 1975). Представителями щелочной Габбро-сиенитовой формации являются оба тела около Свидни и Сеславцев (Д и м и т р о в а, и др. 1975).

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[1] Работа была представлена на XII-ом Конгрессе Карпато-Балканской Геологической Ассоциации, 8—13 сентября 1981 г., Бухарест, Румыния.

[2] Н. Р. Болгария.

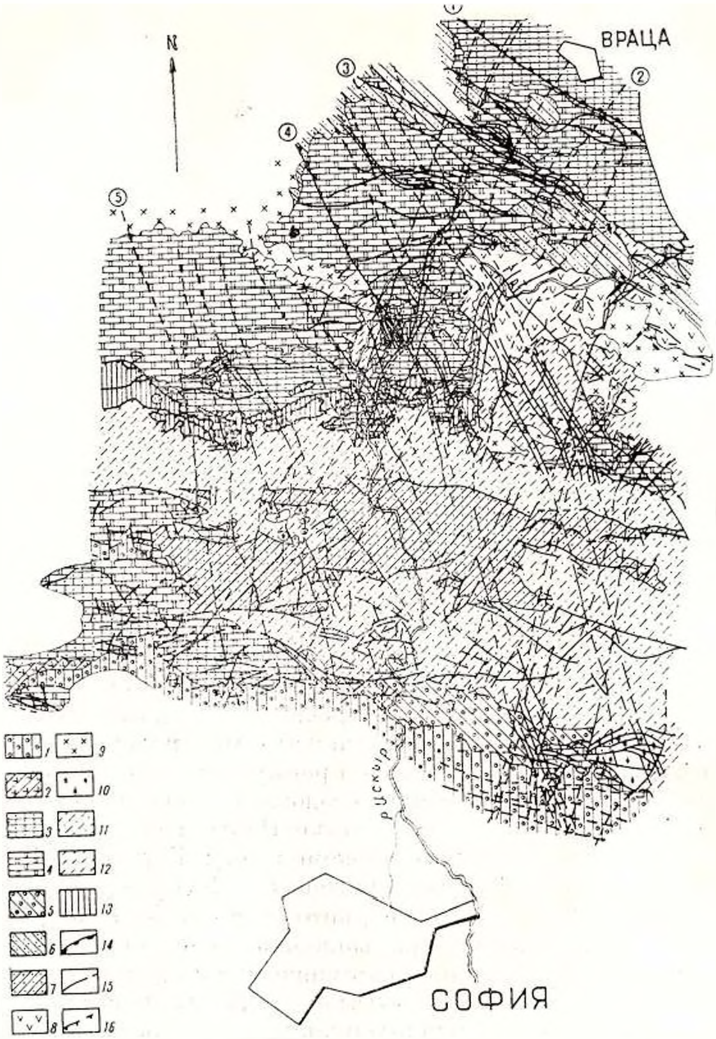

Рис. 1. — Геологическая карта области (составленная автором по данным Тронкова, Кожухарова и др., Пиронкова, Рашкова и др., Динкова и др., автора и др.).

1, плиоцен (галечно-песчано-глинистый геокомплекс); 2, коньяс-кампан (Среднегорский туфо-тефроидно-карбонатный геокомплекс); 3, юра — апт (Пирот-Тырновский мергельно-известняковый — глинисто-песчаниковый геокомплекс, Ниш-Троянский граувакко-алевролитовый флиш; известняковый, регрессивный песчаниково-известняковый, аргиллито-алевролитовый и трансгрессивный мергельно-известняковый геокомплекс); 4, триас (регрессивный брекчеконгломератово-аргиллитовый, доломито-известняковый и базальный песчаниковый геокомплекс); 5, пермь (песчаниково-брекчеконгломератовый геокомплекс); 6, верхний стефан-пермь (лимнический конгломератово-песчаниковый геокомплекс); 7, намюр-вестфаль (лимнический конгломератово-песчаниковый геокомплекс; 8—9, габбро-диорит — гранодиоритовая формация (8, диоритовые порфириты; 9, диориты, гранодиориты, граниты); 10, габбро-сиенитовая формация; 11, ордовик-девон (Кучай-Черногорский и Искырский кварцито-аргиллитовый геокомплекс); 12, кембрий (Берковско-Шипкинский алевролито-аргиллито-филлитовый геокомплекс; 13, аллохтон Издремецкой синклинали (мезозойские геокомплексы); 14, флексуры (1, Врачанская); 15, крутой разлом (2, Косталевский; 3, Плакалницкий; 4, Пребойницкий; 5, Брезенский); 16, линия надвигов.

Исследуемая рудная область охватывает части Берковского (Бончев, 1910, 1930) и Свогенского (Бончев, 1910, 1930) антиклинориев, а также заключенную между ними Издремецкую синклиналь (Белмустаков, 1951) (рис. 2).

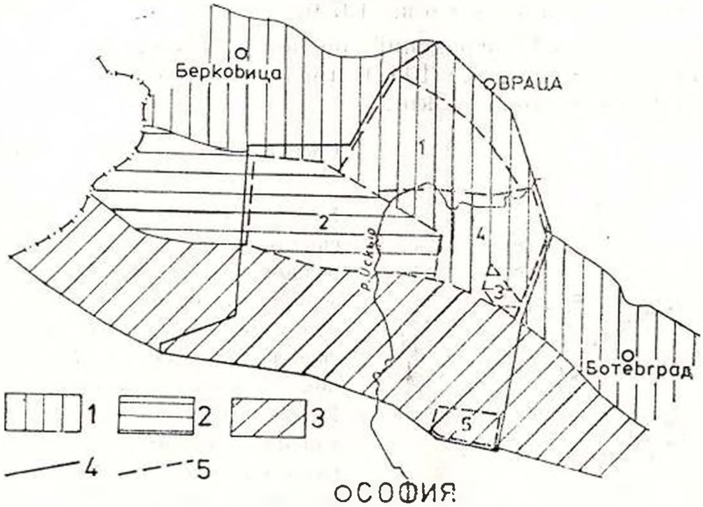

Рис. 2. — Схема главных складчатых структур и рудных полей области Искырского ущелья.

1, Берковский антиклинорий; 2, Издремецкая синклиналь; 3, Свогенский антиклинорий; 4, граница рудной области; 5, граница рудных полей: 1, Плакалницкое; 2, Издремецкое; 3, Осеновлашское; 4, Зверинское (1,2,3 и 4 рудные поля — Врачанско-Искырский рудный район); 5, Кремиковское.

Земная кора региона рассечена несколькими региональными руптурами, заложенными на значительной глубине. Это Балканский, Кюстендилский, Маданско-Михайловградский, Косталевский (Карагюлева, 1961; Бончев и Карагюлева, 1962; Тронков, 1965; Калайджиев, 1982) и Кремиковский разломы. Ориентировка их различная: ЗСЗ—ВЮВ до В—З (Балканский, Кремиковский), СВ—ЮЗ (Кюстендилский), ССЗ—ЮЮВ (Петроханский, Маданско-Михайловградский и Косталевский) (рис. 3).

Балканский глубинный разлом (Хайдутов и Иванов, 1965) представлен разрывами, сгруппированными в два пучка: Плакалницкий (на севере) и Видлицкий (на юге).

Старопланинская лобовая полоса сложена Плакалницким пучком разломов. Отдельные руптуры имеют простирание 110—140°. Ширина пучка достигает 11 Km.

Видлицкий пучок разломов представлен фрактурами В-З и ВЮВ—ЗСЗ направлений, являясь проявлением Забалканского глубинного разлома (Бончев, 1961). Характерны значительные надвиги к северу. Ширина пучка 7—8 km.

Кюстендилский разлом в области Стара Планины (Калайджиев, 1982) представляет собой палеозойскую магмопроводящую структуру. Отчетливо обособлен пучок разломов (Голямопланинский) шириной около 20 km.

Петроханский разлом представляет также магмопроводящую структуру с шириной 2—6 km, расположенную поперечно палеозойскому подвижному поясу (Хайдутов, 1979).

Маданско-Михайловградский разлом (Йосифов, 1976, 1977, 1979; Калайджиев, 1982) представлен Пребойницким пучком разломов шириной около 15 km.

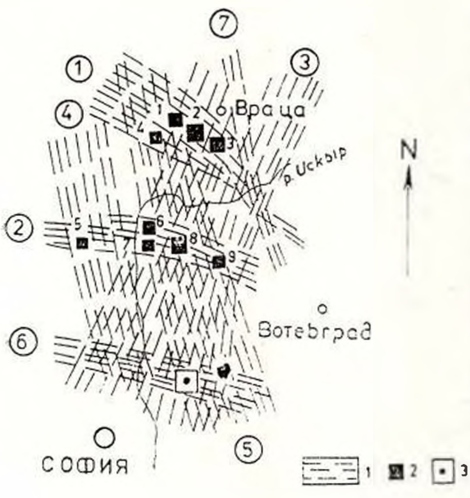

Рис. 3. — Схема регпональных разломов и рудных зон области

1, региональный разлом (1, Староплаништская лобовая линия, Плакалницкий пучок разломов; Салашско-Врачанская рудная зона; 2, Забалканский разлом, Видлицкий пучок разломов; Издремецко-Етропольская рудная зона; 3, Кюстендилский разлом, Голямопланинский пучок разломов; Голямопланинская рудная зона; 4, Петроханский разлом; 5, Маданско-Михайловградский разлом, Пребойницкий пучок разломов; Искырская рудная зона; 6, Кремиковский разлом, Негушевский пучок разломов; Кремиковская рудная зона; 7, Косталевский разлом); 2, полиметаллические и медные месторождения (1, Рупите; 2, Седмочисленици; 3, Плакалница; 4, Соколец; 5, Отечество; 6, Венеца; 7, Бов; 8, Издремец; 9, Рого); 3, железорудное месторождение Кремиковци.

Кремиковский разлом (Велчев и др., 1973) представлен Негушевской зоной разломов (Бончев, 1971) шириной более 5 km. Здесь произошло значительное субгоризонтальное продвижение в меридиональном направлении (Кремиковский надвиг).

Благоприятное положение исследуемой рудной области для локализации оруденения определяется прежде всего ареалом пересечения между Балканским, Маданско-Михайловградским, Петроханским и Кремиковским разломами (рис. 3).

В области выделяются пять рудных зон (рис. 3,4), вытянутых по длине больших разломов, с взаимным пересечением некоторых из них: Салашско-Врачанская, Издремецко-Етропольская, Искырская, Голямопланинская и Кремиковская. Врачанско-Издремецкий рудный район (рис. 2) охватывает части первых двух зон.

Салашско-Врачанская рудная зона (Калайджиев, 1967) приурочена к Старопланинской лобовой полосе. В исследуемом регионе по ее длине располагается Плакалницкое рудное поле, которое сформировалось на месте пересечения Плакалницкого, Голямопланинского и Пре-

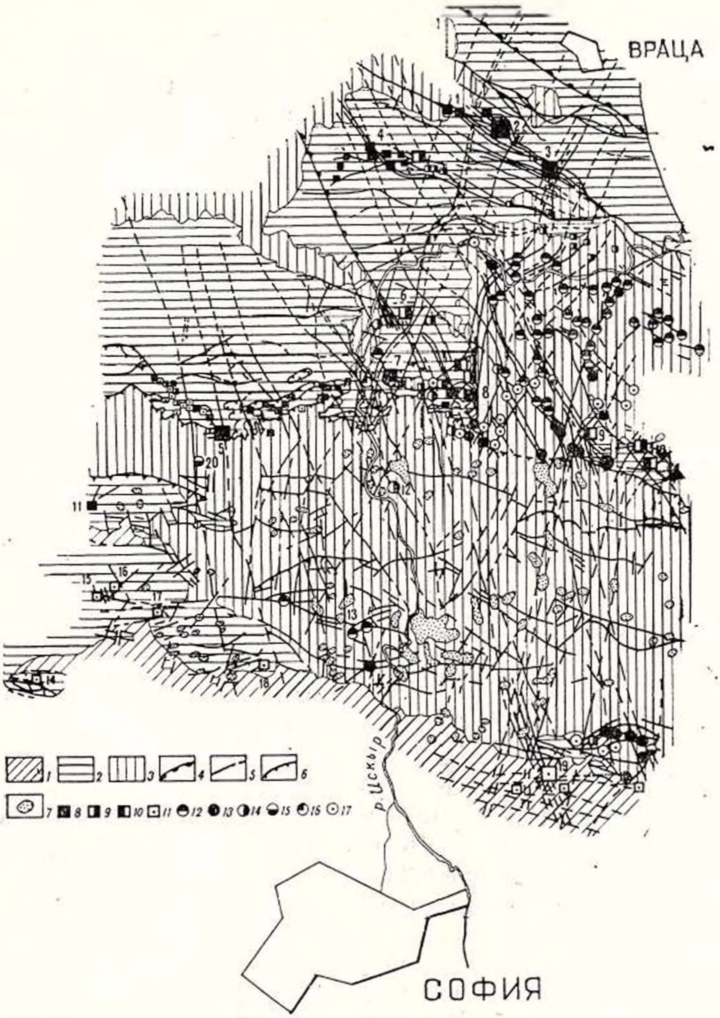

Рис. 4. — Карта эндогенных месторождений и рудопроявлений области Искырского ущелья

1, плиоцен; 2, мезозой; 3, палеозой; 4, флексура; 5, крутой разлом; 6, линия надвигов; 7, литогеохимические аномалии по вторичному ореолу рассеяния свинца, цинка, меди, серебра, бария и др.; 8, стратиформные месторождения и проявления полиметаллических руд (1, Рупите; 2, Седмочисленици; 3, Плакалница; 4, Соколец; 5, Отечество; 7, Бов; 8, Издремец; 11, Ковачица); 9, стратиформные месторождения и проявления медных руд (6, Венеца; 8, Издремец); 10, стратиформные месторождения и проявления свинцово-медных руд (9, Рого;10, Градиште); 11, стратиформные месторождения и проявления железа (14, Сливница; 15, Богьовци; 16, Понор; 17, Беледие хан; 18, Балша; 19, Кремиковци); 12, жильные месторождения и проявления барита (13, Владо Тричков); 13, жильные месторождения и проявления полиметаллических руд; 14, жильные медные месторождения и рудопроявления (12, Болшевик); 15, жильные проявления антимонита (20, Балимор); 16, молибденитовая жила; 17, жильное железное оруденение.

бойницкого пучков разломов. Здесь локализованы самые крупные в стране полиметаллические месторождения стратиформного типа — Седмочисленици и Плакалница. Большинство стратиформных типов орудениния приурочено к доломитам известняковой толщи триасского доломито-известнякового геокомплекса.

Издремецко-Етропольская рудная зона (Калайджиев, 1967) контролируется Забалканским разломом (Видлицкий пучок разломов). Изученная площадь преимущественно охватывает стратиформные месторождения полиметаллических руд Издремецкого и Осеновлашского рудных полей (Калайджиев, 1982). Орудениние локализовано в автохтоне и аллохтоне Издремецкой синклинали. Наибольшей является рудная концентрация в ареале пересечения руптур Видлицким, Пребойницким и Голямопланинским пучками (месторождения Издремец--юг, Издремец-север, Венеца, Бов). Рудовмещающими породами для автохтона являются преимущественно песчаники нижнего триаса и нижней юры, а для аллохтона — доломитизированные известняки и доломиты известняковой толщи триасского доломито-известнякового геокомплекса. Зона охватывает и южную часть Зверинского рудного поля. Рудные проявления в Зверинском рудном поле Врачанско-Издремецкого рудного района относятся к жильному типу и вмещены в палеозойском фундаменте.

Голямопланинская рудная зона (Калайджиев, 1982) вытянута по длине Кюстендилского разлома (Голямопланинский пучок разломов), располагаясь поперечно Салашско-Врачанской, Издремецко-Етропольской и Кремиковской рудным зонам. Она включает самые значительные в стране полиметаллические (Седмочисленици, Плакальница, Издремец-север, Издремец-юг, Венеца, Бов) и железные (Кремиковци) месторождения стратиформного типа.

Искырская рудная зона (Драгов и др., 1976; Калайджиев, 1982) заключена между Пребойницким и Косталевским разломами. Очень четко выражена рудоконтролирующая роль Маданско-Михайловградской дислокации. В ней сосредоточены полиметаллические месторождения Седмочисленици, Плакалница, Издремец-север, Издремец-юг, Венеца, Рого и др., а также и Кремиковское рудное поле.

Кремиковская рудная зона (Канурков, 1980) располагается по протяжению Кремиковского разлома. Характерными являются проявления железа, барита, полиметаллов. Наиболее значительно Кремиковское железорудное месторождение, входящее в пределы одноименного рудного поля. Благоприятное положение Кремиковского рудного поля определяется ареалом пересечения между руптурами Кремиковским, Маданско-Михайловградским и Кюстендилским разломами. Одновременно это рудное поле включается и в Голямопланинскую и Искырскую рудные зоны. Вмещающими породами для стратиформного типа орудиения в рудном поле являются доломитизированные известняки известняковой толщи триасского доломито-известнякового геокомплекса, слагающие Кремиковский надвиг.

Гидротермальное происхождение стратиформных месторождений полиметаллических, медных и железных руд в области Искърского ущелья не вызывает сомнений. Наряду с многочисленными другими доказательствами, выдвинутыми различными исследователями (Минчева-Стефанова, 1961, 1962, 1978 и др.; Атанасов, 1973 и др.; Попов и др., 1979; Канурков, 1980 и др.; Калайджиев, 1967, 1982 и др.; Станев и Панайотов, 1964 и др.), следует сообщить и о том факте, что одни и те же типы оруденения переходят непосредственно из автохтона в аллохтон.

Рудопроводящими каналами растворов явились Балканский, Кюстендилский, Кремиковский, Петроханский и Маданско-Михайловградский разломы. Рудораспределяющими структурами для стратиформных месторождений были надвиги, оперяющие разломы и трещины, древние денудационные поверхности, плоскости наслоений, пласты с повышенной пористостью и трещиноватостью.

Оруденение исследуемой области имеет поздний возраст, о чем свидетельствует прежде всего тот факт, что дорудные рудовмещающие надвиговые структуры формировались после значительной послеверхнемеловой денудации пород.

ЛИТЕРАТУРА

Атанасов Е. (1973) За структурата и естествените граници на медно-оловно-цинковото находище Седмочисленици в Западна Стара планина. *Год. Висш. минно-геоложки инст.*, 19, 2, 1—17.

Балмустаков Е. (1951) Върху геологията на Издремецката синклинала между с. Искрец и Церово. *Изв. Геол. инст.*, 1, 3—16.

Бончев Е. (1961) Бележки върху главните разломни структури в България. *Труд. геол. Бълг., сер. страт. и тек.*, 2,5—29.

— (1971) Проблемы на българската геотектоника. С., „Техника", 204.

— Ю. Карагюлева (1962) Въпроси из тектониката на Предбалкана. *Изв. Геол. инст.*, 10, 119—156.

Бончев С. (1910) Геология на Западна Стара планина. II — Главни линии от геологичния строеж (направа) на Западна Стара планина. *Труд. Бълг. природоизп. д-во*, 4, 1—59.

— (1930) Обяснение на листа Цариброд от геоложката карта на България в мярка 1: 126000. *Универ. библ.*, 114.

Велчев Ц., Р. Димитров, Х. Дачев, Й. Теодосиев, Х. Генчев (1973) Върху дълбочинния строеж на Западния Балкан. *Сп. Бълг. геол. д-во*, 34, 1, 11—18.

Димитров С. (1927) Ерунтивните скали на Балкана между Петроханския проход и Ржана планина. *Сп. БАН*, 36, 93—167.

Димитров Е., Д. Бахнева, Б. Маврудчиев, Б. Каменов, Й. Янев (1975) Магматическия формации Болгарии. *Geologica Balcanica*, 5, 1, 62 — 63.

Драгов П., Д. Йосифов, Н. Банцаров (1976) Металлогенически активные линеаменты и узлы на территории Болгарии. *Geologica Balcanica*, 6, 4, 3—16.

Йосифов Д. (1976) Берковская зона глубинных разломов. *Geologica Balcanica*, 5, 3, 69 — 80.

— (1977) Първоразредни дълбочинни разломи и земнокорни блокове в нагънатите зони и активизираните области на България по геофизични данни. *Геотект. тектоно-физ. и геодин.*, 6, 48—64.

— (1979) О региональных лениналетных узлах в Альпийском подвижном поясе на территории Болгарии и их металлогеническом значении. *Геотект., тектонофиз. и геодин.*, 8—9, 124—136.

Калайджиев С. (1967) Основни закономерности в пространственото разпределение на стратифицираните находища на медни и оловно-цинкови руди в Западна Стара планина. Изв. НИГИ, 4, 103—110.

— (1982) Структурные особенности Врачанско-Издремецкого рудного района (Западная Стара планина). *Geologica Balcanica, 12, 4* (в печати).

Капурков Г. (1980) Върху младоалпийската сулфидно-барит-хематит-сидеритова формация в Западнобалканската структурно-металогенична зона. *Рудообраз. проц. и минер. наход.*, 13, 35—46.

Карагюлева Ю. (1961) Зони на напречно понижаване и разкъсване в Предбалкана. *Труд. геол. Бълг., сер. страт. и тект.*, 3, 275—290.

Минчева-Стефанова Й. (1961) Минералогия и произход на полиметалическото находище в Западна Стара планина. *Труд. геол. Бълг., сер. геох. и пол. изкоп.*, 2, 157—251.

— (1962) Върху алпидските полиметални орудявания тип „Седмочисленици" в Западна Стара планина. *Изв. Геол. инст.*, 10, 227—303.

— (1978) О причинах локализации оруденения стратиформного типа в Искырском ущелье. Докл., Бан, 31, 10, 1329—1331.

Начев Ив., Сл. Янев (1980) Седиментните геокомплекси в България. С., *Изд. Наука и изкуство*, 203.

Попов П., Е. Чернев, М. Антонов (1979) Върху структурата на Плакалнишкото рудно поле. *Год. Висш. минно-геол. инст.*, 25, 2, 43—60.

Станев Ив., В. Панайотов (1964) Нови данни за геоложкия строеж на Локорско-Буховския район на Кремиковското находище. Изв. НИГИ, 1, 33—68.

Тронков Д. (1965) Тектонски строеж и анализ на структурите на Врачанския блок от Западна Стара планина. Пластични дефонмации в съседство с разломните деформации *Труд. геол. Бълг., сер. страт. и тект.*, 6, 217—251.

Хайдутов Ив. (1979) Напречни магмопроводящи структури при формирането на плутонични тела в Западна Стара планина (по примера на Петрохански плутон). *Геотект., тектонофиз. и геод.*, 10, 16—30.

— Ж. Иванов (1965) Некоторые особености магматизма и метамофизма в зоне Балканского глубинного разлома. *Доклады, VII конгресс КБГА*, София, 1, 67—75.

ОСНОВНЫЕ ЧЕРТЫ ГЕОЛОГИЧЕСКОГО РАЗВИТИЯ ПАЛЕОЗОЙСКИХ СЕГМЕНТОВ КАРПАТО-БАЛКАНСКОЙ СКЛАДЧАТОЙ ОБЛАСТИ (НА ПРИМЕРЕ ЗАПАДНЫХ БАЛКАНИД)[1]

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Для Карпато-Балканской складчатой области (КБСО), являющейся составной частью Альпийского (Средиземноморского) складчатого пояса, характерно присутствие крупных блоков, сложенных докембрийскими метаморфическими образованиями , которые являются обломками Африканской и Восточно-Европейской платформ (Муратов, 1963). В течение фанерозоя отдельные части их подвергались тектоно-магматической переработке. При этом в каждый последующий тектонический цикл в переработку включались новые, прежде жесткие участки приподнятых блоков. Наиболее длительным и интенсивным был каледоно-герцинский цикл, который проявился в ряде районов КБСО, в том числе и в Западно-Балканской структурно-фациальной зоне — в Западных Балканидах (ЗБ). Этот район находится в северной краевой части Родопского (Фракийско-Анатолийского) массива, в сложном тектоническом клине — — между субмеридиональным Крайштидно-Вардарским линеаментом и субширотным Предбалканским разломом, отделяющим его от Мизийской плиты (рис. 1). Зона имеет вид удлиненной в субширотном направлении полосы шириной около 50 км., что позволило Бончеву (1971) выделить ее в качестве Балканского линеамента.

Положение ЗБ на границе двух крупных геотектонических элементов — воздымавшегося в течение фанерозоя Родопского массива и погружавшейся Мизийской плиты, обусловило сложное геологическое развитие с интенсивным проявлением тектоно-магматических процессов. Имеющиеся сведения (Добрев, Щукин, 1967) свидетельствуют о том, что формирование ЗБ происходило на коре континентального типа, характеризующейся в настоящее время мощностью 35—40 км, из которой большая часть (~20 км) приходится на гранитный слой.

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Высокая жесткость фундамента и повышенная мобильность обусловили развитие сложной системы разломов, характеризовавшихся различно направленными дифференцированными движениями, следствием чего явилось мозаично-блоковое строение рассматриваемого региона (рис. 2). Главую роль играли разломы северо-западного (310—320°) и

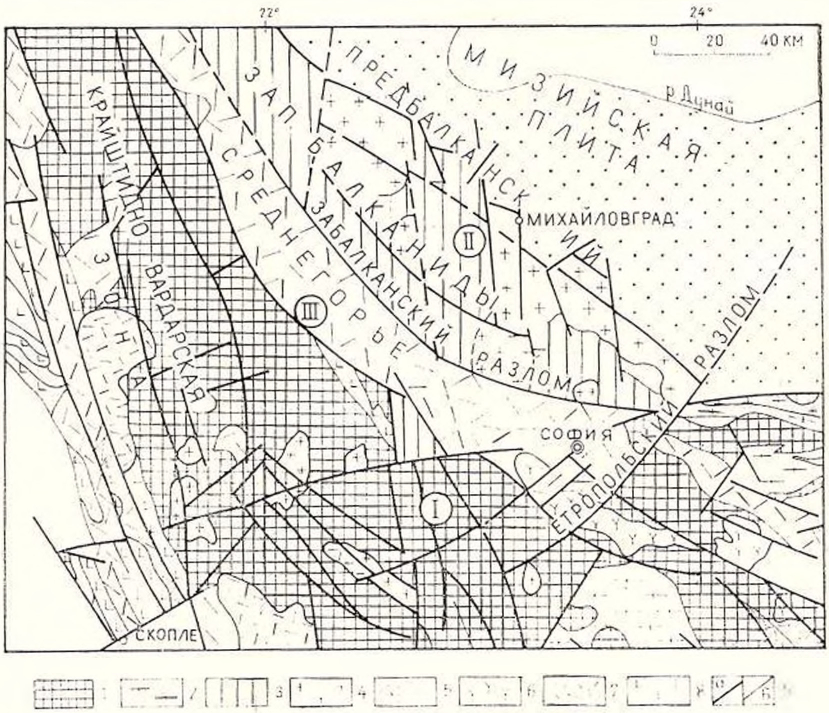

Рис. 1. — Тектоническое положение Западных Балканид. 1. Родопский массив. 1—2, фундамент: метаморфические толщи AR-E (1), гранитоиды (2). II. Западные балканиды. 3—4, каледоно-герцинский структурный этаж (Pz_1—Pz_3): осадочные и вулканогенно-осадочные отложения (3), области развития гранитоидов (по геологическим и геофизическим данным) (4). III, Среднегорье. 5—6, мезо-кайнозойский структурный этаж: осадочные и вулканогенно-осадочные отложения (5), монцонитоиды (6). Кайнозойский структурный этаж (P—N)-вулканогенные образования кислого состава (7). Основные и ультраосновные интрузивы разного возраста (8). Разломы: первого (9а), второго (9б) порядков.

субмеридионального (330—350°) направлений. Наиболее крупные из них принадлежат к категории глубинных и долгоживущих структур, заложение которых предшествовало раннепалеозойскому периоду и предопределило появление и последующее специфическое развитие ЗБ.

Ограничивают и отделяют ЗБ от Мизийской плиты и Родопского палеомассива Предбалканский и Забалканский разломы, характеризующиеся общим северо-западным направлением. Предбалканский имеет вид сложной ступенчатой структуры, образованной сочетанием крупных

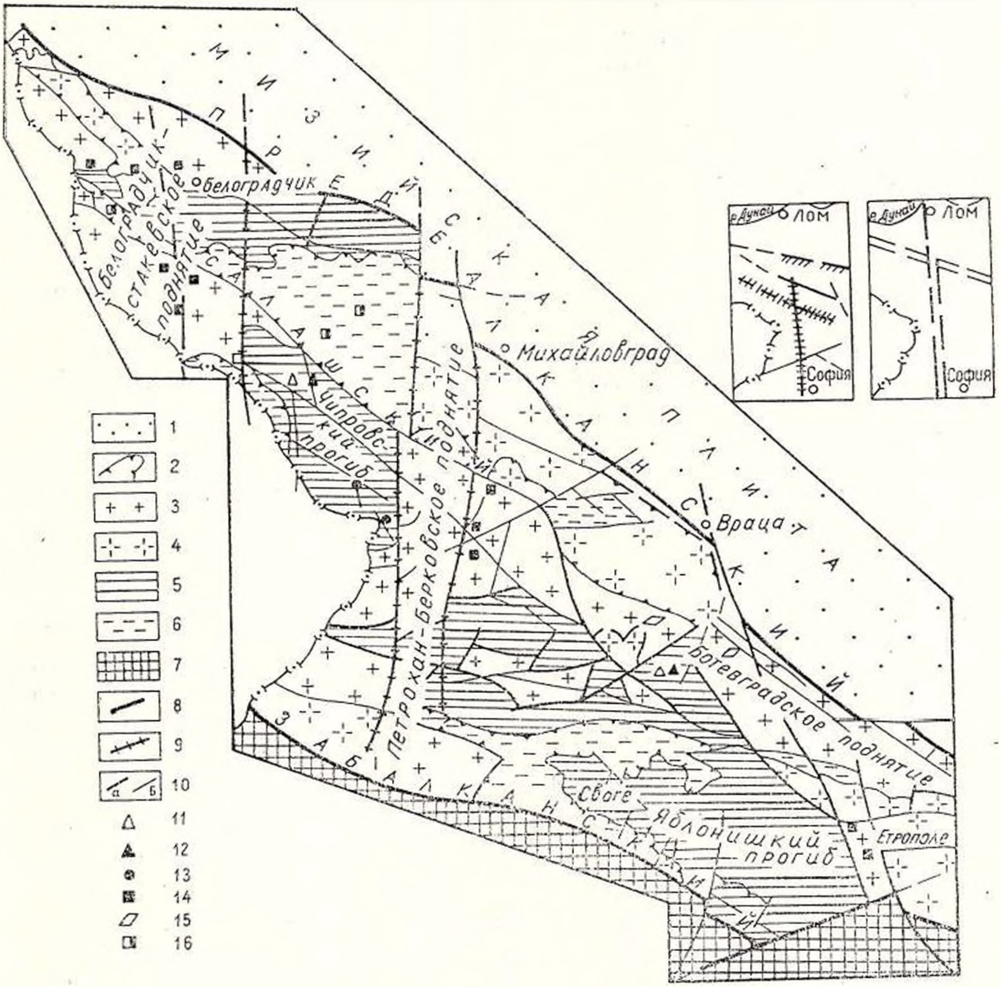

Рис. 2. — Схема блокового строения Западных Балканид (P_z-ий цикл). Мизийская Плита 1, платформенные отложения. Западные Балканиды: 2, межгорные депрессии и впадины орогенного этапа (C_2—P); 3, поднятия, характеризующиеся интенсивным проявлением магматизма, 4, участки поднятий, перекрытые маломощными отложениями P_{z1} или орогенного этапа; 5, прогибы, выполненные отложениями Є;O_1—O_2; O_3—S; 6, участки прогибов, перекрытые отложениями орогенного этапа; 7, поднятия, сложенные допалеозойскими образованиями; 8 — 10, разломы (по геол. и геофиз. данным); 8, глубинные, ограничивающие зону; 9, поперечные (скрытые) разломы фундамента; 10, продольные: а, межблоковые, б, внутриблоковые; 11—16, Эндогенные месторождения: магнетит-пирротиновые и шеелит-молибденитовые (11), полиметаллические (12), золото-полиметаллические (13), медные и медно-пиритовые (14), баритовые (15), гидрогенные-медные (16).

Мелкие схемы (составили Б. Можаев, Н. Кацков и др., 1976): А, разломные зоны по космическим данным; Б, то же по геофизическим данным.

СЗ и субмеридиональных нарушений. По геологическим и геофизическим признакам он отчетливо прослеживается на интервале примерно равном 100 км. В гравитационном поле ему соответствует сложное сочетание линейновытянутых (СЗ) повышенных градиентов (гравитационных ступеней) и того же типа аномалий близмеридионального или СВ направлений. В течение палеозоя отдельные части Предбалканского разлома служили магмоподводящими каналами.

Аналогичными признаками характеризуется Забалканский разлом. Оба разлома, имеющие крутое (местами вертикальное) залегание, создали жесткую раму для всей территории ЗБ, что препятствовало ее крупным горизонтальным перемещениям. Такая же обстановка сохранилась и в последующий (альпийский) цикл развития.

Серия СЗ разломов рассекает и внутреннюю часть площади ЗБ. Они предопределяли пространственное положение прогибов, депрессий и впадин в разные этапы каледоно-герцинского времени.

Существенная (главная) роль в блоковом строении территории принадлежала поперечным (субмеридиональным) разломам, которые относятся к категории скрытых структур — к разломам фундамента. Они разделили в начале цикла ЗБ на крупные блоки, одни из которых в течение всего палеозоя развивались как поднятия (Белоградчик-Стакевское, Петрохан-Берковское). В их пределах формировались разновозрастные интрузивные породы. Другие участки в разное время подвергались прогибанию и являлись ареной формирования палеозойских осадочных и вулканогенно-осадочных толщ. По геофизическим данным субмеридиональные (межблоковые) разломы выражаются границами полей различного уровня (амплитуда 2—4 мгл), цепочками локальных минимумов и максимумов, систематическими изгибами, протягивающимися вдоль прямых линий, и другими деформациями изоаномал. На космических снимках отчетливо фиксируется разломная зона, совпадающая с краевой частью Петрохан-Берковского поднятия (Можаев, Кацков и др., 1976).

Неоднородное (мозаично-блоковое) строение фундамента оказало существенное влияние на последующее развитие ЗБ, которое отличается от типично геосинклинального и, используя терминологию Бончева (1971, 1980), можно охарактеризовать как линеаментно-геосинклинальное.

Раннепалеозойский этап — начальный (Е — S) характеризовался формированием осадочных и вулканогенно-осадочных отложений в пределах изолированных бассейнов, пространственное положение которых определялось разломами северо-западного направления. Поскольку в тектоническую переработку в каждый новый этап вовлекались наиболее жесткие участки, происходила миграция прогибов. Следствием этого явилась пространственная разобщенность их, что вызвало большие затруднения в корреляции выполняющих их отложений и к выделению сложного комплекса, получившего наименование диабаз-филлитоидной формации (Dff). На основании существующих данных и личных наблюдений авторы подразделили сланцевые толщи палеозоя на 3 свиты: чипровскую (Е), средогривскую ($O_1 — O_2$) и грохотенскую ($O_3 — S$) и показали их пространственное распределение. Свиты характеризуются резким преобладанием терригенно-осадочных отложений и отсутствием крупных покровов эффузивов; только в составе двух первых отмечаются диабазы, кварцевые порфиры и альбитофиры, слагающие в основном экструзивные тела, сопровождающиеся туфовыми отложениями. Породы чипровской и средогривской свит метаморфизованы в зеленосланцевой фации, но различаются по степени преобразования. В первой преобладают кварц-хлоритовые, хлорит-эпидотовые и хлорит-актинолитовые сланцы и мрамора. Средогривская свита сложена кварц-хлорит-серицитовыми сланцами (с сохранившимися реликтами первичных минералов) и кварцитами. Мощности свит в глубоких частях прогибов достигают 1500—2000 м, к периферии наблюдается сокращение их и изменение фациального состава.

Отложения грохотенской свиты подвергались очень слабому метаморфизму. В ее составе преобладают глинисто-песчанистые и глинистые сланцы, филлитизированные алевролиты и аргиллиты. В основании свиты — окварцованные песчаники, реже конгломераты. Мощность 1000—1500 м. Среди двух последних свит широко развиты черные (углистые и графитизированные) сланцы — подтверждение того, что области накопления представляли собой неглубокие изолированные бассейны, связанные с открытым морем лишь частично или временно.

К раннему палеозою относится формирование (в пределах поднятых блоков) интрузивов габброидного и гранитного состава (массивы Ком, Стакевский). Эти породы подверглись интенсивному рассланцеванию и сопровождаются признаками, присущими магматическим и анатектическим процессам (Хайдутов, 1979).

Среднепалеозойский, ,,инверсионный,'', этап (Д—C_1). В течение силура, девона и раннего карбона большая часть ЗБ представляла собой поднятый участок земной коры, который подвергался интенсивной тектоно-магматической переработке. Этот период лишь условно можно параллелизовать с инверсионным, характерным для развития типичных геосинклиналей. В рассматриваемом регионе сохранялась прежняя тенденция к воздыманию, что привело к замыканию раннепалеозойских прогибов. Только в пределах небольшой площади (Своге-Яблонишкий прогиб) в течение девона и раннего карбона с неоднократными перерывами отлагались маломощные отложения, представленные аргиллитами, песчаниками, конгломератами, мощностью в первые десятки или сотни м. В пределах ранних поднятий происходило формирование разнообразных т.н. Старопланинских интрузивов гранитоидного состава. Нами они подразделяются на петроханский, святоникольский и мездрейский комплексы. Породы первого представлены габбро, габбро-диоритами, сиенито-диоритами, монцонитами, биотит-роговообманковыми гранитами. Святоникольский и мездрейский — сложены преимущественно гранитами. По петрохимическим признакам они относятся к т.н. кальциево-щелочной группе.

Верхнепалеозойский, ,,орогенный'', этап (C_2—P). Формирование осадочных толщ в течение данного этапа прерывалось тектоно-магматическими процессами, что позволяет подразделить его на 2 периода: ранний (C_2) и поздний (P). В течение раннего периода происходило образование грабенообразных впадин, удлиненных вдоль разломов, которые заполнялись грубообломочными осадками (конглобрекчиями, конгломератами, песчаниками). В наиболее глубоких впадинах формировались песчано-глинистые, местами угленосные, отложения. Мощность колеблется от 200 до 1000 м. Поздний период. Тектонические перемещения по разломам и дальнейшее расчленение рельефа усилилось к началу перми. С этими процессами связана магматическая деятельность, проявившаяся в краевых частях раннепалеозойских поднятий. В результате сформировались субвулканические и экструзивные контрастные по составу (диабаз-липаритовые) тела а также небольшие массивы и дайки субщелочных (калиево-щелочных) пород. К этому же времени относится резкое изменение климата, что способствовало формированию красноцветных отложений озерно-континентальной формации в пределах межгорных депрессий и предгорных впадин, фундамент которых характеризовался гетерогенным геологическим и тектоническим строением. Последнее отразилось на резком изменении фациального состава отложений и на их мощности. В наиболее погруженных частях мощность толщ достигает 1500—1700 м. Среди пород преобладают слабо окатанные и слабо отсортированные отложения (брекчии, конглобрекчии, грубозернистые песчаники) аллювиальных, пролювиальных и озерных фаций. Существование застойных явлений приводило к образованию сероцветных горизонтов песчаников, гравелитов, а также известково-глинистых прослоев, обогащенных углефицированным органическим веществом. Существующие условия способствовали формированию гидрогенных руд.

С нижнего триаса в ЗБ установились субплатформенные условия.

Палеозойские гранитоиды характеризуются присутствием радиоактивных и редких элементов, количество которых возрастает от древных к молодым (U = 2,5—39,0 г/т, Th = 6,5—65,0 г/т, т.е. происходит накопление этих элементов в субщелочных интрузивах орогенного этапа вплоть до образования торийредкоземельного оруденения. Это имеет определенное значение для распознавания магматических комплексов орогенного этапа, с которыми связаны руды различных металлов.

С палеозойским циклом развития (320—240 млн. лет — Минчева-Стефанова, Амов, 1969; Драгов, 1978, 1981; Куйкин, 1975; Димитров, 1977) связано разнообразное эндогенное и гидрогенное оруденение. К наиболее ранним относятся магнетит-пирротиновое и шеелит-молибденитовое, тесно связанные с предшествующими им процессами скарнирования. На флангах этих месторождений развивается полиметаллическое оруденение, наложенное на скарны. Все они располагаются в краевой части раннепалеозойского прогиба, выполненного наиболее древними (кембрийскими) отложениями, в пределах структурного узла, образованного крупными разломами субмеридионального и северо-западного направлений. Здесь же наблюдаются массивы гранитов святоникольского типа. В противоположной краевой части того же прогиба, выполненного более молодыми (ордовикскими) отложениями, находятся месторождения золото-полиметаллической формации. Мощности осадочных толщ здесь меньше, чем в первом случае (по геофизическим данным под ними располагаются гранитоиды). Вся площадь разбита тектоническими нарушениями, сопровождающимися многочисленными дайками кислого и основного состава, а также экструзивными образованиями липаритов и фельзитов. Медное и медно-пиритовое оруденение по геологической ситуации подразделяется на два типа: 1, приуроченное к интрузивам гранитоидов „инверсионного" этапа — медно-порфировый тип; 2, к субвулканическим телам орогенного этапа — к андезитовым (диоритовым или авгитовым) порфиритам — жильный тип. В этих же телах нередко присутствует баритовая минерализация.

Все они приурочены к периферическим частям раннепалеозойских поднятий и тяготеют к ограничивающим их субмеридиональным разломам.

К гидрогенному типу относятся скопления медных (Янев, 1970) и ванадиевых руд, которые приурочены к континентальным красноцветным отложениям и располагаются в краевых частях депрессий орогенного этапа.

Кратко освещенные особенности палеозойского цикла развития Западных Балканид, характеризующихся проявлением разнообразного оруденения, могут оказаться полезными при оценке перспектив аналогичных районов КБСО.

ЛИТЕРАТУРА

Бончев Е. (1971) Проблеми на българската геотектоника. *Техника*, София.

— (1980) Основни геодинамични проблеми на нашето пространство. *Геодинамика на Балканите*, *Техника*, София, 83—94.

Димитров Р. (1977) Изотопный состав свинца и серы и генезис некоторых месторождений цветных металлов Болгарии. *Тр. КБГА*, XI.

Добрев Т. Б., Щукин Ю. К. (1967) Исследования земной коры и природы сейсмичности Карпато-Балканского региона. *Сов. геол.* 5.

Драгов П. (1978) Рудните минерализации в палеозоя на част от Западна Стара планина. *Изв. геол. ин-та, сер. рудни и нерудни пол. изкоп.*, XIX—XX.

— Арнаудов В., Амов Бл. (1981) Последователност на палеозойските металогенни процеси в Зап. *Стара планина*, *ГИ БАН*, София, 23.

Йовчев Й. (1960) Полезни изкопаеми на Н. Р. България. *Техника*, София.

Минчева-Стефанова Й, Амов Бл. (1969) Генетично тълкуване на вариацините в изотопния състав на оловото в полиметалните орудявания тип „Седмочисленици". *Изв. геол. ин-та, сер. геохим., минер., петр.*, 18.

Можаев Б. Н., Кацков Н. и др. (1976) Результаты геологического дешифрирования космических снимков Болгарии, полученных с ИСЗ „E R T S" (рукопись).

Муратов М. В. (1963) Структурные комплексы и этапы развития геосинклинально-складчатых областей. *Изв. АН СССР, сер. геол.*, 6.

Хайдутов И. (1979) Магматични и анатектични явления в Старопланинската палеозойска евгеосинклинала (по примера на Бързийския мигматитов масив). *Геотект., тектонофиз. и геодинам.*, *БАН*, София, 8—9.

Янев Сл. (1970) Палеография на северозападна България през младия палеозой. *БГД*, XXXI, кн. 2.

ЗАКОНОМЕРНОСТИ КАЙНОЗОЙСКОЙ МЕТАЛЛОГЕНИИ НА ТЕРРИТОРИИ СРЕДНЕСЛОВАЦКИХ НЕОВУЛКАНИТОВ [1]

ЛАДИСЛАВ РОЗЛОЖНИК [2]

Среднесловацкая неовулканическая область Западных Карпат является областью субсеквентного и даже финального альпийского вулканизма в смысле Штилле (1953), или по Твалчрелидзе (1977, 1978) можно ее считать частью вулканического пояса центральной области альпийской тектономагматической активизации средиземноморской складчатой системы. Речь идет о металлогенетической области известной в мире своими историческими рудными районами золота, серебра, свинцово-цинковых и медных руд Баянской Штявницы, Банской Годруши и Кремницы.

Тектономагматическая активизация внутренних Западных Карпат, подвигающихся в квазиплатформное состояние в течение верхнего мела, начиналась уже во время палеогена-эоцена. Следы вулканизма эоцена находятся и на территории Средней Словакии, но типичные его проявления находим в Венгрии (Речк, горы Веленце, окрестность Балатона). Точное определение продуктов этого этапа вулканизма представляет определенные трудности поскольку они перекрыты вулканитами миоцена похожего характера. ,,Будинский'' бассейн палеогена по Vass (1976) отличается по характеру от молассовых впадин возникших в течении неогенового вулканизма, хотя по мнению этого автора, как палеогеновый бассейн, так и плиоценовый паннонский пространственно связаны с областью уточнения коры внутри карпатского изгиба. Но уместно заметить, что подобно тому, как банатиты тянутся по направлению вардарской зоны, также и эоценовые вулканиты в Венгрии направлены вдоль линии Балатона в основном восточно-австро-алпийско-западнокарпатском структурном направлении. Напротив, неовулканиты размещены по блокам, как бы ,,хаотически''. Это свидетельствовало бы в пользу того, что на территории среднесловацких неовулканитов развитие моласс происходило в двух стадиях. Нижный этаж — эоценовый, по сравнению с верхним неогеновым имеет до известной степени особенные геотектонические черты развития. Представителем металлогенетической деятельности в

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[2] Горный факультет ВШТ Кошице, Польская Н.Р.

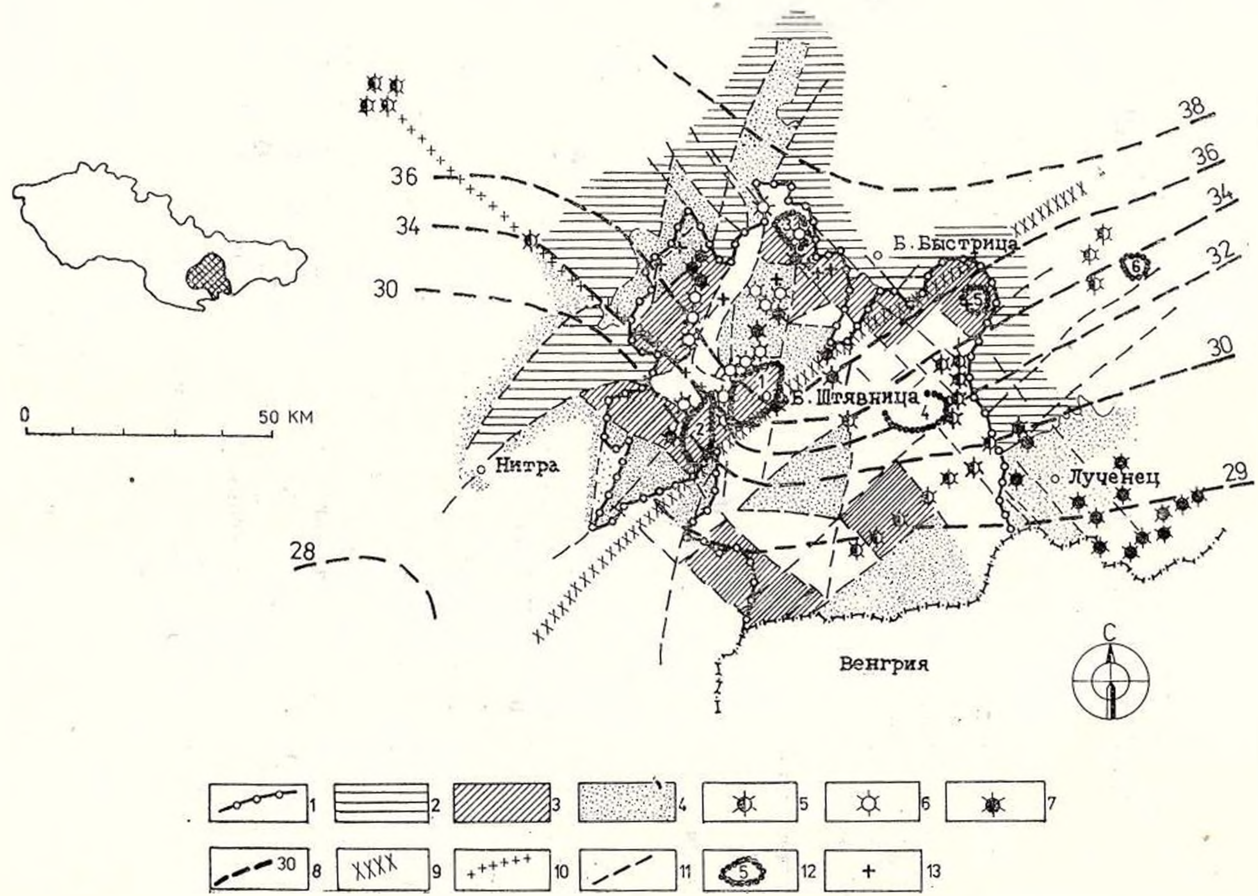

Рис. 1. — Область среднесловацких неовулканитов.

1, контур расширения среднесловацких неовулканитов; 2, донеогеновый фундамент на поверхности; 3, блоки выраженной элевации на территории неовулканитов; 4, депрессии; 5, центра выходов андезитов; 6, центра выходов риолитов; 7, центра выходов базальтов; 8, изогипсы плоскости — M; 9, вепорский глубинный разлом; 10, глубинный разлом пшеровско-штявницкий; 11, разломы; 12, интрузивные центра-рудные районы (1, Банскоштявницко-Годрушский; 2, Пуканецко-Новобанский; 3, Кремницкий; 4, Яворя; 5, Поляны; 6, Тисовца); 13, проявления ртути.

рамках нижней молассы является месторождение Речк в Венгрии с меднопорфировым оруденением, Твалчрелидзе (1977) считает его представителем первой стадии тектоно-магматической активизации Паннонского массива.

Неогенная тектоно-магматическая активизация на территории среднесловацких неовулканитов начинается в начале миоцена (Konečný и др., 1969) после распада упомянутого общего бассейна палеогена образованием многочисленных межгорных впадин и горстов, размещенных мозаично. Вулканизм неогена начинается на территории Венгрии риолитами эгра (Pantó, 1969), и оттуда постепенно расширяется к северу в области средней Словакии. Активность вулканизма кульминирует в бадене — сармате и оканчивается в начале плейстоцена базальтами. Тектоника этих блоков была очень интенсивна, неовулканическая деятельность отличалась пестростью петрографически и петрохимически разнообразных видов, как глубинных, так и поверхностных. Соответственно разнообразной и многофазовой вулканоплутонической деятельности развивались и различные металлогенетические процессы. Рудообразование развивалось в четырех главных формациях:

1. высоко термальные контактно-метасоматические и вкрапленные меднопорфировые руды;
2. гидротермальное жильное оруднение полиметаллического типа;
3. гидротермальное золоторудное оруднение;
4. низкотермальное оруднение Hg, As прожилково-вкрапленного характера.

Магматические и геотектонические основы развития металлогении

Магматическая основа рудообразования на территории Средней Словакии развивалась не в классическом порядке: риолит-андезит-базальт. Хотя вулканическая деятельность начинается в конце олигоцена — начале миоцена риолитами и оканчивается базальтами, в начале плейстоцена, но между этими экстремальными членами магматического ряда выступали попеременно средние и кислые продукты. После широко распространенного андезитового вулканизма изменчивого состава приходят не базальты а в начале риолиты, и только потом — базальты.

Следующей особенностью магматической деятельности является значительная самостоятельность развития отдельных вулканоплутонических центров: Банской Штявници — Годруши, Кремници, Яворья, Поляны и Тисовца. Дифференциация развития магматического процесса временно и пространственно настолько выразительна, что отдельные центры нельзя взаимно различить. Согласно с этим проявляется и металлогенетическая специальность отдельных центров. Взаимное сравнение вулканических продуктов затрудняет отсутствие достаточных стратиграфических критериев корреляции. Определение возраста вулканитов или плутонитов отдельных формаций по методам абсолютного возраста затруднено наложением влияния гидротермальных процессов.

Петрохимически среднесловацкие вулканоплутонические формации проявляют общую тенденцию повышения щелочности от более старых к более молодым (Karolus, 1970) и направление измененения состава кислых и средних членов идет по кривой прототипа Монт Пеле и отличается экспериментальным содержанием Ca — alk.

Комагматичность в рамках вулканоплутонических формаций не так выразительна. Например, в годрушском интрузивном комплексе (R o z l o-

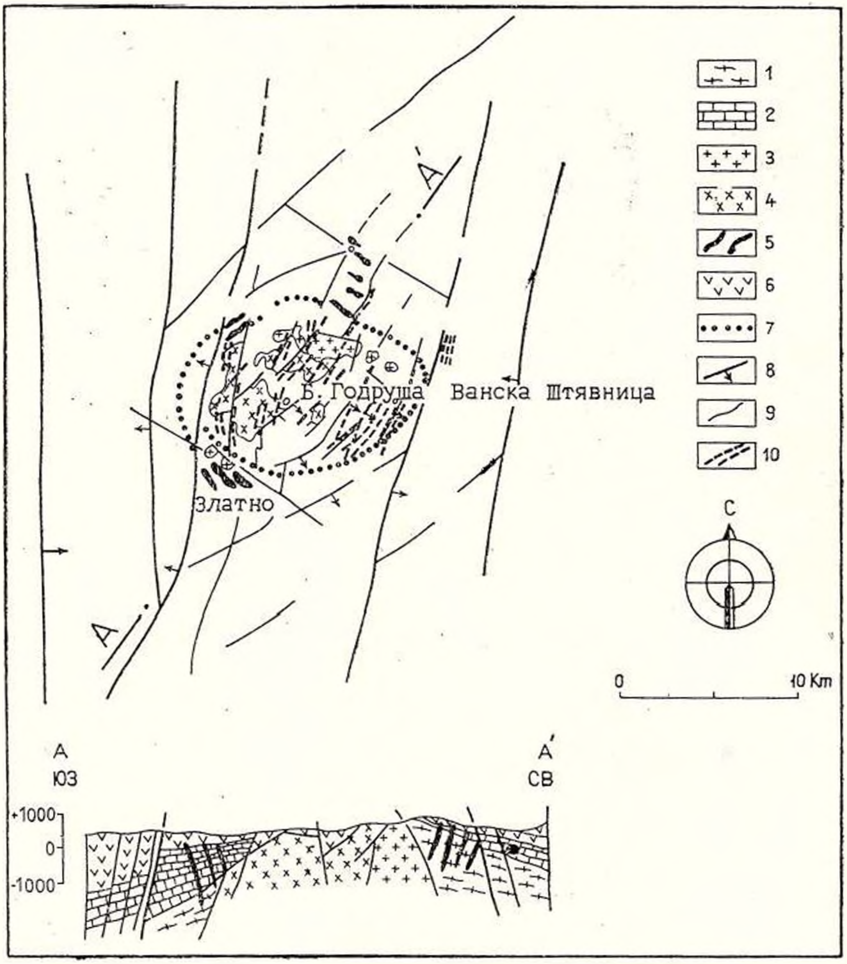

Рис. 2. — Банско-Штявницко-Годрушский рудный район.

1, кристаллический фундамент (герцинникум); 2, толщи палеозоя и мезозоя; 3, габбро-диорит — кв. диорит (неоген); 4, годрушский гранитодиорит (неоген); 5, кольцеобразные дайки гранодиоритовых порфиритов — носители скарного-медно-порфирового оруднения; 6, неовулканиты (андезиты, дациты, риолиты); 7, контура годрушского-интрузивного комплекса; 8—9, разломы; 10, гидротермальные рудные жилы с полиметаллическим оруднением.

ž n i k, Я. Ш а л а т, 1979) группа диоритовых пород со своим широким спектром от леукотоналита до габбро не проявляет магматический ряд сравнимый с какой нибудь знакомой магматической ассоциацией. Следующие более кислые члены комплекса, такие, как гранодиориты, аплитические граниты, аплиты, кварц-диорит, порфириты, хотя и образуют вместе с диоритами общий плутонический комплекс, тем не менее петрохимически не проявляют знаки родства, и химизм плутонических форм нельзя с успехом сравнивать с вулканитами. Пульсационный, многофазовый характер, пестрая петрография образуют общую черту неогенной вулканической деятельности. Вышеприведенную характеристику дополняют результаты некоторых исследований, как например исследования цирконов в плутонитах годрушского интрузивного комплекса (J a k a b s k á, 1980). Из них вытекает, что цирконы отдельных членов четко различаются и цирконы находящиеся в гранодиорите и кварц диоритпорфиритах, своими свойствами отвечают более основным — габбронорнтическим типам.

В общем развитии неовулканитов на территории Средней Словакии свидетельствует о подъеме теплого диапира и с этим связанного плавления коры. Кислые продукты являют собой результаты плавления сиалического вещества, андезиты — гибридное происхождение, а базальты поднялись из мантии.

Кайнозойский магматизм простиряется внутри карпатской дуги в местах стремительного опускания поверхности прерывности Мохоровичича. По данным F u s a n (1979) от центра Паннонского бассейна плоскость Мохо опускается в направлении к внешним Карпатам от 26 км до 50 км. Кайнозойские вулкано-плутонические центры выступают именно в местах, где опускание более стремительное. Через Банскую Штявницу переходим четкий предел густоты в направлении СВ—ЮЗ — — т.е. в направлении Комарно-Попрад, так называемый глубинный разлом — Вепорский. Одновременно этот разлом интерпретируется границей между двумя большими тектоническими единицами Западных Карпат — Вепорика и Татрика. При Банской Штявнице изограды Δg меняют направление от СВ—ЮЗ до дирекции СЗ—ЮВ дуговым способом и индицируют следующий четкий предел густоты — глубинный разлом штявницко-пшеровский. Упомянутые разломы ограничивают блок с более мощной сиалической корой — клинообразный фатранско-татранский блок. Этим способом позиция Банскоштявницкого рудомагматического центра совпадает с выразительным геофизическим и тектоническим узлом.

Штявницко-пшеровский глубинный разлом контролируют тела вулканитов при Бановцах и Угорском Броде. Вдоль Вепорского глубинного разлома в направлении Банской Штявницы — Поляна наблюдаются гравиметрические аномалии, свидетельствующие об элевациях фундамента и существовании плутонических тел.

Следующим доказательством закономерности локализации вулканоплутонических центров является морфотектоническая карта докайнозойского рельефа. Из нее вытекает пестрая мозаика блоков, представляющих поднятия и углубления рельефа, как и протекание неогенной разломной тектоники. Плутонические центры находятся в местах поднятий фундамента, что можно объяснить ригидным характером интрузивных тел. Большинство элеваций фундамента являются интрузивными.

Картина элеваций и депрессий указавает, что при их образовании основную роль сыграли разломы следующих направлений.

Первое направление СВ—ЮЗ является направлением упомянутого Вепорского глубинного разлома и одновременно главным структурным направлением докайнозойского фундамента и носителем вулкано-тектонических зон. Значительные вулкано-плутонические центры — Новобанское, Банско-Штявницкое, Яворие, Поляна идут в направлении Вепорского разлома. Сантовско-абеловский вулканический хребет растянут в этом же направлении, но он является безрудным.

Второе направление Штявницко-Пшеровского глубинного разлома (СЗ—ЮВ) не образует такие четкие вулкано-тектонические зоны, но в нем отражаются древние структуры фундамента, и в местах, где он пересекает тектонические зоны СВ—ЮЗ направления, образовались такие вулкано-плутонические центры, как Кремница и Банска Штявница.

Третье направление Ю—С является молодым направлением, появившимся только в начале неогена. Данное направление образует выраженную поперечную тектоническую зону (по Š t o h l, 1976 — центрально-карпатский линеамент) содействует образованию впадин и горстов, носит дайки риолита и играет важную роль при образовании рудолокализующих жильных структур полиметаллической и золоторудной формации Банско-Штявницкого, Кремницкого и других рудных полей. По центрально-карпатской зоне разломов вулканизм проник более глубоко к северу. Таким образом глубокие разломы направления С—Ю сыграли важную роль в развитии неовулканической активизации Словацкого блока.

Территория среднесловацких неовулканитов является не только областью тонкой коры и областью повышенной тектономагматической деятельности, но и областью термической аномалии. По исследованиям M a r u s i a k и др. (1979) термическая аномалия достигает 50—60°С на 1000 м и полностью покрывает область распространения среднесловацких неовулканитов а также вдоль выше упоминаемого центрально-карпатского разлома внедряется до самой впадины Турца. На целой территории среднесловацких неовулканитов находятся источники термальных вод. Это обстоятельство также говорит в пользу диапирового поднятия мантии, что является типичными признаками областей редукции коры и тектоно-магматической активизации в смысле Т в а л ч р е л и д з е (1977).

С точки зрения металлогении наиболее важны вулканоплутонические центры локализованные в местах поднятия теплового тока, с тонкой корой, с стремительным опусканием поверхности прерывности Мохо, в местах пересечения глубинных разломов, образованных в основном до неогена. Интрузивные центры находятся в местах подъема фундамента формирующих интрузивные элевации. Благоприятными являются места, где донеогенный фундамент состоит из карбонатовых пород, пригодных для скарнообразования. Рудоносные интрузивные центры являются сложными по составу и развитие их многофазовое.

Рудные формации кайнозойского металлогенезиса на территории среднесловацких неовулканитов

В соответствии со сложным, многофазовым тектоническим и магматическим развитием формировались рудообразующие процессы на территории среднесловацких неовулканитов. Рудные поля совпадают с местами вулкано-плутонических центров. Их керном являются плутониты-малые интрузии, дайки, внедренные в породы фундамента различного состава и возраста и часто в более древние вулканические аппараты. Развитие магматизма и рудообразования у отдельных центров имеет отдельные различия но в общем можно наблюдать следующие:

— образование высоко термальных контактно-метасоматических руд и прожилково-вкрапленных (медно-порфировых) руд;
— среднетермальное жильное орудение полиметаллического типа;
— прожилково-вкрапленное низко-термальное орудение Sb—Hg,As.

Другие типы минерализации, например: сольфатарная минерализация серы, вулканоосадочные и другие руды не имеют большого значения.

1. Высокотермальные контактно-метасоматические руды Fe — скарного типа и прожилково-вкрапленные руды (Cu ± Pb — Zn) в скарнах и порфиритах, находятся в тесной связи с малыми телами и дайками диоритов, гранодиоритов и их порфиритов. Эта формация носит парагенетические свойства и околорудные изменения, типичные для месторождения порфировых руд. Проявление этой формации находим во всех вулкано-плутонических центрах, т.е.в районе Банской Штявницы, Кремницы, Яворя и др., но эта формация наиболее изучена в районе Банской Штявницы — именно на примере месторождения Злaтно — южнее от Банской Годруши. Здесь после внедрения малых тел диоритовых пород очень изменчивого состава, поднялся плутон гранодиорита. Плутоны внедрились в различные породы донеогенного фундамента и вызвали интенсивный контактный метаморфизм. Далее вероятно в связи с подъемом плутона, уже после его застывания внедрились и кольцевые, маргинальные дайки гранодиорит-порфирита, несущие скарнообразование с рудами магнетита и прожилково-вкрапленных руд Cu (±Pb—Zn) с пиротином, пиритизацией и другими парагенетическими свойствами и изменениями, характерными для месторождения медно-порфировых руд (Р о з л о ж н и к, Zabranský, 1971), но молибден отсутствует.

2. Гидротермальное жильное орудение полиметаллического типа (Cu, Pb — Zn, Ag — Au) является более молодым чем предыдущее, с широкой парагенетической шкалой и гидротермальными изменениями (пропилитизация, силицификация, адуляризация). Рудные жилы сопутствуют разломам горст-грабеновой структуры, образованным молодыми поднятиями плутонических комплексов. Жилы часто контролируют дайки дацитов и риолитов ССВ—ЮЮЗ направления. В отдельных рудных районах а именно в Банско-штявницком, можно наблюдать зональность, (сверху-вниз): Au → Ag → Pb → Zn → Cu. В Кремницком рудном районе преобладает золоторудное, а в годрушском рудном поле серебряное орудение, полиметаллическая формация появилась в заключительных фазах вулканической деятельности и вероятно, связана с риолитами.

3. Низкотермальное орудение Sb, Hg, As прожилково-вкрапленного типа рассеяно в различных породах в неясной связи с большими структурами, приурочено к заключительному этапу неогенного вулканизма, вероятно, базальтовому.

Заключение

Магматические, геотектонические основы парагенетические разновидности, общий характер металлогении в течении неогена на территории среднесловацких неовулканитов отвечают определению, выскаванному

Твалчрелидзе (1977) о металлогении ,,вулканических поясов областей тектономагматической активизации''. Однако в случае среднесловацких меднопорфировых руд появляются некоторые особенности: орудненне в рассматриваемом нами примере произошло не в приповерхностных условиях (500—150 м) а в глубинах минимально 3 км. Орудненне порфировых руд, по-видимому связано с более основным очагом, как это предполагается у типа медных порфировых руд в активизированных областях.

ЛИТЕРАТУРА

Böhmer M., Kodĕra M. (1975) The paragenesis of subvolcanic deposits of the Wes Carpathians and its development in Time and Space. Proceedings of the Xth Congres GBGA 1973, Section VI.

Forgáč J. (1970) Geochémia ryolitov Slovenska. *Zbornik geol. vied Záp. Karpaty*, 12, s. 137—192, Bratislava.

Fusan J., Ibermajer J., Plančár J. (1979) Neotectonic Blocks of the West Carpathians Geodynamic Investigation Czechoslovakia Final Report Veda Bratislava.

Fusan O., Kuthan M., Ďuranty S., Plančár J., Zbořil (1969) Geologická stavba podložia stredoslovenských neovulkanitov. *Zbornik geol. vied. rad. Zk zv.*, 10. Bratislava.

Jakabská K. (1980) Výskum zirkónov a biotitov. Manusckript VI. BF VŠT Košice.

Karolus K. (1970) Chemizmus eruptivnych fáz stredoslovenských neovulkanitov. *Zbornik geol. vied Západ. Karpaty*, 12, str. 85-136, Bratislava.

Konečný V., Bagdasarjan G. P., Vass D. (1969) Evolution of Neogene volcanism in Central Slovakia and its confrontation with absolute ages. *Acta geol. Acad. Scie. Hung.*, 13, Budapest.

— (1971) Evolutionary Stages of the Banská Štiavnica Caldera and its Postvolcanic structurs Bulg. Volcan. XXXV—1.

Lexa J., Konečný V. (1979) Relationship of the Carpathians volcanic arc to the geodynamic evolution of t he Pannonians Basin Geodynamic Investigation Czechoslovakia Final Report Veda Bratislava.

— Konečný V. (1974) The Carpathians volcanic Arc a Diskussion. *Acta geol. Acad. Scie. Hung.*, 18, 279.

Marušiak I., Lizoň I., Plancár J. (1979) The geothermal field of the West Carpathians. Geodynamic investigations Czechoslovakia, str. 215—221.

Pantó G. (1969) Geology of Northern Hungary 9th KBGA Cangr., Budapest.

Rozložnik L., Šalát J. (1979) Chemizmus hodrušského intruzivneho komplexu z hladiska metalogenézy. Symp. o petrogenéze a geochémii geologických procesov. Veda, str. 215—222, Bratislava.

— Zabranský (1971) O výskyte žilnikovo-impregnačného zdrudnenia medzi obcami: Banská Hodruša, Vysoká a Uhliská. *Mineralia Slovaca*, III, č. 10, p. 85—94, Spišská Nová Ves.

Šefara J. (1980) Tektonická stavba predterciérneho podložia stredoslovenských neovulkanitov, jej vztah k vulkanizmu a zrudneniu. *Geofyz. výskum neovulkanitov Záp. Karpát, Geofyzika*, str. 51—55, Bratislava.

Štohl J. (1976) Zrudnenie stredoslovenských neovulkanitov spojené centrálnokarpatským lineamentom. *Záp. Karpaty ser. min. petr. geoch. lož.*, a, str. 7—40, Bratislava.

Твалчрелидзе Г.А. (1977) Металлогенетические особенности главных типов вулканических поясов. Москва, Недра.

— (1978) Металлогения центральной части альпийского средиземноморского складчатого пояса. *Геолог. руд. мест.* 5, 3–15, Москва.

Vass D. (1979) Genesis of inner-mollasse basins in West Carpathians in light of leading function of mantle in Earth's crust development Czechoslovak geology and global tectonics. *Proced. Conf. Smolenice 1976*, p. 183–199, Veda, Bratislava.

ТИПЫ ВУЛКАНИЧЕСКИХ СТРУКТУР И ИХ ОТРАЖЕНИЕ НА ГЕОФИЗИЧЕСКИЕ ПОЛЯ СМОЛЯНСКОЙ ДЕПРЕССИИ ЦЕНТРАЛЬНЫХ РОДОП [1]

Н. СТЕФАНОВ [2], Н. КРЪСТЕВ [3]

При изучении внутреннего строения вулканических районов, сложенных кислыми вулканитами в Центральных Родопах, возникают большие трудности, связанные с недифференцированным составом пород, слабым проявлением протоструктур течения и небольшим эрозионным срезом. Это обусловливает необходимость использования комбинированных геолого-геофизических методов, среди которых наиболее эффективными оказались структурно-вулканологическая и магнитная съемка Δz в масштабе 1:25000. В работе обсуждаются некоторые вопросы связи вулканических структур и магнитных аномалий по Δz с целью выяснения их взаимной обусловленности. Выделение главных центров извержения, вулканоструктур и магмопроводящих каналов основано на фациальных, структурных и петрографических исследованиях, как и на данных структурного бурения (Бахнева, Стефанов, 1975). Наземная магнитная съемка Δz выполнена в маршрутном варианте вследствие тяжелых условий рельефа. Преобладающая ориентация маршрутов — с севера на юг. При интервале измерения 50 m достигнута плотность около 60 m/km^2.

Смолянская депрессия представляет перворазрядную блоковую структуру, заложенную в южной части Родопского массива к концу верхнего эоцена. Ее положение определяется наличием нескольких куполовидных поднятий в докембрийской метаморфической раме, а морфология — разломными зонами субширотного, субмеридионального, северо-восточного и северо-западного простирания (Стефанов, 1973). В плане депрессия имеет слабо вытянутую форму в субширотном направлении и площадь около 550 km^2. В пределах депрессии выделяются две структурные зоны — периферическая, сложенная кристаллическим и палеогеновым обрамлением и центральная, представляющая собой Переликскую вулканогенную структуру. Депрессия имеет трехярусное внутреннее строение — докембрийский метаморфический цоколь, верхнеэоцен-нижнеолигоценовый вулканогенно-осадочный чехол и олигоцен-миоценовый (?) риолитовый покров (рис. 1).

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[3] ММР, Н. Р. Болгария.

Вулканиты Смолянской депрессии представлены двумя типами риолитов (порфирокластическими и фельзитоидными), а также субвулканическими телами латито-андезитов. В свою очередь они характеризуются собственными вулканическими структурами. Наиболее широко распространены порфирокластические риолиты, образующие комплексный массив, который охватывает центральную и западную часть Смолянской депрессии.

Фельзитоидные риолиты приурочены к двум магмопроводящим структурам — Момчиловской и Левочевской, причем первая линейного; а вторая центрального типа. Момчиловская структура представлена множеством штокообразных тел и даек, протянувшихся в виде непрерывной цепи в субширотном направлении. Ее длина достигает почти 70 km при максимальной ширине 2 km. Левочевская вулканическая структура расположена восточнее г. Смолян, в узле пересечения разломных зон северо-западного, северо-восточного и субширотного направления. Она имеет неправильную форму и охватывает площадь около 7 km². В ней выделены два основных и несколько сателлитных центров извержения, с которыми связаны реликты потоков и значительное количество пирокластитов. Наблюдается и несколько даек и штокообразных тел.

Латито-андезиты концентрируются в субширотной разломной зоне, прослеживающейся южнее г. Смолян, на протяжении 4,5 — 5 km. В пределах насчитывается около 15 даек и штокообразных тел, распространяющихся в том же направлении.

В пределах Смолянской депрессии магнитное поле ΔZ отчетливо разделяется на два типа. Первый тип спокойный, с относительно низкой интенсивностью, не превышающей 250 γ4. Он распространяется на метаморфиты, образующие рамку депрессии и палеогеновые вулканогенно-осадочные образования. Интересно отметить, что экструзивы и субвулканические тела фельзитоидных риолитов и латито-андезитов олигоценового возраста не выражены в магнитном поле ΔZ. Это можно связать с интенсивным кислотным выщелачиванием кварц-серицитового типа, которым затронуты вулканические породы, как и с окислением магнитных минералов. Второй тип магнитного поля является сильно дифференцированным и обладает повышенной интенсивностью. Он связан с олигоцен-миоценовыми (?) порфирокластическими риолитами, слагающими центральную и западную части Смолянской депрессии (Переликская структура). Основная интенсивность магнитного поля ΔZ создается этим типом риолитов в интервале от 250 до 500 γ.

Переликская вулканогенная структура сложена почти исключительно кислыми порфирокластическими вулканитами, в меньшей степени вулканокластитами риолитового до риодацитового состава. Они представлены экструзивно-эффузивными, реже субвулканическими фациями. Эффузивные лавы составляют около 60—70% современной поверхности вулканического ареала, мощность которых варьирует в пределах 100 — 800 m, в среднем составляя около 500 — 600 m. Текучие лавы образуют короткие лавинообразные, часто наслаивающиеся один на другой потоки. Экструзивные образования наблюдаются в виде куполов, обелисков, игл и цепей. Наиболее слабое развитие имеют продукты жерлового фациеса, которые на современном эрозионном срезе встречаются очень редко. Они концентрируются в основном в сильно тектонизи-

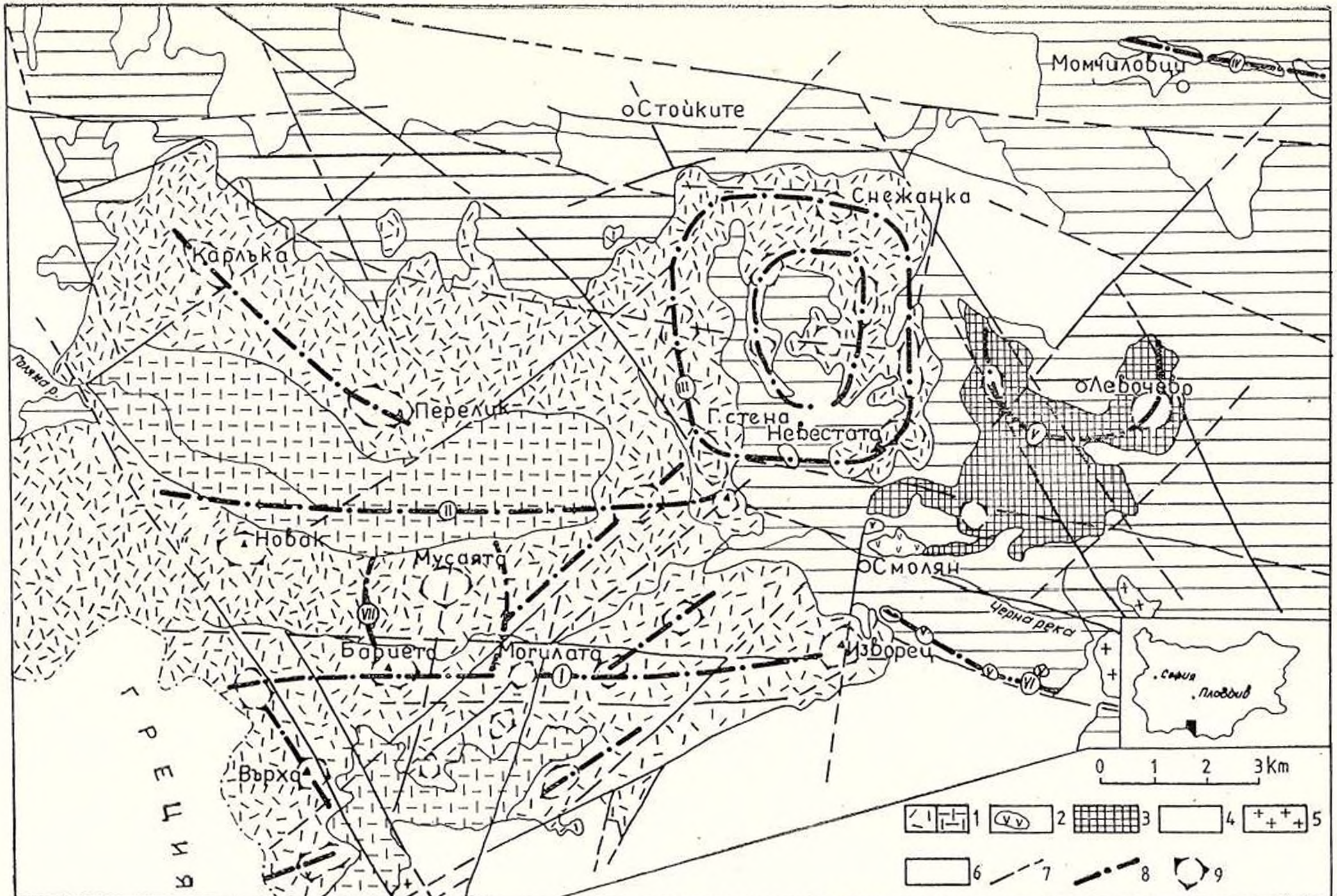

Рис. 1.—Геолого-структурная схема палеогеновых вулканитов Смолянской депрессии.

1. Порфирокластические риолиты: а, эфузивная фация; б, субвулканическая фация: 2, латито-андезиты; 3, фельзитоидные риолиты; 4, терригенно-осадочные отложения палеогена; 5, граниты и гранодиориты палеогенового (?) возраста; 6, докембрийские метаморфиты; 7, разломы; 8, линейные и изометрические (кольцевые и полукольцевые) вулканоструктуры: I — Могильская; II — Гоаеморекская; III — Езеровская; IV — Момчиловская; V — Левочевская; VI — Смолянская; VII — Бариевская; 9, основные вулканические аппараты.

рованных участках и имеют эллиптическую или круглую форму в плане, с размерами 80—100 m в диаметре. Субвулканический фациес обыкновенно представлен небольшими штокообразными телами, размером от нескольких сотен и только некоторые из них достигают 1—1,5 km². Пирокластиты имеют незначительное распространение и представляют не более 5—10% от объема вулканического комплекса. Они развиты обычно в периферии вулканических аппаратов, в их интенсивно эродированных и расчлененных по разломам участках. Основная часть пород эксплозивной субфрации перекрыта эффузивными лавами и вскрыта только буровыми скважинами. Мощность пирокластитов 50—150 m.

Переликская вулканогенная структура, размерами около 180 km², имеет почти изометрическую полигональную форму с клавиатурным внутренним строением. Вулканологические исследования и результаты структурного бурения позволили определить ее как близкую к кальдерам обрушения (M c C a l l, 1963) с вертикальной амплитудой по отношению метаморфической рамы на 600—800 m. Последовательность в образовании разновозрастных вулканических продуктов свидетельствует о том, что в ранние стадии развития она испытывала неравномерное мозаично-блоковое поднятие, после максимальных извержений колапсировала, а в поствулканической стадии оформилась как ступенчато-куполовидный свод с центром вершин Голям и Малык Перелик.

В пределах Переликской вулканогенной структуры выделяются элементарные линейные и изометрические структуры, распределение которых контролируется соответственно региональными разломами и тектоническими узлами пересечения или сочленения разрывных нарушений (рис. 1). Основными линейными структурами являются бортовые разломные зоны, которые ограничивают ареал, а также и внутрикальдерные магмопроводящие каналы, маркируемые трещинными экструзивами, линейно расположенными центрами извержения, сериями вулканических гряд и субвулканическими телами. Основная часть вулканических аппаратов приурочена к региональным нарушениям субширотного направления. Лучше всего выражены магмопроводящие разломы Могильской и Голяморекской линейных вулканических структур (рис. 1), которые являются западным продолжением Смолянской разломной зоны (С т е ф а н о в, 1973).

Могильская вулканическая структура расположена южнее Черной реки и представляет собой цепь линейно расположенных в субширотном направлении куполовидных экструзивов, субвулканических тел и линейных центров извержения. Прослеживаются по линии вершин Барнето, Могилата, Харамийски и Изворец на протяжении около 12 km. Высота вулканической цепи над метаморфным цоколем достигает 800—1000 m, а ширина 450—1200 m. Основные центры эффузивной деятельности (Барнето, Спиура, Могилата, Св. Дух и Изворец) сопровождаются рядом сателлитных вулканов.

Голяморекская вулканическая структура протяжением более 10 km, главным образом маркируется одноименным субвулканическим массивом, расположенным севернее Черной реки. Кроме того, к ней принадлежат несколько центров извержения, картируемых южнее (Новак, Мусалата) и восточнее (Малката стена) субвулканического массива.

Разломы северо-восточного, северо-западного и субмеридионального направления играли второстепенную роль при формировании риолитового ареала. Чаще всего к ним приурочены сателлитные вулканические центры или мелкие субвулканические тела, вследствие чего предполагается более позднее проявление этих нарушений как магмопроводящих каналов в отношении нарушений субширотной системы. Они хорошо развиты в южной части района. Длина локальных вулканических гряд обыкновенно несколько сотен метров, а вулканических цепей достигает 4—5 km. Особенно характерной является Карлыкская магмопроводящая структура северо-западного направления, проходящая между вершинами Голям Перелик и Карлыка. К этой вулканической структуре, прослеживающейся на протяжении около 6 km, приурочены куполовидные экструзивы, вытянутые в том же направлении.

Пространственное положение изометрических кольцевых и полукольцевых вулканических структур обусловлено пересечением субширотных и субмеридиональных разломов. Наиболее хорошо выраженными и значительными по размерам являются Езеровская и Бариевская структуры (рис. 1).

Езеровская вулканическая структура, расположенная севернее г. Смолян, имеет форму близкую к изометрической и хорошо выражена в морфоструктуре района. Представлена кольцевыми телескопированными трещинными экструзивами с внешним диаметром 3,5—4 km. К оконтуривающим ее дуговым разломам приурочен ряд моновулканов с удлиненной или изометрической формой. Наиболее важными из них являются Невестата, Соколица, Снежанка, Голямата стена.

Бариевская структура имеет кальдерное строение и почти полностью совпадает с наиболее глубоко затонувшим (>1000 m) блоком Переликской депрессии. Она заключена между Голяморекской и Могильской линейными вулканическими структурами, а с западной и восточной сторон ограничена субмеридиональными разломами.

Характер магнитного поля в значительной степени отражает основные элементы внутреннего строения Переликской вулканогенной структуны (рис. 2). Хорошая сходимость магнитных аномалий с элементарными вулканоструктурами по направлению, форме и размерам свидетельствует о слабо эродированном рельефе и относительной однородности вулканических масс, что подтверждается геологическими и морфологическими данными. На фоне регионального магнитного поля, создаваемого порфирокластическими риолитами (250—600 γ) выделяются три вида магнитных аномалий. В первом виде объединены аномалии изрезанного (пилообразного) характера (положительные, отрицательные и знакопеременные). Аномалии этого вида расположены в южной и восточной частях Переликской структуры, образуя полукольцевую полосу с выпуклой стороной к метаморфической раме. Южная часть полосы представлена групповой аномалией преимущественно негативного характера и с вытянутой формой в субширотном направлении. Она совпадает с Могильской линейной вулканоструктурой. Восточная группа образует кольцевую аномалию с преобладанием положительных локальных аномалий. Она проявлена над Езеровской изометрической вулканоструктурой и в соответствии с ее телескопированным строением в геофизическом поле выделяются вложенные одна в другую магнитные аномалии разной интенсив-

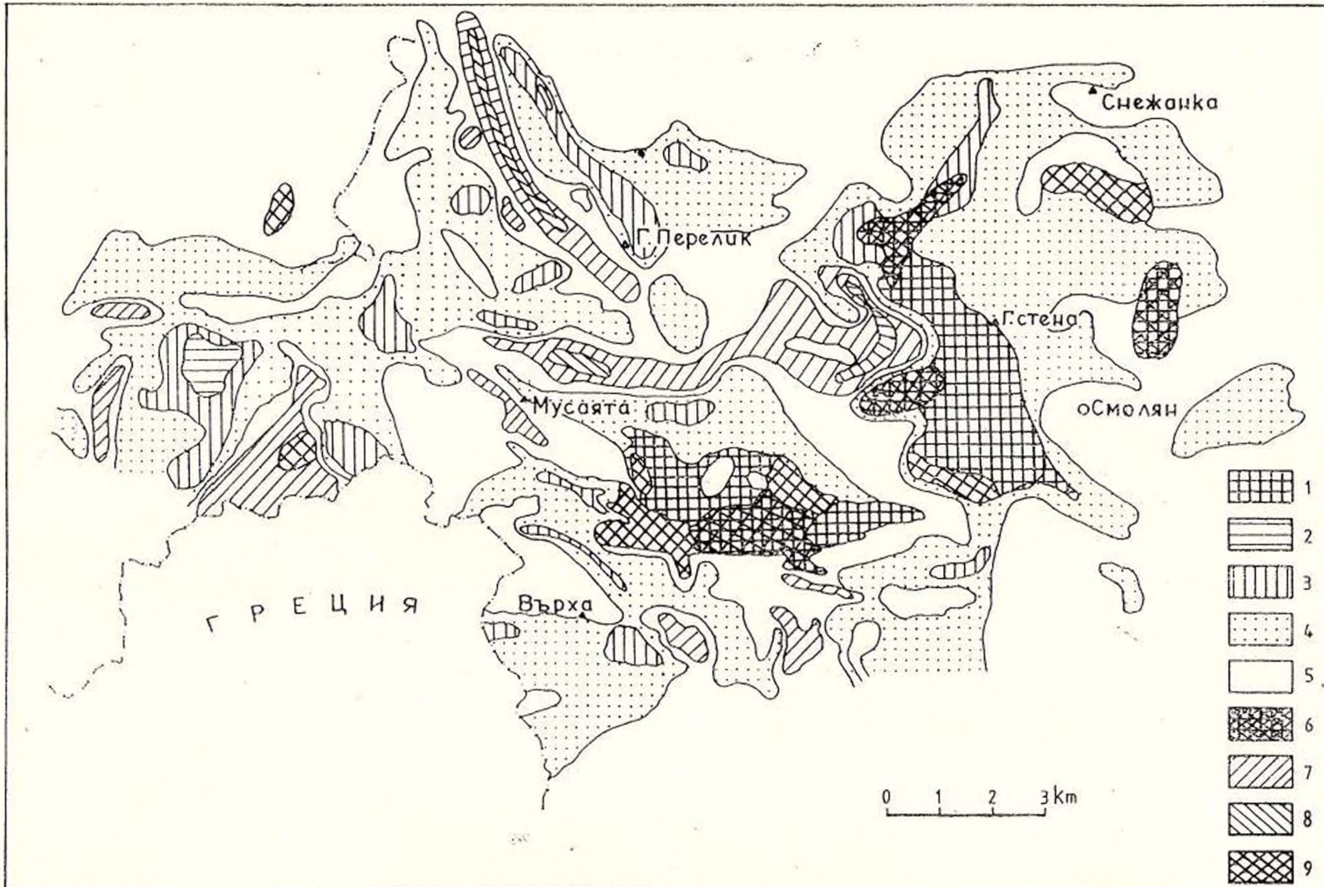

Рис. 2. — Схема геомагнитного поля Δz в пределах вулканогенных образований Смолянской депрессии.

Магнитное поле: 1, изрезанное положительное; 2, $> 750\gamma$; 3, $- 500 - 750\gamma$; 4, $- 250 - 500\ \gamma$; 5, $0 - 250\ \gamma$; 6, изрезанное знакопеременное; 7, $- 250 - 0\gamma$; 8, $- 250 - 600\ \gamma$; 9, изрезанное отрицательное.

ности или знака. Почти полное совпадение аномалий пилообразного характера с участками максимальной насыщенности центров извержения, свидетельствуют о их генетической связи. Имея в виду, что в экструзивных риолитах количество и размеры зерен рудных минералов значительно больше по сравнению с эффузивными лавами (Б а х н е в а и др., 1978), предполагается, что аномалии изрезанного характера обусловлены литофациальными особенностями вулканитов. Кроме того, в эффузивах рудные акцессории затронуты окислением и гидротермальными изменениями, что тоже влияет на интенсивность магнитного поля. Различия по знаку двух перечисленных аномалий, по всей вероятности, связаны с переориентировкой вектора намагниченности, в связи с внедрением субвулканических тел в Могильскую вулканоструктуру.

Второй вид магнитных аномалий включает интенсивные отрицательные аномалии до 1000 γ. Они расположены внутри полукольцевой полосы с изрезанным характером магнитного поля. Четко выраженная аномалия этого вида, вытянутая в субширотном направлении, прослеживается на 9 km по южному борту Голяморекского субвулканического массива. По всей вероятности, она маркирует крутозалегающий магмопроводящий канал Голяморекской вулканоструктуры. Большая магнитная аномалия северо-западного направления, длиной 6 km совпадает с Карлыкской линейной вулканоструктурой. Она отличается асимметрией, выражающейся в появлении с северовостока параллельно ей положительного поля линейной формы. Это устойчивое сочетание магнитных аномалий противоположных знаков, подобно современным вулканам Камчатки (Р и в о ш, Ш т е й н б е р г, 1964), можно объяснить эффективной намагниченностью или наличием крутозалегающего к северо-востоку линейного магмопроводящего канала.

Магнитные аномалии третьего вида характеризуются интенсивностью, превышающей 500 γ. За исключением вышеупомянутой, сопровождающей Карлыкскую аномалию, они имеют ограниченное распространение и небольшие размеры. Прерывистая полоса аномалий этого вида, северо-западного простирания, установлена в западной части Переликской структуры (рис. 2). Она протягивается вдоль Полянской разломной зоны (С т е ф а н о в, 1973) и одноименной вулканической цепи.

Необходимо отметить еще одну особенность магнитного поля, создаваемого порфирокластическими риолитами — это наличие нескольких изометрических полей с низкой интенсивностью до 250 γ. В сочетании с основным полем риолитов (250—500 γ) эти аномалии могут быть объяснены как отражение блокового строения подошвы депрессии, приведшей, в свою очередь, к формированию риолитовых масс с различной мощностью. Мы предполагаем, что именно эти аномалии с относительно низкой интенсивностью соответствуют поднятым блокам метаморфитов. И наоборот, интенсивные аномалии с различными структурами магнитных полей, описанные выше, наряду с информацией об условиях формирования риолитов, показывают и более мощные массы вулканитов, образовавшихся на опустившихся блоках фундамента. Геологические доказательства этих суждений находим в том факте, что периферийные блоки южной части вулканического ареала с низкой интенсивностью (ниже 250 γ), по данным структурного бурения и развитию жерловых и субвулканических тел, приподняты по сравнению с Барневской кальдерой.

Выводы

1. Магнитное поле над третичными вулканитами Смолянской депрессии зависит не столько от состава, сколько от поствулканической гидротермальной и термообработки. Поэтому фельзитоидные риолиты и латито-андезиты отличаются спокойным магнитным полем и низкой интенсивностью (не более 250 γ), а порфирокластическим риолитам Переликской структуры соответствует сильно дифференцированное, с высокой интенсивностью магнитное поле ΔZ (обычно варьирующее от 250 до 700 γ). Оно зависит преимущественно от фации и мощности вулканических пород.

2. Переликский вулканический блок является структурой инверсионного типа, о чем свидетельствуют ступенчато-центриклинальное погружение кристаллического фундамента с палеогеновым чехлом и горстово-периклинальное с риолотовой надстройкой.

3. Прикальдерное пространство Переликской структуры, максимально насыщенное центрами извержения, маркируется магнитными аномалиями изрезанного характера. Высокая интенсивность аномалии объясняется повышенной концентрацией рудных акцессориев в породах экструзивов и субвулканических тел. Внутрикальдерная площадь вмещает преимущественно линейные вулканические структуры, выраженные в магнитном поле вытянутыми отрицательными аномалиями, иногда с асимметричным строением.

4. Предполагается, что поднятые блоки фундамента соответствуют участкам с низкой интенсивностью в пределах сильно дифференцированного магнитного поля.

[4] $1\ \gamma = 10^{-5}$ Oe B CGSM = 79,58 A/m BSI.

ЛИТЕРАТУРА

Бахнева-Стефанова Д., Стефанов, Н. (1971) Формационна принадлежност на риолитите от Смолянската депресия. *Год. ВМГИ*. 12,2.

Стефанов Н. (1975) Фации и вулкано-структурная обусловленность кислого вулканизма Центральных Родоп (на примере Переликской структуры). *Geologica Balcanica*

Ножаров Н., Стефанов Н., Кръстев Н. (1978) Вулканоструктури, магнитно поле и магнитни свойства на риолитите от северната периферия на Брацигово-Доспатската структура. *Сб. докл. науч. -техн. конференция „25 год. ВМГИ"*.

Стефанов Н. (1973) Основни разломни структури в Смолянската депресия. *Сб. докл. науч.-техн. конфер. „20 год. ВМГИ"*, 2.

Донских В. В. (1980) Структурный анализ вулканогенных образований. Методика геологической съемки древних вулканов. Ленинград, Недра.

Ривош Л. А., Штейнберг Г. С. (1964) Геофизическое изучение Камчатских вулканов. *Геология и геофизика*, 7.

McCall G. J. (1963) Classification of calderas: "Kracatoon" and "glencoe" Tyepes. *Nature*, 197: 4863.

GÉOCHRONOLOGIE ISOTOPIQUE
ISOTOPIC GEOCHRONOLOGY
ИЗОТОПНАЯ ГЕОХРОНОЛОГИЯ

CHRONOLOGY OF MIOCENE PYROCLASTICS AND LAVAS OF HUNGARY [1]

BY

E. ÁRVA-SÓS[2], KADOSA BALOGH[2], GEZA HÁMOR[3], ARON JÁMBOR[3], L. RAVASZ-BARANYAI [3]

The most important application of the K/Ar method in Hungary is the study of rocks younger than Mesozoic. This is justified on the one hand by the major role played by Miocene volcanic rocks in the geological structure of Hungary and by the search for mineral deposits associated with them : on the other hand, by the fact that these rocks cannot be dated with and acceptable accuracy by other radiometric methods.

The essential purpose of our organizing a K/Ar laboratory has been the accomplishment of a systematic radiometric chronologic study of Miocene volcanic rocks. The work was carried out by cooperation between the Hungarian Geological Institute, Budapest (MÁFI) , and the Institute of Nuclear Research of the Hungarian Academy of Sciences,Debrecen(ATOMKI). This project has been initiated and is directed by H á m o r .

The analyses were carried out mainly on biotite and feldspar separated from Miocene pyroclastics, and on lava whole rock samples.

In addition to dating in the ATOMKI there has been a systematic development of experimental methods and instruments. A magnetic mass spectrometer, capable of static operation has been built in 1976 (B a l o g h, M ó r i k , 1978). A high capacity argon extraction and purification system has been developed in 1977 (B a l o g h, M ó r i k, 1979) : this work has been sponsored by the Central Office of Geology. In 1979 a microprocessor unit for automatic recording and evaluating the argon mass spectra has been put in operation (M o l n á r, P a á l, 1980). With these instruments the radiogenic argon content of terrestrial material can be determined with an accuracy and precision, which meet the modern requirements. The quoted age data have been calculated using atomic constants suggested by the Geochronological Subcommission at the Sydney meeting in 1976 (S t e i g e r, J ä g e r, 1977). Because of the great number of age data and shortness of time we are not entering into a detailed critical

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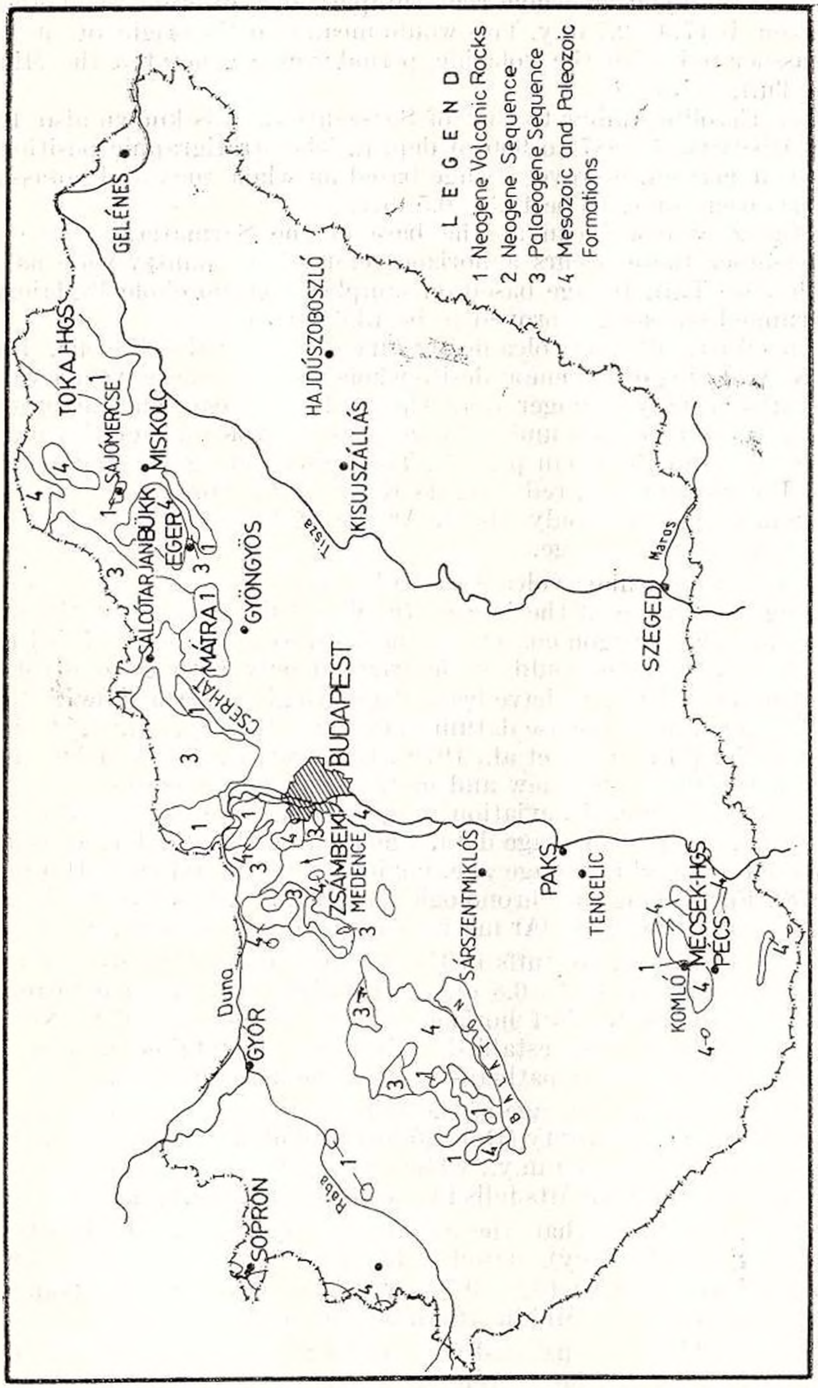
evaluation but we present only the most essential results progressing from one region to the other.

The areas investigated are shown in Figure.

In the Mecsek Mts the Miocene volcanism is represented by hypersthene andesites, rhyolitic and rhyodacitic tuffs. In the vicinity of Komló the andesite occurs even at the surface, though it is known from boreholes of other regions too (for example Nagymányok). However undecomposed samples suitable for age determination could be collected only from the locality of Komló. According to the publication of Árva-Sós and Ravasz (1976), its K/Ar age is 20.5±0.8 m.y. Later on, the control measurements carried out by more advanced methods, did verify this result, with the only change that the actual value of K/Ar age has been indicated to be in the 19—20 m.y. range. The rhyolite tuffs of the Mecsek Mts can be assigned to the Lower Rhyolite Tuff horizon of the country, the rhyodacitic tuffs to the so-called Middle Rhyolite Tuffs. The K/Ar dating of the Lower Rhyolite Tuff is particularly difficult as tectonic movements extending well into the Pleistocene resulted in some samples of a younger radiometric age. In the Mecsek Mts the age of the Lower Rhyolite Tuff horizon does not differ more from that of the andesites than the standard deviation of our age data. More reliable results may be expected only when the $^{39}Ar/^{40}Ar$ method will be used. The average age of rhyodacitic tuffs in the Mecsek Mts agrees, within standard deviation range, with the 16.4±0.8 m.y. value that we published earlier for the Middle Rhyolite Tuff horizon (Hámor et al., 1979).

Borehole Tengelic-2 NE from the Mecsek Mts crossed several hundred meters of uninterrupted pyroclastics, most of which is perlitic rhyodacite, though some rhyodacitic flood-tuff and rhyodacitic rheoignimbrite does also occur. Lithologically the pyroclastics may be assigned to the Middle Rhyolite Tuff horizon. The K/Ar age was determined on separated biotite fractions at 11 points of the interval between 872.5 and 1165.5 m. The obtained average age has been 15.6±0.7 m.y. No decrease in age towards the higher horizons has been observed. The foot-wall of the volcanic complex is Ottnangian variegated clay, while the overlying sediments — regarding their microfauna — may be placed in the NN5 zone (A. Nagymarosi, pers. com.). Thus the K/Ar age may be regarded as a datum, which is somewhat younger than the Carpathian — Badenian boundary. Borehole Paks-2 crossed again volcanic rocks overlying calcareous sandstones in a thickness of several hundred meters. The lower layers are built up of decomposed biotite-hornblende dacito-andesite. The age of this complex has been analyzed with biotites separated from samples at 6 different points in the interval of 1296.5—1502.5 m. The age of the slightly decomposed and oxidized biotites could be determined only with greater error. On the average their age is 18.5±1.7 m.y., which means that the dacito-andesitic volcanic complex in borehole Paks-2 is similar in age to the andesite of Komló. The andesitic complex is overlain in several horizons by what could be petrologically identified as redeposited rhyolitic tuff. Its average age based on biotite from the interval 943.9 to 1094.6 m is 18.7±1.2 m.y., but the interpretation of this datum, given the redeposited character of the samples, is uncertain.



Neogene volcanic areas in Hungary.

At Sárszentmiklós alkali rhyolite is exposed at the surface. Its apparent age obtained on whole rock samples and different magnetically split fractions is 17.0±0.7 m.y. This would mean that the origin of the rock may be associated with the volcanic period which generated the Middle Rhyolite Tuff.

Alkali rhyolite similar to that of Sárszentmiklós is known also from borehole Albertirsa-1 (887 to 900 m depth). The stratigraphic position of this rock is uncertain, however its age based on whole rock and potassium feldspar has been found to be 14.3±0.5 m.y.

In the Zsámbék Basin at the base of the Sarmatian, above the Szomor member there occurs a horizon referred to country-wide as the Upper Rhyolite Tuff. Its age based on samples from borehole Budajenö-3 and determined on biotites proved to be 13.6±0.6 m.y.

In the Mátra Mts the volcanic activity started in the Eocene. However the K/Ar dating of Eocene andesite whole rock samples gave age values that are substantially younger than the geological age. This divergence may be due to rejuvenation under the influence of volcanic activity during the Carpathian and Badenian periods. This rejuvenating affected also the biotite of Eocene andesite, reducing its K/Ar age to about 30 m.y. According to our investigations, only the K/Ar age of hornblendes is in agreement with the geological age.

Carpathian-Badenian volcanism had a rejuvenating effect also on the K/Ar age of biotites of the Lower Rhyolite Tuff horizon. At the same time the atmospheric argon content of the biotites also increased and thus the age of some samples could be determined only with great error. In view of these difficulties, we have been unable to give for the Lower Rhyolite Tuff horizon a more precise datum than 19.6±1.4 m.y., which has been published earlier (H á m o r et al., 1979). However, on the basis of measurements carried out with a new and more advanced technique a younger value within the standard deviation range seems to be more probable as some of the earlier published age data, which had great standard deviation and indicated a too old average age, could not be reproduced (H á m o r et al., 1978). For solving the chronological problems of the Lower Rhyolite Tuff horizon the use of $^{39}Ar/^{40}Ar$ method seems to be absolutely necessary.

The age of rhyodacite tuffs in the Cserhát and Mátra Mts does not differ from the value of 16.4±0.8 m.y., which has been earlier determined for the Middle Rhyolite Tuff horizon (H á m o r et al., 1979). No age difference could have been established in these mountains between the rhyodacite tuffs and the Carpathian-Lower Badenian andesites.

In the Cserhát Mts the age of the subvolcanic andesites representing the end of the volcanic activity (Dobogó-hill, Hollókö, Bercel-hill, Szanda-hill) varies from 13.5 to 15.0 m.y., while the age of the uppermost, covering, andesites in the Mátra Mts falls in the 13.5 to 14.0 m.y. range.

We wish to remark that the age of the rhyolite in the Mátra Mts (Gyöngyössolymos, Kishegy), dated with separated biotites and whole rock samples, proved to be 15.8±0.5 m.y., thus its origin corresponds to the volcanic period of the Middle Rhyolite Tuff horizon.

The age of biotites separated from rocks petrologically identified as Lower Rhyolite Tuff in the Cserépváralja quarry, the borehole Cserépváralja-1 and at Répáshuta in the foothills of the Bükk range, falls in the 16.2 to 19.5 m.y. range. A more accurate age determination may be expected again from the $^{39}Ar/^{40}Ar$ method.

In borehole Alsóvadász-1 in the Cserehát the age of the rhyolite tuff situated directly above the Carpathian-Badenian boundary has been examined. The biotite-based average age, 15.6±0.7 m.y., is in agreement with the results expected on the basis of the stratigraphic position.

North of the Bükk Mts near Sajómercse, above the Lower Pannonian (?) sediments, basaltic andesites with a K/Ar age of 9.6±0.8 m.y. are exposed at the surface. In the Tokaj Mts the analyses revealed that the formation of a considerable part of the dacites and andesites took place already in the Pannonian.

Along the Hungarian-Soviet frontier, borehole Gelénes-1 crossed the rhyolite tuff complex in a thickness of 1417 m and stopped in it at 2002 m depth. At the top of this volcanic complex, at 631 m, the biotite-based age of the tuff has been 11.0±0.6 m.y., which may be regarded as the end-date of the rhyolite tuff volcanism.

In the Pannonian Basin biotite — and feldspar — based age of rhyolite from the borehole Kisujszállás-EK-1 from 1664—1682 m and 1863—1880 m depths has been on the average 18.25±0.3 m.y. This rock is extremely suited for age-determination, though the precise K/Ar age is of reduced value since the exact stratigraphic position of the rock is still unknown.

Conclusions

On the basis of our investigations the following conclusions can be drawn :

1. In Hungary the rhyolitic-andesitic (basalto-andesitic) products of Miocene volcanism extruded during three longer volcanic cycles in the interval between about 20.0 and 9.6 m.y.

2. The age of rhyolitic products of the earliest volcanic activity has been approximated with the datum of 19.6±1.4 m.y. ; however, in many regions, where volcanism was active even later, a more precise age can be expected only from the $^{39}Ar/^{40}Ar$ method.

3. The middle volcanic cycle was of the greatest extent, its rather long period was a continuous activity, well identified within the 14 to 17 m.y. interval, its most intensive phase is approximated with the datum of 16.4±0.8 m.y.

4. The nation-wide Upper Rhyolite Tuff is dated between 13.6 and 11.0 m.y., this volcanic cycle in NE Hungary lasted longer and was more powerful.

In spite of the great number of available measurements our work still has the character of reconnaissance chronology. Detailed age measurements in the volcanic mountains of Hungary and comparison of the age patterns of the individual volcanic areas are tasks to be achieved in the future. More precise data are expected from the $^{39}Ar/^{40}Ar$ method.

REFERENCES

Árva-Sós E., Ravasz Cs. (1978) K-Ar dating of the andesite of Komló (SE Transdanubia) (in Hungarian). *Ann. Rep. of the Hung. Geol. Inst. of 1976. Müszaki Kiado.* Budapest, p. 201–208.

Balogh K., Mórik Gy. (1978) Magnetic mass spectrometer for K-Ar dating (in Hungarian). *ATOMKI Közl.*, 20, p. 215–228.

— Mórik Gy. (1979) High capacity argon extraction and purification system (in Hungarian). *ATOMKI Közl.*, 21, p. 363–375.

Hámor G., Ravasz-Baranyai L., Balogh K., Árva-Sós E. (1979) K-Ar dating of Miocene pyroclastic rocks in Hungary. *Ann. Géol. Pays Hellén*, p. 491–500.

— Balogh K., Ravasz-Baranyai L. (1978) Radiometric age of the Tertiary formations in North Hungary (in Hungarian). *Ann. Rep. of the Hung. Geol. Inst. of 1976. Müszaki Kiadó.* Budapest, p. 61–76.

Molnár I., Paál A. (1980) Microprocessor evaluating system of a magnetic mass spectrometer for the determination of the geological age of K-Ar (in Hungarian). *Mérés és Automatika*, XXVIII, 12, p. 469–472.

Steiger R. H., Jäger E. (1977) Subcommission on geochronology: convention on the use of decay constants in geo- and cosmochronology. *Earth. Planet. Sci. Lett.*, 36, p. 359–362.

CHRONOLOGY OF GRANITOID AND METAMORPHIC ROCKS OF TRANSDANUBIA (HUNGARY)[1]

BY

KADOSA BALOGH[2], E. ÁRVA-SÓS[2], G. BUDA[3]

The granitoid rocks of Transdanubia (West-Hungary) have been classified into two main groups (B u d a, 1975) :

I. Synkinematic granitoids with close association with regional metamorphic rocks formed by anatexis and K-metasomatism occurring in and around the Mecsek Mts (South-Transdanubia).

II. Postkinematic granitoids occurring along the Balaton-Velence fault zone with thermal-metamorphic aureole (Fig. 1).

Chronology of the Synkinematic Anatexite and Metamorphic Rocks

Chronology of anatexitic granitoids and metamorphic rocks were studied by U/Pb, Rb/Sr, and K/Ar methods in the Mecsek Mts and in the southern foreground of the Mecsek Mts (Göresöny horst). The radiometric ages determined by different methods reflect the time of various events of formation of granitoids and metamorphic rocks. A few number of K/Ar age determinations were made by O v c h i n n i k o v et al. (1965) and a greater number of K/Ar mica and amphibole ages were published by Á r v a - S ó s and B a l o g h in 1979. Additional K/Ar mineral ages were obtained later with improved analytical methods. These were in accordance with the previously published data, and will be jointly evaluated here. Results of Rb/Sr studies on whole rock and biotite samples were presented by K o v á c h and S v i n g o r at the XIth Congress of the CBGA in Kiev in 1977.

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[1] Paper presented at the 12th Congress of the Carpatho-Balkan Geological Association, September 8—13, 1981, Bucharest, Romania.

[2] Institute of Nuclear Research of the Hungarian Academy of Sciences, H-4026, Debrecen, Bem tér 18/c.

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U/Pb dating was made on titanite and zircon fractions of different density, mesh size and magnetic susceptibility eparated from the granitoid rock of the quarry at the village of Mórág M esek Mts) by K o u v o.

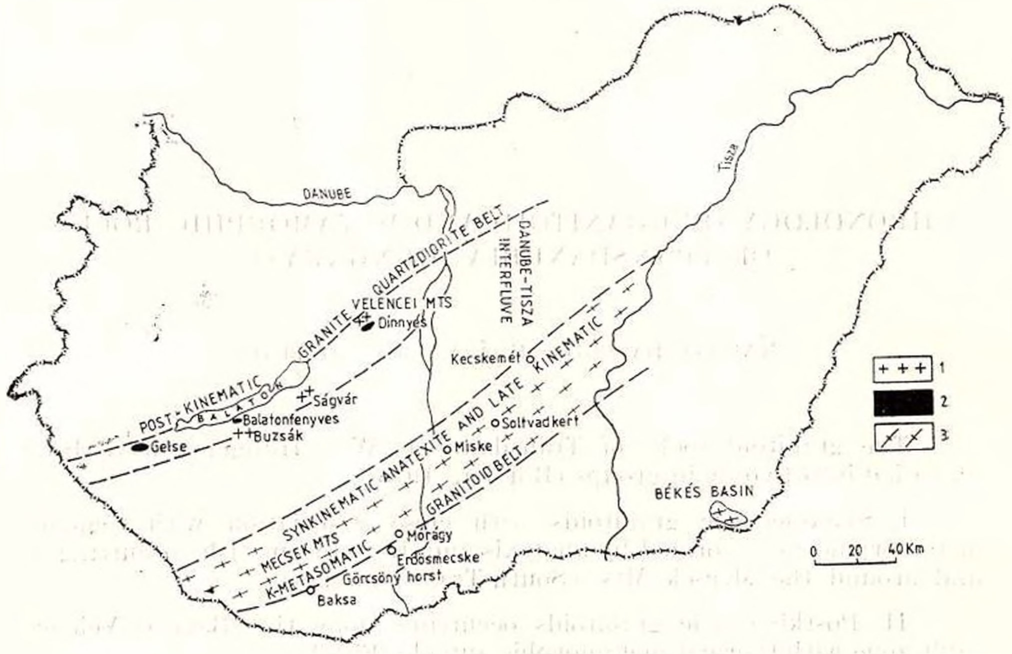

Fig. 1. — Map showing the granitoid occurrences of Hungary.
1, granite and granodiorite; 2, quartzdiorite; 3, anatexitic granodiorite and granite.

The U/Pb ages are summarized in Table 1. The discordance of $^{238}U/^{206}Pb$, $^{235}U/^{207}Pb$ and $^{206}Pb/^{207}Pb$ data show clearly an open system behav-

TABLE 1

U/Pb ages of granitoid rocks from the quarry at Morágy, Mecsek Mts

Sample No	Zircon fractions of different mesh size, density and magnetic susceptibility g cm^{-3}/mesh size	^{238}U ppm	Radiogenic ^{206}Pb ppm	Isotopic ratios $\frac{206}{204}$	$\frac{207}{204}$	$\frac{208}{204}$	Radiometric ages, m.y. $\frac{^{206}Pb}{^{238}U}$	$\frac{^{207}Pb}{^{235}U}$	$\frac{^{207}Pb}{^{206}U}$
A565A	+4.2/+200	1850	65.02	164.7	23.80	57.10	256±2	269±3	380±13
A565B	+4.2/+200 NM	1302	52.42	249.7	28.30	64.03	293±9	300±4	354±17
A565C	+4.2/−200	1536	61.15	225.9	27.09	63.02	289±2	298±3	368±13
A565D	4.0−4.2/+200	2828	76.40	75.57	18.82	48.84	198±2	214±2	398±18
A565E	titanite	241	11.47	33.81	16.66	45.08	344±2	341±6	319±40
A565F	+4.6/+100	1315	52.30	315.7	32.03	69.45	289±2	299±4	378±22
A565G	+4.5/−200	1627	60.39	245.9	28.18	65.53	270±1	281±3	369±20
A565H	4.2−4.6/−200	2498	78.58	138.3	22.37	55.11	230±2	243±2	377±17
A565K	3.6−3.8	1948	68.18	172.7	24.13	69.21	255±1	265±6	348±43

iour, characterized by greater radiogenic lead loss in higher uranium content mineral fractions. The data plotted in the $^{207}Pb/^{235}U - ^{206}Pb/^{238}U$ diagram define a straight line (Fig. 2), suggesting a simple cogenetic sequence. An intersection age of 365±8 m.y. has been established by using potash feldspar lead for common lead correction. This is interpreted as

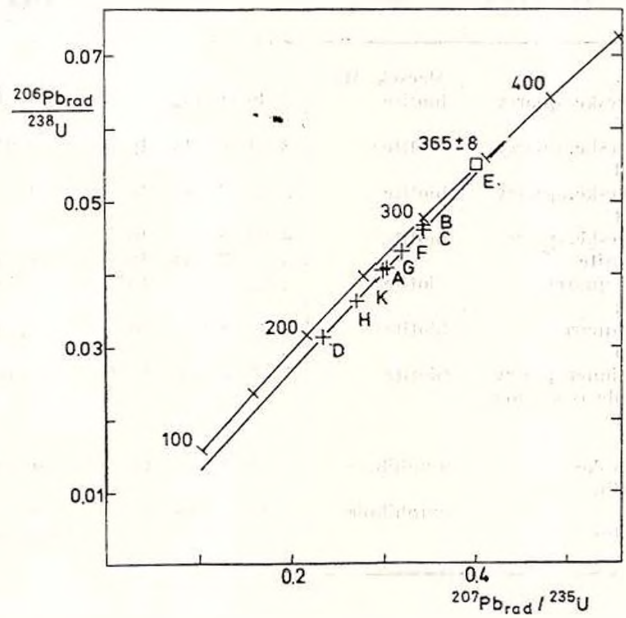

Fig. 2. — Concordia diagram and U-Pb isotopic ratios for zircon and titanite samples from the granitoid rocks of the Mecsek Mts (at village Mórágy). Analysis by O. K o u v o.

the oldest age of the anatexitic granitoids, because a subsequent recrystallization resulting in the complete loss of radiogenic lead is considered to be unlikely. The U/Pb isochron ages are the least sensitive to low-grade metamorphic events and reflect the time of primary magmatic crystallization even when the total rock Rb/Sr ages are lowered (P a g e, 1978).

According to the Rb/Sr studies of K o v á c h and S v i n g o r (1977), the anatexitic granitization can be confined to the 403—273 m.y. interval. Considering the analytical error of individual Rb/Sr ages and the possible inhomogeneity of the initial $^{87}Sr/^{86}Sr$ ratios, a significant deviation between the U/Pb and Rb/Sr data cannot be established.

The K/Ar biotite ages obtained on granitic rocks from the quarries at villages Erdősmecske and Mórágy (Mecsek Mts) are summarized in Table 2. The scatter of data in the 318—352 m.y. interval is due mainly to analytical error, a shorter real time span can be supposed for the closure of the biotite argon and potassium system. The average biotite age is 334±11 m.y. Metamorphic rocks from the neighbouring area have been dated with separated amphibole. An average age of 337±18 m.y. has been obtained, which is in good agreement with the mean biotite datum.

TABLE 2

K/Ar age of granitoid and metamorphic rocks from the Mecsek Mts

Sample No	Locality and rock type	Dated	K	$^{40}Ar_{rad}$ / ccSTP/g	$^{40}Ar_{rad}$ / $^{40}Ar_{tot}$	Age m.y.
		Mecsek Mts				
143	Erdősmecske, quarry granitoid	biotite	7.49	1.052×10^{-4}	0.91	330±11
242	Erdősmecske, quarry granitoid	biotite	6.61	9.943×10^{-4}	0.97	352±12
187	Erdősmecske, quarry granitoid	biotite	5.41	7.880×10^{-5}	0.95	341±16
209	Erdősmecske, quarry microgranite	biotite	6.94	9.358×10^{-5}	0.95	318±16
			5.08	7.453×10^{-5}	1.00	343±19
185	Mórágy, quarry granitoid	biotite	6.27	8.930×10^{-5}	0.99	334±15
141	Mórágy, quarry granitoid	biotite	7.68	1.057×10^{-4}	0.50	323±13
535	Mórágy, inner quarry at the railway station, granitoid	biotite	7.57	1.061×10^{-4}	0.91	329±11
					Average age :	334±11
321	Erdősmecske amphibolite	amphibole	0.246	3.388×10^{-6}	0.79	324±34
332	Bátapáti amphibolite	amphibole	0.407	6.099×10^{-6}	0.86	350±20
					Average age :	337±18

TABLE 3

K/Ar ages of metamorphic rocks from the Görcsöny horst

Sample No	Locality and rock type	Dated mineral	K %	$^{40}Ar_{rad}$ / ccSTP/g	$^{40}Ar_{rad}$ / $^{40}Ar_{tot}$	Age m.y.
	Borehole Baksa-2 biotite-muscovite gneiss					
646	115 m	biotite	8.05	9.706×10^{-5}	0.99	287±11
645	268 m	biotite	6.92	8.747×10^{-5}	0.95	299±11
		muscovite		1.007×10^{-4}	0.95	309±11
647	366 m	biotite	6.90	8.536×10^{-5}	0.95	293±11
		muscovite	7.97	9.994×10^{-5}	0.96	297±11
648	752.9 m	biotite	6.41	7.701×10^{-5}	0.95	286±11
		muscovite	7.75	1.017×10^{-4}	0.94	310±11
644	1094 m	biotite	7.63	8.789×10^{-5}	0.98	275±11
		muscovite	7.79	1.014×10^{-4}	0.96	307±11
184	Borehole Gyód-3 amphibolite	Average ages : biotites : 288±10 Ma ; muscovites 306± Ma				
		amphibole	0.485	$6.372 . 10^{-6}$	0.68	310±14

Both the biotite and amphibole mineral ages are interpreted as the datum when the rock cooled below its blocking temperature, which is 1–200°C higher for amphibole. The concordance of biotite and amphibole ages implies that anatexitic granitization was followed by an uplift 334±±11 m.y. age. By this time the granitoids reached a lower temperature region of the crust.

Metamorphic rocks occurring in the Görcsöny horst (southern foreground of the Mecsek Mts) are reached by several deep drillings. The chronologic study was concentrated on the cores of borehole Baksa-2; dating was done on biotite and muscovite mineral fractions (Tab. 3). The average biotite and muscovite ages are 288±10 m.y. and 306±6 m.y. respectively, the radiometric ages do not change systematically with depth. Allowing for a possible lowering of the biotite ages in consequence of low temperature leaching or hydrothermal processes, no uplift rate is given from the difference of age and blocking temperature of the dated minerals. The significant difference of mineral data in the Mecsek Mts and in the Görcsöny horst indicates that the emergence of the Mecsek Mts took place earlier than the uplift of the Görcsöny horst.

Chronology of Postkinematic Granitoids

K/Ar data measured on granitoid rocks from along the Balaton-Velence Mts main fault line are presented in Table 4. With the exception of samples No. 744 and 745 all the rocks were collected from the Velence Mts.

Monzogranite was reached by borehole Ságvár-3; its K/Ar datum of 259±10 m.y. obtained on a chloritised biotite, is considered only as a

TABLE 4

K/Ar ages of plutonic rocks along the Balaton-Velence Mts main fault line dated minerals: biotite and chloritized biotite

Sample No	Locality and rock type	K %	$^{40}Ar_{rad}$ / ccSTP/g	$^{40}Ar_{rad}$ / $^{40}Ar_{tot}$	Age m.y.
744	Borehole Ságvár-3 monzogranite	2.86	3.091×10^{-5}	0.95	259±10
741	Borehole Székesfehérvár-4. 57.8 m granite	6.24	7.932×10^{-5}	0.96	280±11
649	Székesfehérvár, quarry granite	5.13	4.033×10^{-5}	0.70	192±8
602	Székesfehérvár, quarry granite	6.34	4.485×10^{-5}	0.85	246±10
690	Székesfehérvár, quarry pegmatite	5.66	3.805×10^{-5}	0.61	165±7
650	Pákozd, quarry granite	4.80	2.648×10^{-5}	0.58	137±6
536	Sukoró, quarry granite	5.95	7.294×10^{-5}	0.82	291±11
689	Sukoró, quarry granite	4.65	5.242×10^{-5}	0.77	271±11
691	Nadap, granite	3.69	2.959×10^{-5}	0.63	196±9
743	Borehole Dinnyés-3, 777.7 m granodiorite	5.91	6.737×10^{-5}	0.94	272±11
742	Borehole Sukoró-1. 63.85–63.95 m beforsite	6.92	2.132×10^{-5}	0.80	77.6±3.0
745	Borehole Balaton-fenyves-1, 603 m tonalite	7.50	$9.026 . 10^{-6}$	0.62	30.7±1.0

minimum age of plutonism. The most likely assumption is that this rock is coeval with the Carboniferous granite intrusions of the Velence Mts. This additionally implies that chloritisation resulted only in a relatively small lowering of the K/Ar age, therefore chloritization alone is not a satisfactory explanation of strongly reduced K/Ar data.

The granite intrusions of the Velence Mts crystallized at a depth of 4—5 km (B u d a, 1980), where a fast cooling below the blocking temperature of biotite can be expected. Therefore in case of undisturbed potassium and argon system the K/Ar data should well approximate the time of granite emplacement. As indicated by the highly discordant data the granite was probably affected by the younger Eocene andesitic volcanism occurring in the Mountains. Since partial argon loss is proved by unequivocal geologic evidence, even the oldest data are only minimum ages of granite intrusion.

The age of granodiorite reached by borehole Dinnyés-3 shows that this rock is a more basic derivative of the granite magma.

Unaltered beforsite dykes (I. H o r v á t h, personal communication) in granite, penetrated by borehole Sukoró-1, are considered to be the products of a younger magmatic activity of the Velence Mts, which is not in genetic connection with the granitoids.

Tonalite, reached by borehole Balatonfenyves-1, resulted in a very young age of 30.7±1.0 m.y. Although the possibility of some argon loss cannot be excluded, the real age of this rock is considered younger than the time of granite intrusion in the Velence Mts, since in the Velence Mts even the most altered rocks are of much older age than the tonalite at Balatonfenyves.

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REFERENCES

Á r v a - S ó s E., B a l o g h K. (1979) Investigations of granites and associated metamorphic rocks by the K-Ar method in the Mecsek Mountains. *Földt. Kut.* XXII, 4, p. 33—36.

B u d a G y. (1975) Classification of the Hungarian granitoid rocks on the basis of feldspar investigation. *Proceedings of the X th Congress of the CBGA. Vol. Tect. (sec. III)*, p. 67-74.

— (1980) Report on the granitoid and contact metamorphic rocks of Velence Mts (Unpublished Report).

K o v á c h Á., S v i n g o r E. (1977) Geochronology of the Chrystalline basement of the Pannonian Basin. *Proceedings of the XI th Congress of the CBGA*, Kiev.

P a g e R. W. (1978) Response of U-Pb zircon and Rb/Sr Total Rock and Mineral Systems to Low-grade Regional Metamorphism in Proterozoic Igneous Rocks, Mount Isa, Australia. *J. Geol. Soc. Austr.*, *25*, p. 141—164.

О в ч и н н и к о в Л. Н. И а н о в а М. В., Ш а н г а р е е в Ф. Л. (1965) Абсолютный возраст некаторых геологических образований Венгрии. *Acta Geol. Hung.* *9*, p. 305—312.

PETROGRAPHY AND K/AR DATING OF TERTIARY AND QUATERNARY BASALTIC ROCKS IN HUNGARY [1]

BY

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Since the biostratigraphic correlation of the Pannonian *s.l.* formations in and around the Carpathian Basin is still a difficult task, the application of the radiometric chronologic methods has a special importance.

The systematic K/Ar dating of the Pannonian basaltic rocks started in 1978 on the initiative of G. H á m o r, as a joint work of the Hungarian Geological Institute (MAFI), the National Trust for Oil and Gas Industry (OKGT) and the Institute of Nuclear Research of the Hungarian Academy of Sciences (ATOMKI). Sampling, petrographic as well as bio- and lithostratigraphic data are provided by geologists of the MAFI and OKGT, while dating is undertaken in the ATOMKI.

Up to now about 80 determinations were accomplished on about 50 samples (J á m b o r et al., 1980). These samples represent all the stratigraphic levels where basaltic rocks occur in Hungary.

During this lecture the stratigraphic classification shown on Figure 1 will be used. The time interval between the Sarmatian and Pleistocene will be regarded as Pannonian. The Pannonian *s.l.* sequences were formed from 11—12 m.y. to about 2.4 m.y. B. P. and on the basis of geologic considerations and the present radiometric data one of the authors (J á m b o r) has concluded that 5.5 m.y. is a convenient datum for the Lower-Upper Pannonian boundary, which thus coincides with the Miocene-Pliocene boundary. These statements rely on the bio-and lithostratigraphic data of late years (J á m b o r, S o l t i, 1975; B e n c e et al., 1979; J á m - b o r et al., 1981). The basalts in Hungary appear in the Lower Pannonian in one level, in the Upper Pannonian in three levels and in the Pleistocene in one level (Bár).

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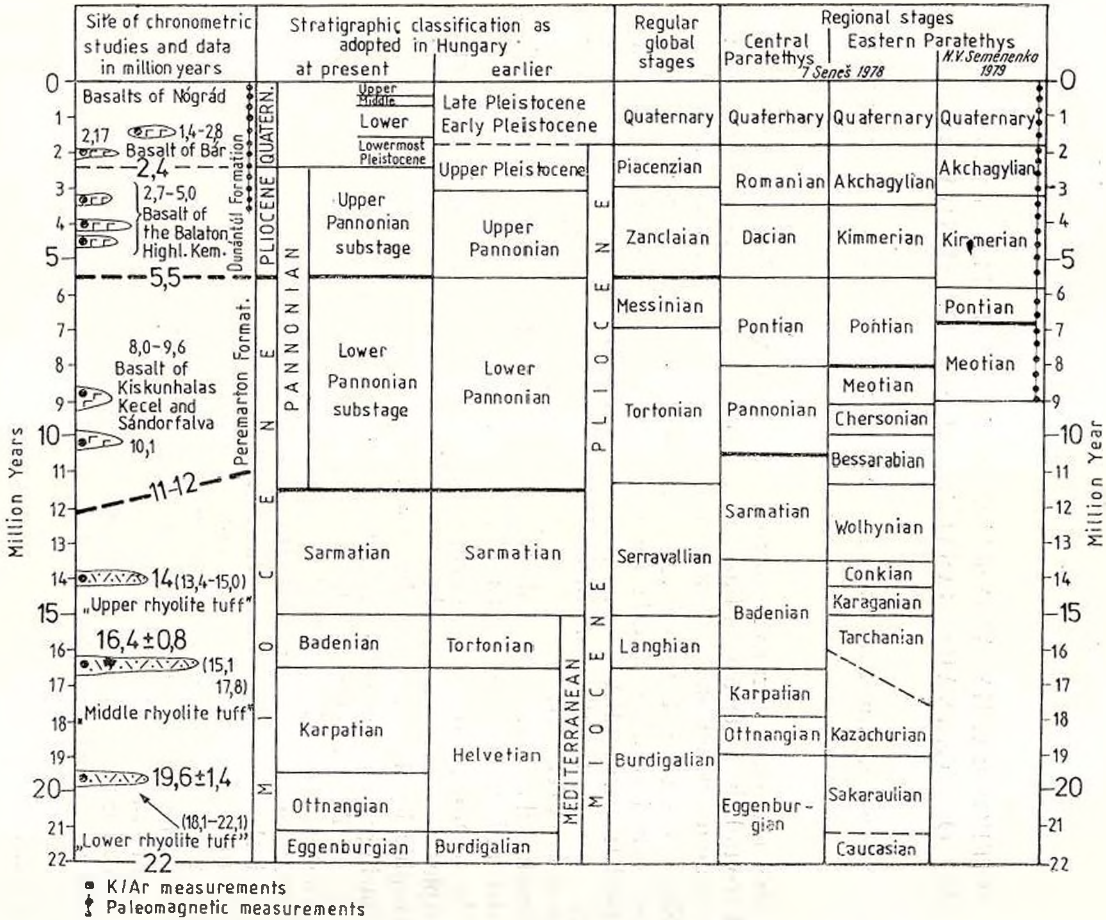


Fig. 1. — Chronostratigraphic position of the Neogene and Quaternary formation, i.e. time units in Hungary (1981) (Compiled on the basis of data from G. Hámor, K. Balogh, B. Cserepes-Meszéna, A. Jámbor, A. Nusszer, L. Ravasz-Baranyai, A. Rónai).

In Hungary the following basaltic occurrences of Pannonian *s.l.* and Pleistocene age are known (Fig. 2) (J u g o v i c s, 1953 a, 1953 b, 1971 a, 1971 b, 1972 ; C s e r e p e s - M e s z é n a, 1978 : K u l c s á r, G u z y - S o m o g y i, 1962 ; M a u r i t z , H a r w o o d, 1936, 1937 ; V i c z i á n, 1965).

1. In the Danube-Tisza interfluvial area near the communities Kiskunhalas, Kecel and Ruzsa basalt pyroclastics and lava rocks were found in the Lower Pannonian (Peremarton Formation) by hydrocarbon exploratory boreholes.

2. In the southwestern part of the Bakony Mts, according to recent deep drillings, the basalts are associated mostly with the middle level and in a smaller part with the lower and upper levels of the Upper Pannonian substage (Transdanubian Formation). The basalts are intercalated or underlain by the Upper Pannonian sediments. The lower, middle and upper levels of the Upper Pannonian substage are characterized by *Congeria ungulacaprae*, *Congeria balatonica* and *Unio wetzleri*, respectively.

3. In the Little Plain the basalts overflow the upper level of the Upper Pannonian and are covered discordantly by fluvial sediments which bio- and lithostratigraphically belong to the Pleistocene.

The basalts in the Bakony Mts and also in the Little Plain are slightly alkalic, two types being distinguished. The first type is of a slightly tholeiitic composition, shows a more explosive character and a $Na_2O > K_2O$ relation is typical for it. The second type is represented by olivine rich varieties, the phenocrysts of which are olivine alone. They show usually a $K_2O > Na_2O$ relation. Within a single volcanic cycle the rocks of type 1 can be regarded as older.

4. A basaltic rock is known southwest of the village Bár, in the southeastern part of Transdanubia. According to the observations of Gy. H ő n i g it overlies the Early Pleistocene red clay, which is in a lithostratigraphically fixed position, and it is covered by Wurmian loess. This rock is an intermediate type between the alkalic basalts and alkalic intermediate rocks and it is classed as jumillite.

5. Northwest of the Mátra Mts, in the northern part of the Nógrád Basin a great number of basalt occurrences, extending also to South Slovakia across the state boundary, are known. These basalts overflow the denuded surface of Oligocene, Ottnangian and Carpathian rocks and they are covered with Wurmian loess. Thus only a poorly defined geological age can be determined by traditional stratigraphic methods. According to their mineralogical and chemical composition these rocks are nepheline basanites.

During the volcanic activity in a number of cases the basaltic lavas did not exchange completely their argon content with the atmosphere ; in this case the initial radiogenic argon content provides a K/Ar age, which is older than the geologic one. This possible deviation may be recognized and usually corrected for by applying the isochron methods. We used the isochron methods for sets of comagmatic whole rock samples and for gravimetrically and magnetically split fractions of a single sample Unfortunately in case of several basalt occurrences the similarity of potassium and argon contents of the collected samples baffled the determination of accurate isochron ages.

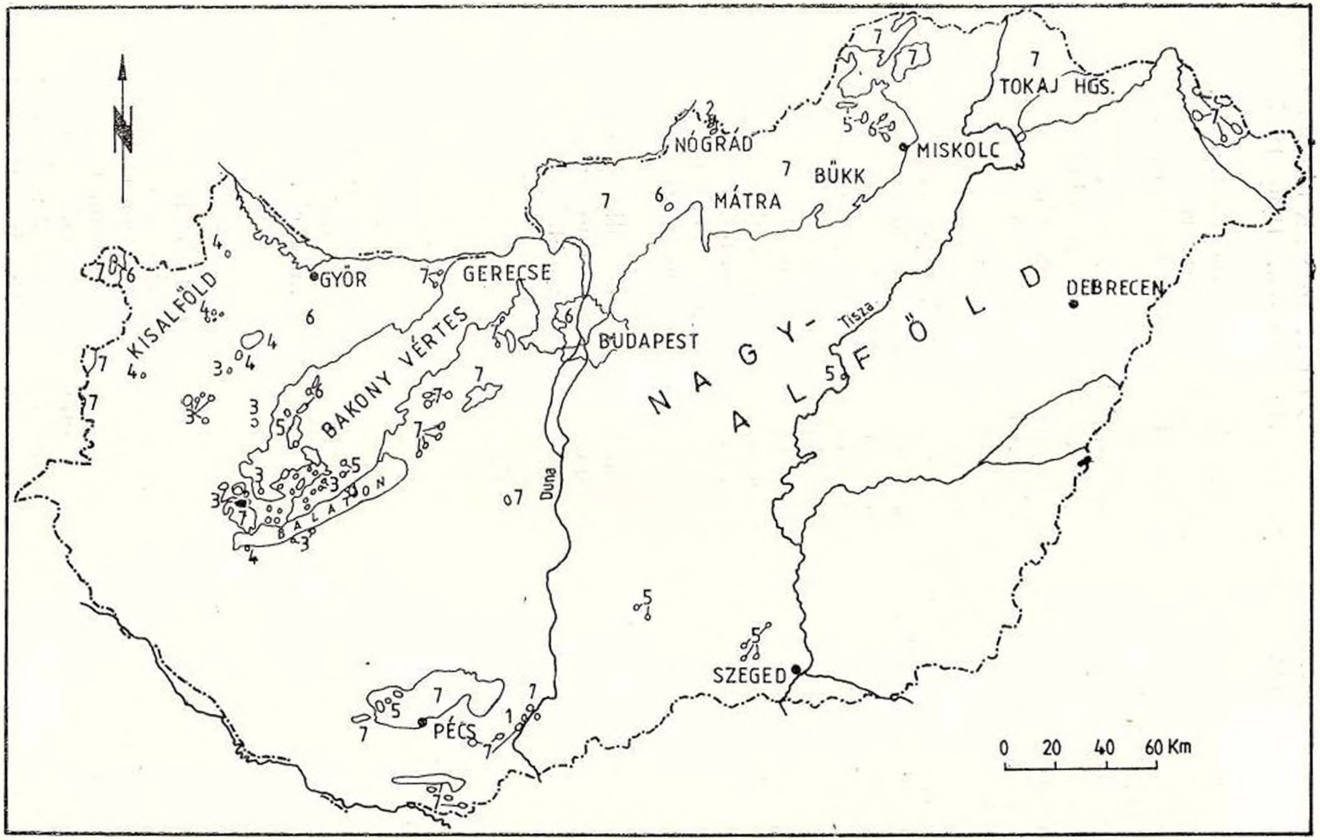


Fig. 2. — Distribution of the Pannonian *s.l.* and Quaternary basalts and basaltic tuffs in Hungary.
1, Quaternary basalt and basaltic tuffs, uncovered; 2, Upper Pannonian-Quaternary basalt and basaltic tuffs, uncovered; 3, Upper Pannonian basalt and basaltic tuffs, uncovered; 4, Upper Pannonian basalt and basaltic tuffs, covered; 5, Lower Pannonian basalt and basaltic tuffs, covered; 6, Pannonian sedimentary rocks; 7, Older sequences, uncovered.

TABLE

K/Ar age of Lower Pannonian basaltic rocks from deep-drillings in the Danube-Tisza interfluvial area

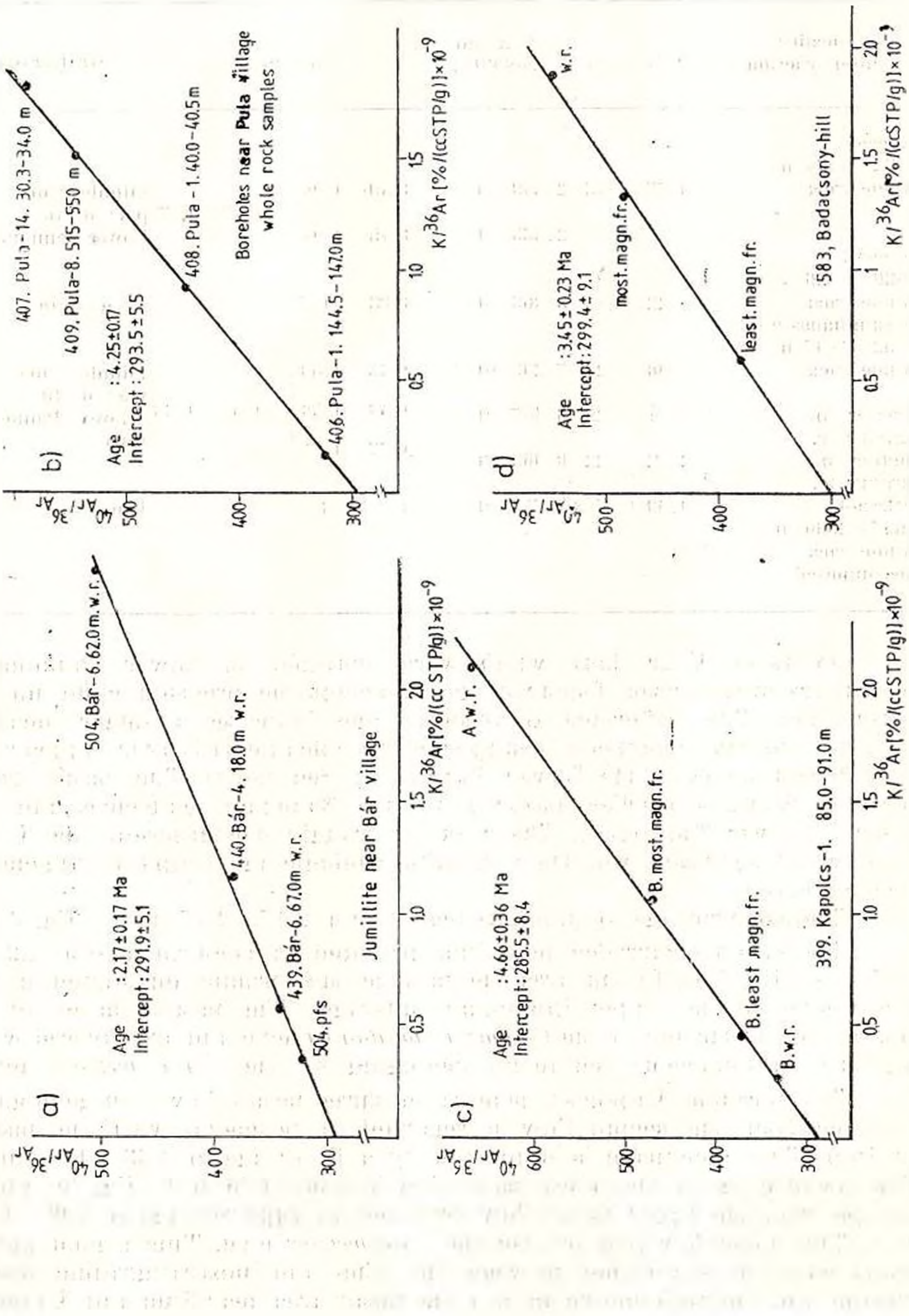
Locality Dated fraction	K %	%	^{40}Ar rad ccSTP/g	Age m.y.		Stratigraphy
Kecel-1 1432–1434 m whole rock	0.77	13	$2.548 . 10^{-7}$	8.50±0.94	8.47±0.77	Middle-Upper part of the Lower Pannonian
			$2.533 . 10^{-7}$	8.45±0.94		
Kecel-2 1426–1426.5 m whole rock	1.22	13	$3.863 . 10^{-7}$	8.13 ±0.71		Uncertain
Kiskunhalas-Ny-3 1162–1167 m whole rock	1.98	24	$7.205 . 10^{-7}$	9.35±0.68	9.61±0.38	Middle-Upper part of the Lower Pannonian
greater sp. gravity fr.	1.71	25	$6.447 . 10^{-7}$	9.68±0.58		
lighter sp. gravity fr.	2.12	22	$8.065 . 10^{-7}$	9.77±0.71		
Ruzsa-4 2657–2666 m whole rock decomposed)	0.68	7.8	$2.753 . 10^{-7}$	10.4±1.8		Uncertain

On table K/Ar data which were obtained on Lower Pannonian basalts are summarized. Isochron ages of acceptable precision could not be determined. The difference of absolute ages indicates a longer volcanic activity, and this conclusion is supported by volcanic tuff layers appearing in different levels in the Lower Pannonian Sediments. The basalt from borehole Ruzsa-4 overlies discordantly the Badenian sequences and its cover is Lower Pannonian. The rock is strongly decomposed, the K/Ar data, which approach well the assumable geologic age, imply a syngenetic decomposition.

The isochron age of jumillite near Bár is 2.17±0.17 m.y. (Fig. 3 a).

The basalt occurrence near Pula provided an isochron age of 4.25±0.17 m.y. (Fig. 3 b). In this area the basaltic tuff production started in the lower level of the Upper Pannonian substage. The basalts dated by us poured out at the end of the *Congeria balatonica* level and are covered with alginitic diatomaceous sediments belonging to the *Unio wetzleri* level.

The borehole Kapolcs-1 penetrated three basalt flows. On geological considerations the second flow is regarded to be coeval with the basalt at Pula. This conclusion is supported by a K/Ar age of 4.33±0.44 m.y. The isochron age of the oldest basalt flow is 4.66±0.36 m.y. (Fig. 3c) and a sample from the upper basalt flow provided an apparent age of 3.93±0.35 m.y. This latter flow is in or over the *Unio wetzleri* level. Thus a good agreement could be established between the litho- and biostratigraphic observations and the radiometric data in the basalt area near Pula and Kapolcs.


a)
Age :2.17±0.17 Ma
Intercept: 291.9±5.1
504.pfs
439. Băr-6. 67.0m w.r.
440. Băr-4. 18.0 m w.r.
504. Băr-6. 62.0 m. w.r.
Jumillite near Băr village
$^{40}Ar/^{36}Ar$
300
400
500
0.5
1.0
1.5
20
$K/^{36}Ar[\%/(cc STP/g)]\times10^{-9}$
b)
407. Pula-14. 30.3-34.0 m
409. Pula-8. 515-550 m
408. Pula-1. 40.0-40.5 m
406. Pula-1. 144.5-147.0 m
Age :4.25±0.17
Intercept: 293.5±5.5
Boreholes near Pula village
whole rock samples
$^{40}Ar/^{36}Ar$
300
400
500
0.5
1.0
15
$K/^{36}Ar[\%/(ccSTP/g)]\times10^{-9}$
c)
Age :4.66±0.36 Ma
Intercept: 285.5±8.4
B.w.r.
B. least magn.fr.
B. most. magn.fr.
A.w.r.
399. Kapolcs-1. 85.0-91.0 m
$^{40}Ar/^{36}Ar$
300
400
500
600
0.5
1.0
1.5
2.0
$K/^{36}Ar[\%/(ccSTP/g)]\times10^{-9}$
d)
Age :3.45±0.23 Ma
Intercept: 299.4±9.1
least magn.fr.
most. magn.fr.
w.r.
583, Badacsony-hill
$^{40}Ar/^{36}Ar$
300
400
500
0.5
1
1.5
2.0
$K/^{36}Ar[\%/(ccSTP/g)]\times10^{-9}$

4

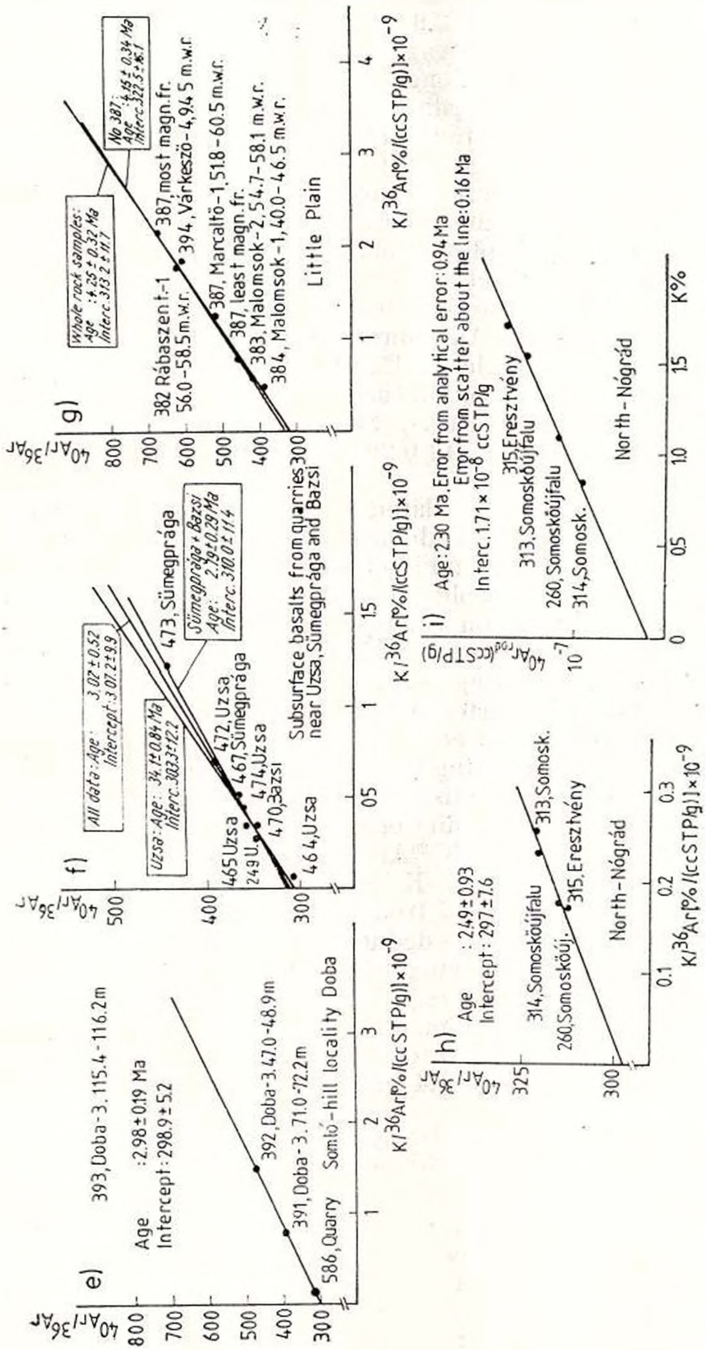


Fig. 3. a, b, c, d, e, f, g, h, i. – Isochron diagrams of K/Ar data.

According to an isochron age of 3.45±0.23 m.y. (Fig. 3 d) the basalt of Badacsony hill has been formed in the *Unio wetzleri* level.

A single sample age of 2.94±0.34 m.y. has been obtained on the basalt of Haláp hill. This young age confronts with stratigraphic and petrographic data (its basement belongs to the *Congeria balatonica* level and it is an olivine rich variety), according to which an older age has been expected.

4 samples have been dated from the Somló hill. 3 points fit a straight line defining an isochron age of 2.98±0.19 m.y. (Fig. 3 e). The fourth sample, coming from the lower part of borehole Doba-3, provided an older apparent age. This sample may contain excess argon or it may as well be the product of the older volcanic activity, since its basement belongs to the *Congeria balatonica* level.

Subvolcanic basalts were dated from Uzsabánya and the quarries of Sümegprága and Bazsi. An approximate value of 3.41±0.84 m.y. has oeen obtained for the samples of Uzsabánya; this is not an isochron age bince the scatter of points around the straight line is greater than it would sollow from the analytical errors. The joint isochron age of rocks from Sümegprága and Bazsi (2.79±0.29 m.y.) is interpreted as the time of basaltic magmatism (Fig. 3 f).

In the Little Plain the dated basalts come from boreholes near Rábaszentandrás, Marcaltő, Malomsok and Várkesző. Out of 7 samples 5 fit a straight line and define an isochron age of 4.25±0.32 m.y. (Fig. 3 g). This age appeared to be too old, since the basalts are in or over the *Unio wetzleri* level. As a control, on a sample from borehole Marcaltő-1 a single sample isochron age has been determined. A similar age of 4.15±0.34 m.y. has been obtained, indicating that, at least in this area, the formation of the *Unio wetzleri* level started before these dates. This is in full contradiction with paleogeographic considerations, since also in the Upper Pannonian the real basin was being in this area and not in the Bakony Mts.

The age of basalt volcanism in the Nógrád Basin has been studied by dating 4 samples in the vicinity of Somoskőújfalú. Due to the chemical composition the $^{40}Ar/^{36}Ar-K/^{36}Ar$ isochron age is unprecise (2.49±0.93 m.y. fig. 3 h). The $^{40}Ar_{rad}-K$ isochron age is 2.3 m.y. (Fig. 3 i). The error of this age, as calculated from the analytical errors, is 0·94 m.y., but it is only 0.16 m.y. when it is deduced from the scatter of points around the fitted line. It is likely that analytical errors were overestimated in this case, and an age of 2–2.5 m.y. is advocated for the basalt volcanism. This is supported by the radiometric study on neighbouring Slovakian territories, where similar K/Ar data have been obtained. In Slovakia, however, basalts as young as 1.4 m.y. have been found too (Balogh et al.).

Summarizing the previously presented data it can be concluded that this K/Ar study contributed to a better age determination of the Pannonian *s.l.* sequences (i.e. sequences between the Sarmatian and Quaternary) in the Carpathian Basin. The radiometric data do not contradict the opinion of paleontologists, who correlate the Sarmatian-Pannonian boundary with the Lower-Upper Bessarabian boundary and estimate this age as 11–12 m.y. The radiometric data do not contradict the supposed coincidence of Lower-Upper Pannonian and Miocene-Pliocene boundary. Owing to the difficulties of collecting suitable samples and the interpretation of radiometric data and to the uncertainties of stratigraphic correlation, the age determination of faunal and lithostratigraphic levels of the Upper Pannonian could not be accomplished up to now.

REFERENCES

Balogh K., Mihaliková A., Vass A. D. (in press) Radiometric dating of basalts in Southern and Central Slovakia. *Geol. Prace.*

Bence G., Jámbor Á., Partényi Z. (1979) Exploration of alginite (oil shale) and bentonite deposits between Várkeszö and Malomsok. *Ann. Rep. of the Hung. Geol. Inst. of 1977*, p. 257–267.

Cserepes-Meszéna B. (1978) On the Lower Pannonian basalts and Proterozoic migmatites uncovered by the hydrocarbon-exploratory borehole Kiskunhalas-Ny-3. *Földt. Közl.* p. 53–64.

Jámbor Á., Partényi Z., Solti G. (in press) Geologic features of basalt volcanites in Transdanubia. *Ann. Rep. of the Hung. Geol. Inst. of 1979* (1981).

— Solti G. (1975) Geological conditions of the Upper Pannonian oil-shale deposit recovered in the Balaton Highland and at Kemeneshát. *Acta Miner. Petr.* 22, p. 9–28. (Szeged).

— Partényi Z., Ravasz-Baranyai L., Solti G., Balogh K. (1980) K/Ar dating of basaltic rocks in Transdanubia. *Hungary ATOMKI Közl.*, 22, p. 173–190.

Jugovics L. (1953 a) Les occurrences de tuf basaltique dans les environs de Tapolca. *Ann. Rep. of the Hung. Geol. Inst. of 1944*, p. 13–25.

— (1953 b) Les occurrences de basalte du Bakony meridional et des environs du lac Balaton. *Ann. Rep. of the Hung. Geol. Inst. of 1953*, p. 65–87.

— (1971) Über den Bau der Basaltgebiete des Balatonhochlandes und des Tapolcaer Beckens. *Ann. Rep. of the Hung. Geol. Inst. of 1968*, p. 223–244.

— (1971) Geologischer Bau der basaltvorkemmen am Kab-Berg und Umgebung/Kabberger Basaltgruppe. *Ann. Rep. of the Hung. Geol. Inst. of 1968*, p. 245–255.

— (1972) Die Basalt- und Basalttuffvorkommen der Kleinen Ungarischen Tiefebene. *Ann. Rep. of the Hung. Geol. Inst. of 1970*, p. 79–101.

Kulcsár L., Guzy-Somogyi A. (1962) Le volcan de Sághegy de Celldömölk. *Acta Geogr. Debrecina*, 9, p. 33–83.

Mauritz B., Harwood H. F. (1937) Die basaltischen Gesteine der Tátikagruppe im Plattenseegebiet (Ungarn). *Min. Petr. Mitt.* 48, p. 373–401, Wien.

— Harwood H. F. (1937) Das basaltische Gestein des Ságberges bei Celldömölk in Ungarn. *Földt. Közl.* 67, p. 241–256.

Viczián J. (1965) Basalt aus dem Komitat Baranya. *Föld. Közl.* 95, p. 448–452.

DES DATATIONS K-AR CONCERNANT SURTOUT LES MAGMATITES SUBSÉQUENTES ALPINES DES MONTS APUSENI [1]

PAR

MARIA LEMNE [2], ELEONORA VÎJDEA [2], MIRCEA BORCOŞ [2], ANCA TĂNĂSESCU [2], OLIVIA ROMANESCU [2]

Les résultats des études géochronologiques des dernières cinq a nnée s'accordent partiellement à l'image (position temporelle des roches érup tives examinées) offerte par les colonnes stratigraphiques testées, en four nisant de nouvelles données sur la présence d'autres étapes (phases) magmatogénétiques alpines sur le territoire des Monts Apuseni (fig.).

Les déterminations faites sur des roches plus anciennes — préalpine (schistes cristallins, granites, pegmatites) appartenant au même aréa géologique-structural sont utilisées comme éléments démonstratifs pour relevant à l'échelle du temps, sans aucun doute et manifestement, les conséquences des effets régionaux de la régénération magmatique, tectono-magmatique ou exclusivement tectonique alpine.

Les déterminations éxécutées par la méthode de la dilution isotopique, dans le laboratoire de l'IGG, sur la roche en totalité, biotite et muscovite (tabl. 1, 2, 3, 4, 5, 6), ont aidé à estimer et reconsidérer l'âge :

1. des produits volcaniques tertiaires des Monts Métallifères :

2. des roches intrusives banatitiques crétacé supérieur-paléocènes du cristallin du Gilău (Monts Apuseni du nord) ;

3. des roches éruptives mésozoïques (?) occurrant comme blocs dans les formations mésozoïques du bassin de l'Ampoi (Monts Métallifères) ;

4. des roches éruptives intrusives mésozoïques de l'aréal à l'activité magmatique du type Vorța-Dealul Gliganul-Dealul Mare (Monts Métallifères — zone méridionale central-occidentale).

On relève aussi les déterminations de l'âge conventionnel de certaines roches métamorphiques, granites et pegmatites du cristallin du Gilău d'âge cambrien et des volcanites permiennes des Monts Codru-Moma.

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[1] Note présentée au 12ème Congrès de l'Association Géologique Carpatho-Balkanique 8—13 septembre 1981, Bucarest, Roumanie.

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1. *Les volcanisme tertiaire*

Les déterminations sur les volcanites badéniennes et post-badéniennes (fig., tabl. 1) placent les produits entre des limites rapprochées de l'intervalle d'âge établi par des informations géologiques directes. Nous envisageons partiellement la phase des andésites quartzifères de Barza et premièrement la phase des andésites sarmatiennes de Săcărîmb

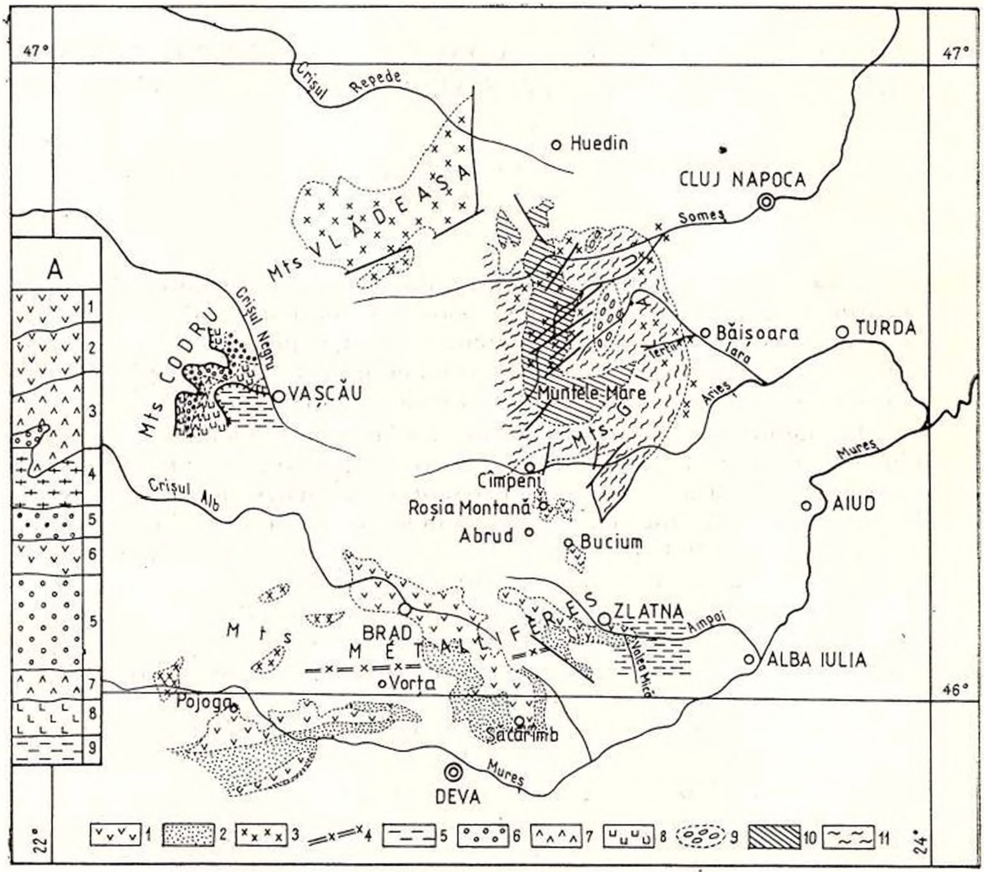

Fig. — Monts Apuseni — la localisation des formations datées par la méthode du K-Ar. 1, volcanites tertiaires ; 2, molasse tertiaire ; 3, banatites ; 4, l'alignement tectonomagmatique Vorţa-Dealul Mare-Vălişoara ; 5, formations sédimentaires mésozoïques ; 6, molasse permienne ; 7, rhyolites permiennes ; 8, basaltes permiens ; 9, aréal avec les occurrences de pegmatites ; 10, granites ; 11, roches métamorphiques. A, le schéma simplifié de l'évolution du volcanisme néogène dans les zones Brad-Săcărîmb et Zlatna : 1, andésites quartzifères du type Cetraş ; 2, andésites-andésites quartzifères du type Săcărîmb ; 3, andésites-andésites quartzifères du type Barza ; 4, formation volcano-sédimentaire andésitique badénien supérieur-sarmatienne ; 5, molasse tertiaire (conglomérats, graviers aux intercalations de marnes, argiles et sables) ; 6, andésites-andésites quartzifères des types Ciinel et Pleşa ; 7, andésites du type Faţa Băii ; 8, rhyolites, rhyodacites-ignimbrites ; 9, formations sédimentaires crétacées.

(~13,5 m.a.) et celle des andésites quartzifères de Cetraş qui semblent atteindre la partie supérieure du Pliocène (5,36 m.a.). Les effets superposés de réactivation tectono-volcanique exercés à l'intérieur des mêmes structures volcaniques complexes y résultées — le régime thermique et la composition des roches étant modifiés aussi par l'altération hydrothermale et la minéralisation — restreignent d'une manière variable et souvent confuse, la valeur des déterminations; les âges isochrones sont acceptés comme valeurs minimums. Un exemple édifiant dans ce sens est offert par les andésites quartzifères de Cîinel et les dacites de Pleşa. L'âge conventionnel (9,4—12,9 m.a.) ne correspond pas à leur position stratigraphique dans le Badénien moyen. Il en résulte aussi que celles-ci sont antérieures aux andésites quartzifères de Barza qui les surmontent et qui sont, à leur tour, intensément réactivées au point de vue magmatotectonique et métallogénétique par la génération d'une série de corps sous-volcaniques, vers la fin de la phase. Les âges conventionnels des andésites quartzifères de Barza, déterminés sur un nombre réduit d'échantillons, varient entre 9,6 et 22,4 m.a. et marquent en général le domaine de leur position stratigraphique — Badénien supérieur-Sarmatien inférieur.

Particulièrement intéressants sont les résultats des déterminations faites pour les produits qui représentent le début de l'activité volcanique tertiaire, c'est-à-dire les rhyolites et les andésites couvertes par l'horizon des conglomérats de Faţa Băii, considérés d'âge badénien inférieur. L'âge isochrone obtenu pour les andésites est d'approximativement 30 m.a. et pour les rhyolites ignimbritiques est de ~ 39 m.a. On remarque ainsi dans les bassins du Techereu et de Pătrînjeni-Zlatna la présence d'une phase volcanique (magmatique) oligocène-éocène et en même temps la nécessité de reconsidérer la position stratigraphique de ce qu'on appèle "l'horizon des conglomérats de Faţa Băii supposé d'âge badénien inférieur". Celui-ci pourrait représenter également un niveau rouge plus ancien de conglomérats éocène-oligocènes.

2. *Les roches intrusives banatitiques et les effets de la magmatogenèse banatitique du Cristallin du Gilău*

La plupart d'échantillons analysés (tabl. 2) provient de la zone de Băişoara, située au contact est du Cristallin du Gilău (fig.); celle-ci contient des roches éruptives très variées (dacites, andésites quartzifères, diorites quartzifères et granodiorites) formées durant une étape courte d'évolution crétacé supérieur-paléocène ou parfois simultanément dû aux processus de différentiation du stade final de consolidation des corps intrusifs. L'âge isochrone de ces roches-là qui englobent aussi deux autres échantillons à position plus éloignée, vers l'intérieur du massif cristallin (les vallées Ierţii et Someşul Rece) indique 64,79±0,97 m.a. Par rapport aux banatites du massif de Vlădeasa, dont l'âge isochrone est de 78,89 m.a. et dont les âges conventionnels varient entre 74,89—41,68 m.a., indiquant une activité magmatique subséquente subhercynienne et laramienne de longue durée — crétacé supérieur-paléocène (R o m a n e s c u et al., 1981), les banatites du Cristallin du Gilău sont plus jeunes — paléocènes — et ont été générées vers la fin de la magmatogenèse laramienne.

L'influence de la magmatogenèse subséquente banatitique et de la tectogenèse laramienne sur le massif cristallin du Gilău a été d'une grande ampleur. Les nombreuses occurrences de roches éruptives (rhyolites, dacites, andésites, diorites), la présence des zones de fracturation régionale y associées, les valeurs très réduites obtenues pour les roches métamorphiques, les granites et les pegmatites de cette unité-là (fig.) attestent sans doute l'existence des processus surtout magmatiques, suggérés d'ailleur par les données géophysiques récentes (A n d r e i et al., 1980). En 1966, en utilisant la méthode du K-Ar — la variante volumétrique on a obtenu dans les laboratoires de l'IGG (L e m n e et al., fide I a n o v i c i et al., 1969, 1976) un âge de 522 m.a. Des investigations ultérieures (M i n z a t u et al., 1967 ; S o r o i u et al., 1969) faites sur la biotite, la muscovite et le feldspath potassique des granites, ont indiqué des âges conventionnels variant entre 70 et 184 m.a. et entre 89 et 232 m.a. respectivement. I a n o v i c i et al. (1976) concluent que "le rajeunissement" alpin des granites est dû aux charriages crétacés bien qu'aux intrusions banatitiques. Les valeurs des déterminations obtenues (tabl. 3) pour les granites de Muntele Mare indiquent des âges qui varient entre 70,5—109,0 m.a., tandis que celles pour les produits des roches métamorphiques de la zone banatitique de Băişoara attestent un âge de 63,1—76,2 m.a. Les mêmes observations sont valables pour les pegmatites de cette unité : les âges conventionnels obtenus sur 11 échantillons de pegmatites varient entre 68—140,9 m.a.

Les déterminations mettent en évidence des effets thermiques de très grande ampleur et au développement régional, liés aux processus magmatiques subséquents alpins. Le rang des valeurs déterminées suggère que ces processus-là pourraient être liés à la phase subséquente banatitique laramienne ou même à celle subhercynienne. Il en résulte aussi qu'on ne peut pas déterminer l'âge initial des granites, pegmatites et des roches métamorphiques par la méthode du K-Ar à cause des pertes partielles de l'argon radiogénique.

3. *Remarques préliminaires sur l'âge des blocs de granites, granodiorites, quartzdiorites et diorites remaniés dans les dépôts sédimentaires mésozoïques du bassin de l'Ampoi (Monts Métallifères) (fig.)*

La présence des blocs de roches éruptives mentionnés ci-dessus, quelquefois à dimensions métriques, reste un problème à résoudre. Jusqu'à présent on n'a pas identifié ni leur source ni leur âge. Les âges conventionnels déterminés sur des roches entières de Valea Mică (Zlatna) varient entre 75,7 et 150,2 m.a. (tabl. 4). Sans qu'on exclut la possibilité d'une magmatogenèse subhercynienne comme source génératrice, il y a aussi la possibilité que cette valeur soit plus grande et qu'elle correspond à une phase magmatique crétacé inférieure. La position actuelle des roches renvoie aux transformations mécaniques et chimiques qui finissent par des pertes en argon radiogénique. La détermination de l'âge par l'angle d'extinction du quartz des blocs de granites indique des valeurs de 130—140 m.a. (M. B ă l a n fide B o r c o ş et al., 1980) attribuant cette phase au début du Néocomien.

4. *Données sur l'âge des roches éruptives du type Vorţa, Dealul Gliganul et Dealul Mare (Monts Métallifères — zone méridionale central-occidentale)*

Il y a certains repères géologiques et géophysiques (B o r c o ș et al., 1979) qui marquent une relation cogénétique entre les types de roches mentionnés ci-dessus. Ainsi, les résultats des déterminations isotopiques obtenus jusqu'à présent pour les andésites quartzifères et les dacites de Dealul Gliganul pourraient être généralement valables pour les quartz-diorites et granodiorites aux indices métallogénétiques des zones de Vorța et Dealul Mare. L'âge isochrone calculé pour les dacites et les andésites quartzifères de Dealul Gliganul (tabl. 5) est de 123,32±1,76 m.a. Ainsi on mentionne, pour la première fois, à base des données géochronologiques (K-Ar), des manifestations magmatogénétiques post-kimmériennes nouvelles dans les Monts Métallifères, appartenant à l'intervalle Valanginien-Hautérivien.

Par corroborant ces observations-là avec l'image structurale pré-néogène des Monts Métallifères, récemment obtenue par la corrélation des données géologiques-géophysiques (B o r c o ș et al., 1979), on arrive, à la conclusion que l'alignement tectonomagmatique orienté EO (fig.) contenant les corps intrusifs détectés par la géophysique à Vorța, Dealul Mare et Vălișoara à l'est, n'appartient pas au système banatitique. Ces données-là indiquent aussi d'autres intervalles de temps qu'il faut accepter et analyser même si à première vue ils sont incompatibles avec les schémas classiques d'évolution magmatogénétique réalisés pour le territoire en discussion.

5. *Informations préliminaires supplémentaires*

On a analysé deux échantillons de rhyolites et un échantillon d spilite (tabl. 6) en vue de vérifier les effets des processus de rajeunissemen- subis par les volcanites permiennes acides et basiques entraînées, à côt de la molasse permienne qui les englobe, dans la tectonique de grande amé pleur des Nappes de Codru. Les âges conventionnels déterminés, relativement rapprochés, varient entre 88,4–105,4 m.a. Celui-ci est un exemple manifeste de régénération alpine des roches éruptives pendant une phase autrichienne, peut-être aux effets subhercyniens, dont les effets mécaniques et les modifications physico-chimiques (fracturation, lamination, métamorphisme de début — récrystallisation) ont changé le contenu initial en argon engeandrant une différence d'environ 154 m.a. Selon leur position dans la colonne stratigraphique, l'âge de ces valeurs-là serait d'environ 250 m.a.

Conclusions

En tenant compte de ce que nous venons de présenter, on envisage les suivants :

— l'existence d'une phase magmatique aux implications métallogénétiques pendant l'intervalle de temps correspondant au Néocomien dans les Monts Métallifères ;

— d'éventuelles implications magmatogénétiques de la phase subhercynienne dans les Monts Métallifères similaires aux situations constatées dans les Monts Apuseni du nord ;

— l'âge éocène-oligocène des rhyolites et des andésites de Fața Băii (considérées badéniennes) ;

TABLEAU 1

Les valeurs de l'âge K-Ar pour les roches alpines et préalpines des Monts Apuseni

No	Echantillon	Localisation	Pétrotype	Matériel analysé	% K	^{40}Ar rad moles ($\times 10^{-10}$)	^{40}Ar rad (%)	L'âge conventionnel (m.a.)	L'âge d'isochrone (m.a.)
Phase de Cetraş (1)									
1	24	Piatra Bulzului	andésite quartzifère	biotite	6,42	1,2550	43,44	$10,9 \pm 0,5$	
2	25	Hărţăgani	andésite quartzifère	biotite	6,07	0,7804	58,95	$7,2 \pm 0,3$	$5,36 \pm 0,81$ $(^{40}Ar/^{36}Ar)_0 = 403,84 \pm 45,76$
3	26	Duba	andésite quartzifère	biotite	5,88	0,7085	31,49	$6,7 \pm 0,3$	
4	27	Zîmbriţa	andésite quartzifère	roche entière	2,38			$11,5 \pm 0,4$	
Phase de Săcărîmb (1)									
5	1	Zukerhut	andésite quartzifère	biotite	6,40	1,0024	49,26	$8,7 \pm 0,4$	
6	2	Zukerhut	andésite quartzifère	biotite	6,46	0,6983	31,01	$6,5 \pm 0,3$	
7	3	Zukerhut	andésite quartzifère	biotite	6,34	0,6755	36,89	$6,0 \pm 0,3$	
8	4	Zukerhut	andésite quartzifère	biotite	6,35	0,8503	40,21	$7,4 \pm 0,3$	$13,57 \pm 1,74$ $(^{40}Ar/^{36}Ar)_0 = 121,98 \pm 40,88$
9	5	Zukerhut	andésite quartzifère	biotite	6,31	0,7418	39,88	$6,6 \pm 0,3$	
10	6	Zukerhut	andésite quartzifère	biotite	6,24	0,1129	5,13	$1,0 \pm 0,2$	
11	7	Zukerhut	andésite quartzifère	biotite	6,35	0,4200	19,02	$3,7 \pm 0,3$	
12	8	Zukerhut	andésite quartzifère	biotite	6,32	0,6074	26,15	$5,4 \pm 0,3$	
13	9	Zukerhut	andésite quartzifère	biotite	6,53	0,3392	15,43	$2,9 \pm 0,2$	

14	1	Goruniștea	andésite quartzifère	biotite	6,46	1,4034	70,36	12,1 ± 0,5	
15	2	Goruniștea	andésite quartzifère	biotite	6,62	0,4183	15,73	3,5 ± 0,3	
16	3	Goruniștea	andésite quartzifère	biotite	6,69	1,4802	69,49	12,4 ± 0,5	
17	4	Goruniștea	andésite quartzifère	biotite	6,96	1,0565	36,39	8,5 ± 0,4	13,63 ± 1,26 $(^{40}Ar/^{36}Ar)_0$ = 208,61 ± 53,65
18	6	Goruniștea	andésite quartzifère	biotite	7,03	0,8394	16,22	6,7 ± 0,6	
19	7	Goruniștea	andésite quartzifère	biotite	6,84	1,5957	73,48	13,0 ± 0,5	
20	8	Goruniștea	andésite quartzifère	biotite	6,88	1,2527	68,34	10,1 ± 0,4	
21	2	Sarcău	andésite quartzifère	biotite	6,22	1,3380	52,77	12,0 ± 0,5	
22	4	Sarcău	andésite quartzifère	biotite	6,38	1,2265	53,69	10,8 ± 0,4	
23	5	Sarcău	andésite quartzifère	biotite	6,19	1,3630	56,51	12,3 ± 0,5	13,61 ± 0,96 $(^{40}Ar/^{36}Ar)_0$ = 253,63 ± 46,81
24	6	Sarcău	andésite quartzifère	biotite	6,22	1,5339	72,42	13,7 ± 0,5	
25	7	Sarcău	andésite quartzifère	biotite	6,31	1,4344	78,31	12,7 ± 0,5	
26	16	Calvaria Mare	andésite quartzifère	biotite	5,71	1,3578	11,00	13,0 ± 1,6	
Phase de Barza (1)									
27	83	Dealul Neagra	andésite quartzifère	roche entière	2,09	0,3790	45,21	10,1 ± 0,4	
28	85	Gal. Podul Ionului	andésite quartzifère	roche entière	1,84	0,5815	67,38	17,6 ± 0,8	
29	70	Vîrful Jidovului	andésite quartzifère	roche entière	0,76	0,3045	33,53	22,4 ± 0,9	
30		Brad sanatorium	andésite quartzifère	biotite	6,52	1,1196	67,87	9,6 ± 0,4	

(Continuation tableau 1)

No	Echantillon	Localisation	Pétrotype	Matériel analysé	% K	^{40}Ar rad moles ($x10^{-10}$)	^{40}Ar rad (%)	L'âge conventionnel (m.a.)	L'âge d'isochrone (m.a.)
Phase de Ciinel-Pleșa (1)									
31		Ciinel	andésite quartzifère	biotite	6,35	1,4693	51,44	12,9 $\pm$ 0,5	
32	4	Pleșa	dacite	biotite	6,02	1,0108	48,65	9,4 $\pm$ 0,4	
33	6	Pleșa	dacite	biotite	6,42	1,4682	28,35	12,7 $\pm$ 0,7	
Phase de Fața Băii (2)									
34	84	Dealul Teiuș	andésite	roche entière	0,49	0,4408	48,15	51,2 $\pm$ 2,1	
35	115	Vallée Bizaci	andésite	roche entière	2,10	1,1877	81,77	32,3 $\pm$ 1,2	29.89 $\pm$ 0,8 $^{40}Ar/^{36}Ar$ = 395,16 $\pm$ 20,99
36	113	Almașul Mare	andésite	roche entière	2,04	1,6351	54,93	45,6 $\pm$ 1,8	
37	91	Pojoga	rhyolite	roche entière	4,13	2,4145	38,29	33,4 $\pm$ 1,5	
38	111	Pătrinjeni	rhyolite	roche e tière	2,80	1,9361	87,65	39,5 $\pm$ 1,5	
39	110	Pătrinjeni	rhyolite	roche entière	3,58	3,2510	68,78	49,9 $\pm$ 1,9	38,85 $\pm$ 2,03 $(^{40}Ar/^{36}Ar)_0$ = 351,10 $\pm$ 54,11
40	50	Pătrinjeni	rhyolite	roche entière	2,66	1,9738	66,33	42,4 $\pm$ 1,5	
41	49	Pătrinjeni	rhyolite	roche entière	2,27	1,7193	74,54	43,1 $\pm$ 1,6	

TABLEAU 2 (2)

1	2	3	4	5	6	7	8	9	10
1	49	Mașca	andésite	biotite	6,67	7,0763	85,65	58,7±2,2	
2	46	Someșul Rece	granodiorite	roche entière	3,00	2,6849	64,50	49,5±1,9	
3		Băișoara	granodiorite	biotite	6,45	8,4366	12,57	71,0±7,8	64,79±0,97 ($^{40}Ar/^{36}Ar)_0$ = 197,33±39,31
4	F94	Mașca	granodiorite	biotite	6,90	8,0166	95,84	64,1±2,4	
5		Cacova	granodiorite	biotite	6,81	7,5648	81,97	61,4±2,3	
6	F83	Mașca	dacite	roche entière	2,45	2,8166	56,87	63,4±2,4	
7	F46	Mașca	andésite quartzifère	roche entière	2,15	2,1283	65,14	54,8±2,1	
8		Valea Ierții	granodiorite	biotite	6,80	6,5591	79,15	53,4±2,0	

TABLEAU 3
Granites (1)

1	2	3	4	5	6	7	8	9	10
1	95	Călățele	granite gneissique	muscovite	8,50	12,7640	92,02	82,4±3,1	
2	90	Călățele	granite	muscovite	8,46	8,9873	79,22	58,7±2,2	
3	93	Călățele	granite	muscovite	8,80	17,5692	84,97	109,0±4,1	
4	89	Călățele	granite gneissique	muscovite	8,60	11,0043	90,11	70,5±2,6	
5	91	Călățele	granite gneissique	biotite	6,68	10,1225	87,30	83,1±3,1	
6	102	Vallée Someșul Cald	granite	muscovite	8,20	9,5408	81,95	64,0±2,4	
7	77	Vallée Someșul Cald	granite	biotite	6,21	9,7773	89,25	86,5±3,2	
8	41	Vallée Someșul Rece	granite	biotite	6,68	8,0289	83,19	82,3±3,1	
9	51	Vallée Someșul Rece	granite	biotite	6,57	10,6334	79,59	88,7±3,3	
10	45	Cime Muntele Mare SE	granite gneissique	biotite	7,44	10,0874	85,66	74,5±2,8	

Pegmatites (1)

11	8	Vallée Someşul Rece	pegmatite	muscovite	9,09	11,3924	79,72	69,0±2,6	
12	25	Muntele Rece	pegmatite	muscovite	8,42	17,0167	94,51	110,0±4,1	
13	15	Muntele Rece	pegmatite	muscovite	7,87	14,3941	72,51	100,0±3,8	
14	18	Muntele Rece	pegmatite	muscovite	8,63	10,8770	72,46	69,5±2,6	
15	19	Muntele Rece	pegmatite	muscovite	7,80	10,1412	82,82	71,6±2,7	
16	109	Gal. Crişani	pegmatite	muscovite	8,48	12,5461	70,00	81,3±8,0	
17	144	Vallée Tîrlova	pegmatite	muscovite	8,49	21,2792	95,00	135,8±9,0	
18	134	Valea Calului	pegmatite	muscovite	8,26	21,5229	68,00	140,9±13,0	
19	135	Valea Calului	pegmatite	muscovite	8,47	20,5704	70,10	131,7±11,0	
20	128	Vallée Ursu	pegmatite	muscovite	8,35	13,7695	68,20	90,4±8,1	
21	238	Vallée Hugeana	pegmatite	muscovite	8,38	11,4174	73,10	75,0±6,0	

Cristallin (1)

22		Valea Ierţii A	micaschiste	biotite	7,04	8,3774	92,41	65,7±2,4	
23		Vallée Iara	schiste séricitеux	roche entière	2,09	2,8937	80,63	76,2±2,8	
24		Valea Micii	micaschiste	roche entière	4,91	5,6121	77,51	63,1±2,4	

TABLEAU 4 (2)

1	2	3	4	5	6	7	8	9	10
1	106	Valea Mică (Zlatna)	granodiorite	roche entière	1,49	2,7514	50,20	103,5±5,0	
2	107	Valea Mică (Zlatna)	granodiorite	roche entière	0,86	2,0520	59,58	132,7±5,7	
3	108	Valea Mică (Zlatna)	granodiorite	roche entière	0,40	0,9364	46,89	130,3±6,6	
4	77	Valea Mică (Zlatna)	granite	roche entière	0,23	0,3082	34,81	75,7±5,0	
5	75	Valea Mică (Zlatna)	granite	roche entière	0,25	0,6787	56,11	150,2±6,0	

TABLEAU 5 (1)

1	2	3	4	5	6	7	8	9	10
1	138	Vallée Bărăşti	dacite	biotite	6,37	13,6601	73,11	116,7 ± 4,3	
2	137	Vallée Bărăşti	dacite	biotite	6,65	16,6712	75,10	135,8 ± 4,2	
3	136	Vallée Bărăşti	dacite	biotite	6,58	14,0210	77,36	116,0 ± 4,3	
4	135	Vallée Bărăşti	dacite	biotite	6,61	13,1883	65,07	108,9 ± 4,2	123,32 ± 1,76
5	133	Dealul Gliganul	andésite quartzifère	roche entière	2,11	4,8212	77,88	124,2 ± 4,6	$(^{40}Ar/^{36}Ar)_0$ = 284,82 ± 54,83
6	105	Dealul Gliganul	andésite quartzifère	roche entière	4,01	9,2520	94,27	125,3 ± 4,5	
7	132	Vallée Bărăşti	andésite	roche entière	2,55	5,7284	96,42	122,1 ± 4,5	

TABLEAU 6 (1)

1	2	3	4	5	6	7	8	9	10
1	94	Vaşcău	rhyolites	roche entière	5,74	10,3326	95,62	98,5 ± 3,7	
2	96	Apuseni du nord	rhyolites	roche entière	6,31	10,1658	91,26	88,4 ± 3,5	
3	97	Vallée Tărcăiţa Codru Moma	spilites	roche entière	0,97	1,8812	80,24	105,4 ± 4,1	

$\lambda = 5,305 \times 10^{-10}$ an^{-1}

$\lambda_{\varepsilon} = 0,585 \times 10^{-10}$ an^{-1}

$\lambda_{\beta} = 4,72 \times 10^{-10}$ an^{-1}

$K^{40}/K = 1,19 \times 10^{-4}$ moles/moles

(2) $\lambda = 5,543\ 10^{-10}$ an^{-1}

$\lambda_{\varepsilon} = 0,581\ 10^{-10}$ an^{-1}

$\lambda_{\beta} = 4,962\ 10^{-10}$ an^{-1}

$K^{40}/K = 1,167 \times 10^{-4}$ moles/moles

$^{40}Ar/^{36}Ar = 295,5$

± représente la détermination analytique exacte au niveau de confidence de 68%

— les effets régionaux tectoniques et magmatiques de rajeunissement alpin des roches préalpines ; les âges obtenus se situent au niveau de principales phases de tectogenèse et magmatogenèse subhercynienne et laramienne.

Nous remarquons aussi que les âges conventionnels pour les roches qui ont subi des processus tectonomagmatiques et métallogénétiques, diminuent graduellement. La régénération (rajeunissement) est conditionnée aussi par l'intensité des phénomènes. Dans ces cas-là, l'effet final rend impossible l'obtention des âges initiaux des roches préalpines par la méthode isotopique du K-Ar et mène à prendre les précautions nécessaires pour interpréter les valeurs ainsi obtenues pour les roches alpines.

BIBLIOGRAPHIE

Andrei J., Bălan M. (1978) Rapport, archives de l'Institut de géologie et géophysique. Bucarest.

— Calotă C., Cristescu T., Roth M., Ciucur E. (1979) Rapport, archives de l'Institut de géologie et géophysique, Bucarest.

Borcoş M., Andrei J., Lupu M., Berbeleac I., Cristescu T. (1979) Rapport, archives de l'Institut de géologie et géophysique, Bucarest.

— Andrei J., Lupu M., Cristescu T., Berbeleac I., Ştefan A., Lazăr C. (1980) Rapport, archives de l'Institut de géologie et géophysique, Bucarest.

Cioflica G., Istrate G., Popescu G., Udubaşa G. (1966) Contribuţii la cunoaşterea vîrstei produselor vulcanice din regiunea Hărţăgani-Trestia (Munţii Metaliferi). *Stud. cerc. geol. geofiz., geogr., seria geologie*, 11, 1, Bucureşti.

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

— Borcoş M., Bleahu M., Patrulius D., Lupu M., Dimitrescu R., Savu H. (1976) Geologia Munţilor Apuseni. Ed. Acad. RSR, Bucureşti.

Mînzatu S., Lemne M., Tănăsescu A., Ioncică M., Vîjdea E., Colios E. (1967) Rapport, archives de l'Institut de géologie et géophysique, Bucarest.

Romanescu O., Lemne M., Vîjdea E., Mînzatu S., Iosipenco N. (1981) Rapport, archives de l'Institut de géologie et géophysique, Bucarest.

Soroiu M., Popescu G., Kasper V., Dimitrescu R. (1969) Contributions préliminaires à la géochronologie des massifs cristallins des Monts Apuseni. *An. Şt. Univ „Al. I. Cuza", sect. IIb (geol.)*, Iaşi.

SOME VIEWS ON THE ISOTOPIC AGE PATTERN (K/AR; RB/SR; U/PB) OF THE SOUTHERN CARPATHIANS METAMORPHIC AND MAGMATIC COMPLEXES [1]

BY

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The Southern Carpathians display a very complex structural framework, with differences in individual zones, dominated by the Getic Nappe. This large scale structural zone of the Romanian territory consists dominantly of Precambrian and Low Paleozoic metamorphic rocks associated with many granitoid and ultrabasic rocks intrusions. A crustal thickness of about 50 km results from the axial zone gravity data, by comparison with the Transylvanian Depression — about 35 km thick — and the Moesian Platform — about 30-35 km thick. The most significant geological element of the Southern Carpathians is expressed by the coincidence of different structural elements (faults, folds, cleavages and foliations) and by the main trend of this range regardless of the complexes age (either old or new). Also the granitoid intrusions (Assyntic, Caledonian) exibit in most cases an elongated form paralleling the main range trend.The Carpathians main trend is strictly followed by the main fold axes. It is worth noting the presence of the syncline infilled with the obvious schistous Tulișa Series. Regarding the above statements one must concede that old structures were affected by a radical readjustment and a basic rejuvenation during the Alpine movements. Some tracks of an oceanic crust zone are concealed under the marginal basin (Getic Depression) in the transition zone, between the Moesian Platform and the Danubian Autochthon. The presence of an oceanic crust zone, which could have possibly disappeared by subduction, is suggested by the great structural differences of the Moesian Platform and the Danubian basement, proving an independent geological evolution. The transition from an oceanic zone existing between the Danubian Authochthon and the Getic Domain is attested by the presence of the Severin Nappe deposits (flysch, Wildflysch and diabases). The

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oceanic crust zones disappeared following the subduction processes, though sometimes the ophiolites slices overthrust the continental basement. The compressive forces accounted for the development of the Getic Nappe. In order to point out these parallel structures and textures one must concede that, beside block rotation movements, there are also tectonic flows during the Alpine movements. In between the intensive flow zones there are sections where old structures are mostly preserved. It is the case of some isometric contours and massive texture granitoid bodies or the case of some metamorphic rocks whose absolute age is at least synchronous with the age of the piercing granitoid bodies. The Alpine tectonic movements in the crystalline basement are proved by the results of radiometric measurements. These measurements illustrated the pre-Assyntic, Assyntic or Caledonian age rejuvenation, which took place exclusively during the Alpine orogeny.

The Danubian Autochthon

From the point of view of the lithostratigraphic, metamorphic and tectonic regimes, three metamorphic series were separated in this tectonic unit : the Drăgșan Series including the two complexes (the amphibolites and the sericite-chlorite complex) ; the Lainici-Păiuș Series and the Tulișa Series. One considers that the first two lithological series formed during the Upper Precambrian B ($850\pm50-375$ m.y.) i.e. during the Assyntic-Baikalian orogeny ; they formed in the metamorphic regime of the almandine-amphibolite facies, albite-epidote-amphibolitic facies respectively.

Within the Drăgșan Series amphibolitic complex one notices massive or banded amphibolites, garnet and epidote amphibolites, biotite or garnet amphibolitic schists, amphibole gneisses, marble-like limestones with many serpentinite concordant intercalations, metadiorites and metagabbros. The granitization of these complexes occurs as the granitoid rocks cut them across as veins. The sericite-chlorite complex is made up of sericite-chlorite schists, quartzites, chlorite gneisses and epidote-albite-actinote-chlorite-schists. The structures of these rocks point out epiclastic relicts — occurring in the lowest horizons of the complex — and fine pellitic pyroclastic relicts interbedded with epiclastic rocks — occurring in the upper horizons of the complex. The rocks belonging to the two complexes are obviously overprinted by retromorphism.

The Lainici-Păiuș Series is made up of metamorphic rocks ; their structures, textures and mineral assemblage point out a genesis from epiclastic deposits in the amphibolite facies metamorphic regime. By migmatization and granitization the micaceous schists, the quartzites and the granitic gneisses enrich in feldspar and gradually pass to grano-gneisses. Within these metamorphic rocks one notices many apophyses of the granitoid bodies and aplite-pegmatite intercalations. The Lower Paleozoic Tulișa Series is made up in the lower part of metaconglomerates followed by quartzites, crystalline limestones, tuffogenous green rocks with cipoline limestone intercalations and various phyllites (sericite, graphite, etc types). Within the first two metamorphic series one notices many synkinematic, postkinematic granitoid bodies of Baikalian-Caledonian age.

Figure 1 illustrates the isotopic ages of the granitoid rocks intruding the metamorphic basement made up of the Drăgșan and Lainici-Păiuș metamorphics (Fig. 1).

These range — according to the K-Ar method of dating — between 656 m.y. (Assyntic) and 104 m.y. (Neokimmerian). Within the bodies one notices that the Petreanu granitoids (656 m.y.) are the oldest, followed in chronological order by the Tismana granites (580 and 110 m.y.), the Șușița type (524—256 m.y.) the Sfîrdinu granite (372—148 m.y.), the Ogradena granite (335—186 m.y.), the Cherbelezu granite (321—150 m.y.), the

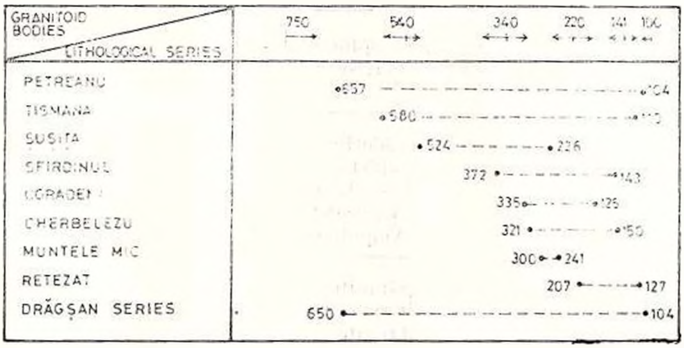

Fig. 1. — The isotopic age evolution in some granitoid bodies and crystalline schists of the Danubian Domain.

Retezat (207—137 m.y.) and the Muntele Mic granites (300—241 m.y.). From the geostructural elements and the mutual relations point of view the Tismana type porphyroid granite is thought to be much younger than the intruded Șușița granite. If the Tismana granite type is related to the Baikalian orogeny, the Șușița type must be related to the Assyntic one. The 524—256 m.y. ages are due to the thermal effects undergone by rocks during both the Caledonian and the Hercynian orogenies. The other bodies, namely Sfîrdinu, Cherbelezu and Retezat are all overprinted by Hercynian orogenies. The Ogradena granitoid also presents Old Kimmerian isotopic ages, and the Cherbelezu and the Retezat granitoids present Neokimmerian ages. The Petreanu granitoid body shows a great range of ages —656—104 m.y. (from Assyntic to Kimmerian) — and the Pietrii Peak body presents a range between 280—94 m.y. (from the Hercynian to the Austrian phase). Figure 1 illustrates the long time activity of the Petreanu and Tismana bodies, the more restricted one of the Șușița body (between Caledonian and Hercynian) and the most restricted one of the Retezat body (Old-Neokimmerian).

The presentation of Figure 1 is not at its best because only 1o determinations were effected for the Petreanu body, 25 for the Tismana body, 26 for the Șușița body and merely 3 for the Retezat body. Therefore it is reasonable to hold that all granitic bodies were strongly affected by the Alpine orogeny. Concerning the Danubian Autochthon metamorphic rocks (Tab.) one notices that only a few of them were subject to absolute age determinations — the Drăgșan Series respectively, especially in the Pe-

TABLE

Types of isotopically analysed rocks

Unit or eruptive body	Analysed Sample	Sample location	Rock	The isotopic geological age K/Ar	Rb/Sr	U/Pb Pb/Pb
The Danubian Domain						
Mesometamorphic crystalline	29	Eşelniţa, Mraconia, Iardaşiţa Valley	Amphibolite	103–305		
			Gneiss	104–656		
Epimetamorphic crystalline	30	Iardaşiţa, Mraconia, Sfirdinu Valley	Chlorite-sericite schists	89–306	266	
			Quartz-albite-chlorite schists	123–128		
			Amphibole schists	103–448		
Synorogen magmatism	112	Muntele Mic, Retezat, Sfirdinu, Cherbelezu, Ogradena, Şuşiţa, Olteţ Mts	Granites	150–580	256–393	
			Granodiorites	94–370		
			Diorite	127–411		
			Adamellite	241		
			Porphyroids	148–162		
			Pegmatite	362–400		
			Lamprophyres	181–229		
			Kerssantite-Spessartite	153–183		
			Diabases	160–239		
Postorogen magmatism	53	Summit Pietrii Şucu Summit, Tismana, Novaci (Paring) Lotru, Vilcan	Granites	110–518	140	605
			Granodiorites	223–537		
			Diorites	94		
			Metadiorites	155–183		
			Pegmatites	115–207	140	
The Getic Domain						
Crystalline schists	169	Semenic, Sebeş, Lotru, Godeanu, Cibin, Făgăraş, Leaota, Iezer-Păpuşa, Poiana Ruscă Mts	Gneiss	124–425		1,000–1,100
			Paragneiss	64–426	838	
			Micaschists	149–321		
			Amphibolites	168–330		
			Pegmatite	115–273		
			Migmatite	154–313		
			Chlorite schists	150–370		
			Quartz schists	102–317		
Magmatism		Sicheviţa, Poneasca, Albeşti, Timiş, Criva, Bîrsa, Ciclova-Oraviţa	Granite	302–475	116–185	328–620
			Metabasalt porphyry		66–68	
			Granodiorite		67–72	

treanu-Retezat and Almaj areas. The oldest rocks are the Poiana Mărului-Caransebeș and the Almaj Mts amphibolites (448 and 388 m.y.); these ages correspond to the Caledonian orogeny; the youngest rocks would be the chlorite-sericite schists (102 m.y.) and their ages would correspond to the Neokimmerian orogeny. It is worth noting that the Drăgșan Series metamorphic rocks underwent intensive changes during the Caledonian, Hercynian and Kimmerian orogenies.

The data in Figure 1 and Table point out that within the Danubian Autochthon geological complexes (granitoid bodies, the Drăgșan Series metamorphic rocks) the last K/Ar ratio changes took place during the Alpine orogeny (Neokimmerian).

The Getic Nappe Metamorphic Rocks

On the maps of the Geological Institute the Getic Nappe metamorphic rocks (Tab.) of the Sebeș-Lotru Series are included in the Upper Precambrian A (1600—850 m.y.) and the epimetamorphic series is included in the Lower Paleozoic (320 m.y.). The Sebeș-Lotru metamorphic rocks consist of various gneisses and micaschist types interbedded with quartzites, amphibolites, pegmatites, aplites, migmatites, serpentinites, eclogite and granitoid rocks. The epimetamorphic series is made up of sericite-chlorite schists interbedded with marble-like limestones, quartz-schists, amphibolites and others. Within the Sebeș-Lotru Series metamorphic rocks the K-Ar gneiss age determination reveals values ranging between 426—150 m.y., compatible with the Caledonian and respectively Alpine Old Kimmerian orogenies. The Rb/Sr absolute age determination would be equivalent to 838 m.y., that is the Dalslandian. An U/Pb analysis of the Bota Valley gneiss type zircon crystals indicated an age ranging between 1,000—1,100 m.y., thus ascertaining the Dalslandian age of these schists. The 206 m.y. age indicates that they were affected by different orogenic phases, undergoing various K-Ar ratios changes. The last phase took place during the Old Kimmerian orogeny. Regarding the epimetamorphic series schists, the oldest ones are notably the Muntele Mic-Poiana Ruscă chlorite sericite schists (370 m.y. age), belonging to one of the Caledonian orogenic phases. The youngest ones would be some chlorite-sericite schists, occurring on the Mănăstirii Valley and at Iazuri-Poiana Ruscă (135 m.y. age), belonging to the Alpine orogeny (Neokimmerian). One should mention that some metallic sulphide mineralizations included in the chlorite-sericite schists, exhibit a Rb/Sr determined absolute age ranging from 780 m.y. to 280 m.y. (the Teliuc galena, the Muntele Mic galena respectively). The Boul Summit iron ores associated with the banatitic metasomatosis indicated, by U/Pb-Pb/Pb determination, an age ranging between 185—152 m.y. The granitoid bodies dispersed among the Iezer-Păpușa and Leaota metamorphic rocks exhibit a discrepancy. Concerning this assumption one might remark that the Albești granite determined by the K/Ar method, indicates an age of 477—302 m.y. (that is Caledonian-Hercynian), while the intruded rocks are 352—175 m.y. old (the Iezer amphibole schists and the Bădeanca Valley paragneisses respectively), therefore belonging to the Hercynian, Alpine orogeny respectively (Old Kimmerian). The granitoid bodies intruding the Sebeş-Lotru Series metamorphic rocks (Criva-Poiana Ruscă-Slatina, Sicheviţa, Poneasca) show various ages. Thus the K/Ar determination of the Criva body revealed a 329 m.y. age, and the U/Pb and Pb/Pb determinations of the Poneasca and Sicheviţa bodies revealed an age of 620 m.y. and 350—328 m.y. respectively, evidence for the Assyntic and Baikalian orogenic phases. The Rb/Sr determination of the Bîrsa-Fierului Valley, Morişoara, Răchiţii granite bodies revealed an age of 185—116 m.y.—evidence for the Alpine orogeny (Kimmerian).

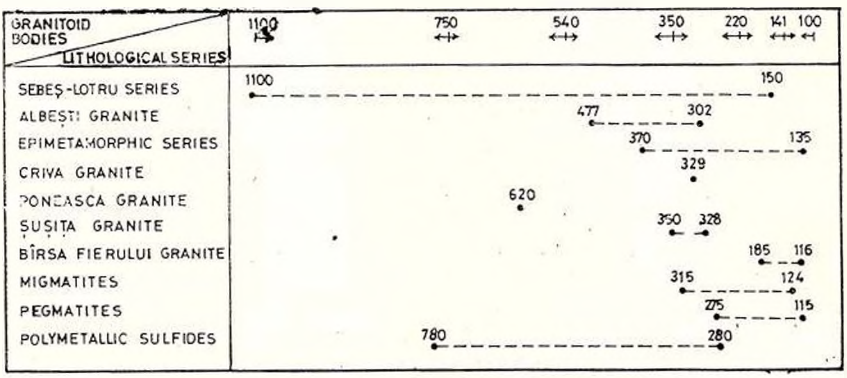

Fig. 2. — The isotopic age evolution in some geological formations of the Getic Domain.

The different magmatic rocks significant for the Sebeş-Lotru Series showed the following ages: 346—124 m.y. (the Cozia ocellar gneiss); 313—285 m.y. (the Brebu migmatites); 273—115 m.y. (the Lotru Valley pegmatites): 235 m.y. (the Căpîlna-Sebeş ocellar gneisses); 135 m.y. the Dobreiaşu Valley pegmatoids (Fig. 2).

Figures 1 and 2 obviously illustrate that the Sebeş-Lotru Series metamorphic rocks are the oldest — exhibiting a very long development (from the Dalslandian to the Alpine-Old Kimmerian), while the Drăgşan Series developed since the Assyntic till the Alpine (Neokimmerian). Regarding the granitoid bodies, the Petreanu body mostly overlaps the Drăgşan Series metamorphic rocks — in both emplacement and development. A slight drift value is noticed for the Tismana body, estimated as younger than the Şuşiţa body. The latter developed along 260 m.y. compared to the Tismana and Petreanu bodies which developed along 470 m.y. and 550 m.y. respectively. The Ogradena and Cherbelezu bodies recorded an almost parallel development (470 m.y. and 170 m.y. respectively). The Muntele Mic and Retezat bodies recorded shorter developments (50 and 80 m.y. respectively). The isotopic dating reveals that the internal and external bodies had an early emplacement and a long evolution (Petreanu in the northern part and Tismana in the southern part) and the central bodies were emplaced later and generally had shorter evolutions. In the Getic Domain, the Sebeş-Lotru metamorphic rocks had a 900 m.y. evolution and the epimetamorphic series metamorphic rocks had a 200 m.y.

evolution. The migmatization and pegmatization processes developed along 120 m.y. and 150 m.y. respectively. A subject still open for discussion is that of the metamorphic mineralizations of the epimetamorphic series — especially the polymetallic sulphides-emplaced 400 m.y. before their host rock and developing along a period of 500 m.y. — a longer period of evolution than that of the host rocks. The discrepancy is mainly due to different methodologies. The age of the epimetamorphic series metamorphic rocks was determined by the K/Ar method, and the age of the mineralizations was determined by the Rb/Sr method.

REFERENCES

Airinei Ş. (1980) Radiografia geofizică a subsolului României. Ed. Ştiinţifică şi Enciclopedică. Bucureşti.

Bagdasarian P. G. (1972) Despre vîrsta absolută a unor roci eruptive şi metamorfice din masivul Ditrău şi munţii Banatului din România. *Stud. cerc. geol. geofiz. geogr., Seria geol.*, 17, 1, p. 13–21. Bucureşti.

Codarcea-Dessila M., Savu H. (1967) Cu privire la interpretarea datelor obţinute prin metoda K/Ar în Carpaţii Meridionali. *Stud. cerc. geol. geofiz., geogr., Seria geol.*, 12, 2, p. 351–365, Bucureşti.

Ene I. (1980) Report, the archives of the Enterprise for Rare Metals.

Giuşcă D., Ianovici V., Mînzatu S., Soroiu M., Lemne M., Tănăsescu A., Ioncică M. (1967) Asupra vîrstei absolute a formaţiunilor cristaline din vorlandul orogenului carpatic. *Stud. cerc. geol. geofiz., geogr.*, Seria D (Prosp. geof.) III, p. 67–173, Bucureşti.

Pavelescu L., Pop G., Ailenei G., Ene I., Soroiu M., Popescu G. (1975) K/Ar age determinations from the Apuseni and the Banat Mountains. *Rev. Roum. Geol. Geophys. et Geogr.,Geophysique*, 19, p. 67–79, Bucureşti.

— Pop G., Catilina R., Ene I., Ailenei G., Popescu G., Soroiu M., Lubenescu D. (1978) Report, the archives of the State Committee of Nuclear Energy, Bucharest.

Soroiu M., Popescu G., Gherasi N., Arsenescu V., Zimmermann P. (1976) K/Ar dating by Neutron activation of some igneous and metamorphic rocks from the Southern Branch of the Romanian Carpathians. *Eclogae geologicae Helvetiae*, 63, 1 Birkhäuser printers, Basel (Switzerland).

ГЕОХРОНОЛОГИЯ ЮГО-ЗАПАДОГО КРАЯ ВОСТОЧНО-ЕВРОПЕЙСКОЙ ПЛАТФОРМЫ [1]

Н. П. СЕМЕНЕНКО [2]

Восточно-Европейская платформа закончила свое развитие 1100—1200 млн. лет тому назад. К западу от края Восточно-Европейской платформы, проходящего параллельно Предкарпатскому прогибу, начиная от фундамента Карпат и до Арморикапского массива и побережья Атлантического океана во Франции, залегают в фундаменте складчатых сооружений кристаллические породы относящиеся к структурным ярусам пятого докембрийского мегацикла возрастом от 1100 до 550 млн. лет. В фундаменте Карпат они слагают галицийские ярусы складчатости, котоные подразделяются на Раховский цикл возрастом 550—700 млн. лет и Черноморский цикл возрастом 700—1100 млн. лет. Древнейшими являются геологические формации пятого докембрийского мегацикла представленные Черноморским ярусом раннегалицийской складчатости возрастом 650/700—1100 млн. лет, синхроным раннебайкальскому времени орогенеза или даласландской складчатости, выходящей на поверхность в юго-западной окраине Балтийского щита. Пояс этой складчатости прослеживается в меридиональном направлении дугообразно изгибаясь.

В фундаменте Карпат самой северной точкой, где установлены кристаллические образования этого времени, являются Ржешатары возле г Кракова, возраст которых по роговой обманке (по нашим определениям) 870 млн. лет и метаморфизм отвечает гнейсово-амфиболовой ступени. Это пояс проходит на территории Советских Карпат к западу от Раховского массива. Дугообразно изгибаясь (по данным румынских геологов Д. Джушкэ и др.) в районе гор Апусень на территории Румынии пояс прослеживается в широтном направлении на территории южной половины Венгрии, где по определениям А. Ковача установлен возраст 950—1000 млн. лет. Далее он прослеживается в меридиональном направлении на территории Югославии в Вардарской зоне, где самой южной точкой является Пелагонский массив, в котором Делеоном установлены возрасты 800—1100 млн. лет. Самой восточной точкой развития Черноморской складчатости является фундамент мегаантиклинория

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[2] СССР.

Крыма, где нами установлены возрасты 800—1100 млн. лет. Этот южный широтный пояс Черноморской складчатости опоясывает, очевидно, и Мизийскую платформу. Здесь на территории Румынии в Мизийской платформе выступают более древние ярусы докембрия возраст которых для формации Палазу Маре установлен 1800 млн. лет. В северной части этот пояс поворачивает в субширотном направлении на территорию Словакии и за пределами рассматриваемой карты к западу от Карпат этот ярус складчатости прослеживается в пределах Чешского массива, где установлены возрасты от 565 до 1100 млн. лет, а на территории ГДР: в Гранулитовом, Виндельфельдском массивах и Шварцбургской антиклинали они прослеживаются возрастом (по нашим определениям) 750—790 млн. лет.

Для северной зоны Арморикапского массива во Франции установлены возрасты от 600 до 950 млн. лет, а в Центральном французском массиве и фундаменте Парижского бассейна установлены возрасты 670—650 млн. лет. На юге Европы в Пиринеях для чарнокитов и гнейсов установлен возраст 580 млн. лет.

В северной части Европы на Скандинавском полуострове в юго-западном обрамлении Балтийского щита установлены возрасты Даласландской складчатости — 800—1000 млн. лет. К западу от нее в районе Телемаркена в южной Норвегии развиты интрузии и метаморфизм гранулитовой ступени этого же возраста 900—1000 млн, лет, накладывающийся наподстилающие Готские структурные ярусы амфиболитовой ступени метаморфизма возрастом 1200—1300—1400 млн. лет.

В Шотландии также установлены выступы древных структурных ярусов докембрия возастом до 2500 млн. лет, которые сходны с докембрием Гренландии.

Таким образом, в пределах Скандинавского полуострова в западной части Балтийского щита западная граница Восточно-Европейской платформы проходит по складчатой зоне даласландского возраста 800—1000 млн. лет. В южной части Скандинавского полуострова в Южной Норвегии развит складчатый ярус даласландской складчатости 800—1000 млн. лет, наложившийся на подстилающие Готские структуры возрастом 1200—1400 млн. лет. Они уходят на запад под Каледонский складчатый пояс. Здесь на юге от Балтийского щита отсутствует ярус Галицийской складчатости Раховского цикла возрастом 700—560 млн. лет, широко развитый в южной части Европы в Карпатах, Чешском массиве, в Арморикапском и Центрально-Французском массивах.

Таким образом выделяется причленная к западу от края Восточно-Европейской платформы узкая субплатформа стабилизированная в постдаласландское время 700—1100 млн. лет, вытянутая в субширотном напра влении вдоль Балтийского моря, где отсутствует позднегалицийский ярус Раховского цикла докембрия V возрастом 700—560 млн. лет, развитый широко в южной части Европы. Пояс этой позднегалицийской складчатости непосредственно окаймляет Украинский щит и широко развит в фундаменте Польских, Советских и Румынских Карпат, прослеживается на востоке вдоль южного края Украинского щита и развивается на западе окаймляя Мизийскую платформу.

На территории Советских Карпат метаморфические толщи, слагающие Галицийский ярус складчатости, вскрыты скважинами в Предкарпатском прогибе на глубине 1000—2000 м и выходят в Чивчинских горах и Раховском массиве, где установлен их возраст 560—630—700 млн. лет. Этот же пояс прослеживается на юге у края склона Украинского щита в скважине в г. Кагуле. Далее на северо-запад этот ярус складчатости вдоль края Восточно-Европейской платформы (Волыно-Подольской плиты) прослеживается скважинами на территории Польских Карпат, где нами установлены возрасты: в Пуще, на глубине 1055 м — 545 млн. лет; в Нечайне на глубине 2265 м—625 млн. лет и в Подбожье на глубине 2172 м—640 млн. лет. Румынскими геологами (Д. Джушкэ и др.) метаморфические формации этого яруса складчатости широко прослежены на территории Южных Карпат в горах Апусень и Добрудже. Параллельный западный пояс Галицийского яруса складчатости развит на территории Словакии, который прослеживается в широтном направлении и в северной части Венгрии.

В Галицийском ярусе складчатости — верхнем финальном ярусе пятого докембрийского мегацикла развиты разные ступени метаморфизма метаморфических толщ: гнейсово-амфиболовая, двуслюдяная филлитовая ступени в Раховском массиве, в г. Кагуле, в Южных Карпатах на территории Румынии и до филлитовой и аспидно-сланцевой ступени в зоне Предкарпатского прогиба Советских и Польских Карпат и др.

В Румынских Карпатах с гнейсовыми ступенями метаморфизма связаны внедрения гранитов и мигматитов.

Рассматривая эти два структурных яруса коры, залегающие в фундаменте Карпато-Балканского региона, которые формировались в докембрии V в отрезке времени от 550 до 1100 млн. лет, мы можем констатировать, что в каждом ярусе наблюдаются разные ступени метаморфизма геологических формаций и они не связаны ни с мощностью отложений, ни с их возрастом; как более древние — более глубокие, так и более молодые — вышележащие структурные ярусы представлены метаморфическими породами разных ступеней метаморфизма (или разных метаморфических фаций). Метаморфизм связан с фазами тектонических складкообразовательных процессов. Метаморфизм гнейсовых ступеней связан с внедрениями гранитов и магматическими замещениями. Вдоль по простиранию складчатых поясов и вкрест наблюдается неравномерность метаморфических циклов, что связано, очевидно, с неравномерными тепловыми потоками и флюидно-магматическими потоками, так же как и неравномерными давлениями в процессе динамического движения горных масс при складкообразовании вдоль складчатых систем. Платформенными аналогами складчатых ярусов кристаллических пород Галицийской складчатости фундамента Европы являются отложения полесской серии возрастом 700—1000 млн. лет и волынской серии возрастом 550—700 млн. лет, залегающие на западном склоне Украинского щита в обрамлении Карпат.

Неравномерность ступеней метаморфизма связана также с повторным наложением новых тектоно-магматических циклов в долгоживущих подвижных зонах каким является Карпато-Балканский регион, начиная с 1100 и до 10—5 млн. лет. Поэтому кристаллический фундамент представляет полиметаморфическое образование. В этом же поясе Галицийской складчатости складчатые движения проходили и в нижнем палеозое; фазы каледонского цикла накладываются параллельно на галицийские метаморфические образования. В каледонском цикле выделяются: древне-каледонская или сандомирско-добруджинская фаза, синхронная салаирской фазе Сибири возрастом 475—550 млн. лет, проходившая в кембрии и нижнем ордовике. Она наблюдается в г. Бажове, в Сандомирском кряже в Польше, в Добрудже в Румынии и в разломах и вулканизме края Украинского щита. В Добрудже установлен возраст зеленых сланцев и амфиболитов 470—530 млн. лет. Кембрийская диабазо-филлитовая формация развита в Болгарии в цоколе Мизийской платформы.

Среднекаледонская фаза возрастом 425—470 млн. проходившая в ордовике и силуре, развита в южной части Галицийского пояса в Кагуле в амфиболо-гнейсовой и двуслюдяной ступенях. Она наблюдается в Польских Татрах в амфиболо-гнейсовой ступени, в Лаузетском массиве в ступени роговиков и сланцев, в Котовицах и Мржигляде в ступени аспидных сланцев. Граниты Лаузетского массива относятся, очевидно, к этому циклу. Наконец, позднекаледонская фаза возрастом 380—420 млн. лет, проходившая в силуре и в нижнем девоне, развита в Татрах и в Украинских Карпатах. В последних она представлена ступенью глинисто-аспидных сланцев.

Метаморфизм каледонского цикла накладывается на галицийские метаморфические образования в ступени глинистых, аспидных и филлитовых сланцев и проявляется в Раховском массиве и в Чивчинах в виде ретроградного метаморфизма филлитовой ступени, наложенного на гнейсовые ступени метаморфизма галицийского времени. В южной части: в Кагуле и в Добрудже, очевидно, наблюдается наложение прогрессивного метаморфизма двуслюдяно-гнейсовой и гнейсово-амфиболовой ступеней. Таким образом, Галицийский метаморфический пояс, окаймляющий Украинский щит, представляет метаморфическое образование Галицийских циклов, прошедшее переработку в Каледонском цикле.

Каледонский цикл метаморфизма проявляется также вдоль южного борта Мизийской платформы.

Герцинский цикл складчатости магматизма и метаморфизма представлен на рассматриваемой территории тремя фазами:

ранне-герцинская фаза возрастом 330—360 млн. лет, проходившая в нижнем карбоне — верхнем девоне, проявилась в Южных Карпатах;

средне-герцинская фаза (возрастом 270—330 млн. лет), проходившая в карбоне, известна в Украинских Карпатах в виде наложенных процессов ретроградного метаморфизма аспидно-филлитовой ступени диафтореза, в Южных Карпатах в гнейсово-амфиболовой ступени, в двуслюдяной гнейсовой и филлитовой ступенях, а также на территории Венгрии в двуслюдяной гнейсовой ступени, связанной с внедрениями гранитов;

поздне-герцинская фаза складчатости магматизма и метаморфизма (возрастом 240—270 млн. лет), проходившая в перми и закончившаяся в конце триаса, развита в наложенных структурах Южных Карпат в Румынии, в Кагуле—Советских Карпатах и Добрудже. Двуслюдяные гнейсы и сланцы гетского покрова в Румынии имеют (по нашим определениям) возраст 215—255 млн. лет.

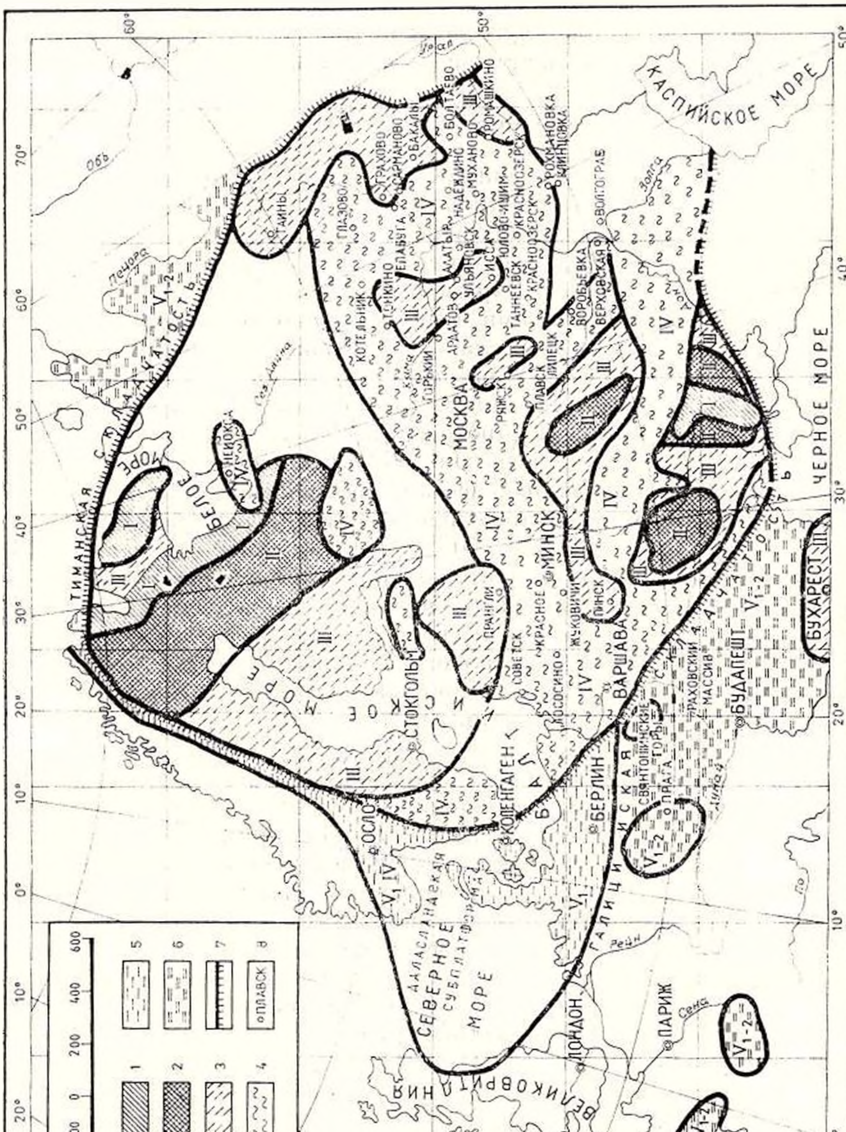
Таким образом на картах отдельных циклов метаморфизма рационально показать и районы развития различных фаз каледонского и герцинского циклов.

Альпийский — мезозойский цикл складчатости магматизма и метаморфизма доларамийского времени, возрастом 100—220 млн лет, проявляется на древних ярусах в большинстве в виде ретроградного динамотермального метаморфизма — аспидной и филлито-сланцевой ступеней, проходивших в сравнительно узких зонах в юрское и меловое время. Эти движения также сопровождались интрузиями габбро-диабазов и лерцолитов юрского и нижнемелового времени возрастом 150—170 млн. лет в Советских Карпатах: интрузиями юрских гранодиоритов сопровождающихся гранитизацией древних сланцев в Южной Венгрии. Эти возрасты наложенного метаморфизма широко развиты в Румынских Карпатах и известны в узких зонах динамометаморфизма в Татрах, наложенных на каледонские и герцинские образования. В Болгарии триас-юрский кристаллический комплекс наблюдается в филлитовой и двуслюдяной гнейсовой или эпидот-амфиболовой фазах метаморфизма.

Южно-Альпийский — постларамийский метаморфизм и магматизм, проходивший в отрезке времени от 10 до 100 млн. лет, проявляется в виде высоких ступеней метаморфизма, в южной части на Балканах. Здесь феноменальными являются гранитные внедрения и магматические замещения, мигматизация и образование гнейсовых ступеней метаморфизма до гнейсово-амфиболовых ступеней в палеогеновое время возрастом 70—80 млн. лет, которые наложены на древние образования. Эти возрасты 70—80 млн. лет устанавливаются как свинцовым, так и аргоновым методами по роговой обманке.

Проявления альпийского магматизма и наложенного метаморфизма на древние ярусы наблюдаются в северной части Югославии в Банате. В балканидах на территории Болгарии характерны процессы ороговикования третичных вулканитов-андезитов под воздействием внедрения монцонитовых тел и других магматических внедрений. Такие явления интенсивного ороговикования широко развиты в андезитах в районе Ропотам. Здесь характерно также развитие процессов метасоматической полевошпатизации — адуляризации и трахитизации в эффузивных толщах разного состава наряду с гидротермальными процессами пропилитизации и др.

В Советских Карпатах характерно развитие роговиков в узких зонах неогеновых отложений в связи с внедрениями штоков гранодиоритпорфиров и диорит-порфиритов субвулканического типа возрастом около 10 млн. лет. лет. Здесь (например у с. Хижы) устанавливаются такие ассоциации роговиков: корунд-силлиманит-кордиеритовая; корунд-андалузит кордиеритовая; гиперстен-плагиоклазовая; гиперстен-биотит -плагиоклазовая и биотит-кордиеритовая, относящиеся к ступени пироксеновых роговиков. Эти области, так же как и корни неогенового вулканизма рационально показать в виде ореолов. Весьма интересен альпийский магматизм и метаморфизм возрастом 30—15 млн. лет, наложенный на древние каледонские и герцинские метаморфические образования в Альпах на территории Швейцарии. Здесь они интенсивно проявляются в виде различных ступеней от аспидно-сланцевой и филлитовой ступени, представленной вторичными фенгитовыми сланцами, до гнейсово-амфиболовой и двуслюдяно-гнейсовой ступени с дистеном.



Карта геохронологии докембрия Европы.

1, Докембрий I 2700—3500; 2, Докембрий II 2000—2700; 3, Докембрий III 1700—2000; 4, Докембрий VI 1100—1700; 5. Докембрий V_1 700—1100; 6, Докембрий V_{1-2} 560—1100: 7, граница в Европейской платформе; 8, Пункты определения возраста докембрия в скважинах на платюорме.

В одних случаях они являются ретроградным, а в других — прогрессивным метаморфизмом по отношению к ранним ступеням метаморфизма каледонского и герцинского циклов. Здесь в Швейцарии известны и интрузии гранитов третичного времени (30—15 млн. лет) развитые на ограниченных территориях, а области развития вторичных сланцев на больших территориях связаны с интенсивными динамическими стрессами и приуроченными к ним термальными потоками.

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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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travaux du XII-ème congrès de l'association géologique carpatho-balkanique

METAMORPHISME - MAGMATISME - GÉOLOGIE ISOTOPIQUE