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POLYMETAMORPHISM, A DISCUSSION BASED ON EXAMPLES FROM THE ROMANIAN CARPATHIANS

BY

ION BALINTONI[1]

Introduction

The present study is a discussion on regional or orogenic metamorphism, as defined by Miyashiro (1975), with special reference to its mineralogical aspects. Regional metamorphism during Alpine time, characterized by low intensity and occurring on small areas in the Romanian Carpathians, is left aside. The main notions used by the petrological study of metamorphics are : mineral association and paragenesis. A mineral association is a group of minerals which constitutes a metamorphite sample ; a paragenesis represents the synchronous minerals of an association. This definition of the paragenesis is in agreement with Vernon's definition (1976) but differs from Winkler's (1976), according to which a paragenesis includes the minerals of an association placed in equilibrium, which implies their tangential position on the one hand and the simultaneous occurrence of several synchronous parageneses in a mineral association on the other hand. We consider that the paragenesis defined by Winkler (1976) should be better called subparagenesis. By taking into account a certain lithostratigraphic unit, the number of mineral associations will correspond, on the whole, to the number of petrographic types encountered and an equal number of parageneses for the simplest case and an indeterminate but limited one for more complicated cases. To make things easier, the term of general mineral association should be used for all the minerals belonging to a certain lithostratigraphic unit and that of general paragenesis for all synchronous minerals of the same lithostratigraphic unit. The polymetamorphism designates the repeated action of regional metamorphism factors on the same rock pile at great time intervals so that the periods of activity belong to different orogeneses (e.g., Hercynian, Caledonian, Cadomian, etc.). Although one may speak of the impetuous development of the study of metamorphism during the last two decades, due to the progress achieved by experimental petrology and the revo-

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lution of earth sciences brought about by global tectonics, the concept of polymetamorphism is not even mentioned by the most representative researchers of this time, such as Turner (1968), Miyashiro (1975), Winkler (1976) or Vernon (1976).

Middle Paleozoic Formations

The data regarding the Middle Paleozoic formations have been recently synthetized by Kräutner (in Săndulescu et al., 1981) for the East Carpathians, by Kräutner et al. (1981) for the South Carpathians and by Dimitrescu (in Rădulescu, Dimitrescu, 1982) for the Apuseni Mts. The Middle Paleozoic formations, proved by paleontological evidence, belong to the Ordovician-Lower Carboniferous interval and underwent metamorphism during Hercynian orogenesis, Sudetian phase ; the metamorphics are unconformably overlain by non-metamorphosed Upper Carboniferous. In case the formations considered by some authors not to belong to the Middle Paleozoic are excluded (most of the Ţibău Series and the Argestru Series of the East Carpathians — Balintoni, 1981, 1982, the Nădrag Series, the Bătrîna Series, the Govăjdia and Ghelar Series in southern facies of the South Carpathians — Hîrtopanu, Balintoni in Udubaşa et al., 1983, unpublished data), the general mineral associations of Hercynian metamorphics include only one general paragenesis, which belongs thermodynamically to the chlorite zone. The characteristic minerals of these general parageneses are : albite, epidote, clinozoisite, actinote, chlorite, stilpnomelane, chloritoid, lotrite (pumpellyite), prehnite, sphene, rutile, sericite, iron oxides. The Middle Paleozoic formations overlie transgressively the Early Caledonian metamorphics or more ancient ones ; within them there is a stratigraphic unconformity between Devonian and older sequences.

Late Precambrian-Lower Paleozoic Formations

These formations are less known than the Middle Paleozoic ones. Organic remains are scarce, ill preserved and have a wide circulation in time. Isotopic ages are few and it is difficult to interpret them. Information about these sequences is given by Kräutner (1983) in connection with the East Carpathians, by Dimitrescu (1983) for the Apuseni Mts and by Kräutner et al. (1983 a) for the South Carpathians. Hârtopanu et al. (1982), Balintoni (1982) and Gheuca, Dinică (1983) argued for the non-inclusion of the Arada Formation, the Codru Complex and the Muncel Formation in the Apuseni Mts as well as of the Lereşti Formation in the South Carpathians, in the Late Precambrian-Lower Paleozoic formations stated by the above mentioned authors. The Cibin Group in the South Carpathians is insufficiently proved paleontologically and petrographically. Thus, we leave it aside. Therefore, the Tulgheş Group in the East Carpathians and the Biharia Group in the Apuseni Mts will be commented upon.

Tulgheş Group. According to new paleontological data and to a new analysis of the existing paleontological and isotopic data, Iliescu et al. (1983) assign the Tulgheş Group to the Cambrian age, marked by possible downward transition to the Vendian and upward one to the Lower Ordovician. Balintoni, Chiţimuş (1973) reported two metamorphic parageneses built up of metapelites of the Tulgheş Group, and considered them an illustration of polymetamorphism. The authors mentioned an older paragenesis made up of biotite and a rutile paramorph after brookite in equilibrium with albite, chlorite, muscovite and a second paragenesis consisting of chlorite after biotite, second generation light-ferriferrous rutile (resulted from the recrystallization of preexisting rutile) and iron oxides as neoformation minerals. To the general parageneses of the Tulgheş Group may be also added epidote, clinozoisite, actinote and sphene, bearing thermodynamic significance. The first paragenesis formed under static conditions. The second, concomitantly with the generation of a sometimes highly penetrating foliation in metapelites. The former paragenesis points to thermodynamic conditions in the biotite zone and the latter in the chlorite zone, being of regressive nature. Due to the prevalence of Hercynian K-Ar ages disclosed by the study of rocks in the Tulgheş Group, the second paragenesis is assigned to Hercynian metamorphism and as far as the paleontologic ages of the Tulgheş Group are not greater than Lower Ordovician, the first paragenesis may be considered Early Caledonian. The lowermost part of the Tulgheş Group is not known due to the Alpine and pre-Alpine overthrusts (Balintoni et al., 1983) and the uncertain occurrence of transgressively overlying Middle Paleozoic formations.

Biharia Group. According to Balintoni (1982), the Biharia Group was proved paleontologically by Visarion (1970), Visarion, Dimitrescu (1971) and Solomon et al. (1981). The Biharia Group represents a mainly metabasite sequence. Giuşcă (1979) reported the complex character of mineral associations of rocks in the Biharia Group, while Balintoni (1983) distinguished in the Highiş-Drocea massif two general parageneses of regional metamorphism : a relict general magmatic paragenesis and a thermic contact paragenesis due to Upper Paleozoic granitic intrusions. The older general metamorphic paragenesis, formed under static conditions includes : hastingsite hornblende, epidote I + albite I (within old plagioclases), sphene I (with relict ilmenite in the middle), biotite, almandine. From thermodynamic point of view, it is characteristic of the almandine zone or of the albite or epidote amphibolite facies (e.g. Miyashiro, 1975). The more recent general paragenesis, formed under dynamic conditions, includes : actinote, chlorite, albite II, epidote II, sphene II, magnetite. The second generation albite is of porphyroblastic nature, epidote is iron poorer (frequently clinozoisite) and sphene does no longer include ilmenite relics. The minerals belonging to the second general paragenesis of regional metamorphism are iron poorer and the latter occurs as magnetite ; from thermodynamic point of view they enter the chlorite zone and are related to a penetrative foliation in which they range. As far as the Biharia Group alone is transgressively overlain by the Belioara Subgroup and probably the Păiuşeni Subgroup, of Middle Paleozoic age, the second general paragenesis is assigned to Hercynian metamorphism ; according to paleontological data, the first general paragenesis may be assigned to the Early Caledonian age, the metamorphic history of the Biharia Group being similar to that one of the Tulgheş Group. The lowermost part of the Biharia Group is not known ; this group forms the most important part of the Alpine Biharia Nappe. As regards the Late Precambrian-Lower Paleozoic formations, the following characteristics are to be noted : two general parageneses of regional metamorphism ; the regressive aspect and the local occurrence of the second related to a penetrative foliation ; the essentially static character of the former ; the variable intensity, but not too high, of the first metamorphism ; the omission of some zonations of the first metamorphism.

Precambrian Formations

Recent syntheses on the Precambrian Carpathian formations have been carried out by Kräutner (1980, 1983), Kräutner et al. (1981), Kräutner, Balintoni (in Săndulescu et al., 1981), Kräutner et al. (1983 a, b), Dimitrescu (1983), Iancu (1983). According to Kräutner et al. (1983 a, b) they are of Middle Proterozoic age (1650 ± 50 to 850—1000 m.y.), their initial metamorphism belongs to Grenvillian orogeny and lower and upper Middle Precambrian formations can be distinguished by means of superposition. The same authors consider that the following metamorphism belongs to Early Caledonian orogeny and there is a pre-metamorphic sequence Upper Precambrian-Lower Paleozoic in age. Salop's (1983) division of the Precambrian points to the Neoproterozoic as part of the former sequence of Precambrian formations mentioned above, while the Epiproterozoic is conformable with the Eocambrian (Vendian) and the Lower Paleozoic. Iancu (1983) reports even a pre-Grevillian age for some South Carpathian metamorphics and also admits other Grenvillian and Assyntic ones. The existing objective data render hypothetical the pre-Grenvillian metamorphics, while the Grenvillian ones are supported by two isotopic ages only : Rb-Sr age of 842 m.y. (Bagdasarian, 1972, fide Kräutner et al., 1983 a) for the South Carpathian Sebeş-Lotru Group and K-Ar age of 748 m.y. (Semenenko et al., 1969, fide Kräutner, 1983) for the East Carpathian Bretila Group. As regards the division of Middle Proterozoic into lower and upper (Kräutner et al., 1983 a, b) by means of conformity between Cumpăna Group and Făgăraş Group in the Făgăraş Mts (Kräutner, 1980), recent studies (Balintoni, 1983, unpublished data) have pointed to a tectonic relationship between the two groups all over the massif and thus to the non-validity of the dividing criterion. Finally, in the Danubian metamorphic formations in the South Carpathians and the Rebra Group in the East Carpathians, the U-Pb ages from granite zircons and K-Ar ages of 601—667 m.y. were reported (Grünfelder et al., 1981, fide Kräutner et al., 1983 a ; Mînzatu et al., 1975, fide Kräutner, 1983). These account for Cadomian orogenesis.

Parageneses of the Rebra Group. Balintoni, Gheuca (1977) have described in the general mineral associations of the rocks in the Rebra Group from the Bistriţa Mts, three parageneses which account for different thermodynamic fields during their genesis : (1) staurolite, kyanite pointing to Barrovian-type metamorphism ; (2) andalusite, cordierite generated under static conditions at the expense of the former, posterior to the generation of a highly penetrative foliation which affects the paragenesis (1) mechanically. Paragenesis (2) is a high temperature and low pressure one. In some places, posterior to the mentioned penetrative foliation, one notices fibroblastic sillimanite generated on biotite. The relationship sillimanite-paragenesis (2) is however unclear ; in the Lăpuş Mts where the sillimanite is of the same generation as the one in the Bistriţa Mts ; andalusite and cordierite are lacking. Paragenesis (2) shows limited extension. At the top of the formations belonging to the Rebra Group it is to note a general paragenesis (3) including chlorite, albite, epidote, actinote, characteristic of the chlorite zone and formed on preexisting minerals. The authors cited above have assigned the first paragenesis to Cadomian metamorphism, the second to Early Caledonian metamorphism and the third to Hercynian metamorphism.

Parageneses of the Sebeş-Lotru Group. Bercia (1975) and particularly Hârtopanu (1982) have shown the remarkable paragenetic complexity of the rocks belonging to this group. From some areas of the South Carpathians, mainly from metapelites, they reported some parageneses including the following index minerals for specific fields of thermodynamic factors : (1) staurolite, kyanite, prior to strong deformational stage ; (2) staurolite, kyanite ± sillimanite, posterior to the mentioned deformational stage ; (3) andalusite, cordierite ± sillimanite, also posterior to the same deformational stage ; (4) chlorite, albite, epidote, actinote, chloritoid as general paragenesis of late regional metamorphism. According to these authors, the first three parageneses correspond to two metamorphic events ; the high temperature and low pressure paragenesis (3) shows limited extension within the area of paragenesis (2), where it forms by substituting paragenesis (1). Concomitantly with parageneses (2) and (3) took place the regional migmatisation subordinately accompanied by palingenetic magmas at present-day erosion levels. Except for migmatisation on wide areas of paragenesis (2) of barrovian type, it is to note a similar metamorphic history of Rebra Group and Sebeş-Lotru Group. However, it is worth mentioning that the rocks of the Rebra Group from Preluca Lăpuşului exhibit a general reorganization of the paragenesis including staurolite, kyanite, without any important mineralogical transformations, posterior to strong deformation (Balintoni, 1982, unpublished data). Because of limited space we present no other Precambrian piles. Further on, their general characteristics are treated upon. The Precambrian formations do not exhibit inner unconformities, their outcropping lower parts resting on tectonic planes.

General Characteristics of pre-Alpine Carpathian Metamorphics

1. The Hercynian metamorphics exhibit a sole general paragenesis, the Early Caledonian ones two and the Precambrian ones two or three. That is to say that pre-Hercynian metamorphics are polymetamorphic and to one orogenesis may be assigned only one general paragenesis. Thus, the instances from the Carpathians decrease the importance of multi-stage metamorphism model and plead for a polycyclic one.

2. Obvious zonations have been reported so far for the second general paragenesis of Precambrian metamorphics (Bercia, 1975 ; Hârtopanu, 1982). It means that the initial general parageneses were not zoned for any of pre-Alpine Carpathian metamorphics piles. In the past, isogrades were stated between the index minerals of different general parageneses.

3. The Hercynian general paragenesis belongs to the chlorite zone both of super- and infrastructures, where it formed from top to bottom to the limit reached by water. The initial general paragenesis of Precambrian metamorphics is of low pressure type in places of high temperature (andalusite, cordierite, sillimanite) and in case it formed during Early Caledonian orogeny (this could be accounted for by the fact that the general paragenesis of Early Caledonian metamorphics formed under static conditions), then it differs thermodynamically from the latter.

4. Precambrian metamorphics alone exhibit autochthonous granitizations associated with extended migmatisation, as well as regional migmatisations independent of granitoids. This accounts on the one hand for the deep erosion of Precambrian metamorphics, and on the other hand for their great thickness, opposed to the film-like aspect of most Hercynian metamorphics.

5. As regards some Precambrian metamorphics with only two general parageneses, a regional migmatisation posterior to initial metamorphism was revealed.

6. No high pressure paragenesis has been reported so far for the pre-Alpine metamorphics in the Romanian Carpathians.

7. The thickness of metamorphics decreases obviously from Precambrian to Hercynian ones. This characteristic is accompanied by the decrease of intensity of metamorphism. As compared to the above mentioned characteristics, the metamorphics which have not been assigned to the orogenies in which they were classified (most of the Tibău Series and the Argestru Series in the East Carpathians, the Arada Formation, the Codru Complex and the Muncel Formation in the Apuseni Mts, the Nădrag Series, the Bătrîna Series, the Ghelar and Govăjdia Series in southern facies and the Lereşti Formation in the South Carpathians) are Precambrian in age.

Physical Features of Metamorphic Processes

1. The examples given above show that repeated metamorphic processes lead to mineralogical changes of pre-existing metamorphics in case : the baric field alters within a mineral association the equilibrium of which depends on pressure at a given temperature ; fluids or magmas,

independently of their movement direction, enter another thermodynamic field than the one in which the pre-existing metamorphics were generated ; the crystalline networks of pre-existing minerals become deformationally unstable with great rock volumes and a thermodynamic field characteristic of the metamorphic domain.

2. The pre-existing metamorphics are not always proved to have been retromorphosed as infrastructures of some sedimentary or volcano-sedimentary sequences.

3. For some general parageneses one may state high grade cystallization conditions, even in the absence of some minerals such as staurolite or aluminium silicates, on the condition that chlorite be lacking ; the absence of the latter from a general paragenesis where it may be chemically present, points to a zone above the almandine zone, namely the staurolite zone. The initial metamorphism of all Carpathian Precambrian metamorphics enters the present-day erosion level of chlorite-out field.

4. Some Precambrian piles exhibit eclogites. As compared to the initial general paragenesis they show relict features.

5. Hercynian metamorphics and probably a part of the Cibin Group (Kräutner et al., 1983 a) overlie a sialic basement. By considering the Alpine metamorphics, such instances are related to overthrust fields in collision areas between sialic plates.

Pre-Alpine Metamorphics and Global Tectonics

We note the following : Hercynian nappes are known on the Carpathian territory (Balintoni et al., 1983, Berza et al., 1983) ; pre-Alpine high pressure metamorphics are not known, but one encounters pre-Alpine low pressure parageneses as well as Cadomian, Early Caledonian and Hercynian granitoids within delimited areas that point to certain alignments ; metamorphosed sequences overlie a sialic basement or represent thick sialic piles ; there are also some alignments delimited by pre-Alpine ophiolites. All these remarks account for different metamorphic processes related to converging contacts between crustal plates, ended by collisions, starting from Cadomian to Hercynian times. Owing to their characteristics, most of Carpathian pre-Alpine metamorphics represent sequences of the compartment situated above the Wadati-Benioff plane.

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CURVES OF MINERAL ISORELATIONS, A CONCRETE METHOD TO RESEARCH METAMORPHISM CONDITIONS

BY

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Introduction

The assessment of physical conditions of metamorphism in various areas is usually obtained by tracing the isograde curves with the help of index minerals. This method was used for a long time in the Carpathian areas. In mose cases, anyhow, it appeared that assemblages containing index minerals are in an evident chemical unbalance, taking into consideration the fact that the number of real phases often exceeds by the one foreseen by the mineralogical phases rule. More than that, the microscopical observations on reciprocal relationships among the minerals composing these complex assemblages show a clear paragenetic superposition as some minerals are relict and others of neoformation. Even the index minerals which are classic have either relict or neoformation appearances. Having in view these considerations, we think that the isograde lines traced now on the map refer to heterogeneous events and conditions ; they mask the real physical conditions from the climax period of metamorphism. Taking into account this difficulty, we agree to another alternative of studying the mineral assemblage. The main idea is : by microscopical analyses we underline for each area the relict and the neoformation minerals and when considering them, we delimit the mineral reaction of adaptation which took place in the respective area. The area outline where the same type of relationship took place is shown on the map as a line which we denominate as the mineral isorelation curve. Thus, instead of isogrades which should have designated the same metamorphism conditions (usually equilibrium monovarying conditions of metamorphism), within the metamorphic space, some mineral isorelation curves are traced, which separate geological spaces with different mineral reactions. In this way, we think a better observation basis can be outlined, in order to understand the evolution sense of variable conditions of metamorphism.

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Reaction Identification

The mineral reactions are practically identified under microscope, by following the space relationships among them. This method has been used for a long time but it needs a careful precaution, because :

— a thin section not always detects a whole mineral reaction ;

— the chemical composition of many metamorphic minerals is variable, as for the isomorphous series, so that it is often difficult to delimit the chemism of the reaction terms. For this reason we are often obliged to attribute a standard composition to many minerals with a variable chemism ;

— the relative age of minerals is sometimes ambiguous and thus the old reaction term cannot be separated from the neoformed one.

It is also to be understood that a mineral reaction can be really "read" under microscope only where the reactions were incomplete, namely only where some transition states were preserved from the initial reaction member to the final one.

Under microscope, the minerals belonging to the initial reaction term are metasomatically replaced by the minerals of the second reaction members. A reaction is completely established ("read") when there appear some coupled substitutions, such as :

a) A substituted by B ; b) C substituted by D, and when the general chemism of the assemblage A + C is the same as the general chemism of the assemblage B + D. In this case, the real and complete mineral reaction is the following :

$$A + C \longrightarrow B + D$$

Taken as a whole, this is an isochemical mineral reaction, but when regarded at a crystal scale, it is expressed as a metasomatic substitution.

In most cases, by direct observations in the same thin section only a side of the substitution couple is remarked. For example, only the substitution of A by B is evident, while the substitution relationship of C by D is absent. The causes can be different, but the most frequent one seems to be the relative amount of reacting minerals. If the complete reaction supposes :

$$aA + cC \longrightarrow bB + dD$$

where a, c, b and d are stoechiometric coefficients and if within the initial assemblage the relative amount of A is larger as compared to the coefficient a asked by the reaction, the final result of the reaction would be :

$$A + C \longrightarrow B + D + A \text{ (in excess)}$$

In this case, the initial phase A survives as a relict, coexisting with the reaction products B and D. In exchange, the phase C completely disappears and is going to be deduced. Anyway, it is known that in each metamorphic area the relative ratios among minerals from an assemblage vary from one point to another, so that if in a certain space from the metamorphic area it is an excess of A, it is possible that in another space to be an excess of C, the latter surviving as a relict. In this case, for each space separately, we "read" in thin sections only partial reactions : A + x → B + y or C + x → D + y.

Here x is the old mineral which disappeared or which is not seen under microscope in a relationship of substitution with one of the neoformation minerals. Such reactions, with a hidden term (x), which is going to be deduced, are called cryptical.

Isorelation in the South Carpathians

The question of replacing metamorphism isogrades by mineral isorelation curves was recently discussed by Hârtopanu (1982) who traced this kind of curves in the Mehedinţi Mts ; they were traced as well in the Godeanu Mts by the authors of the present paper (Figs. 1a and b).

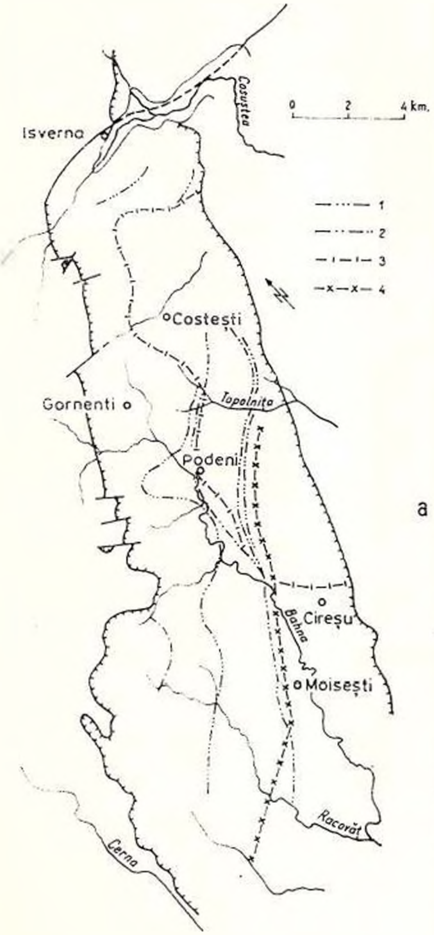

Fig. 1 a — Mineral isorelation curves in the crystalline of the Bahna Outlier (scale 1 : 200 000).
Substitution relationships. 1, biotite garnet ; 2, staurolite' andalusite : 3, biotite andalusite ; 4, kyanite muscovite.

Fig. 1b — Mineral isorelation curves in the crystalline of the Godeanu Mts.
Substitution relationships. 1, kyanite muscovite ; 2, staurolite andalusite ; 3, biotite/garnet.

Each isorelation curve delimits an area in which a certain type of cryptical mineral reaction was found. Here below we give the main lines :

1) The staurolite-andalusite isorelation line. This one separates the area with a clear substitution of andalusite by staurolite ; it corresponds to the following cryptical reaction :

$$\text{staurolite} + x \longrightarrow \text{andalusite} + y$$

The optical constants of staurolite relicts from andalusite neoformation phenoblasts correspond to a Fe-staurolite. The absence of Fe in andalusite makes us suppose that y must be a complementary Fe bearing mineral. The complete mineral reaction is similar to that proposed by Wenk et al. (1974) :

$$6\ \underset{\text{Fe staurolite}}{Fe_2Al_9Si_4O_{23}\ (OH)} + 11\ \underset{\text{quartz}}{SiO_2} = 4\ \underset{\text{almandine}}{Fe_3Al_2(SiO_4)_3} + 23\ \underset{\text{andalusite}}{Al_2SiO_5} + 3\ H_2O \quad (I)$$

2) Another line outlines the area where kyanite is replaced by muscovite, corresponding to the following cryptical reaction :

$$\text{kyanite} + x \longrightarrow \text{muscovite} + y$$

The kyanite substitution by muscovite is easily to remark under microscope. At first sight, it would seem to be a side of the classical regressive reaction :

$$\text{kyanite} + \text{potassium feldspar} + H_2O \longrightarrow \text{muscovite} + \text{quartz}$$

Anyhow, in most cases there is no sign to indicate the presence of potassium feldspar within the primary assemblage with kyanite. In exchange, the cryptical reaction area is partly superposed over another reaction area, namely :

3) The substitution relationship area of biotite by garnet (almandine), namely the area with biotite relicts in garnet corresponding to the following cryptical reaction :

$$\text{biotite} + x \longrightarrow \text{almandine} + y$$

It clearly appears that cryptical reactions 2) and 3) are complementary sides of the same isochemical mineral reaction :

$$\begin{aligned} Al_2SiO_5 + KFe_3AlSi_3O_{10}(OH)_2 + 2\ SiO_2 \longrightarrow KAl_3Si_3O_{10}\ (OH)_2 + \\ + Fe_3Al_2\ (SiO_4)_3 \end{aligned} \quad (II)$$

4. Another isorelation line outlines the area with an evident substitution of biotite by andalusite. The cryptical reaction in this case is :

$$\text{biotite} + x \longrightarrow \text{andalusite} + y.$$

By deduction we conclude that this cryptical reaction is the complementary side of a complex mineral reaction which implies the simultaneous participation of reactions (I) and (II) :

Thus, by the help of isorelation lines we can conclude that within the same metamorphism space there coexist some terms of two mineral parageneses belonging to two different conditions.

The mineral isorelation lines tracing in the Mehedinţi and Godeanu mountains (South Carpathians) ends a long research period.

In the Godeanu Mts, Bercia (1975) described a Barrovian type metamorphism in the western part of the massif and an intermediary type metamorphism of low pressure, in its eastern part. The question whether the two metamorphism types (located in neighbouring, partly superposed areas) are synchronous or in succession, is still unsolved.

In the Mehedinţi Mts, in the Bahna Outlier area, a Barrovian type metamorphism was described as well, over which there is a partly superposed Pyrenean type metamorphism (Hârtopanu, 1975). The direct substitution relationships among minerals characteristic for the two types of metamorphism demonstrate their succession in time and therefore the polymetamorphic character of the metamorphic area of the Mehedinţi Mts.

The isorelation lines underline the polymetamorphism by the pointing out of the two superposed parageneses, both in the Mehedinţi Mts and in the Godeanu Mts. The two parageneses show that the same metamorphic area successively passed from higher pressure conditions to lower pressure ones.

Final Remarks

The analysis of the structural relationships among minerals is the main way to discover the chemical reactions and polymorphic transformations which took place during the polycyclic metamorphism. Thus, we can distinguish the old minerals (paleominerals) and the new ones (neominerals). The paleominerals survival is probably due to two causes :

— the relatively low speed of some complex reactions which implies the participation of many minerals, widespread on a relatively large area. The slow diffusion of components towards new germs could be the main reason of such a slow reaction ;

— the variability of minerals ratio from an assemblage which causes the excess of one or more minerals. The mineral in excess is not entirely consumed and it is preserved as a relict among the other neoformation minerals.

The discovery of paleominerals is sometimes a difficult job which supposes a very careful observation under microscope. But even in this case the microscope can seldom notice relationships among minerals implied in reaction, either because the reaction space is by far larger than the area of a thin section, or because some paleominerals were totally consumed in the metamorphic reaction. The study of mineral relationships on a large metamorphic area, by tracing the isorelation lines, compensates this disadvantage. The sketches below exemplify in a better way the process by which the isorelation lines help to the discovery of a reaction, taking as a model the following reaction :

kyanite + biotite + quartz ⟶ almandine + muscovite

Figure 2 shows the initial paragenesis area (kyanite + biotite + + quartz), indicating the spaces where kyanite and biotite respectively are in excess. Figure 3 shows the same area after suffering a new metamorphic process. The isorelation line apparently separates two parageneses : muscovite + garnet + biotite (on the right) and muscovite + + garnet + kyanite (on the left). But the mineral assemblage on the right has biotite as a paleomineral and the assemblage on the left contains kyanite as a paleomineral. Only near the isorelation line the coexistence of the two paleominerals is possible and only here the mineral reaction can be relatively correctly "read". Even the mineral reaction kyanite + biotite + quartz → muscovite + garnet, which in our opinion is very widespread in the polymetamorphic areas of many orogene zones, is hardly noticed without the help of isorelation lines.

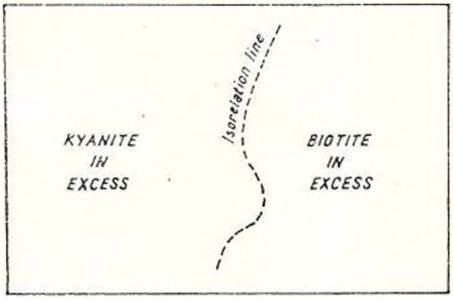

Fig. 2 — Widespreading area of the metamorphic assemblage : kyanite + biotite + quartz (before polymetamorphism).

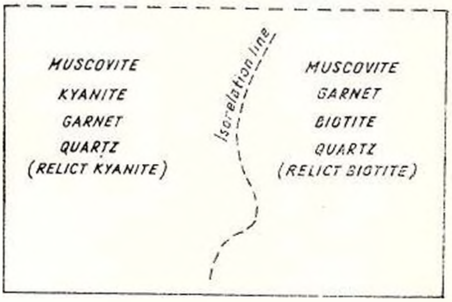

Fig. 3 — Post-polymetamorphic metamorphic area.

It is true that the tracing of an isorelation line supposes a very hard work. But we think that with the help of these lines, some otherwise unnoticed mineral reactions can be discovered. We also think that the mineral isorelation line has the quality to express in a concrete way a mineral reaction in the field and at the same time the quality to materialize on the map the statistical character of the perspective reaction.

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DYNAMIC AND RETROGRADE METAMORPHISM : EXAMPLES FROM THE ROMANIAN SOUTH CARPATHIANS

BY

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Recent progress achieved in the study of the present geological structure and past evolution of the Romanian Carpathians results from a better approach to the metamorphic petrology.

Thus, large areas of long time called "greenschist facies rocks" have proved to be polymetamorphic, with at least one medium-grade relict paragenesis, variously obliterated by low-grade minerals. Moreover, some "phyllite zones" sandwiched between gneissic zones are in fact mylonitic or blastomylonitic belts, marking the border between distinct tectonic units. The aim of this paper is to review such key zones in the South Carpathians, familiar to the authors, after a brief discussion of the terminology and genetic models most widely accepted.

The State of the Problem

Textbooks and review papers generally make a distinction between elongate zones of highly deformed rocks and large areas of downgraded crystalline schists, by discussing them either in distinct sections, or in independent contributions. However, a simultaneous discussion on the terminology for the main processes and products of dynamic and regional metamorphism, from both the points of view of metamorphic petrology and structural geology, is useful.

Dynamic metamorphism is a solid state rock transformation in which directed pressure (stress) has the dominant role and corresponding important strain is obvious (Harker, 1950 ; Turner, 1968 ; Winkler, 1976). By tradition, its thermal régime was considered to be low, and the deformation of coarse grained rocks to be of crush type (seismic cataclastic metamorphism), while fine grained rocks show plastic behaviour, flowing along slaty cleavage planes (Harker, 1950 ; Spry, 1969). Recent papers claim for higher temperatures in the deepest-seated segments of some fault zones with consequent involvement of

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different mechanisms of deformation (aseismic quasiplastic creep — Sibson, 1977).

From the beginnings of petrology, regional metamorphism has been regarded as a dynamothermal solid state process, in which both temperature and various types of pressure play essential roles. Extensive retrogressions occur usually due to subsequent distinct orogenic cycles.

Theoretical considerations (Harker, 1950) and a gradational series of resulted rocks may justify the consideration of dynamic metamorphism as a special case of dynamothermal metamorphism, symmetrical to thermal metamorphism, the opposite end member. However, mapping data reveal distinct outcrop patterns for the products of dynamic metamorphism — elongate thin belts interpreted as shear zones — and for the products of dynamothermal "regional" metamorphism — large areas with a more or less belt-like shape.

The terminology used for the products of dynamic metamorphism is rather unsettled due to gradual variations of physical conditions and distinct mechanical behaviour of various rocks in the same environment. Following (partly) Higgins (1971), and especially Spry (1969) and Sibson (1977), we use the following classification :

Cataclastic rocks (random fabric)	incohesive	fault breccia — fault gouge
	cohesive	protocataclasite-cataclasite-ultracataclasite
Rocks with fluxion structure	matrix with minor neomineralization-recrystallization	protomylonite-mylonite-ultramylonite (phyllonite when phyllosilicates are abundant)
	matrix with strong neomineralization-recrystallization	blastomylonite

Cataclastic Rocks in the Olt Valley

In the Olt Valley, two orthogonal faults (30—50 km long) dissect both the uppermost Alpine nappes of the South Carpathians and their posttectonic cover : the N-S trending Olt Fault and the E-W trending Cozia Fault, considered synchronous (post-Lower Miocene) by Hann and Szász (1984).

The Olt Fault is distentional, with the eastern side down-thrown. This fault is marked by both cataclastic (protocataclasites, cataclasites, ultracataclasites, pseudotachylites) and mylonitic rocks, but geomorphologic effects are minor as both walls are built up of similar metamorphic rocks.

The Cozia Fault is also distentional, with the northern side down-thrown, as the posttectonic cover is offset over 1500 m. The contrast between the northern sedimentary wall and the southern gneissic wall is marked by a steep slope 1000 m high. The dinamically deformed rocks may be seen only in the southern wall, where extensive cohesive crush breccias are conspicuous.

The southern slope of the Cozia gneissic horst was also affected by extensive cataclasis — the Brezoi Breccia, long time disputed by Romanian geologists whether of tectonic or sedimentary origin. Hann and Szász (1981) have pointed out that in this area a sedimentary Senonian breccia comes in contact, in places, with a cohesive crush breccia formed on the Cozia Augengneisses. This crush breccia extends 50 km eastwards and is the result of pre-Senonian tectonics.

Most of the faults and fault rocks in the Olt Valley are typical of the elastic-frictional behaviour of brittle rocks, reflecting a shallow (< 10 km) crustal environment (Sibson, 1977). They involve Precambrian to Senonian rocks, being the result of Neoalpine (Tertiary) tectonics.

Cataclasites and Mylonites at the Sole of the Getic Nappe

The Getic Nappe is not only the first nappe recognized in the South Carpathians (Murgoci, 1905), but also the most important one as areal extent (15 000 km^2), observed thickness (several kms) and minimal horizontal displacement (100 km). Its sole is marked by a layer of dynamically metamorphosed rocks, tens or hundreds of meters thick, described for the first time as mylonites-ultramylonites by Gherasi (1937) in the Godeanu Outlier. The medium-grade polymetamorphic Sebeș-Lotru Group rocks (paragneisses, mica schists, migmatites, amphibolites, leptynites, etc.) may show either cataclastic random fabric, or foliated mylonitic fabric (fluxion structure), according to the rock type or to the place of sampling. As a rule, protocataclasites-cataclasites-ultracataclasites derive from quartzo-feldspathic rocks, while mylonites-ultramylonites are found in more mica or amphibole rich rocks. Medium-grade minerals are usually subject only to physical actions, since biotite, garnet, hornblende are not frequently chloritized. They occur as more or less eye-shaped porphyroclasts enclosed in a fine grained matrix, mainly resulted from crushing, with extensive recrystallization but minor neomineralization. The presence of migmatic veins in paragneisses may produce similar aspects in outcrops, as the quartz-feldspathic ptygmatic layers are boudinated in a mica-rich quasiplastic deformed matrix.

The nature and dimensions of the dynamically metamorphosed layer are highly variable, depending both on the angle between the thrust plane and the lithological layering in the crystalline schists of the nappe and on the metamorphic or sedimentary nature, in a given area, of the top of the underlying Danubian Unit.

Chlorite Blastomylonites in the Culmea Cernei-Vîlcan-Parîng Mountains

One of the lowest Alpine nappes in the South Carpathians (the Main Lower Danubian Unit, Berza et al., 1983) exhibits, over a length of 120 km, a thin (≈ 1—2 km) strip of blastomylonites, exposed in the Culmea Cernei, Vîlcan and Parîng Mts. The blastomylonites belong to two distinct lithostratigraphic units : Drăgşan Group (amphibolites, leptynites, mica gneisses ± staurolite ± kyanite) with a few associated plutons, to the north ; Lainici-Păiuş Group (quartzites, biotite gneisses, crystalline limestones, graphitic mica gneisses ± sillimanite ± andalusite ± cordierite and various types of migmatites) with several associated granitoid plutons and porphyry diorite dykes, to the south. Medium-grade Precambrian metamorphism, of Barrowian type in the Drăgşan Group and of lower pressure type in the Lainici-Păiuş Group, was first followed by a static regional retrogression, attested by plagioclase → epidote + albite, biotite → chlorite, cordierite → pinnite, diopside → tremolite alterations, called "autoretromorphism" by Savu (1970).

The boundary zone between Drăgşan and Lainici-Păiuş rocks (or associated granitoids) stands out by the progressive development, from both directions inwards, of a north dipping penetrative mylonitic foliation, deformed in places by centimetric to metric tight kink folds. Medium-grade metamorphic or igneous minerals are more and more distorted, broken and replaced and occur as relict porphyroclasts (mainly feldspars, muscovite and hornblende). Strong neomineralization (albite, chlorite, epidote, tremolite, calcite, fine grained white mica, stilpnomelane) and recrystallization (quartz) do occur and yield to the rocks a blastomylonitic aspect. It is obvious that in this case the dynamic metamorphism developed in a thermal regime typical of low-grade metamorphism, in good agreement with the quasiplastic mechanism of deformation inferred from fabric aspects.

This blastomylonitic strip is the result of the overthrust of the Drăgşan Group rocks (and associated granitoids) on the Lainici-Păiuş Group rocks, or associated igneous rocks. The relations with various Paleozoic and/or Mesozoic formations in the Vîlcan, Parîng and Retezat Mts point, for this thrust, to a surely pre-Alpine (pre-Jurassic, possibly even pre-Namurian) but post-Devonian age (Kräutner et al., 1981).

Biotite Blastomylonites in the Mehedinţi Plateau

In the Porţile de Fier outlier of the Getic Nappe, an old tectonic contact of metamorphic nappe type occurs within the Precambrian Sebeş-Lotru Group rocks (Iancu, Hârtopanu, 1982). Underlying the typical Sebeş-Lotru sequences (with polymetamorphism testified by superimposed generations of mineral parageneses and/or folds), a quartzitic unit (Jidoştiţa Formation) with specific metamorphic (paragenetic and fabric) features occurs in apparent structural continuity. The transition is marked by a set of common surfaces — S_3 in the Sebeş-Lotru rocks (where they obliterate obviously the older S_1 and S_2 planes), S_1 in the Jidoştiţa rocks, respectively.

The increasing density of S_3 planes is accompanied by a neomineralization in the biotite zone superposed on the kyanite zone mineral association of the Sebeş-Lotru rocks. The latter appear as relict porphyroclasts (kyanite, plagioclase, muscovite, biotite, hornblende) in a neoformation matrix of quartz, albite-oligoclase, new biotite. These features of the blastomylonites prove that the thermal régime required by the dynamic reorganization of the rocks corresponds to the biotite zone.

The nature of the boundary between the two mentioned units is subject to discussion, as it may represent either a metamorphic (biotite zone) overthrust (post S_2 for the Sebeş-Lotru Group and post S_0 for the Jidoştiţa Formation), or (less probably) an overturned unconformity obliterated by metamorphism (Iancu, Hârtopanu, 1982). Whatever its origin, this plane was subsequently folded, therefore the related biotite blastomylonites exhibit a complex outcrop pattern and are locally found within the Sebeş-Lotru sequence.

Corbu Garnet Blastomylonites — Almaj Mountains

In the South Almaj Mts a N-S trending shear zone extends over 15 km north of the Danube (Mărunţiu, Seghedi, 1983 a). This shear zone, 300—500 m wide, involves the Plavişeviţa Gabbros to the west and the polymetamorphic rocks of the Neamţu Group (Upper Precambrian) — plagioclase gneisses, mica schists, amphibolites, crystalline limestones — to the east. Mylonitisation increases progressively westwards in the Neamţu Group rocks, towards the tectonic contact with the Plavişeviţa Gabbros, but usually unaffected rocks grading into highly deformed ones (ultramylonites, phyllonites, blastomylonites) occur even in the most highly deformed area. The rocks of the shear zone underwent intense deformation, recrystallization and neomineralization. A penetrative mylonitic foliation is dominant in the shear zone, highly obliterating the earlier structural elements. Medium-grade metamorphic minerals — garnet, staurolite, sillimanite and andalusite, variously altered to pinnitic aggregates — are deformed and bent in microfold hinges preserved as relics of a previous metamorphic foliation ; the porphyroclasts obviously suggest a complex metamorphic history of rocks foredating mylonitisation. The mineral associations in mylonites (muscovite, chlorite, albite, quartz, garnets as minute, idioblastic or atoll crystals in terrigenous rocks ; albite-epidote-actinolite-chlorite-quartz in amphibolites ; uralite-actinolite-epidote-albite-quartz or actinolite-epidote-quartz in the Plavişeviţa "Epigabbros" and blastomylonites respectively) point to garnet grade temperature prevailing during mylonitisation (Mărunţiu, Seghedi, 1983 b).

Field and geochronological evidence suggests a Caledonian age for these deep seated mylonitic processes.

In the Cornereva area, the Corbu zone mylonites have been involved in Alpine thrusts (Mărunţiu, Seghedi, 1983 b) ; Alpine deformations resulted in cataclastic rocks with random fabric and thus polycyclic dynamometamorphosed rocks are common in areas where the Neamţu Group rocks occur.

Regional Retrogression in the South Carpathians

Excepting the middle part of the pile of Alpine nappes building the South Carpathians (the Getic Nappe respectively), the Precambrian medium-grade rocks are frequently retrogressed over large areas. In the Supragetic Units, many lithostratigraphic sequences long time described as epizonal have been proved to be polycyclic (Codarcea, 1931 ; Balintoni, 1969 ; Giuşcă et al., 1977 ; Hann, Szász, 1984 ; Iancu, in press ; Gheuca, Dinică, in press). Symmetrically, at the bottom of the nappe pile, some of the Precambrian formations of the basement of the Danubian Units were described as polycyclic (Pavelescu, 1953 ; Gherasi, Dimitrescu, 1969 ; Savu, 1970 ; Bercia, Bercia, 1970 ; Mărunţiu, 1976 ; Gunnesch, Gunnesch, 1978 ; Berza, Seghedi, 1983). Even in the Getic Nappe — the Borăscu retrogressed zone (Gherasi, 1937) and the North Sebeş Mts retrogressed belt underlying the Cibin Group (Kräutner, 1980) — or in scales at its top (Uria Unit — Hann, Szász, 1984) or its sole (Borăscu Unit — Gherasi et al., in press), medium-grade crystalline schists are retrogressed to low-grade associations.

The regional retrogression is marked by both the development of low-grade minerals (quartz, albite, chloritie, sericite, tremolite-actinote, stilpnomelane, clinozoizite-epidote, prehnite, calcite, magnetite, pyrite) and superimposed folding, frequently cozonal with the preexistent structures. Index minerals (sillimanite, andalusite, kyanite, staurolite, cordierite) or paragenesis, preserved as relict metamorphic porphyroclasts, account for the preexistent metamorphic events of medium or high grade (Barrowian or low-pressure type). On a mesoscopic scale, retrogressed areas are characterized by the development of new folds and foliations, overprinting earlier structures.

Low-grade superimposed metamorphism is also responsible for an incipient metamorphic differentiation, as testified by quartz ± chlorite ± tremolite ± epidote ± adular veins. Some authors also claimed its significance as metallogenetic factor, as magnetite (Hann, Szász, 1984) and base metal sulphide (Giuşcă et al., 1978) concentrations in the Cibin and Făgăraş Mts show paragenetic and structural features common with the retrogressive event.

The degree of retrograde regional adaptation is unhomogeneous and diversified : in some areas the retrogression is uncomplete, older minerals and structures coexisting with the new ones, and in other areas the predominance of neoformation minerals associated with new very penetrative foliations lends an obvious greenschist facies appearance to the rocks. This type of retrogression was ascribed to regional low-grade Paleozoic (Early Caledonian or Variscan) metamorphism, progressive in respect to the Paleozoic formations, but regressive as regards the basement (Balintoni, 1969 ; Savu, 1970 ; Berza, 1975 ; Kräutner, 1980 ; Iancu, in press).

The regionally retrogressed Precambrian sequences were frequently involved in Alpine or pre-Alpine thrusts, locally undergoing additional dynamic metamorphism (Uria Formation, Rîuşorul Formation, parts of the outcrop area of the Drăgşan Group, Corbu and Vodna "phyllites", etc.). These areas are difficult to study, but modern devices of metamorphic petrology and structural geology have enabled contemporary researchers to reconstitute their complex history, thus achieving more realistic structural models of the South Carpathians.

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Pétrologie-Minéralogie

MINERALOGY OF ALPINE VEINS FROM THE ROMANIAN CARPATHIANS

BY

EMIL CONSTANTINESCU [1], GAVRIL SĂBĂU [2]

The Alpine veins, raising interest for their mineralogical and textural peculiarities, are well developed in the Alps (Parker, 1954), the Sudetes (Ansilewski, 1958), the Rhodopes (Kostov, 1965). This paper describes some new occurrences within the Romanian Carpathians. The Alpine veins which were examined consist of some various mineralogical assemblages, being encompassed within different metamorphic rock types.

1. The assemblage: quartz-adularia-chlorite-apatite-actinolite, within migmatites and laminated granites (Parîng Mts : the Jiu Strait, the Gruniu Brook).

2. The assemblage : quartz-chlorite-albite-actinote-calcite-hematite within amphibolites and amphibolic schists (Făgăraş Mts : the Cheia Valley, the Gălăşescu Peak, Cîineni).

3. The assemblage: quartz-albite-chlorite-actinolite-epidote ± rutile

4. The assemblage : quartz-chlorite-pyrophyllite-paragonite/muscониu), and paragneisses (Leaota Mts : the Brusture Brook).

The assemblage : quartz-chlorite-pyrophyllite-paragonite/muscovite-chloritoid within chloritoid bearing pyrophyllitic schists (Parîng Mts : Izvorul, Jieţ).

The Alpine veins are lenticular in form or appear as irregular nests usually transversally oriented as compared to the schistuosity planes and very seldom along schistuosity (Fig. 1). Cavities are partly or rarely almost completely filled up, sometimes exhibiting a zoning structure (Fig. 2).

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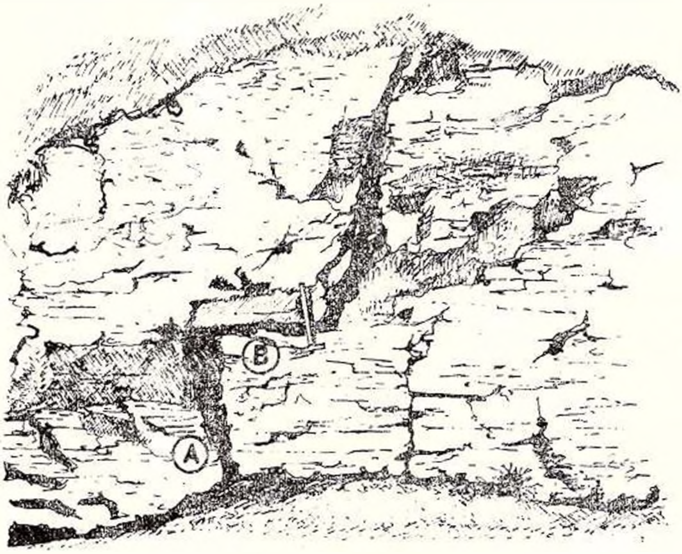

Fig. 1 — Alpine vein "en échelon" with partly unconformable setting: (A), partly concordant (B) in migmatites (Parîng).

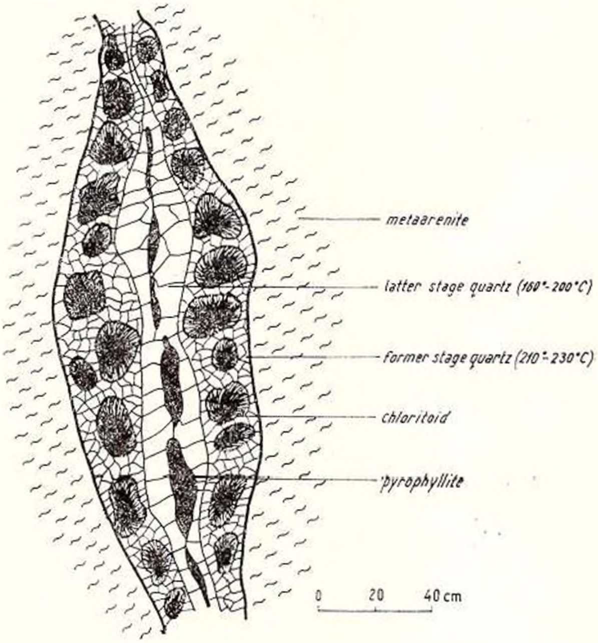

Fig. 2 — Quartz, chloritoid and pyrophyllite vein with zoning structure in the Schela Formation (Parîng), after Popescu (1983) (unpublished data).

Description of Main Minerals

The composing minerals were analysed by a complex mineralogical study in order to find out their morphological aspects, their optical characters and their chemico-structural peculiarities.

Quartz is present in all assemblages, quantitatively dominating the other minerals. Quartz crystals can reach remarkable sizes, up to 20 cm long. Its habitus can be columnary — 20/5 cm (Gruniu) or shortly prismatic — 20/15 cm. Faces were identified (10$\bar{1}$0), (10$\bar{1}$1), (01$\bar{1}$1), (3$\bar{1}\bar{2}$1), (2$\bar{1}\bar{1}$1), as well as some rare forms (05$\bar{5}$2), (30$\bar{3}$1), (03$\bar{3}$1) (Fig. 3). Face combinations are richer (Gruniu) or simpler (Leaota). Crystals are colourless and glasslike, milky white, smoky or green. Under microscope, some chlorite inclusions, marginally disposed and in parallel to crystal faces (Figs. 4, 5), or fibrous actinote in the crystal mass were noticed within green crystals. In some cases they can show fluid inclusions (Jieţ) ;

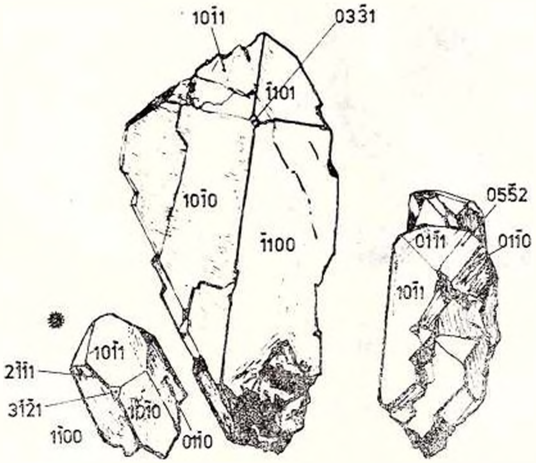

Fig. 3 — Morphology of quartz crystals from Alpine veins (Parîng) (coll. R. Strusiewicz).

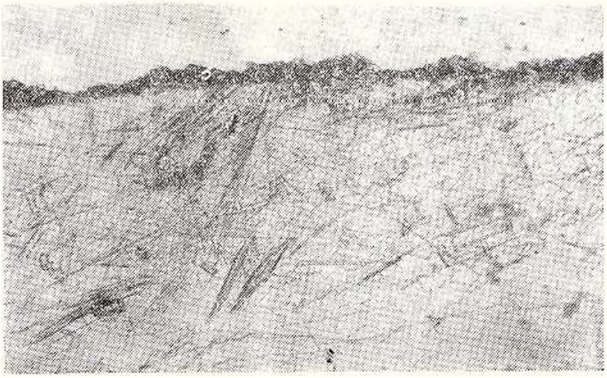

Fig. 4 — Setting of chlorite and actinote inclusions within greenish quartz (thin section, N //, 10 ×).

here, the temperature of the fluid phase homogenization was measured on these inclusions (150 determinations) ; for formation temperatures there were obtained some values between 160—180°C (Popescu, Constantinescu, 1982).

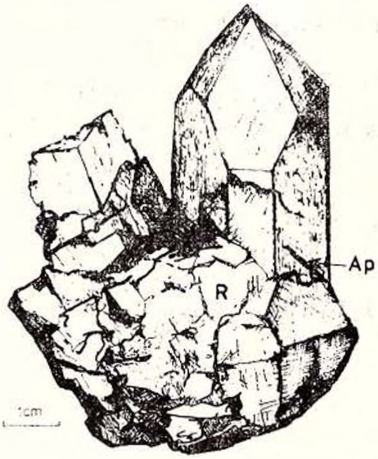

Fig. 5 — Assemblage (1) quartz-adularia-ripidolite-apatite (Parîng).

Chlorite is the only mineral found together with quartz in all assemblages. Its crystal dimensions are between 1 mm and 1 cm. It shows a lamellar habitus and is disposed in radial aggregates (Fig. 6). Its colour varies from yellowish-green to dark green. Taking into account optical properties and the chemico-structural determinations, the ripidolite and clinochlore could be separated.

Ripidolite is characteristical for assemblage 1 (Parîng Mts, the Jiu Strait and the Gruniu Brook) (Fig. 5), where it appears within agglo-

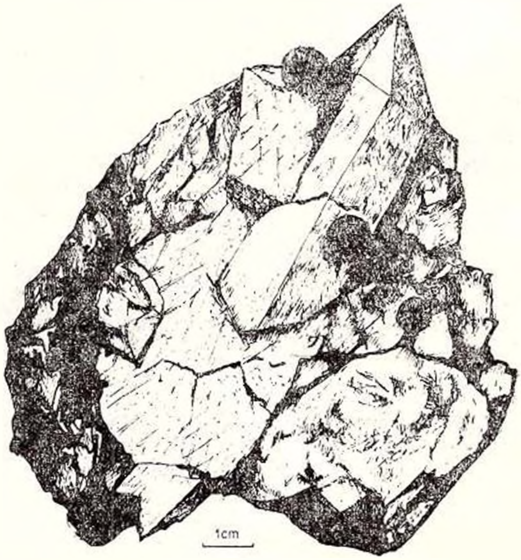

Fig. 6 — Chlorite rosettes associated with quartz and calcite (assemblage 2) (Făgăraș).

merations of crystals with a helminthic structure (Fig. 7), single crystals having a prismatic habitus. It is weakly pleochroic from yellowish to green and its birefringence colour is dark grey, often anomalous with bluish shades. Clinochlore which appears within assemblages 2 and 3 is set out in rosettes with undulose extinction (Fig. 8) or in dispersed isolated crystals. Its pleochroism is more intense with chlorites in assemblage 2 and weaker with those in assemblage 3. Its birefringence

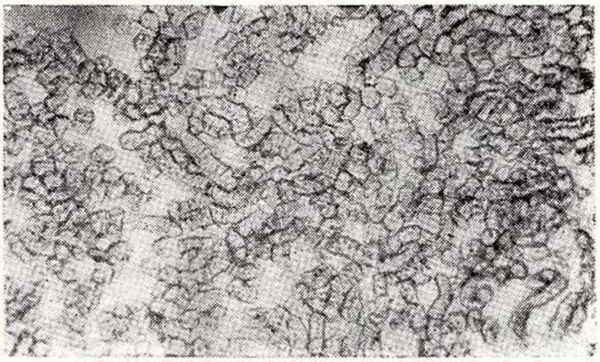

Fig. 7 — Heimintic aggregates of ripidolite (thin section, N //, 80 ×).

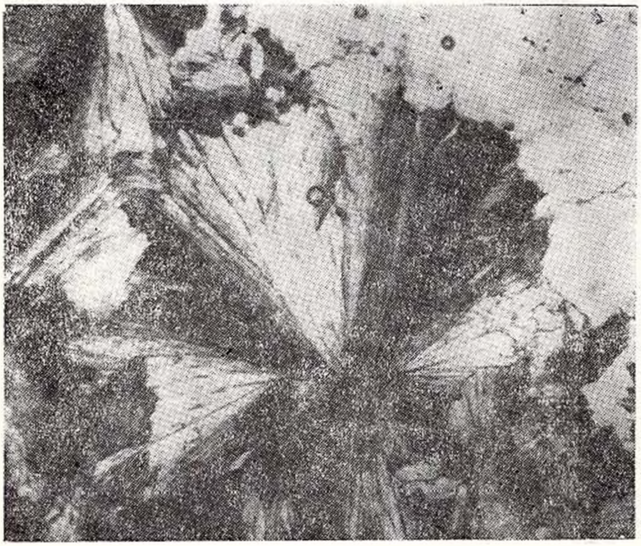

Fig. 8 — Radial aggregate of chlorite with undulose extinction (thin sectionn, N+, 40 ×).

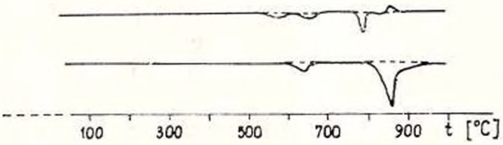

Fig. 9 — Thermodifferential curve of ripidolite (Parîng).

colours are usually anomalous (yellow-brown, violaceous), rarely normal but very low (dark grey).

The precise identification of mineral species could be carried out by means of RX diffraction, IR spectrography and thermodifferential analyses (Fig. 9).

Adularia appears only in assemblage 1 (Parîng Mts, Gruniu) associated with quartz, ripidolite, apatite (Fig. 3). Its crystals have a shortly prismatic habitus and dimensions between 1—4 cm. Faces are well represented $(1\bar{1}0)$, (110), (001), (201), with a larger development of face (201), in prejudice of face (001) (Fig. 10). Twins often appear after Manebach, Baveno and pericline+albite laws. Microscopical observations have indicated an optical non-homogeneity marked by a variation of the 2V angle, between 67—83° and of the extinction angle $\widehat{CNm}$ between 12—17°, in different crystal zones. There were noticed some zonal structures, both macroscopically by colour variations, and microscopically by the existence of some lamellae having a different optical behaviour (Fig. 11).

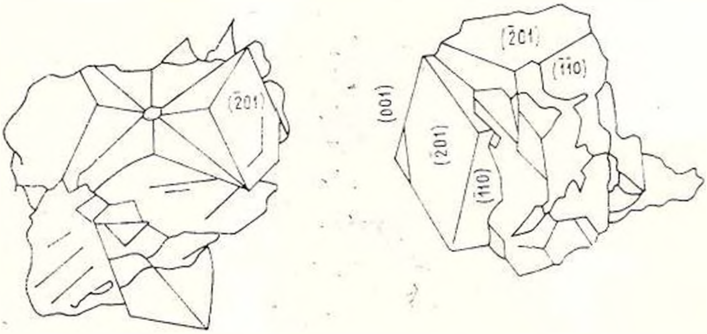

Fig. 10 — Morphology of adularia crystals (Parîng).

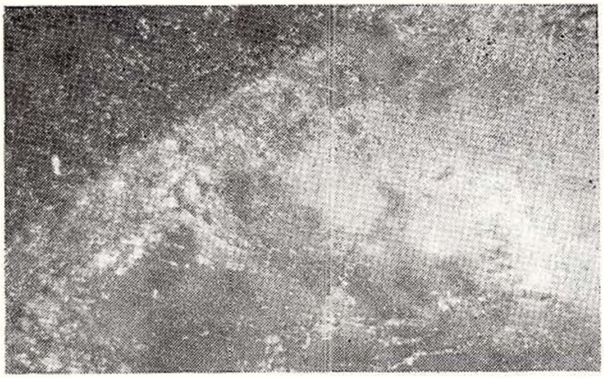

Fig. 11 — Lamellar and diffuse optical non-homogeneities in adularia (thin section, N +, 40 ×).

Chemical analyses (Tab. 1) allowed the calculation of the structural formula $K_{3.092}$ $Na_{0.245}$ $Ba_{0.100}$ Ca_{0022} $Si_{11.901}$ $Al_{4.427}$ O_{32} and the participation ratio of final terms from the feldspar group : F_k 89%, Ab 7%, Cs 3%, An 1%.

The d/n values and the I/100 corresponding intensities, calculated according the main diffraction lines are included in Table 1. The triclinity calculation has given the value = 0.8 and the sometimes diffuse character of 130 reflexes points out the non-homogeneous character of the adularia from the Alpine veins from the Paring Mts with monoclinic and triclinic symmetry zones.

TABLE 1

Chemical analyses of minerals Alpine veins

Minerals Oxides	Adularia Gruniu, Paring Mts	Pyrophyllite Jiet, Paring Mts
SiO_2	64.20	64.19
TiO_2	—	0.01
Al_2O_3	19.26	28.77
Fe_2O_3	0.11	0.40
FeO	0.09	—
MgO	0.05	0.14
CaO	0.12	0.16
BaO	1.42	—
Na_2O	0.70	0.14
K_2O	13.04	0.20
H_2O	0.70	5.73
	99.69	99.74

The main absorption bands in infrared which were obtained on the analysed adularia are characteristic for the stretching frequences Si(Al)-0 (1045 cm^{-1}) ; Si-Al-Si (728 cm^{-1}) and for the deformational frequences Si-O-Si (433 cm^{-1}). The value of transmission minimum from 645 cm^{-1} occupies an intermediary position among characteristical values for sanidine 639 cm^{-1} and microcline 650 cm^{-1}, indicating an intermediary state of order, as the other band positions.

Albite appears in assemblage 2 and 3 in association with quartz-chlorite-actinote-calcite-hematite, and quartz-chlorite-actinote-epidote respectively. Its crystals are up to 4—6 mm and form monomineral aggregates where only the terminal faces and the prismatic habitus, elongated after the lateral pinacoid (010) can be distinguished. Under microscope it shows a subhedral outline.

The determined optical constants : the refraction indexes, the extinction angle and the 2V angle show a content of 5—10% An. While comparing the values of the d/n parameter and those of the I/100 intensity ratio calculated from the RX diffraction analyses which were made in the domain 2∅: 10—55°, with data obtained by Borg and Smith (1969) there comes out that our values correspond to those obtained for the low albite.

This is confirmed by projection of values of wavelengths of absorption bands in IR from domains 18.2—18.9 and 15.3—16.2 on Hafner and Laves' diagram (1957) which are included in the characteristic low temperature domain.

Actinote appears as macroscopically visible crystals (2—3 mm) in assemblage 2 and 3, associated with quartz-chlorite-albite-calcite or epidote, as microscopical inclusions within quartz and adularia in assemblage 1 and as asbestiform aggregates (assemblages 2 and 3).

Prophyllite appears in assemblage 4 associated with quartz, chlorite and paragonite/muscovite. Its colour is pearly white. In thin sections it appears as scales and elongated leaflets with an anhedral outline, generally having submillimetric dimensions. It is colourless or with a slightly greenish shade. Its birefringence colours vary from grey to greenish-yellow. The extinction angle c^Np = 8—10°. The angle 2V determined in basal sections varies between 54—58°.

Its chemical composition is given in Table 1, providing the following structural formula : $K_{0.031}$ $Na_{0.033}$ $Ca_{0.021}$ $(H_3O)_{0.199}$ $Mg_{0.025}$ $Fe^{3+}_{0.036}$ $Al_{3.803}$ $Si_{7.728}Al_{0.272}O_{20}(OH)_{3.801}/24(O)/$. The values of interplanar distances determined by RX diffraction are shown in Table 2 and the thermodifferential curve in Figure 12. The crystallinity index, considered to be a sensible indicator of phyllosilicate genesis (Dunoyer de Segonzac, 1969) was calculated according the ratio between the height of the peak corresponding to the (002) reflex and its width at half-hight. The obtained value, —122, is higher for the crystallinity index of pyrophyllite from pyrophyllitic schists —47 (representing the average of five analyses) after Popescu and Constantinescu (1982) and very high as compared to the value obtained for hydrothermal pyrophyllites from Romania —8 (Ianovici et al., 1981).

Together with the above mentioned minerals, the Alpine veins contain as well some locally spread or quantitatively subordinated minerals :

Calcite appears in assemblage 2 as crystals with rhombohedral habitus up to 2 cm with a cleavage perfectly following the colourless

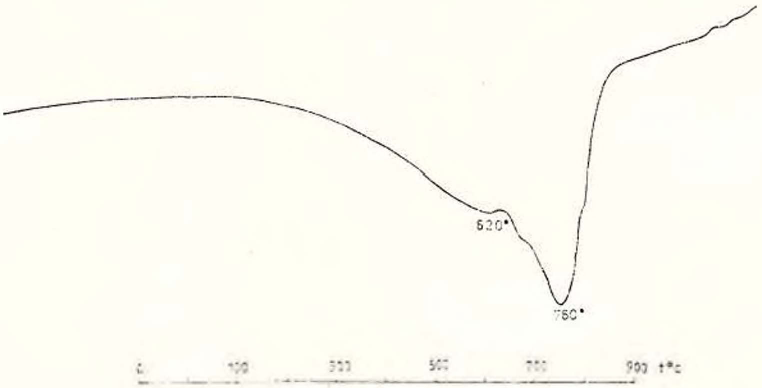

Fig. 12 — Thermodifferential curve of pyrophyllite (Parîng).

TABLE 2

d/n in Å and I/100 values obtained by RX diffraction for minerals of the Alpine veins

1. Adularia			2. Pyrophyllite				3. Ripidolite			
Nr.	d/nÅ	I/100	Nr.	hkl	d/nÅ	I/100	Nr.	hkl	d/nÅ	I/10
1	3 .93	10	1	002	9 .0544	49 .5	1	001	14 .10	40
2	3 .766	15	2	004	4 .5344	7 .2	2	002	7 .08	100
3	3 .45	10	3	006	3 .0455	100	3	003	4 .71	10
4	3 .29	20	4	132	2 .4012	3	4	004	3 .54	20
5	3 .229	100	5	008	2 .2905	10	5	005	2 .82	10
6	2 .985	35	6	0 .0 .10	1 .8391	22	6	131 .20$\bar{2}$	2 .59	10
7	2 .889	10	7		1 .5289	23	7	132 .202	2 .54	10
8	2 .760	10	8		1 .3799	6	8	132 .20$\bar{3}$	2 .45	15
9	2 .55	25	9		1 .3644	7	9	13$\bar{2}$ .202	2 .38	10
10	2 .319	10					10	133 .20$\bar{4}$	2 .25	10
11	2 .154	10					11	135 .204	2 .00	15
12	2 .119	10								
13	2 .049	10								
14	1 .887	45								
15	1 .795	35								
16	1 .615	10								
17	1 .449	15								
18	1 .406	10								
19	1 .382	10								

rhombohedral faces or in masses having a granular aspect, greyish-white in colour.

Epidote appears as grains or shortly prismatic crystals, sometimes with curved faces (Fig. 13).

Hematite appears in assemblage 2 as lamellar crystals, iron-grey coloured, sometimes disposed in aggregates with a radial aspect.

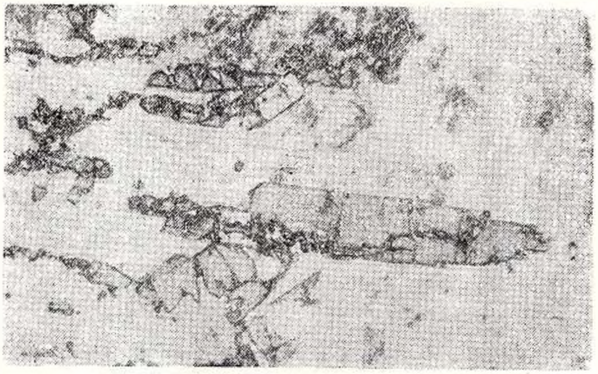

Fig. 13 — Epidote and apatite crystals included in adularia (thin (section, N //, 10 ×).

Conclusions

The mineralogical composition of the Alpine veins from the Romanian Carpathians is a simple one and in all the examined assemblages it is closely related to the mineralogical composition of the host rocks. The development of certain minerals is controlled by a specifical petrographic environment. Thus, the adularia exclusively appears in gneissic rocks (migmatites, laminated granites) and the pyrophyllite exclusively appears within pyrophyllitic schists. Chlorites, which appear in all assemblages are characterized by the presence of the ferriferous species (ripidolite) within the Alpine veins of gneissic rocks and of the magnesian species (clinochlore) within the Alpine veins encompassed in amphibolites and amphibolic schists. Hematite appears as well only in veins of the amphibolic schists as it is formed according to iron levigated from amphiboles. These observations, correlated to the partial depletion of the host rocks in chemical elements which form the minerals of the Alpine veins, support the idea of the formation of these minerals by lateral secretion processes.

The reciprocal relationships among minerals of the Alpine veins which were macroscopically noticed, indicate a depositional succession of these ones, which allows separation within certain veins (assemblages 1—3) of an initially alkaline stage, represented by the crystallization of quartz, adularia and albite, followed by a calc-alkaline stage, when actinote and epidote are formed. Two crystalization stages can be distinguished as well in assemblage 4; a) chloritoid + quartz and b) pyrophyllitie ± quartz of lower temperature (Fig. 2).

The crystal morphology for quartz and adularia, the chemico-structural characteristics (order-disorder etc.) for albite and adularia, and the values of homogeneity temperatures for fluid inclusions of quartz indicate medium to low temperatures of formation.

The mobilization and redeposition of these minerals are associated with the action of metatectical fluids which have circulated through open fractures within metamorphic rocks, after their deformation.

Therefore, the Alpine type veins represent the final manifestations of regional metamorphism processes from the Romanian Carpathians.

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* * * X-ray Powder Data File. *ASTM*.

Pétrologie-Minéralogie

EQUILIBRIUM GEOSURFACES DYNAMICS OF MINERAL PARAGENESES

BY

ION HÂRTOPANU[1], MARIN ŞECLĂMAN[2]

Study Problems

The metamorphic zones with sillimanite and kyanite are often found in superposition relationships within the South Carpathians : the kyanite zone overlies the sillimanite zone (Savu, 1970 ; Bercia, 1975) (Fig. 1).

The limit separating the two zones is a surface which turns from the horizontal position mainly because of foldings and numerous fractures belonging to the alpine orogenesis. The most plausible explanation of this superposition seems to be due to the peculiar control of depth over the metamorphic factors, mainly over temperature and pressure. Miyashiro (1973) seems to be the first who has drawn the attention upon the fact that peculiarities of different series of facies found in various parts of the world are due to the different values of geothermal gradients during the metamorphism time. As we agree that the vertical succession of the kyanite and sillimanite zones is essentially caused by the geothermal gradient, we can see that the metamorphic zone succession from the Godeanu and Semenic Mts is possible only if the geothermal gradient had a value up to 50°C/km during metamorphism (Fig. 2). A higher gradient would have caused an interposition of the andalusite zone between the kyanite and the sillimanite zones, a fact that has not been noticed yet within the above mentioned massifs. The minimum value of the geothermal gradient must have been of about 35°C/km. Without this value, the rocks must have been found in a partly melting state even in the kyanite zone, which is not found in field. The conclusion is that a geothermal gradient having reasonable values between about 35°C/km and 50°C/km explains well enough the superposition of kyanite and sillimanite zones in some alpine massifs from the South Carpathians.

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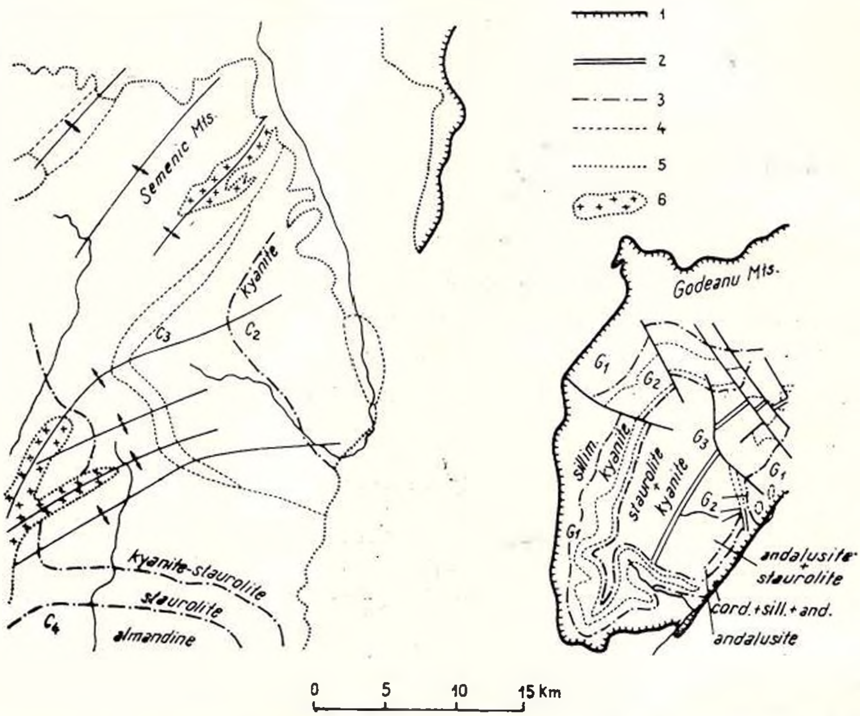

Fig. 1 — Relationships of metamorphism isogrades with lithostratigraphic limits (after Savu, 1970 ; Bercia, 1975).
1, overthrust line ; 2, limit between different baric types of metamorphism ; 3, isogrades of metamorphism ; 4, limit of lithologic complex (formation) ; 5, transgression limit ; 6, granitic body.

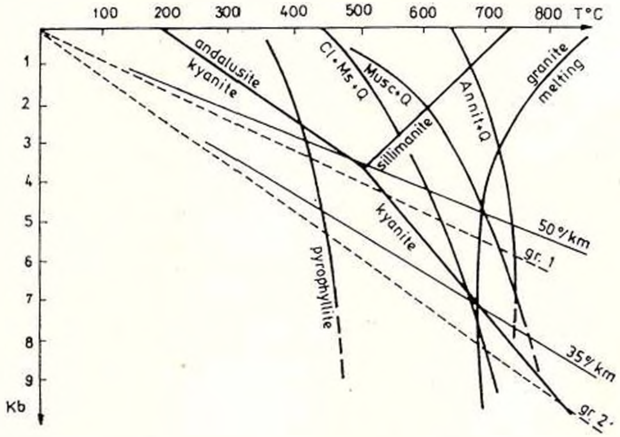

Fig. 2 — P-T space of metamorphism from the Godeanu and Semenic Mts overlined by mineral equilibrium curves and by geothermal gradient of 35°C/km and 50°C/km.

A general conclusion is that if temperature and pressure — which are the main factors within many metamorphic complexes — depend on depth, the various metamorphic zones which are formed must be found in spatial superposition relationships, namely these zones must be found ones under the others, so that the metamorphic degree, in a given place, would grow according to depth. Within the tridimensional geological space, the separating limits among zones could represent some more or less irregular surfaces. According to Tilley (1925) we could call these limits as isogrades. Anyhow, if we admit that a certain limit separates metamorphic zones with stable parageneses, we think that the term of "equilibrium geosurface" is more suitable. Within a metamorphic zone, the mineral assemblage is in a bivariant equilibrium (if the main factors of equilibrium are temperature and pressure), while the equilibrium geosurface marks the monovariant condition from the geological space and represents the geometric space of all points of the geological space where the metamorphic assemblages of the two superposed zones can coexist in equilibrium.

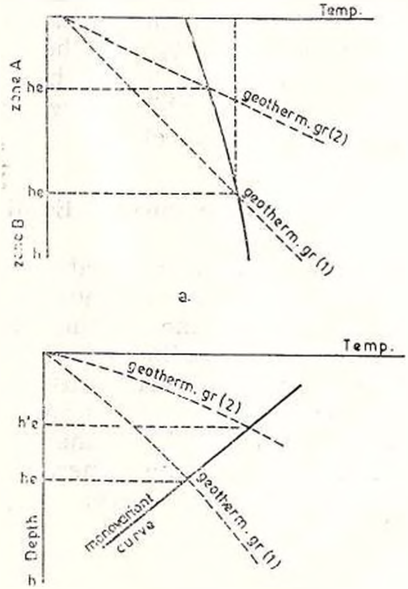

Fig. 3 — Change of equilibrium depth according to the slope sign of the equilibrium monovarying curve ; a, positive slope ; b, negative slope.

The equilibrium geosurfaces form and position are mainly controlled by the geothermal gradient.

In a certain point of the metamorphic area, the equilibrium depth between two zones is determined by the intersecting place of the monovarying equilibrium curve with the geothermal gradient curve. In Figure 3a this depth is noted with h_e. At $h > h_e$ only the assemblage B is stable and at $h < h_e$ only the assemblage A is stable (in the limit case when the geothermal gradient is the same in the whole metamor-
cated on a horizontal equilibrium geosurface). But the geothermal
phic space, the h_e depth on a given vertical corresponds to a point lo-
gradient's inconstancy within the metamorphic area (which is expressed

by the variation of temperature in the horizontal plane) entails the changing of the h_e depth from one place to another. In Figure 3a one can see that at a higher geothermal gradient (noted by 2), the equilibrium depth is lower (h_e). In exchange, in Figure 3b, where the monovariant equilibrium curve has a negative slope, it is shown that the equilibrium depths diminish as the geothermal gradient increases. Therefore, where the geothermal gradient changes from one point to another, the equilibrium depths are inscribed on an irregular equilibrium geosurface. In this case, the equilibrium geosurface is turned away from the horizontal ; as the deviation amplitudes are directly proportional to the horizontal geothermal gradients from the metamorphic space (it is to be noted that the equilibrium geosurface is not parallel to the geothermal surface).

Dynamics of Equilibrium Geosurfaces

A lot of observation data show that within one and the same point of the crust the geothermal gradient was modified with time. The variability in time of the geothermal gradient has large consequences upon the equilibrium geosurface, determining thus the vertical upward or downward shifting of this surface, according to the variation sense of the geothermal gradient and to the monovarying curve slope. Figure 4 shows the effect of the geothermal gradient lowering as a result of the general cooling of the metamorphic system from the metamorphic space. At the initial gradient, the equilibrium depth corresponds to h_1 but by the lowering in time of the geothermal gradient the depth got down to h_2. Within the space between h_1 and h_2 the assemblage A becomes metastable and must pass, by reaction, into the assemblage B. Thus, the mineral reactions can take place only within this depth interval which, in the bidimensional space is encompased between the old equilibrium geosurface (a fossil geosurface) which passed through h_1 and the new surface (real surface) which passes through h_2. We agree to call this space “reacting” (reacting space), taking into account the fact that only in this space a mineral reaction of thermodynamic accomodation to the time variation of the geothermal

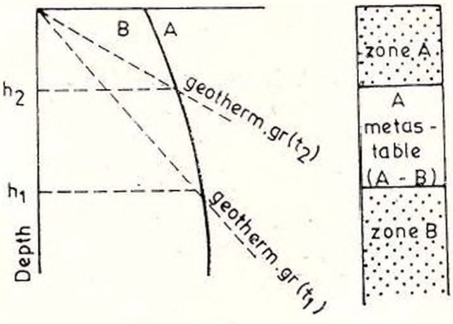

Fig. 4 — Geothermal gradient increase determines a prograde metamorphism for equilibrium curves with positive slope.

gradient is possible. As during the geological time geothermal gradients were variable, it is very likely that, in reality, the separating limit between the neighbouring metamorphic zones would not be a surface (an isograde plane), but a reacting space having variable thicknesses where some reactions to replace one assemblage to another there took place. The separating limit between the kyanite and the sillimanite zones from the Godeanu and Semenic Mts is a good example to confirm the real existence of the reacting space. Close to the equilibrium geosurface of the two zones (to the isograde line) a paragenetic superposition is remarked, showing clear relationships of partial replacement of the kyanitic assemblage with the sillimanitic assemblage and inversely, a proof of the two ways oscillation of the geothermal gradient.

Final Remarks

It is very important that between the metamorphic zones there interposes an equilibrium geosurface with monovarying conditions. Any oscillation in time of the caloric fluxes affect in the first place the space near the equilibrium geosurface which becomes thus a reacting space. The sense of the metamorphic reactions within the reacting space depends not only on the variation sense of the geothermal gradient, but also on the mineral origin of reactions. In most cases, the mineral reactions have positive enthropies and reaction volumes and correspond to a positive slope of the equilibrium curves in T-P diagrams. In this case, the increase of the geothermal gradient determines in the reacting space a prograde metamorphism (the lower temperature assemblage passes into the higher temperature assemblage) (Fig. 4). But there are some cases when the reaction curves slopes are negative (e.g. the dehydration reactions at very high pressures or the probable reactions of transformation between andalusite and sillimanite). In these situations the increase of the caloric fluxes determines retromorphous mineral reactions (Fig. 5) within the reacting space.

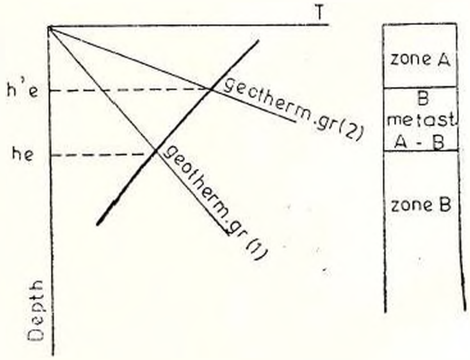

Fig. 5 — Geothermal gradient increase determines a retrograde metamorphism for equilibrium curves with negative slope.

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COMPLEX CRITERIA OF SEPARATING WEAKLY METAMORPHOSED FORMATIONS. AN EXAMPLE : THE SOUTH CARPATHIANS

BY

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The present paper deals with some lithostratigraphical entities which were affected by a low grade metamorphism and which could be well individualized within the Paleozoic-Mesozoic succesion from the Mehedinţi-Retezat Danubian Unit (Stănoiu, 1973) of the South Carpathians (Fig. 1).

These paleontologically dated entities point out the effects induced by deformation and metamorphism and constitute some guide marks to clear up the lithostratigraphy of the Paleo-Mesozoic metamorphosed formations.

The lithostratigraphical separation of weakly metamorphosed formations from the Carpathian Orogene area implies some hard difficulties caused both by the discontinuous development on a Precambrian basement reactivated during the Paleozoic, and by the non-homogeneity of the deformation and metamorphic degree within the Alpine structures.

The areas with a progressive, orogenic, low grade metamorphism are isolated and point out some non-homogeneities and discontinuities during the same phase. At the same time, some formations show clear effects of overlapping folding (either polyphasic or polycyclic) sometimes accompanied by a partial mineralogical reorganization or by shearing due to overthrusts.

Taking into consideration the complex character of the implied phenomena, the synchronous formations can show a different deformational and metamorphic evolution and therefore some different structural-mineralogical aspects. At the same time, some differently aged formations (either Paleozoic or Mesozoic) but with a similar premetamorphic lithology, can get, by metamorphism and deformation, some convergent structural-petrographic aspects.

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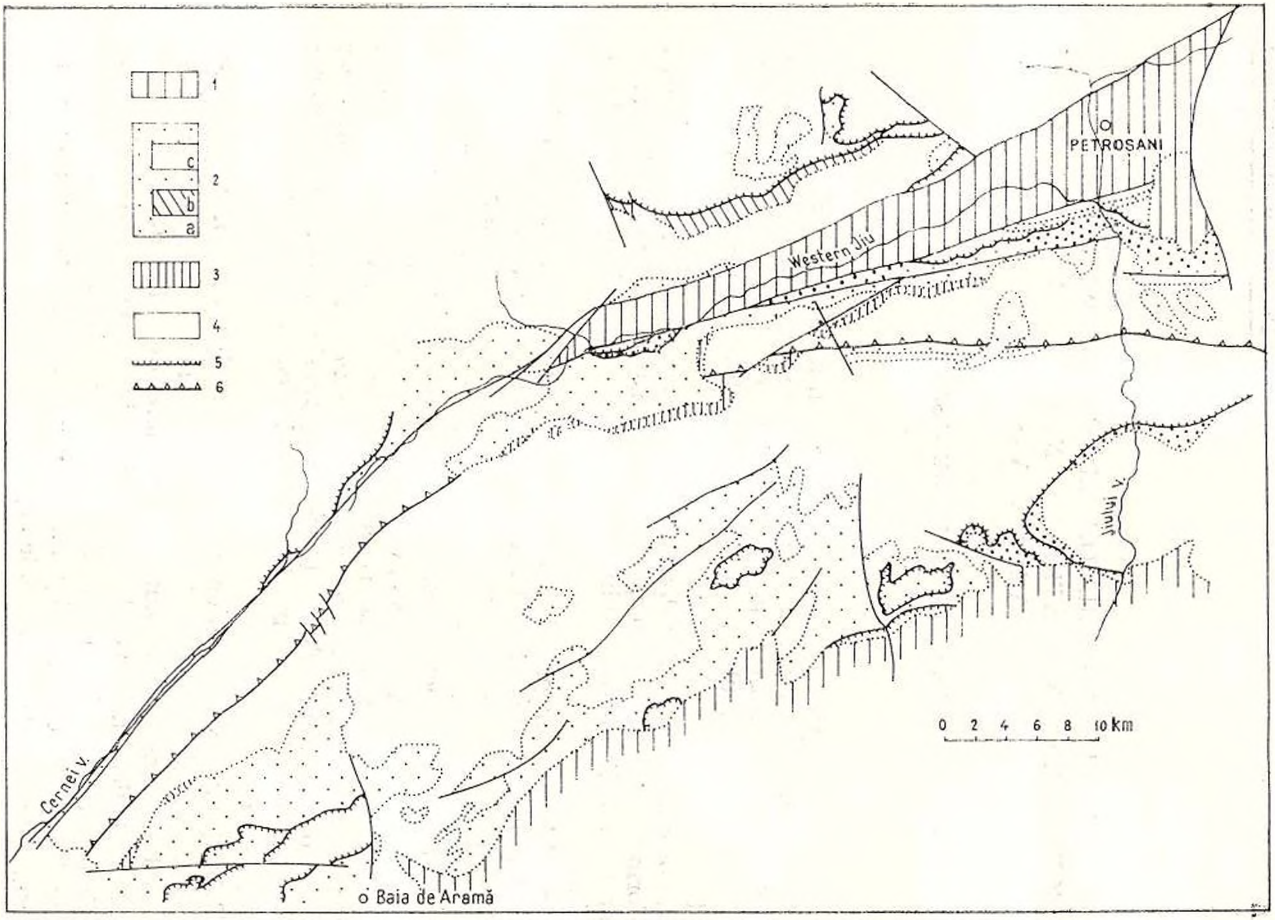


Fig. 1 — Geological sketch of the Vîlcan Mts (South Carpathians).
1, post-Laramian cover ; 2, a, non-differentiated Paleo-Mesozoic cover (post-Ordovician-Silurian) ; b, Upper Paleozoic formations ; c, Lias formations with anthracite and metamorphic minerals (chloritoid, pyropyllite, graphite) ; 3, Lower Paleozoic formaions ; 4, Precambriar basement ; 5, overthrusting line ; 6, pre-Alpine tectonical line.

These aspects explain why some Paleozoic and Mesozoic rock sequences were considered together as being unitary in time and space (e.g. the "Tulişa Series", the "Schela Formation", etc.).

As the subject of our paper is not a revision of the existing stratigraphical schemes and of the evolution in the terminology used for Paleo-Mesozoic formations, we shall deal with the low grade metamorphic formations which have paleontological dating.

1. Stratigraphical Criteria

Among the known and applied stratigraphical criteria for individualizing lithostratigraphical entities, we must mention :

— the facies unit and the establishment of lithostratigraphical successions and of real thicknesses (mainly within multistratified sequences) as compared to the initial stratification (So), which in most cases is modified or blurred by deformations, sometimes superposed.

— the paleontological and micropaleontological content.

When there are some evident variations in the same main structural unit, it is necessary to follow the facies variations, the initial thicknesses and the areas with visible effects of orogenic compression (by folding) accompanied by a mineralogical reorganization under stress conditions.

This paper deals with some formations with low grade metamorphism of Paleo-Mesozoic age.

1.1. *Lower Paleozoic*

Within the Lower Paleozoic (pre-Devonian), the Valea Izvorului and Coarnele formations were lithologically individualized and paleontologically dated (Stănoiu, 1976).

Lithologically, they consist of a lower, mainly quartzitic member and of an upper, mainly schistous member (chloritous ± sericitic ± quartzous schists with lenticular intercalations of graphitic schists and limestones).

The Valea Izvorului Formation has supplied a tabulata fauna (favositins and halisitids), tetracorals, bryozoans (fenestelids), brachiopods, trilobites (*Flexicalymene* sp., *Encrinus* sp. or *Cromus* sp.) and crinoids (*Caleidocrinus artifex*) which belong to the interval Ordovician-Lower Silurian (Stănoiu, 1976).

From both formations, Iliescu (in Stănoiu, Iliescu, 1976) and Visarion have determined a palyno-protistological association with acritarches "scolecodonts" and microspores of Ordovician-?Silurian type. Within the rocks belonging to the Coarnele Formation, Visarion (in Solomon et al., 1976) quotes some spores within the Middle Cambrian-Ordovician.

1.2. *Upper Paleozoic*

There are some continental, detrital formations representing the Variscan molasse facies attributed by Stănoiu and Lejal Nicol (in press) to the Westphalian-Autunian interval by reconsidering a flora assemblage previously attributed to the Upper Devonian (Stănoiu, 1976). There were separated (Stănoiu, Lejal-Nicol, in press) : the Valea de Brazi Formation with Westphalian plant remains and the Culmea Bradului Formation which was attributed to the Autunian because of its stratigraphical superposition and of its facial similarity with some other zone formations.

The Valea de Brazi Formation (Westphalian-Saephanian) is well represented in the eastern part of the Retezat Mts ; it is about 50—100 m thick and it consists of sericito-graphitic metapelites, metapsammites and metapsephites. Some spores characteristical for the Upper Carboniferous (*Florinites* sp., *Endosporites* sp., *Alisporites* sp., *Potoneisporites* sp.) together with Devonian spores (probably reworked) were identified by Visarion within the rocks of this formation.

The Culmea Bradului Formation (Autunian) is formed of metapsephites, metapsammites and reddish or grey metapelites, associated with acidic metavolcanics.

1.3. *Mesozoic*

For the purpose of this paper only certain Mesozoic formations were selected, i.e. those showing a low metamorphic grade with unreliable paleontological remnants ; they can be hardly separated from the similarly featured Paleozoic formations.

Within the formation of the Jurassic-Lower Cretaceous sedimentary cycle the Schela Formation (Manolescu, 1932) has been separated, i.e. a Lias formation of the Gresten type facies.

On the largest outcropping area of the Mehedinți-Retezat Unit, the Lias formation mainly contains psammo-psephitic deposits with a maximum stratigraphic thickness of about 50 m. A pelitic-psammitic facies, characterized by the presence of carbonaceous matter (anthracite) is developing in some allochthonous units (northern border of the Vîlcan and Parîng Mts and southern border of the Vîlcan Mts, Schela zone). Here, there were identified some Lias plants (Manolescu, 1937 ; Semaka, 1963 ; Stănoiu, 1982).

2. **Petrological Criteria**

The petrological criteria utilized by us are : 1) the petrographic and mineralogical assemblages, 2) the identification of metamorphic neoformation parageneses and 3) their correlation to some differentiated structural elements. More specialized studies were made concerning clay and opaque minerals, which can bring some indications about the way and the degree of metamorphic adaptation.

According to the above mentioned criteria, there were obtained some conclusive and convergent data for the previously presented formations.

Other Paleozoic and Mesozoic low grade metamorphic formations do not have unequivocal datings which makes difficult the time location of metamorphic events (e.g. the Oslea Formation, the Tusu Formation pro parte, etc.).

2.1. *Lower Paleozoic*

The rocks of the Valea Izvorului and Coarnele formations show a synmetamorphic penetrative foliation accompanied by stable minerals in the chlorite zone of the greenschist facies. The metamorphic neoformation minerals oriented in the S_1 plane are the following : sericite (illite), quartz, chlorite, albite, tourmaline, rutile. The pre-metamorphic sedimentary structures and the bedding (S_0) are obliterated by the metamorphic paragenesis, mainly within the Coarnele Formation. The Valea Izvorului fossiliferous Formation is less adapted to metamorphism (from the mineralogical point of view) and preserves more pre-metamorphic minerals.

X-ray data give a simpler assemblage of clay minerals : illite, chlorite. The abundance of chlorite is remarkable as compared to the other low grade metamorphic formations. The crystallochemical parameters of illite indicates a very good crystallinity, but associated to a variable ratio of intensity for reflexes 002/001. The chlorite parameters show a more ferriferous term.

Reflected light microscopy study of some polished sections showed the presence of rutile and graphite ; the latter shows a slightly variable optics and exhibits a plastic behaviour. The carbonaceous matter is lacking ; it is typical for the Lias formation only.

The deformations on mesoscopic scale in the Lower Paleozoic rocks are dominated by a general, very penetrative foliation (S_1) accompanied by a prograde mineral assemblage. The Precambrian basement synchronously underwent a retrograde adaptation which can be attributed to a late Caledonian phase.

2.2. *Upper Paleozoic*

The Valea de Brazi and Culmea Bradului Upper Paleozoic formations show simple microscopic deformations marked by decimetric to metric folds which affect a unique foliation (S_1) within the coarse grained rock sequences. This foliation is marked by the strong flattening — within the bedding plane (S_0) — of the polygenous elements from metaconglomerates. The matrix is metamorphically partly reorganized.

The associated microfolds which are often asymmetrical and flattened show overturned axial planes, marked by penetrating plane-axial cleavages within finer, interstratified sequences. These features suggest a superposed folding. The fine, metapelitic sequences show a "slate-like" cleavage, rarely with microfolds preserving bedding (S_0), and incipient transpositions of this one after S_1. Graded bedding are sometimes preserved within multistratified sequences. The assemblages contain premetamorphic minerals and metamorphic neoformation minerals (illite, quartz, chlorite, rutile, graphite) which are quantitatively subordinated.

The prevalence of sedimentary mineral assemblages, the small amount of neoformation minerals associated with superficial folds (specifical to the upper structural level) separate them both from the Lower Paleozoic formations and from the Mesozoic ones (mainly the Lias ones), suggesting the effects of a late Variscan deformation.

2.3. *Mesozoic*

The Lias formation (Schela) show mixed assemblages formed of pre-metamorphic minerals (sedimentary relicts — either clastic, reworked from the basement, or formed in the sedimentary environment or by diagenesis) and metamorphic neoformation minerals. The most frequent clastic minerals are : quartz, feldspars, micas (muscovite, biotite), tourmaline. These minerals are distinctly dominating within coarse rocks (metapsephites, metapsammites) and have visible effects of intracrystalline deformation. Metamorphic reorganizing of the matrix can partly be observed in the bedding plane. The sedimentary and diagenetic minerals are dominant within metapelitic and metasilty rock sequences ; under conditions of orogene metamorphism these rocks are mineralogically more intensely adapted and structurally featured after maximum stress directions.

The clay minerals of the Lias formation give the possibility to follow the transition phases from the burial diagenesis to metamorphism. The Lias formation is characterized by a wide mineralogical variability of the argillaceous fraction, although illite is predominant. All samples contain small amounts of kaolinite (a sedimentary "residual" mineral) and a lot of samples contain chlorite-vermiculite (a mixed-layer reflecting a transition to metamorphism). Illites have a wide variation domain of crystallinity indexes (Kubler, 1967). The ratios of intensity for reflexes 002/001 have a relatively narrow variation interval with a tendency to concentrate values within the aluminous illite domain (Fig. 2a), which allow a direct correlation with the metamorphic degree. The chlorites of the Lias formation have a variable ratio Fe/Mg which could suggest some progressive crystallo-chemical modifications connected to the metamorphism intensity (Fig. 2b).

The variation domain of the relatively wide crystallinity index can be the result of an incomplete and selective adaptation, when some minerals specifical to metamorphism (chloritoid, pyrophyllite) have a high frequency within the Lias formation. The microscopic observations have pointed out the mineralogical heterogeneity of petrographic assemblages (at a thin section level). Microstructural data underline an incomplete mineralogical restructuration and a "step by step" crystallization with static and dynamic phases. Chloritoid appears as rosettes and well individualized crystals, then slightly flattened and penetrated, suggesting a static crystallization before orogene compression deformations. This can be the result of a quick growing of lithostatic pressure in a pre-kinematic "orogene burial" phase (Iancu, unpublished data). The continuation of the mineralogical neoformation under dynamic conditions is underlined by synkinematically crystallized minerals (pyrophyllite, high cristallinity illite, quartz) in the S_1 foliation which is plane-axial and associated to mesoscopic folds (B_1). The S_1 foliation is then affected by shear cleavages (S_2) ; the chloritoid is sometimes occurring as lens-shaped aggregates between S_1 and S_2 planes (Fig. 3).

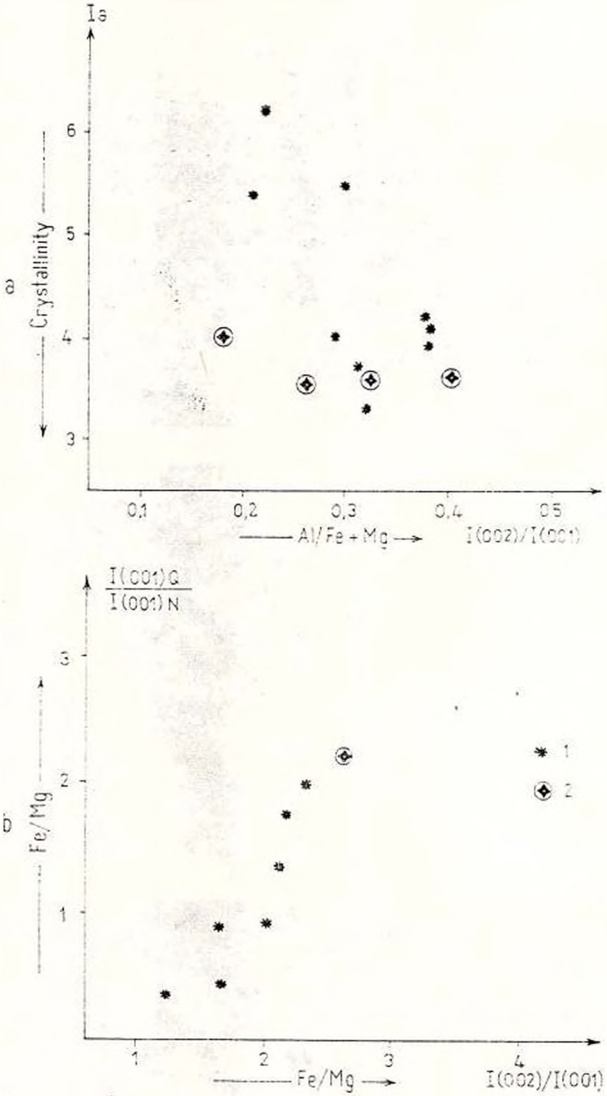


Fig. 2 — Relationships between crystallinity of illites and intensity ratio of reflexes 002 001 (a) ; crystallo-chemical characterization of chlorites (b).
1, Lias formation ; 2, Coarnele Formation. Microscopic aspects of the rocks of the Schela Formation (Lias).

A detailed study concerning the carbonaceous matter has pointed out the same transition and gradual adaptation of the carbonaceous matter from brittle anthracite (Fig. 4) to graphite through the intermediary stages of meta-anthracite and pre-graphite. The carbonaceous fragments have variable (morphological and optical) aspects, connected with their association to other minerals and with their position as compared to the S_1 plane. Some frequently homogenized petrographic components of anthracite (vitrinite, fusinite) can be found within thicker bands as well. Even in these cases, the colour type of bireflexion and anisotropy corresponds to graphite ; the diffraction peak

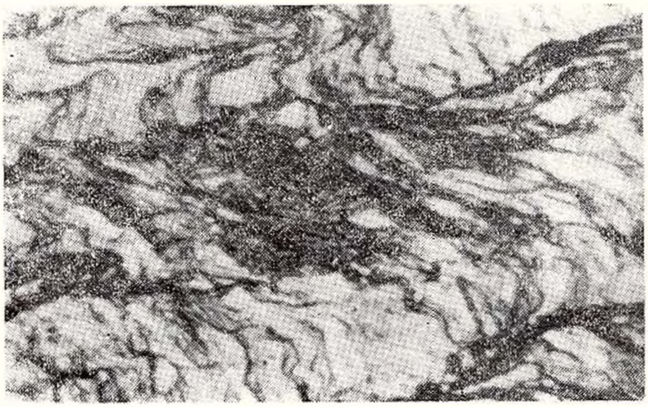
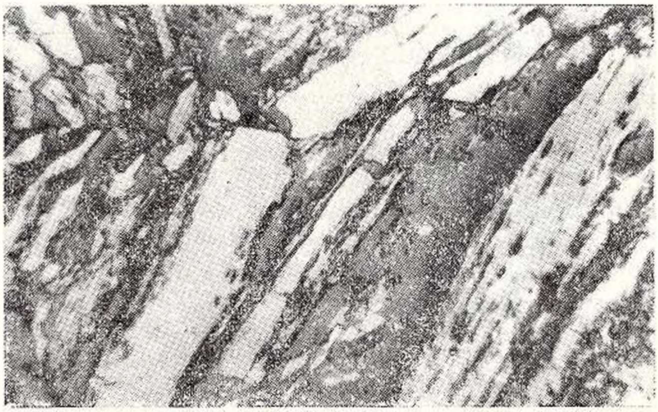


Fig. 3 — Pre-S_1 chloritoid rosette (Cd), between S_1 metamorphic foliation and S_2 cleavage. Thin section, N //, × 50.

Fig. 4 — Brittle behaviour of anthracite bands. Polished section (PS), N //, × 130.



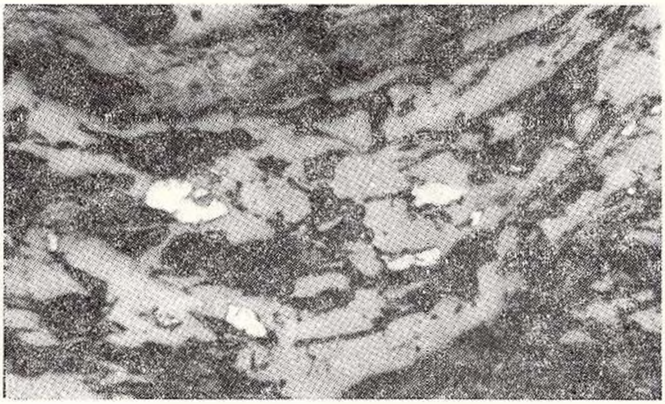
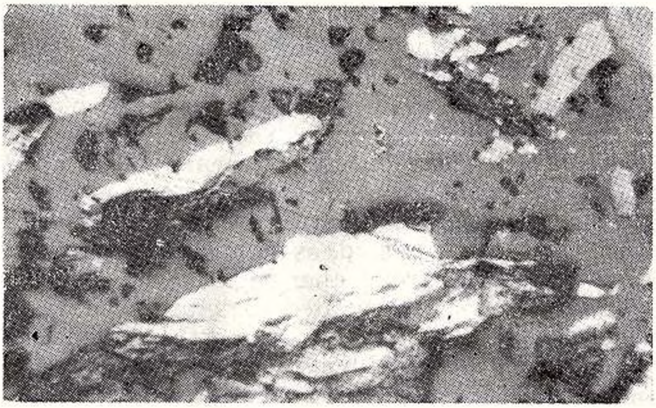


Fig. 5 — Anthracite "grains" enclosed in quartz remain optically unchanged. PS, N //, × 130.

Fig. 6 — Meta-anthracite showing marginal bands of advancing "lamellization" which behave somewhat plastically and become optically anisotropic. PS, N //, × 130.



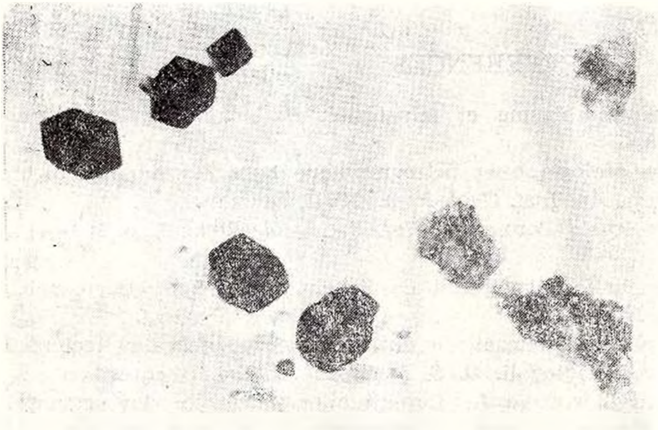


Fig. 7 — Transmission electron micrograph of the newly formed graphite. × 6 000.

at d = 3.34—3.36 Å is sharp, which confirms the beginning of a graphite-like structural organization of coal, confirmed as well by the electronomicroscopical images (Fig. 7). The thinner bands or the lenticular separation of coal are more deformed, getting some plastic properties (Fig. 6), their optics is already characteristical for pre-graphite (the stage following the meta-anthracite). The optics of the carbonaceous mass is highly variable even on the same polished section ; together with lens-shaped bodies exhibiting a plastic behaviour and the pre-graphite optics, there appear some fragments of meta-anthracite with very weak bireflexion and anisotropy, protected by quartz (Fig. 5).

The carbonaceous material is constantly associated with pyrite (sometimes pyrrhotine and rutile, relict or post-diagenetically crystallized minerals).

Thus, the Lias formation shows mineralogical proofs of reorganization after structural directions imposed by folding which are attributed to a pre-Laramian phase (Austrian phase). The existence of the post S_1 shear deformations as well as the allochtonous position (either within cover nappes, or as olistoliths within Upper Cretaceous formations) of formations from the western flank of the Vîlcan and Parîng massifs, can represent some effects of the Laramian phase.

We can conclude that the lithostratigraphic entities, separated according to stratigraphical and petrographical criteria, have specifical characters obtained through orogene deformation accompanied by a partial restructuration and by a metamorphic neoformation with a variable intensity, but at the same time in connection with some inherited characters (lithology, mineralogical composition, chemism, structures, etc.). On the other hand, as compared to the unmetamorphosed synchronous formations from other region, within the South Carpathian Danubian there can be separated some areas with a first low grade orogene metamorphism, due either to pre-Alpine phases (as for Paleozoic formations), or to Alpine phases (as for Mesozoic formations).

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INTERSTRATIFIED CLAY MINERALS IN THE HARGHITA MOUNTAINS, ROMANIA

BY

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Introduction

The argillization processes are very widespread in the Harghita Mts zone.

Most of the clay minerals are formed by hydrometasomatic alteration of volcanic rocks (andesites and their pyroclastites) and some of them by direct crystallization from hydrothermal solutions.

In the Harghita Mts, within the largely widespread hydrometasomatic transformations, there are all stages of transformation from fresh rock to completely argillized rock.

Neacşu and Urcan (1975), while studying the argillization processes of phenocrystals, drew the conclusion that in the initial argillization stages the solution supply was small, as the rock nature was the main factor in the formation of clay minerals.

In a subsequent study, Neacşu and Urcan (1978) deal with the argillization processes of volcanic rocks from the Harghita Mts in the intense alteration stages when the solution nature determined the type of the transformation products.

In a more recent paper (Setel et al., 1982) nacrite, pyrophyllite and zunyite are mentioned for the first time in this region.

The present paper is a study by X-ray diffraction of about 200 samples which allowed a new interpretation of argillization mechanism and a detailed study of mixed-layer minerals.

Hydrometasomatic Argillization Processes

The hydrometasomatic argillization processes from the Harghita Mts are marked by an intense activity of solutions, the formation of argillization zones being determined by their pulsating character ; this

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leads to three argillization stages : a kaolinic one, a hydromicaceous one and a mixed-layer one.

The Kaolinic Stage

The first stage of hydrometasomatic argillization is kaolinization in which the most frequent mineral is kaolinite T, associated with secondary quartz.

The paragenesis is formed of stratified minerals without alkali and with a high Al/Si ratio. Sometimes, nacrite, pyrophyllite and zunyite (which is probably a regular mixed-layer mineral with alternations of pyrophyllite layers with alumina layers and with a high substitution of F^- for OH^-) occur in paragenesis with kaolinite T.

These solutions had higher temperatures, with an acid pH, and allowed the alkali and alkaline-earthy levigation from the system.

The presence of nacrite, but most of all that of pyrophyllite is connected to the crater zones ; previously, there were mentioned some similar parageneses in other zones with a crater hydrometasomatic activity, namely at Talagiu (Apuseni Mts) and Cavnic (Maramureş Mts) (Ianovici, Neacşu, 1969 ; Ianovici et al., 1981). The hydrometasomatic activity from the crater zones was characterized by higher temperatures, while in the farther zones, although solutions had the same chemical composition, they had lower temperatures ; in this case, the formation of nacrite and pyrophyllite was no longer possible.

The Hydromicaceous Stage

Subsequently, solutions change their character and become neutral or weakly alkaline and K^+ rich, which allows the transformation of the previous minerals into hydromica. According to 02 1— 11 1 reflections (4.4—2.6 Å) it was concluded that hydromica polymorphy of a hydromuscovite is 2 M_I, and by maintaining the 001 reflection by calcination at 950°C and according to the very sharp form of basal reflections, hydromica was considered as well crystallized.

The hydromicaceous hydrometasomatic argillization process is the most important argillization process from the Harghita Mts.

The Mixed-Layer Stage

In this last stage solutions change again their character, they become Mg^{2+} and Ca^{2+} rich and with a more alkaline pH.

The intensity of this hydrometasomatic process is lower and has as a result the formation of random mixed-layer minerals, twocomponent or threecomponent, clearly dominated by hydromica.

Although the zone with argillizations formed of mixed-layer minerals from the Harghita Mts is the most widespread geological formation made up of mixed-layer minerals in Romania, one can conclude that the alteration process has a low intensity, because the smectite or the chlorite ratios are low and the complete transformation into montmorillonite or chlorite is never reached.

Hydrothermalism

The hydrothermal clays, which were formed on fissures by direct crystallization from hydrothermal solutions having the same origin as hydrometasomatic solutions, were generated by the successive crystallization of kaolinite, in the first stage, sometimes in paragenesis with nacrite and sulphides and finally of smectite and chlorite. Hydrothermal clays are frequently formed of montmorillonite and chlorite, minerals which stand for the last hydrothermal phase from the region.

Mixed-Layer Minerals

The argillization zones formed of mixed-layer minerals are very widespread in the Harghita Mts and are crossed by drillings down to 900 m deep. They are exploited, under the commercial name of "kaolin"

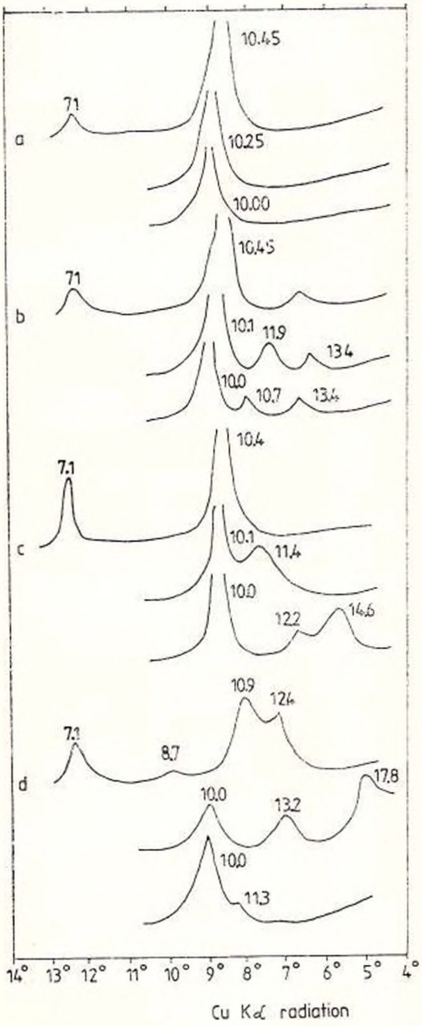

Fig. — Diffractograms of twocomponent random mixed-layer minerals (a) and threecomponent ones (b, c, d) between 4—14° 2 θ , $Cu_{K\alpha}$ radiation, Ni filter. For each sample the first diagram represents the natural, untreated sample, the second diagram represents the glycolated sample and the third diagram represents the sample calcinated for two hours at 550°C.

which after chemical beneficiation is used in the paper industry and in the ceramics industry (Neacşu, Neacşu, 1980).

About 80% from the analysed samples are formed of mixed-layer minerals which show a basal reflection of first order at 10.4—10.5Å and an irregular sequence of basal reflections.

Twocomponent Mixed-Layer Minerals

In 90% of cases, the random mixed-layer mineral has the basal reflection 001 at 10.4—10.5Å, does not expand when saturated with glycol or ethylene-glycol, but on the contrary, it diminishes its position at 10.2—10.3 Å and when calcinated at 550°C it contracts, reaching 9.9—10 Å, due to the collapse of the montmorillonite, which is present in a small amount in interstratification (Fig., a). The basal spacing 001 in the sample heated at 550°C, smaller than 10 Å, shows the absence of chlorite.

The 10.4—10.5 Å peak is very sharp (as the other basal reflections) indicating a well crystallized hydromuscovite.

The behaviour at treatments of the random mixed-layer mineral, especially by glycolation, corresponds to Brindley's observations (1951), who shows that under 20% montmorillonite in disordered interstratification with 10 Å mineral, the mineral no longer expands by glycolation. This behaviour is similar to the behaviour of random mixed-layer minerals from semiarid and desert soils from South Africa, which give 10.3—10.6 Å on an untreated sample, 10.1—10.3Å by glycolation and 10 Å by calcination at 550°C (Merve, Heystek, 1960, 1961).

The twocomponent mixed-layer mineral treated with Mg^{2+} and then glycolated, modifies its position at 10.2—10.3 Å and when treated with K^+ maintains its position at 10.4—10.5 Å. This indicates that the swelling mineral from the interstratification is montmorillonite, not vermiculite.

The use of the graphic method Mering (1949), in order to obtain the main coefficients of probability, indicates $P_M = 0.15$, while the use of Brown and Mac Ewan's curves (Brindley, 1951) indicates $P_M = 0.20$.

The twocomponent random mixed-layer mineral is characterized by the following main coefficients of probability : $P_H = 0.80—0.85$; $P_M = 0.15—0.20$ (where H = hydromica, M = montmorillonite).

Drits and Sakharov's curves (1976) for the probability domain $P_H > 0.75$ (with spacings 10.9—11.9 Å by glycolation) can be used to obtain the junction factor and in the conditions $P_{MM} = P_{MHM} = 0$ it is obtained a junction factor g higher than 0 (1 or 2).

Threecomponent Mixed-Layer Minerals

About 10% from the random mixed-layer minerals from the Harghita Mts are threecomponent : they are formed mostly of hydromica, subordinately montomorillonite and chlorite.

Figure (b, c and d) shows diffractograms of three threecomponent mixed-layer minerals.

In order to study the structure of these random mixed-layer minerals, Weaver's method (1955) was used to reduce them to two two-component systems 10—14 Å (hydromica-montomorillonite + chlorite, and hydromica + dehydrated montmorillonite-chlorite) on samples saturated with Mg^{2+} or calcinated at 550°C.

By using Weaver's method (1955) and Jonas and Brown's graphic method (1959) for threecomponent disordered interstratifications we obtain the following main coefficients of probability for the three samples from Figure (b, c and d) :

Sample 1d : $P_H = 0.60$; $P_M = 0.35$; $P_C = 0.05$;
Sample 1c : $P_H = 0.85$; $P_M = 0.10$; $P_C = 0.05$;
Sample 1b : $P_H = 0.80$; $P_M = 0.13$; $P_C = 0.07$.

(where H = hydromica, M = montmorillonite, C = chlorite).

By using Drits and Sakharov's method (1976) in order to find out the junction factor g, it results that for threecomponent mixed-layer minerals as well, in the conditions $P_H > 0.75$ and $P_{MM} = P_{MHM} = 0$, the junction factor g is higher that 0, being 1 or 2.

Thermodynamic Considerations

Both the twocomponent random mixed-layer mineral and the threecomponent one, with a junction factor g = 1 or 2, indicate the presence of some more stable mixed-layer minerals, as the free energy of the system is lower and the pressure and temperature of the formation environment are higher. These random mixed-layer minerals have a larger regularity tendency and allow the determination of physico-chemical conditions of thermodynamic transformations in the hydrometasomatic argillization stages from the Harghita Mts (Drits, Sakharov, 1976).

The presence of basal peaks for chlorite (Fig., b) or those for montmorillonite (Fig., d) can be interpreted not only as a discrete presence of these minerals, but also as their segregation tendency.

Conclusions

In the Harghita Mts zone, the hydrometasomatic argillizations are very widespread and they are represented by the kaolinic stage, the hydromicaceous stage and the mixed-layer stage.

The mixed-layer minerals are twocomponent in 90% of cases and threecomponent in 10% of cases ; they are of a random type with a regularity tendency, seldom with a segregation tendency and represent an incipient stage of incomplete transformation of hydromica into montmorillonite and chlorite.

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Pétrologie-Minéralogie

ERUPTIVE BRECCIAS ASSOCIATED WITH SOME TERTIARY MAGMATITES FROM ROMANIA

BY

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The areas with Tertiary volcanics in Romania exhibit eruptive breccia occurrences mainly associated with magmatic bodies in subvolcanic facies. Their investigation implies several aspects of petrological and metallogenetic importance.

There are several categories of breccias which, according to the classification of Wright and Bowes (1963), belong to the following types : (1) intrusion breccia ; (2) explosion breccia ; (3) intrusive breccia.

(1) The intrusion breccia is the result of the mechanic effect of magma intrusion in the country rock and occurs frequently in the contact area of subvolcanic magmatic bodies, as well as in some marginal areas of volcanic necks in the Oaş (Tarna Mare), Gutîi (Ilba, Herja, Cavnic, Băiuţ, etc) and Metaliferi Mts.

(2) The explosion breccia resulted from the explosive action of vapours and gases of magmatic nature within a confined space below the surface, is frequently encountered in Neogene volcanics ; it is better developed in mineralization areas in the Metaliferi Mts (Baia de Arieş, Roşia Montană, Deva, etc.). In places it may resemble the collapse breccia, such as the Baia de Arieş breccia bodies (Cochet, 1958).

(3) The intrusive breccia is the result of rock fragmentation and mobilization by magma or magmatic gases, with or without magmatic matrix (Wright, Bowe, 1963) ; it is also widespread in the Oaş Mts (Băile Turţului), the Rodna Massif (Izvorul Roşu, Cobăşel Mt), the Metaliferi Mts (Căraciu volcano, Măgura Ţebii, Baia de Arieş, Băiţa, etc.).

The breccia may often have a combined explosive-intrusive genesis and the breccia bodies exhibit a complicated "architecture".

Former geological studies were concerned with the morphology and some aspects of breccia genesis (Cochet, 1958 ; Socolescu et al., 1977 ; Ghiţulescu et al., 1979).

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The present note regards the geological and petrological features of explosion and intrusive breccias from the Metaliferi Mts and the Rodna Massif. One may also infer some aspects related to the metallogenetic significance of eruptive breccia formations.

Geological Emplacement and Morphology of Eruptive Breccia Bodies

The volcanic explosion and intrusive breccias, generally called eruptive breccias, associate with the Neogene magmatism of subduction areas inside the Carpathian arc. The breccia bodies, studied by us, associate mainly with quartz andesites and dacites belonging to the second Neogene volcanic cycle, Upper Badenian-Pontian in age (Rădulescu et al., 1981).

The best developed and most interesting eruptive breccias belong to the subvolcanic and volcanic complexes from areas with consolidated, "rigid" rocks of Mesozoic or older age, which underwent Neozoic tectogenesis. The igneous complexes including eruptive breccias occur in elevation areas of the basement (Baia de Arieş) which may resemble a horst by shape (Rodna Massif) or within posttectonic depressions over-

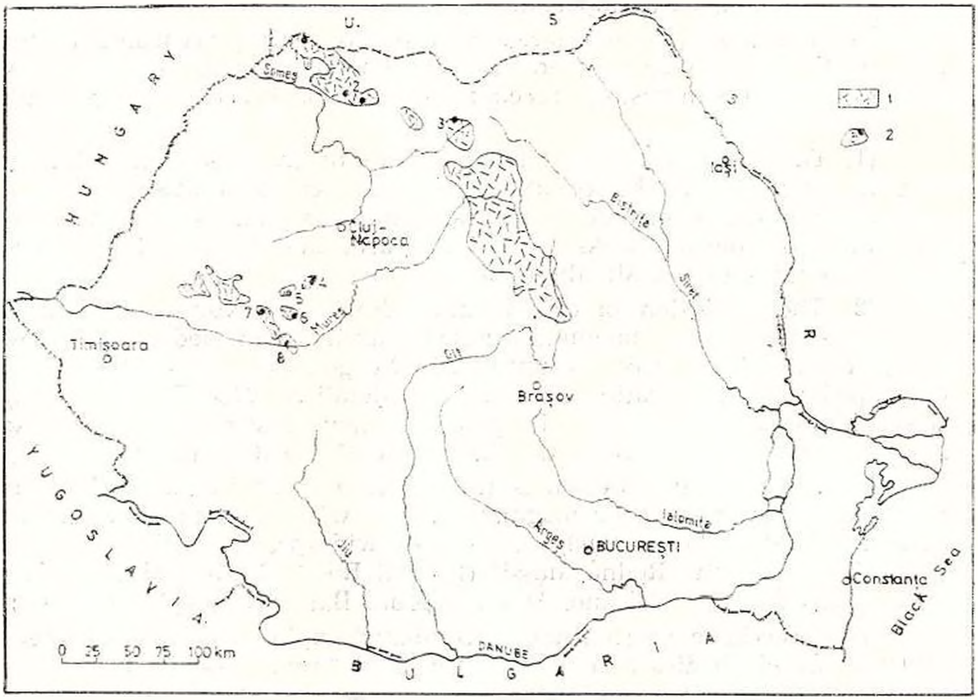

Fig. 1 — Distribution of Tertiary magmatites with eruptive breccias from Romania. Neogene igneous rocks (1) ; eruptive breccia occurrences (2).
1, Tarna Mare, Turţ (Oaş Mts) ; 2, Herja, Cavnic, etc. (Gutîi Mts) ; 3, Izvorul Roşu, Cobăşel (Rodna Mts) ; 4, Baia de Arieş ; 5, Roşia Montană and Bucium ; 6, Almaş-Stănija ; 7, Măgura Ţebii, Băiţa ; 8, Deva (Metaliferi Mts).

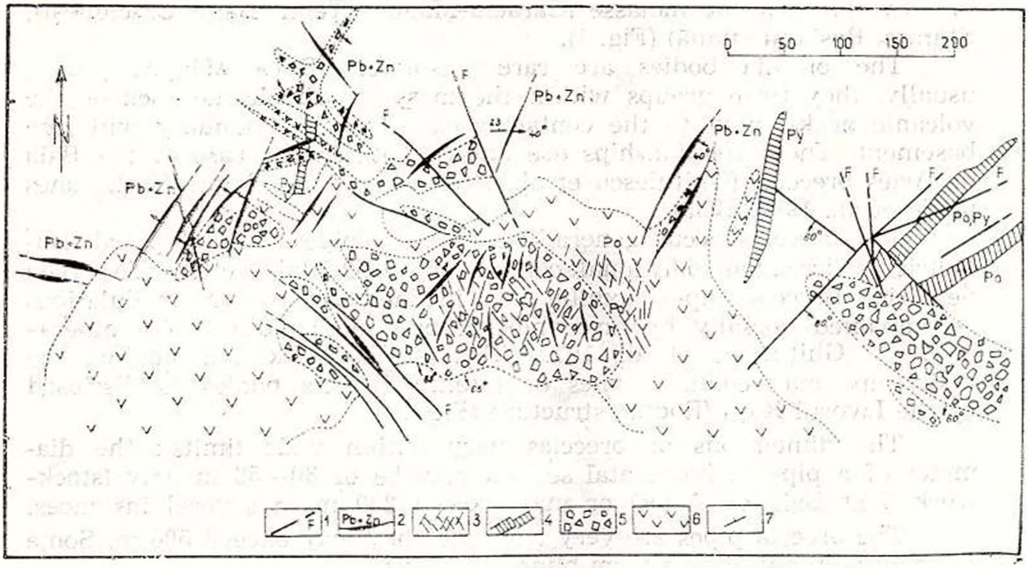

Fig. 2 — Geological setting of the Izvorul Roşu eruptive Breccia (Rodna Mts). 1, tectonic fractures ; 2, vein containing Pb and Zn mineralizations ; 3, sulphide or gold mineralizations of stockwork type ; 4, metasomatic pyrrhotite and/or pyrite mineralizations in limestones ; 5, polymictic eruptive breccias ; 6, amphibole and biotite quartz andesite ; 7, metamorphic rocks : micaschists, amphibolites, crystalline limestones, etc.

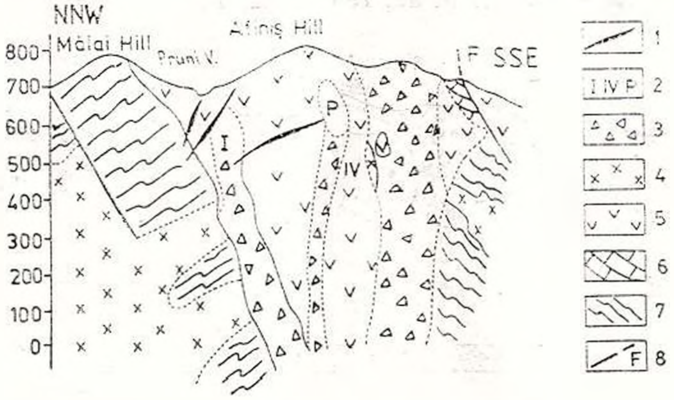

Fig. 3 — Geological section through the Afiniş-Baia de Arieş eruptive structure (Metaliferi Mts) (acc. to Ghitulescu et al., 1979).
1, gold veins ; 2, gold stockworks — I, IV, P ; 3, breccia pipes ; 4, subvolcanic microporphyry andesite ; 5, Afiniş andesite subvolcanic bodies ; 6, crystalline limestones ; 7, crystalline schists ; 8, faults.

lain by the Neogene molasse (Căraciu-Măgura Țebii, Băița Crăciunești, Stănija, Roșia Montană) (Fig. 1).

The breccia bodies are rarely isolated (Deva, Măgura Țebii); usually, they form groups within the massive subvolcanic rock or the volcanic necks, next to the contact area or to the boundary with the basement. These relationships are known both in the case of the Baia de Arieș breccias (Ghițulescu et al., 1979) and of the Rodna Veche ones (Jude et al., 1982, 1983).

The breccias occur generally as breccia pipes with elliptical, isometric or irregular polygonal contour in horizontal section. The Baia de Arieș Breccia pipes exhibit an inclined axis with modulations accompanied dorsally by an explosion breccia including gold mineralizations (Ghițulescu et al., 1979) (Fig. 3). The rocks surrounding the main pipe may contain dykes or satellite breccia bodies as disclosed by the Izvorul Roșu (Rodna) structure (Fig. 2).

The dimensions of breccias vary within wide limits; the diameter of a pipe in horizontal section may be of 30—50 m only (stockwork 3 at Baia de Arieș) or may exceed 200 m in several instances.

The breccia pipes are very long and they may exceed 500 m. Some of them crop out, others form blind chimneys.

The Izvorul Roșu (Rodna Massif) eruptive breccias exhibit a complex structure owing to a main pipe with irregular polygonal contour in horizontal section, accompanied by several breccia dykes and satellite bodies. The breccias and the mineralization were controled by fissure systems generated by vertical stress (Fig. 2). There are instances (Măgura Țebii, Băița Crăciunești) in which the main breccia pipe was emplaced prior to the instrusion of younger quartz andesite dykes (Jude et al., 1973, Cioflica et al., 1968) (Fig. 4).

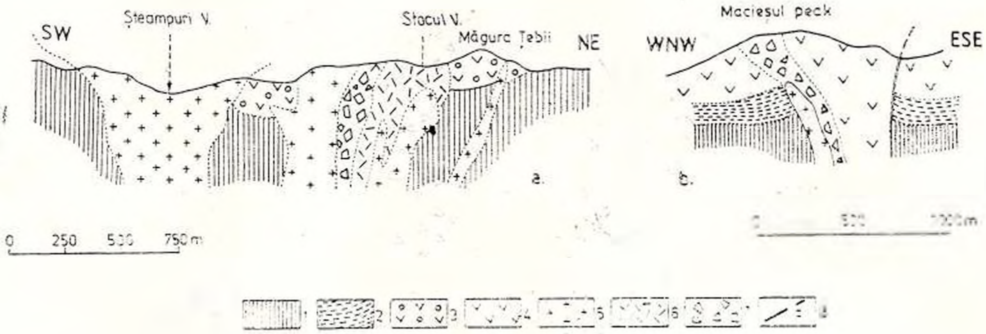

Fig. 4 — a) Geological section through the Măgura Țebii subvolcanic structure with eruptive breccias (Metaliferi Mts) (acc. to Jude, 1973, 1980).
b) Geological section through the Măcieșul-Băița-Crăciunești volcanic structure (Metaliferi Mts) (acc. to Cioflica et al., 1968).
1, Mesozoic eruptive and sedimentary rocks; 2, Miocene molasse formation; 3, Neogene andesite pyroclasts; 4, amphibole andesite (Făerag); 5, quartz andesites; 6. eruptive breccias with felsitic matrix; 7, eruptive breccias with tuffisite matrix; 8, tectonic fractures.

Petrographic Features of Eruptive Breccias

The most common petrographic features are exhibited by the explosion breccia, resulted from the crackling of eruptive rocks under the impulse of magmatic gases without a notable displacement of rock fragments : it is a crackle breccia. These breccias may be deprived of matrix, while the space between the "fragments" may be either free or mineralized, generating ore deposits, such as the Baia de Arieș gold stockworks no. 1 and 2, or base metal mineralizations.

The intrusive and explosive-intrusive polymictic breccias are more frequent and more characteristic. They consist of angular or subangular rock fragments of different lithological nature and varied dimensions (frequently from 0.3 to 20 cm, some blocks being bigger than 50 cm), as well as with different matrix. The polymictic breccia consists mainly of andesite, microdiorite or dacite "fragments" (Roșia Montană) accompanied by rock fragments from the breccia pipe walls : quartzites, crystalline schists, crystalline limestones, hornfelses (Rodna, Baia de Arieș), Mesozoic basaltic rocks (Măgura Țebii), Cretaceous sandstones (Roșia Montană) and even "exotic" fragments (gneisses) which ascended from the basement. The Izvorul Roșu (Rodna) breccia pipe exhibits remnants of pyrrhotite and pyrite mineralization prior to the breccia ; also magnetite fragments are reported by Udubașa (1974).

The breccia fragments are generally unoriented. There are instances in which the fragments exhibit a linear arrangement probably due to the fluidization of the breccia. The matrix of some bodies may be represented by a dense or porous andesite or dacite rock (Măgura Țebii, Metaliferi Mts). More often the matrix is polymictic with "tuffisite" aspect ; it consists of millimetric fragments of different rocks (andesites, microdiorites, quartzites, micaschists, limestones, etc.), crystal fragments and volcanic glass ("cinerite") generally transformed into postmagmatic neominerals, possibly mineralized.

Characteristics of Metasomatic Transformations Underwent by Eruptive Breccias and Adjacent Igneous Rocks

The pyrometasomatic processes are disclosed by the local occurrence of a skarn formation including garnet, diopside, vesuvian and epidote in the contact area of the quartz porphyry microdiorite stock from the Cobășel Mt (Rodna Mts) with the crystalline limestones. The pyrometasomatosis acts partly on the eruptive breccias that occur on the western and northwestern sides of the igneous body, towards its dome, bringing about the concentration of garnet and epidote within the breccia matrix. A barium-rich neoformation feldspar is added.

The high temperature metasomatosis is also characterized by the frequent occurrence of postmagmatic apatite in breccias ; it is a new feature which could be of some petrogenetic significance. Apatite was reported in the breccia matrix, in the Rodna Massif, as xenomorphic crystals and grain agglomerations, associated with tourmaline, rutile and fine casiterite grains in places. In the case of the Măgura Țebii breccias, fine grained apatite is accompanied by clinozoisite, pistacite and quartz ;

at Baia de Arieş, it occurs in highly silicified breccias and altered andesite rocks.

Propylitisation affects the subvolcanic bodies including eruptive breccias almost wholly. The eruptive breccias from the Cobăşel Mt (Rodna) and Măgura Ţebii (Metaliferi Mts) contain clinozoisite associated with pistacite and actinote, chlorite and quartz in places which point to high temperature propylitisation.

K-metasomatosis is marked on the one hand by the presence of adularia and on the other hand by the concentration of micaceous minerals, mainly of sericite. The adularia either forms pseudomorphoses after the plagioclase of adjacent igneous rocks and of breccia fragments or occurs in idiomorphic crystals associated with quartz within the holes of breccias and on fissures. However, adularia is frequently turned into sericite and/or kaolinic argillaceous minerals. Micaceous minerals of sericite type substitute the feldspar, biotite and in places amphibole of breccia fragments and constitute quartz associated massive concentrations within the matrix of the Izvorul Roşu Breccia (Rodna Mts). The optical features, correlated with the chemical analysis data, point to lithium-rich white mica.

It is also worth mentioning the concentration of hydrothermal carbonates is certain breccia bodies (Izvorul Roşu, Rodna, stockwork 5 Baia de Arieş, etc.) and the general metasomatosis with argillaceous, kaolinic minerals.

Some subvolcanic structures including eruptive breccias are characterized also by the occurrence of zeolites which disclose the presence of hydrothermal solutions of alkaline nature. Zeolite metasomatosis is well developed in the Măgura Ţebii andesite subvolcanic complex and in its associated breccias ; it was reported from boreholes at a depth of 500 m. Epistilbite occurs at small depth, while gonnardite occurs at greater depth (Jude et al., 1980).

Relationships Between the Mineralization and the Eruptive Breccias

The network of pyrrhotite and pyrite veins in the quartz andesite occurring to the east of the Izvorul Roşu Breccia pipe (Rodna Mts), similar to the metasomatic concentrations in the Rebra Series crystalline limestones (Fig. 2), points to the occurrence of a mineralization prior to the eruptive breccia. This is also accounted for by the sulphide fragments in the breccia. In all the other instances known so far, the mineralization is posterior to the breccias. Thus, sulphide veins or gold mineralizations occur in breccias and adjacent rocks both in the Rodna and the Metaliferi Mts. At Baia de Arieş and Stănija telluride veins occur also in eruptive breccias (Ianovici et al., 1969).

Another feature is the sulphide and/or native gold impregnation of the breccia matrix ; the limestone pieces may be partly or wholly substituted by lead and zinc minerals such as in the case of breccias from Baia de Arieş (Lazăr, 1966) or Rodna Mts (Udubaşa, 1970 ; Socolescu et al., 1977 ; Jude, 1982). The most interesting mineralizations from breccias were reported at Roşia Montană, in the Cetate Mt and at Cîrnic (Cotreanţa stockwork). One should also note some explosion breccias from Baia de Arieş which include adularia silicified andesite blocks surrounded by centimetric gold quartz crusts (Cochet, 1958 ; Ianovici et al., 1969).

The mineralization of breccias may also occur as ”gold pyrite“ stockworks rich in arsenic, as for instance some breccia bodies from the Rodna Mts (Fig. 2) (Jude et al., 1982, 1983).

However, the breccia bodies are not homogeneously mineralized ; there are instances in which the mineralization concentrates in the marginal areas and others in which it occurs in the central area of the breccia pipe.

On the Genesis of Eruptive Breccias

The geological literature of these last 50—60 years offers numerous hypotheses on the genesis of eruptive breccia pipes. We note the following : explosion due to the vaporization of underground water brought about by ascending magma (Lindgren, Bastin, 1922) ; fluidization of fault breccia and crackled rocks (Farmin, 1941) ; brecciation of rocks at the intersection of fracture zones (Kuhn, 1941, in Mitcham, 1974) ; products of exsolved vapour from magmas (Norton, Chathes, 1973) ; expanding of rocks from the walls of void spaces present in areas with multiple faults (Mitcham, 1974).

The eruptive breccia complexes of the type of those occurring in the Rodna Massif (Fig. 2) and the mineralized and non-mineralized fissure systems point out a vertical stress similar to a volcanic explosion.

The genesis of the main channel and of fissure systems which include the eruptive breccia may be best accounted for by the model of Norton and Chathes (1973). Gases and magmatic vapours exsolved from magmatic melt accumulated temporarily at the top of a magmatic intrusion situated at small depth in lithosphere rocks. At the critical pressure, the magmatic vapours were released by explosion along extension fissures from the intrusion dome. Then followed several phenomena : decrease of system pressure and temperature, further crystallization of magma and rock expanding from the walls of formerly gaseous cavity, the occurrence of eruptive breccia implicitly.

Both the size and features of breccia depend on explosion energy ; a low intensity explosion entails the in situ brecciation of rocks (crackle breccia), without notable shift of rock fragments ; on the other hand, a high intensity explosion generates real breccia pipes, accompanied by fissures and eventually breccia dykes.

For the most cases, the constitution of eruptive breccia should be looked upon as a complex process achieved during several stages. To the former stage of explosive brecciation succeeded the mobilization of breccia material, possibly accompanied by a new low viscosity magma pulsation due to an increased volatile content. This is disclosed by the oriented texture, with fluidal aspect, of some eruptive breccias with tuffisite of felsitic matrix (Izvorul Roşu — Rodna Massif, Măgura Ţebii — Metaliferi Mts). The occurrence of volatile components, fluorine, chlorine (?) and boron is accounted for by the presence of postmagmatic apatite and tourmaline within the matrix of several breccia bodies. Anyhow, the geological literature reports the occurrence of postmagmatic apatite related to explosive brecciation (Nikitina et al., 1971).

The recurrence of brecciation phenomena accounts for the presence of complex structures in which a former breccia generation (felsitic matrix), such as the Măgura Țebii one (Metaliferi Mts), was crossed by a late breccia with tuffisite matrix (Fig. 4). The hydrothermal solution influx, chemically unbalanced as compared to the breccia material, brought about several metasomatic transformations such as propylitisation, K-alteration with adularia and sericite, argillitic alteration and other facies previously mentioned.

The metallogenetic significance of eruptive breccias is accounted for by the fact that they are structures which favour the ascending circulation of mineralizing hydrothermal solutions and they also represent a favourable environment for the deposition of mineral substances. The relationships between the Neogene mineralization and eruptive breccias are very conclusive.

The postmagmatic apatite reported from most eruptive breccia bodies may be considered a significant mineral for geological prognosis. According to Williams and Cesborn (1977) apatite was involved in early stages of hydrothermal systems and concentrated at the top of intrusions including porphyry copper mineralization.

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Pétrologie-Minéralogie

FLUID INCLUSIONS IN HYDROTHERMAL CALCITE AND THEIR SIGNIFICANCE IN CRYSTALLOGENESIS

BY

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Introduction

The typomorphical features of calcite were studied for the first time by Maucher (1914), then by Kalb (1928) and later by Shkabara (1940 — mentioned by Lazarenko, 1979). Also, the general diagrams regarding the variation of habits of calcite crystals depending on temperature conditions are known in literature (Kostov, 1968, 1979 ; Lazarenko, 1979).

In this paper, a diagram of crystallogenetic significance for calcite and its minerals associated on the basis of the 770 determinations on the homogenization temperatures of fluid inclusions in 134 calcite samples, from 8 hydrothermal ore deposits, is presented. The diagram gives a correlation between the habits of calcite crystals and homogenization temperatures of primary fluid inclusions from this mineral.

Calcite Crystals

The calcite samples have been collected from fissures of volcanic rocks and from hydrothermal veins of the following ore deposits : Herja, Baia Sprie, Cavnic, Băiuț, Țibleș (Gutîi-Țibleș Mountains), Stînceni (Călimani Mountains), Rușchița (Poiana Ruscă Mountains) and Băişoara (Gilău Mountains).

The habit of calcite crystals varies from rhombohedron ($02\bar{2}1$) and scalenohedron ($21\bar{3}1$), ($21\bar{3}1$) + ($10\bar{1}1$), ($10\bar{1}0$ + ($01\bar{1}2$) to basal rhombohedron ($10\bar{1}1$).

Usually each simple or composed form of calcite crystals is associated with certain minerals (Tab. 1).

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TABLE 1

The prevailing crystal forms of calcite, homogenization temperatures of primary fluid inclusions and its mineral association

Habit	Ore deposit	Homogenization temperatures (°C)			Association
		no. det.	range	maximum frequency	
(0221)	Herja, Băiuț	20	60–80		C
(2131)	Herja, Cavnic, Țibleș, Rușchița	200	80–140	95–100	C–M–J
(1010) (0112)	Herja, Stinceni, Rușchița	250	170–280	190–205	C–A
				255–270	C–S–G;C–F
(1011)	Cavnic, Stinceni	180	260–375	280–320	C–Py
	Țibleș	120	250–320	300–315	C–Py–Cp

C, calcite; M, marcasite; J, jamesonite; A, antimonite; S, sphalerite; G, galena; Py, pyrite; F, fluorite; Cp, chalcopyrite.

Fluid Inclusions

Depending on the habit of calcite crystals and the minerals with which calcite is associated, several types of fluid inclusions have been identified (Pomârleanu et al., 1967, 1968, 1972, 1981, 1982).

The data on the homogenization temperatures of fluid inclusions in calcite were used to draw up the histograms and frequency curves with which the respective genetic interpretation has been done (Pomârleanu, Pomârleanu, 1982).

Correlation Between Homogenization Temperatures of Fluid Inclusions and Habits of Calcite Crystals

The diagram from Figure represents a correlation between homogenization temperature of primary fluid inclusions in calcite and its crystallographic habits. In agreement with the diagram it results that calcite has been crystallized under various temperature conditions : from below 60—80°C for the habit $(02\bar{2}1)$ up to 370°C for the habit $(10\bar{1}1)$ which is usually associated with pyrite.

Fluid inclusions in euhedral scalenohedron calcite associated with marcasite or jamesonite (Herja and Cavnic) show homogenization temperatures of 80—140°C (Tab., Fig.). Inclusions in the same form of calcite crystals from districts : Korsnäs-Finland (Rehtijärvi, Kinnunen, 1979), Laisvall-Sweden (Roedder, 1968) and Central Siberian Platform (Andrusenko, 1971) have homogenization temperatures in the same range.

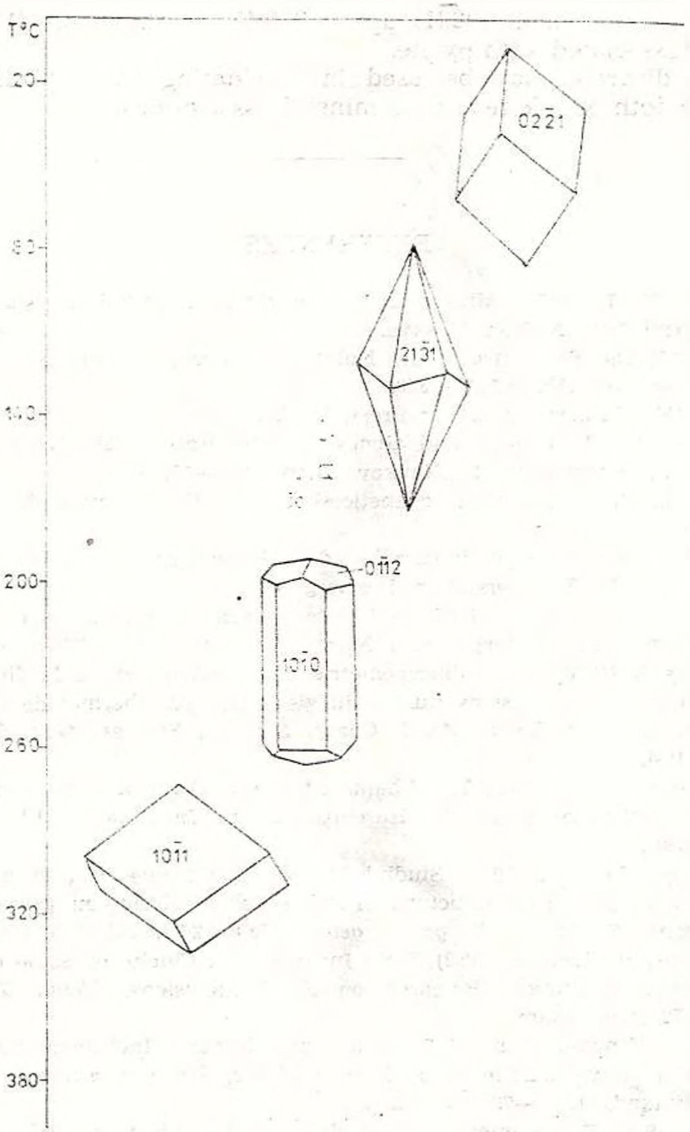

Fig. — Diagram showing the correlation between the habits of calcite crystals and homogenization temperatures of primary fluid inclusions from this mineral.

Conclusions

In this paper is presented a diagram that shows a correlation between homogenization temperatures of fluid inclusions in calcite and its crystallographic habits. The homogenization temperatures vary from

60—80°C for the habit $(0\bar{2}21)$ up to 375°C for the habit $(10\bar{1}1)$ which is usually associated with pyrite.

The diagram may be used in elucidating the crystallogenetical conditions both of calcite and its mineral associations.

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Pétrologie-Minéralogie

A BIMODAL IGNEOUS COMPLEX OF NEOGENE AGE, ŢIBLEŞ, EAST CARPATHIANS, ROMANIA

BY

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Introduction

The Ţibleş igneous complex is located between the Oaş-Gutîi and Călimani-Harghita segments of the Neogene volcanic chain in the East Carpathians. It builds up together with Toroiaga, Rodna and Bîrgău units the so-called subvolcanic zone (Peltz et al., 1972). This paper summarizes the main structural and petrologic features of the Ţibleş igneous complex, which exhibits a bimodal character as compared to the other igneous units of Neogene age in Romania.

Geologic Setting and Structure of the Igneous Complex

The Ţibleş igneous complex is situated southwards of the Bogdan Vodă (E-W) transcrustal fault, where the Oligocene-Miocene sedimentary rocks of the Autochthon underlie the Paleogene ones of the Wildflysch Nappe and of the central tectonic unit (Fig. 1).

The igneous bodies cut sedimentary rocks of Paleogene and Lower Miocene age with contact metamorphic overprints ; the age of magmatites is thus undoubtedly post-Lower Miocene (Edelstein et al., 1981). The Earth's crust is here 30-35 km thick (Socolescu et al., 1975) and the nappe group of the Oriental Dacides constitutes the basement. The igneous complex consists of three units : (a) the SE unit (Arcer-Ţibleş-Măgura Neagră), (b) the central unit (Tomnatec-Stegioara-Hudieş) and (c) the NW unit (Hudin). The SE unit includes the main igneous body consisting of a central stock of composite constitution and an external ring-like zone ; in the vicinity there are many smaller igneous bodies

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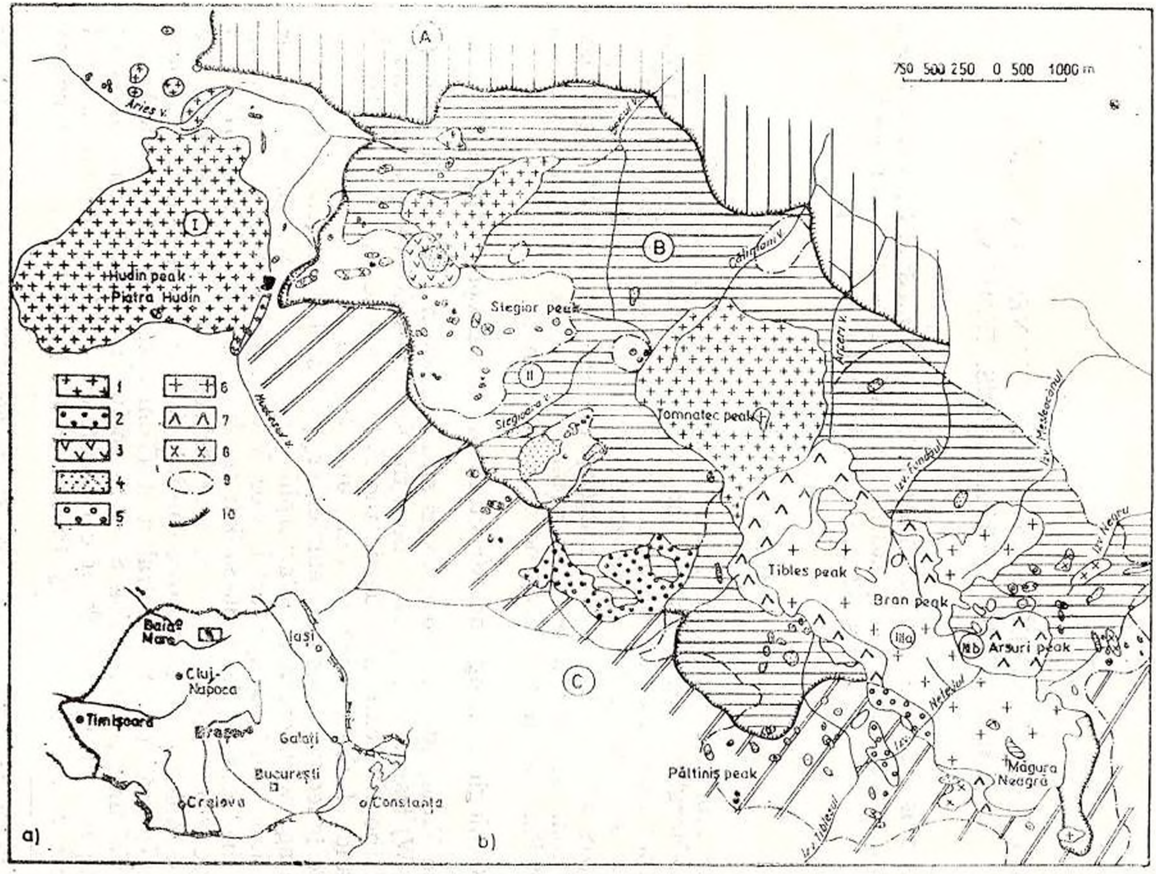

Fig. 1 — a, location of the Țibleș-Hudin Mts on the Romanian territory ; b, geological sketch of the Țibleș igneous complex. Acidic formation (AF) : 1, micrograndiorites, dacites ; intermediate formation (IF) : 2, tonalites ; 3, plagidacites-dacites; 4, quartz-diorites-tonalites ; 5, microdiorites ; 6, quartz monzodiorites ; 7, microgabbrodiorites, two pyroxene latite andesites ; 8, gabbrodiorites ; 9, contour of thermometamorphism aureole ; 10, overthrust.

A, Lăpuș Nappe ; B, main tectonic unit ; C, Autochthon ; I, NW unit ; II, central unit ; III, main igneous body : a, central stock ; b, envelopes.

TABLE 1

Eruptive formation	Area (km²)	Petrogenetic associations	Intrusive events	Petrotypes	Typical occurrences	Morphology	Depth level (supposed)
Acid	10,5 10,8*	Rhyodacites	I	Dacites $\xrightarrow{2)}$ rhyolites ↑	Tomnatec	Intrusive cupola	Subvolcanic
				7. Microgranodiorites→ microgranites	Hudin	Laccolith cupola	,,
Intermediate	11,8 50,8*	Tonalite series	II—III(?)	6. Tonalites a)→ granodiorites b) tonalites s. str.	Grohot Groapa	Microneck	,, ,,
				5. Plagidacites→ dacites	Piatra Hudin Groapa	Neck apophyses	,,
				4. Quartz diorites→ granodiorites→ tonalites: (a) with 2 pyroxenes (b) with px+amphiboles	Hudieş (a) Groapa (b) Stegior (b)	Neck apophyses	,, ,, ,,
		Monzodiorite series Main eruptive body central stock	V	3f. Lamprophyres; aplites; micropegmatites	Bran Izv. Brani	Veins	Subvolcanic
			IV	3e. Menzogranites	Gal. 4 Arcer		↑
				3d. Granodiorites (granogabbro)	Izv. Neted		
				3c. Monzodiorites→ granodiorites	Gal.4 Arcer		
				3b. Monzodiorites→ quartzmonzonites	,,		
				3a. Monzodiorites→ quartzmonzogabbro	Tibleş, Bran		Plutonic
		Envelope	III	2. Two pyroxene latite-andesites (±quartz)	Arcer	Microlaccolith; dykes; apophyses	Subvolcanic
				1. Gabbrodiorites (±quartz) (a) coarse grained	Izvorul Negru, Izvorul Roşu Hudieş	,,	,,
				(b) fine grained	Arsuri	Microlaccolith; neck	,,

* as suggested by geophysical data.

of varying composition (Fig. 1, Tab. 1). At a depth of about 1000 m the geophysical data suggest a unitary and nearly circular form of the entire SE igneous unit. The central unit includes microgranodiorites in the north and many bodies of intermediate composition in the south and in the central part. Some of them are composite as a result of polystadial igneous activity (Pop et al., unpublished). The NW unit contains the Hudin intrusive cupola with dominant microgranodiorites and small dacite bodies around it (Fig. 1).

Petrography

The Ţibleş igneous rocks constitute two main igneous formations : a) acidic and b) intermediate formation. The first one is compositionally homogeneous with small structure and texture variations ; it consists of microgranodiorites and dacites. The intermediate *Fm* is compositionally more complex and shows marked transitional features, as well as a heteromorphic character (according to Rittmann, 1973). It consists of two distinct rock suites : (1) monzodiorites-monzogranites and (2) tonalites (Tab. 1).

The petrotypes have been established by using the classifications of both Streckeisen (1967) and Rittmann (1973). The main minerals are plagioclases *(plg)*, clino- and orthopyroxenes *(cpx, opx)*, amphibolites, ± biotite, quartz, alkali feldspars. The opaque accessories are : magnetite, ilmenite, rutile, chalcopyrite, pyrrhotite and pyrite ; the other include apatite, sphene, zircon, orthite. Corroded quartz phenocrysts and cordierite were observed only in the rocks of the acidic *Fm*.

The typical mineral assemblages are : (1) *plg* (An_{60-90})-*cpx* (magnesian augite)-*opx* (40% $FeSiO_3$)±amphibole, in gabbrodiorites, two pyroxene diorites, latite andesites ; (2) *plg*(An_{35-85})-*cpx*-*opx* (70% $FeSiO_3$) amphibole±biotite-quartz, in quartz diorites, tonalites, plagidacites ; (3) *plg* (An_{40-60})-*cpx* (calcic augite)-*opx* (hypersthene)-amphibole-alkali feldspar (orthoclase, anorthoclase)-quartz, in monzodiorites ; (4) *plg* (An_{45-65})-*px*-amphibole-biotite-quartz (euhedral phenocrysts)-cordierite-alkali feldspars (anorthoclase, sanidine), in the rocks of the acidic *Fm*. For more details see Pop et al. (in press).

Petrochemistry

There were about 130 available chemical analyses ; most of them (112) are made on the rocks of the intermediate *Fm* (the acidic *Fm* contains more altered rocks). The chemical components obey as a rule the normal distribution law. SiO_2, TiO_2 and, to a lesser extent, Al_2O_3 show bimodal distributions (Fig. 2a), the two modes corresponding to the acidic and intermediate formations. The last one exhibits a normal distribution of these components too (Fig. 2b). The average chemical composition (Tab. 2) matches well (excepting the acidic *Fm)* the chemistry of the andesitic rocks (Taylor, 1969 ; Gill, 1981). The CIPW and Rittmann norms suggest the transitional and heteromorphic character of the petrotypes reflecting thus the actual conditions, i.e. plutonic and (shallow) subvolcanic, under which the rocks consolidated. Normative corundum and cordierite appear by calculating CIPW and Rittmann norms, respectively. On the QAP diagram (Fig. 3) the rocks of the acidic *Fm* occupy mainly the granodiorite field, whereas the rocks of the intermediate *Fm* clearly follow two distinct trends : (A) of monzodiorites-monzogranites, and (B) of tonalites-granodiorites. Ac-

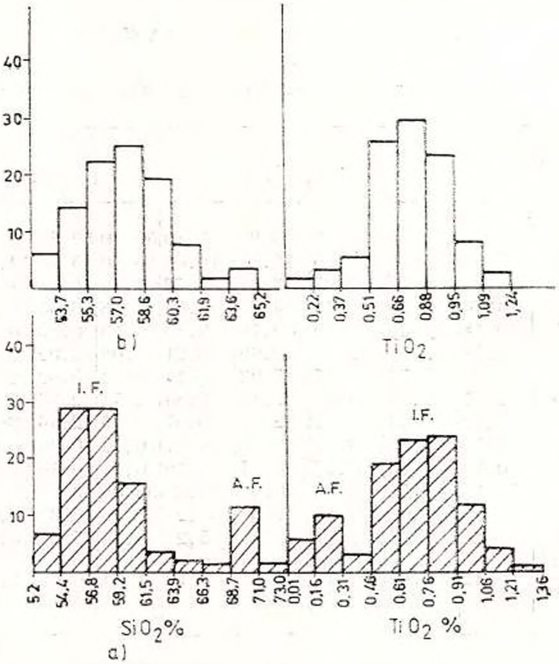

Fig. 2. — SiO_2 and TiO_2 distribution of Ţibleş magmatites. a, global ; 2, intermediate formation.

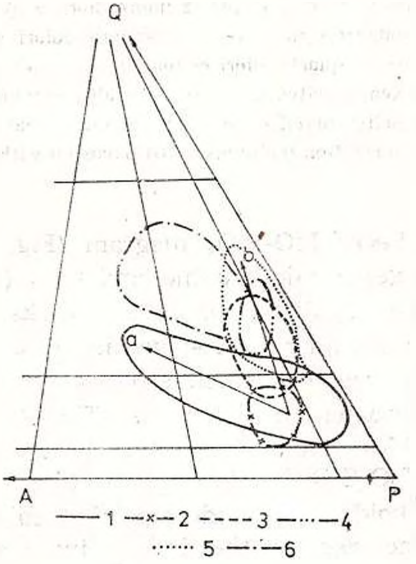

Fig. 3. — QAP diagram.

1, monzodioritic suite ; 2, two pyroxene latite andesites ; 3, two pyroxene quartz diorites (± amphiboles)-tonalite ; 4, tonalites ; 5, plagidacites-two pyroxene and amphibole dacites ; 6, acidic formation ; A, monzodiorite-monzogranite trend; B, tonalite-granodiorite trend.

cording to the K_2O-SiO_2 diagrams of Taylor (1969) and Peccerillo and Taylor (1976) the rocks of the intermediate *Fm* belong mainly to andesites and K-rich andesites (Fig. 4). On this diagram the main evolution trends can also be seen : (1) of K-enrichment (K-rich andesites and dacites), and (2) of maintaining, or even decreasing, the K content

TABLE 2

Mean chemical compositions of Ţibleş magmatites

Oxides	1		2	3					4	5	6	7	8
	a	b		a	b	c	d	e					
SiO_2	53,29	54,95	55,12	56,93	57,89	59,40	59,83	63,53	57,75	59,16	63,65	69,66	57,16
Al_2O_3	17,94	,17,53	,20,00	16,35	16,68	16,00	16,47	14,48	18,14	16,92	17,55	16,45	17,16
Fe_2O^0	3,84	4,13	2,07	3,44	1,72	1,72	3,42	1,37	3,25	3,47	1,67	,531	3,19
FeO	4*74	5,25	3,49	4,74	4,77	4,38	3,55	3,24	4,a2	3,69	2,74	1,36	4,39
MnO	0,18	0,23	0,10	0,18	0,13	0,13	0,13	0,10	0,16	0,17	0,13	0,09	01,7
MgO	4,53	3,71	2,99	3,69	3,21	3,09	3,02	2,29	2,73	2,76	1,92	0,83	3,41
CaO	8,94	7,74	8,47	7,02	6,18	6,44	5,82	2,36	6,70	5,88	4,71	2,30	7,04
Na_2O	2,67	2,69	2,60	2,80	2,84	2,74	2,77	3,00	2,25	2,6	3,33	3,10	2,79
K_2O	1,13	1,24	1,82	2,10	3,67	2,86	2,46	3,35	1,48	1,33	1,58	2,41	1,88
TiO_2	0,73	0,69	0,63	0,78	0,91	0,72	0,49	0,60	0,79	0,89	0,45	0,18	0,71
P_2O_5	0,15	0,16	0,17	0,17	0,19	0,19	0,16	0,15	0,15	0,13	0,13	0,11	0,16
An-n	53,1	51,7	52,5	40,9	32,3	36,0	38	28	49,2	46,8	46,3	17,8	
DI	37,7	41,8	43,0	49,9	54,2	53,7	55,3	62,8	47,5	51,6	58,6	77,0	
N–A	1,2	1,9	2,1	3,4	5,2	5,7	7,0	5,0	3,7	3,6	6,7	10,7	
$\frac{K_2O}{Na_2O}$	0,42	0,46	0,7	0,75	1,29	1,04	0,89	1,12	0,59	0,50	0,47	0,78	

1, gabbrodiorites: a) coarse grained (d 2 mm) (n = 8); fine grained (n = 4); 2, pyroxene latite-andesites (n = 16); 3: a, quartz monzodiorites (n = 30); b, quartz monzodiorites-quartz monzonites (n = 1); c, quartz monzodiorites-granodiorites (n = 4); d, microgranodiorites in satellite bodies (n = 7); e, monzogranites (n = 4); 4, quartz diorites-tonalites-granodiorites (n = 26); 5, plagidacites-amphibole and two pyroxene dacites (n = 9); 6, tonalites-granodiorites (n = 4);7, microgranodiorites-microgranites-dacites-rhyolites (n = 18) (acidic formation); 8, mean chemical composition of the intermediate formation (columns 1—6) balanced with the outcropping area. An-n — normative anorthite.

(normal andesites and dacites). The FeO^*/MgO-SiO_2 diagram (Fig. 5) suggests the congruency with the orogenic calc-alkaline andesites (according to Gill, 1981) as well as some affinities (for certain rocks) to the tholeiitic andesites. However, the dominant calc-alkaline trend is seen on the Na_2O+K_2O/SiO_2 diagram (Fig. 6), which stresses, in addition, the supercalcic and calcic character of the rocks. The "continental" evolution trend of the Ţibleş magmatites may partly be depicted on the $Al_2O_3/CaO+Na_2O+K_2O$-SiO_2[N-A] diagram (Fig. 7)

A normal feature has the Nockolds-Allen index related minor element distribution ; Ba and Zr increase as the index increases, whereas Cr, Ni, Co, Sc and V decrease ; Y, Yb and Ga remain unchanged. The distribution diagrams of the minor elements in the Ţibleş rocks are astonishingly similar to the diagrams of Taylor (1969) for andesites (Udubaşa et al., in press).

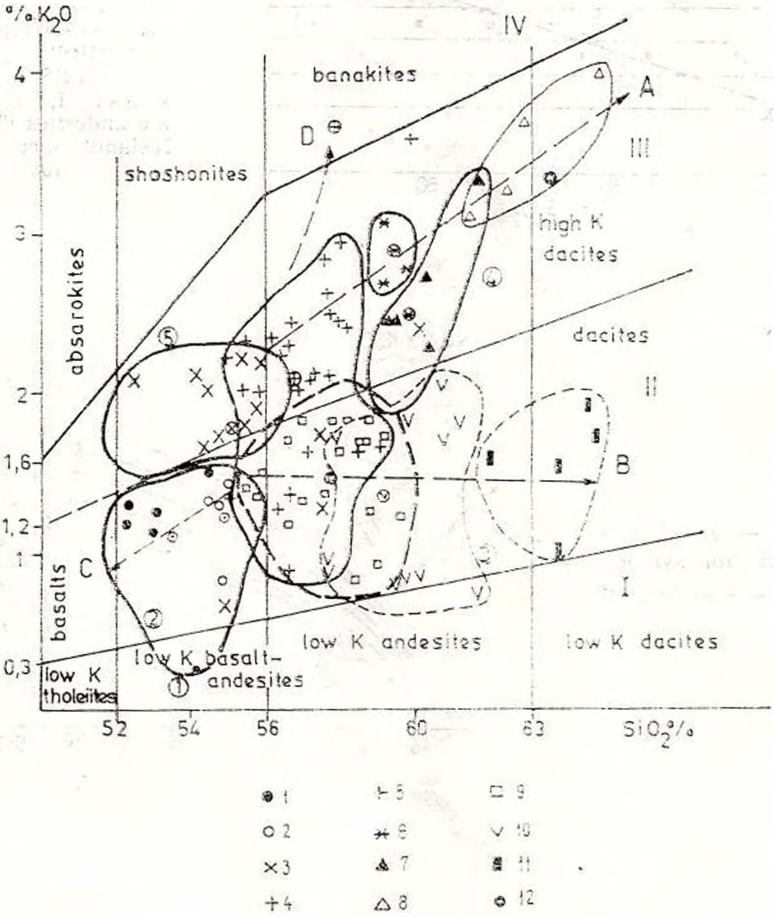

Fig. 4. — K_2O-SiO_2 diagram of intermediate formation.

1, gabbrodiorites : coarse grained ; 2, gabbrodiorites : fine grained : 3, two pyroxene latite andesites ; 4, monzodiorites (± quartz) ; 5. quartz monzodiorites-quartz monzonites ; 6, quartz monzodiorites-granodiorites ; 8, monzogranites ; 9, quartz diorites-tonalites-granodiorites ; 10, plagidacites-two pyroxene and amphibole dacites ; 11, tonalites; 12, mean values of petrotypes.

I, island arc tholeiitic series ; II, calc-alkaline series ; III, high-K calc-alkaline series ; IV, shoshonite series ; 1*, low-K basaltic andesites ; 2*, basaltic andesites ; 3*, andesites ; 4*, high-K andesites ; 5*, high-K basaltic andesites (acc. to Peccerillo and Taylor, 1976).

A, B, C, D, evolution trends (see the text).

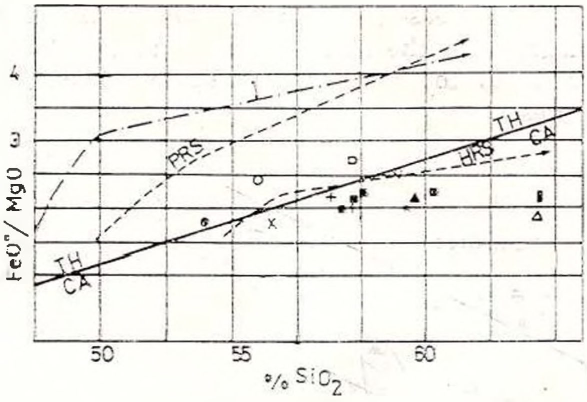

Fig. 5. — Mean petrotypes plotted on FeO*/MgO-SiO$_2$ diagram (for symbols see Fig. 4). TH, tholeiitic andesites; CA, calc-alkaline andesites; HRS, hypersthene series; PRS, pigeonite series; I, non-orogenic andesites (Tingmuli, Iceland) (acc. to Gill, 1981).

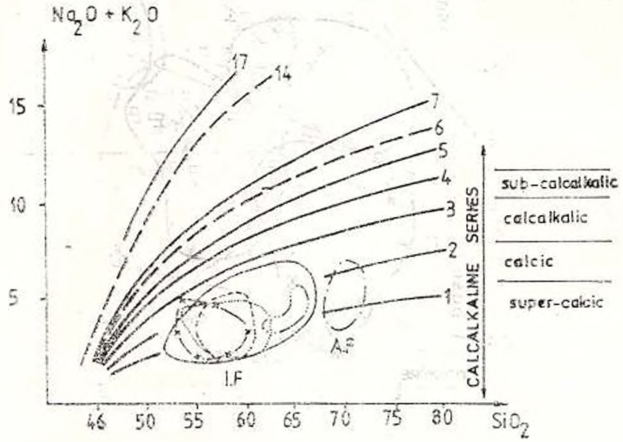

Fig. 6. — Alkali-SiO$_2$ diagram (for symbols see Fig. 3).

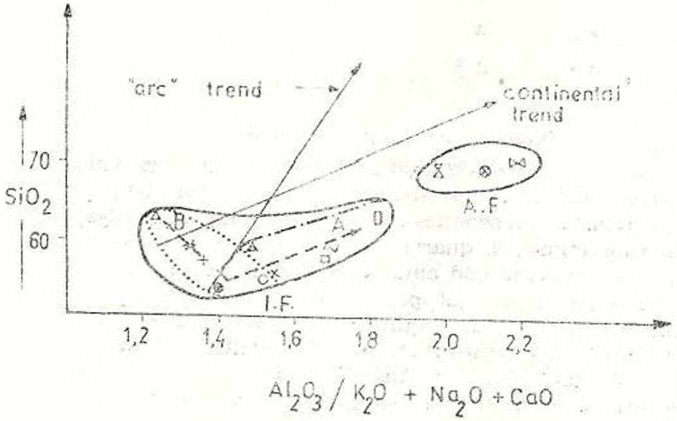

Fig. 7. — Țibleș mean petrotypes plotted on SiO$_2$-Al$_2$O$_3$/CaO+Na$_2$O+K$_2$O diagram. Lines of "arc" and "continental" trends acc. to Feiss (1980). A, tonalite trend; B, monzodiorite-monzogranite trend.

Discussion

The Ţibleş igneous Complex has appeared as a result of two-phase magmatic activity evolved under non-volcanic conditions. The igneous rocks show a clearly bimodal character. The first main phase had a unique magmatic event and gave rise to the acidic *Fm* consolidated at shallow depth (subvolcanic). The second main phase appears to be formed during more than two magmatic events giving rise to the intermediate *Fm* formed at greater depths (subvolcanic-plutonic). The evolution of the latter phase exhibits two distinct trends : (1) of K enrichment (monzodiorites-monzogranites) and (2) of K depletion with SiO_2 increase (tonalites-granodiorites). These two trends are clearly developed on the DI-SiO_2 diagram (Fig. 8), on which a gabbrodiorite trend exists too.

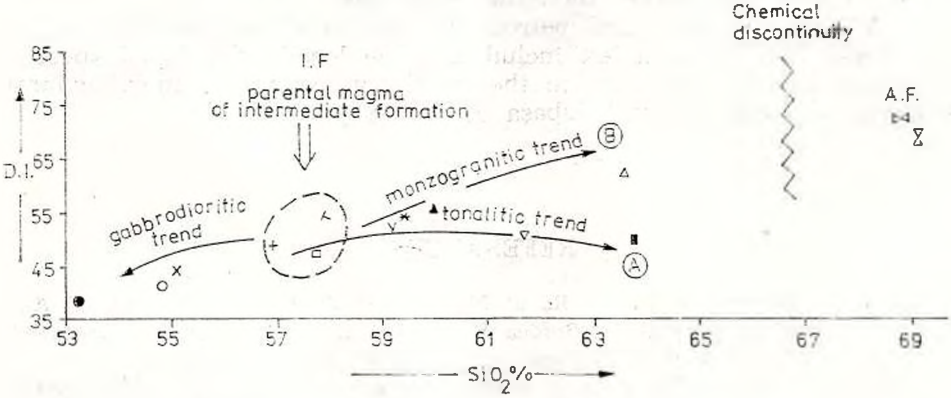

Fig. 8. — Magmatic evolution trends of the intermediate formatin according to the differentiation index (D.I.) and SiO_2 (for symbols see Fig. 4).

The two igneous formations derived from dominant andesitic magmas which have been issued by lithosphere subduction at the western margin of the Eurasian Plate.

The firstly formed acidic *Fm* evolved from a magma supply with simpler evolution and shows more signs of contamination with sialic matter (suggested by the presence of cordierite and corroded quartz phenocrysts). This first magmatic phase lacks in metallogenetic products. Afterwards, probably from a different depth, a new amount of little differentiated magma (compositionally similar to the quartz monzodiorites — see columns 8 and 3A in Table 2 — and to the orogenic andesites) was located in two main magmatic chambers with independent evolution. The first one — situated in the NE unit — has generated, by nearly in situ differentiation, the monzodiorite suite with three igneous events resulting in the formation (1) of the ring rocks, (2) of the central stock, and (3) of the vein rocks. The second magmatic chamber — situated in the central tectonic unit — had a vertical zoning and gave rise to the tonalite suite with two igneous events : (1) formation of plagidacites and two pyroxene amphibole tonalites, from the magma position in the apical part, and (2) formation of two pyroxene diorites from the deeper part. Mainly base metal ore veins are developed in connection with the intermediate *Fm* ; in addition, some disseminated ores (Cu, Mo, etc.) are also known (for details see Udubaşa et al., in press).

Island arc setting on continental lithosphere may explain some mixed characters of this rock association, i.e. of orogenic calc-alkaline andesites with well marked tholeiitic tendencies.

All the structural and petrographic-petrochemical peculiarities of the Ţibleş igneous Complex including some locally developed specific alteration zones are similar to the porphyry (copper or molybdenum) systems as discussed by Udubaşa et al. (in press).

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Pétrologie—Minéralogie

LE MÉTAMORPHISME DES CHARBONS DES CARPATHES MÉRIDIONALES ROUMAINES

PAR

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Le dernier temps, pour la caractérisation des charbons, on a utilisé, de plus en plus, des méthodes optiques quantitatives et la diffraction Rx (Landis, 1971 ; Grew, 1974 ; Kisch, 1974 ; Diesel, Offler, 1975 ; Kwiecinska, Kajizar, 1975 ; Kwiecinska, 1980). Cela a permis une corrélation assez exacte des stades de transformation du graphite et des états prégraphitiques au degré de métamorphisme des roches hôte déduit des paragenèses des minéraux silicatés (Kwiecinska, 1980). En Roumanie de telles recherches ont été entreprises sur l'anthracite de la formation de Schela (Popescu, 1982) et sur les houilles anthraciteuses de Banat (Preda, Nedelcu, 1983).

Les dépôts charbonneux des Carpathes Méridionales se trouvent dans des conditions structurales différentes et ont des âges différents, fait qui donne la possibilité de comparer le degré de carbonification avec le métamorphisme des roches hôte, pour arriver à des considérations concernant les causes du degré différencié de métamorphisme des dépôts synchrones.

Les résultats des recherches qu'on va présenter se basent spécialement sur les investigations de la reflétivité de la vitrinite des échantillons de charbon d'âge Carbonifère de Baia Nouă-Cucuiova et Lupac et de charbon d'âge liasique d'Anina, Pregheda et de la formation de Schela. De plus, nos recherches ont été effectuées aussi sur certains échantillons de graphite de Baia de Fier. Parallèlement aux investigations optiques quantitatives, on a analysé le même matériel par la diffraction Rx.

Considérations géologiques

Dans les Carpathes Méridionales se trouvent des charbons humiques différents en tant qu'âge et degré de carbonification tant sur

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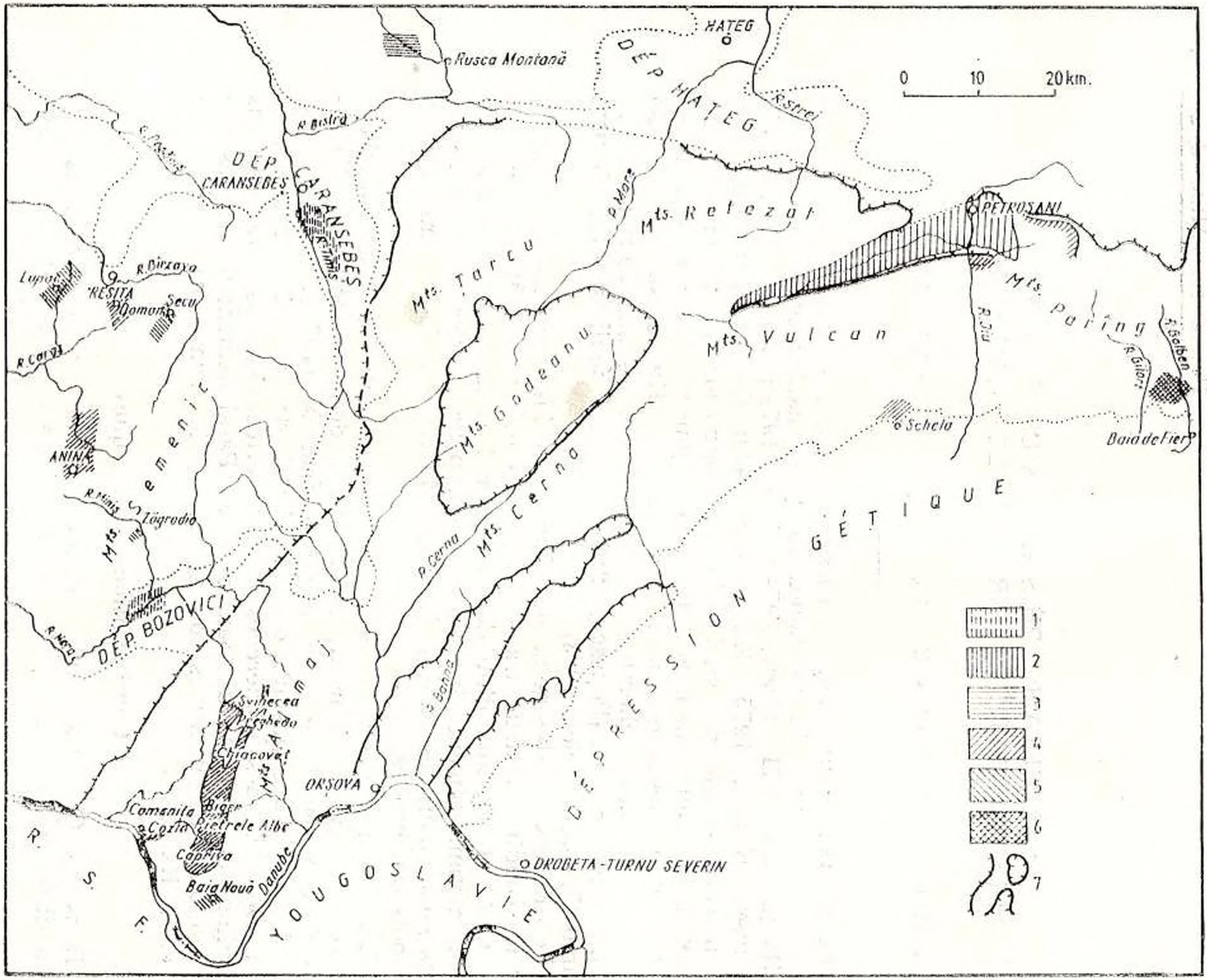

Fig. 1 — Répartition des formations à charbons sur l'esquisse des Carpathes Méridionales.
1, Sarmatien-Tortonien ; 2, Miocène-Oligocène ; 3. Crétacé supérieur ; 4, Lias ; 5, Carbonifère ; 6, cristallin ; 7, ligne de chevauchement (domaine gétique).

le domaine de sédimentation gétique que sur l'autochtone danubien. De même, on trouve des charbons dans les bassins post-tectoniques intermontagneux (fig. 1). Les formations cristallines de l'autochtone danubien contiennent à Baia de Fier, d'importantes quantités de graphite.

Les prémisses de la formation des charbons dans les Carpathes Méridionales ont existé déjà depuis le Précambrien (le graphite du cristallin de Lainici-Păiuș — l'autochtone danubien), mais les gisements d'importance économique se sont formés à partir du Carbonifère, comme suit :

— sur le domaine gétique : Lupac et Secul — houilles anthraciteuses (Carbonifère) ; Anina et Doman — houilles (Lias) ;

— sur l'autochtone danubien : Baia Nouă-Cucuiova — houilles anthraciteuses (Carbonifère) ; Cozla-Camenița — houilles ; Pietrele Albe, Biger, Pregheda, Svinecea Mare — houilles anthraciteuses ; Schela, Valea Izvorului, Jieț — anthracite (Lias) ;

— dans les bassins post-tectoniques : Rusca Montană — charbon brun, houilles (Crétacé supérieur) ; Petroșani — houilles, charbon brun (Oligocène-Miocène) ; Caransebeș-Mehadia, Bozovici — lignite, charbon brun (Miocène).

Les recherches effectuées par Răileanu (1953, 1963), Năstăseanu (1964, 1973, 1978 etc.) ont conduit à la connaissance de la succession géologique des dépôts de charbon. Mateescu (1956-1968) a présenté des données concernant le métamorphisme des charbons. Enfin, Manolescu (1933), Bițoianu (1973) et Semaka (1962) ont mis en évidence la flore génératrice de charbons.

Pendant le Carbonifère les charbons se sont formés dans des bassins lacustres de basse altitude (bassins limniques) instalés après la phase d'orogenèse sudète pendant le Westphalien-Stéphanien par l'accumulation de la végétation de marais (gisements autochtones).

L'orogenèse asturienne a été accompagnée par des éruptions (roches éruptives, agglomérats volcaniques, tufs etc.) intercalées dans les complexes charbonneux de Baia Nouă-Cucuiova, qui ont diminué les conditions favorables au développement et à l'accumulation de la végétation. La sédimentation lacustre reprise pendant le Permien inférieur a cessé dans la phase d'orogenèse saalique accompagnée par de considérables phénomènes volcaniques.

Les phases tardives de l'orogenèse hercynienne ont soumis à l'érosion les dépôts paléozoïques et la base cristalline durant tout le Trias. Ce n'est qu'au début du Jurassique que la mer s'est installée dans les Carpathes Méridionales en occupant au début les zones de dépressions tectoniques dont les marges plus mobiles ont été couvertes temporairement par des marais favorables à la formation des charbons. Ces zones paraliques et aux charbons autochtones se sont superposées sur les zones mobiles du Carbonifère. Le nombre réduit des couches minces de charbons intercalées dans les dépôts, généralement grossiers, prouve une subsidence rapide et saccadée à courtes périodes de calme durant lesquelles se sont formés les schistes argileux à charbons.

A partir du Lias moyen le milieu marin s'élargit et la sédimentation continue jusqu'au Crétacé supérieur y compris. Les tectogenèses alpines (les phases autrichienne, soushercynienne, laramienne) ont plissé et replissé les dépôts mésozoïques et paléozoïques tout en formant les synclinoriums de Reşiţa-Moldova Nouă et de Sirinia-Svinecea, orientés NE-SO, affectés par plusieurs fractures longitudinales et transversales accompagnées souvent par d'importants décrochements. Ont été mis en place les nappes gétique et de Severin, des corps des roches magmatiques connus sous la dénomination de banatites qui donnent des phénomènes intéressants de contact à Rusca Montană.

A Petroşani, bassin paralique à charbons autochtones, les dépôts sont fortement tectonisés dans la partie d'ouest et les charbons présentent un degré avancé de carbonification.

Les bassins post-tectoniques miocènes contiennent des lignites passables aux charbons bruns. Les dépôts sont faiblement plissés et ils présentent des fractures à d'importants dénivellements.

Composants pétrographiques

Les composants pétrographiques identifiés par les analyses effectuées sur les sections polies par Mateescu (1956-1972), Biţoianu (1972), Popescu et al. (1982), Preda et Nedelcu (1983) sont présentés dans le tableau 1.

TABLEAU 1

Les composants pétrographiques des charbons de principaux gisements des Carpathes Méridionales

Gisement	Âge	Degré de carbonification	Microlithotypes, %				Subst. minérales, %	Mat. volatiles, %
			Vitrite	Clarite	Durite	Fusite		
Petroşani	Miocène-Oligocène	Houilles	45–88	5–30	1–7	2	5–7	27
Roman	Lias	Houilles	83–86	abs.	abs.	9–20	3–16	11–27
Anina	Lias	Houilles	18–72	2–3	7–75	2–50	5	28–36
Pregheda	Lias	Houilles anthraciteuses	61–86	abs.	abs.	6–31	6–14	2–4
Schela	Lias	Anthracite	75–93	abs.	abs.	8–25	5–10	2–6
Secul	Carbonifère	Houilles	62–86	0–10	abs.	1–8	8–27	18
Lupac	Carbonifère	Houilles anthraciteuses	90	abs.	abs.	5	5	3–4
Baia Nouă Cucuiova	Carbonifère	Houilles anthraciteuses	60–70	5–7	10–20	1–2	3–5	9–10
Baia de Fier	Précambrien	Graphite	—	—	—	—	—	—

Données optiques quantitatives

Les déterminations de reflétivité (R) pour la vitrinite ont été réalisées au microscope Amplivale Palphotométrie Karl Zeiss Jena pour les longueurs d'onde de 487 m, 552 m, 591 m, 658 m. en air à l'étalon de silice. A base de ces résultats on a élaboré les courbes de dispersion pour chaque type de charbon (fig. 2). Il en résulte que :

— les valeurs de R et de la biréflexion (△R) augmentent à mesure qu'on passe de l'houille à l'houille anthraciteuse-anthracite, ces valeurs étant plus grandes pour l'anthracite de Schela-Gorj ;

— on ne constate point de relation entre l'âge des charbons et le R, fait bien illustré si on compare les houilles de Petroşani (Oligocène-Miocène) à celles d'Anina (Lias) ;

— dans le cas des charbons du Carbonifère, les courbes sont sensiblement semblables. Mais ceux du Lias diffèrent beaucoup en fonction du gisement. Cela est visible tant dans le cas des charbons liasiques de Banat autant que dans celui de l'anthracite de la formation de Schela.

Les investigations Rx se sont réalisées sur des échantillons prélevés des gisements de charbons appartenant au Carbonifère, Lias et Oligo-Miocène à l'aide du diffractomètre Phillips à l'anticathode de Cu, filtre de Ni à 35 Kv, 15 mA, vitesse du goniomètre de 1°/min, vitesse de déplacement du papier de 600 m/h, étalon interne KCl.

Il en résulte que :

— les diffractogrammes des charbons carbonifères n'indiquent aucun reflet pour les plans basaux (002) et (004). Un comportement similaire ont aussi les houilles liasiques d'Anina et celles oligocène-miocènes de Petroşani. En échange l'houille anthraciteuse de Pregheda, toujours du Lias, est le seul type de charbon qui présente un reflet diffus pour le plan basal (002) ;

— les diffractogrammes de l'anthracite de la formation de Schela révèlent pour tous les échantillons analysés un caractère diffus des reflexions des plans basaux (002) et (004) en comparaison avec les diffractogrammes du graphite de Baia de Fier.

Pour le reflet (002) on a découvert deux ou trois valeurs qui sont plus grandes que d_{002} du graphite (3,36 Å) et varient entre 3,43-3,59 Å. Ces valeurs se superposent sur l'intervalle 3,40-3,75 Å, caractéristique aux phases de passage de charbons à graphite (Landis, 1971 ; Kisch, 1974 ; Kwiecinska, 1978).

Commentaire sur les résultats

La corrélation des données concernant la reflétivité de la vitrinite avec celles résultées de l'examination des diffractogrammes révèle que :

— les charbons carbonifères — houilles anthraciteuses — présentent une faible tendance de rangement de la vitrinite, fait soutenu par la baisse valeur du R ($<9^0/_0$) et de son anisotropie très réduite (△ R $<$ 1). Cette caractéristique relève le fait que les diffractogrammes

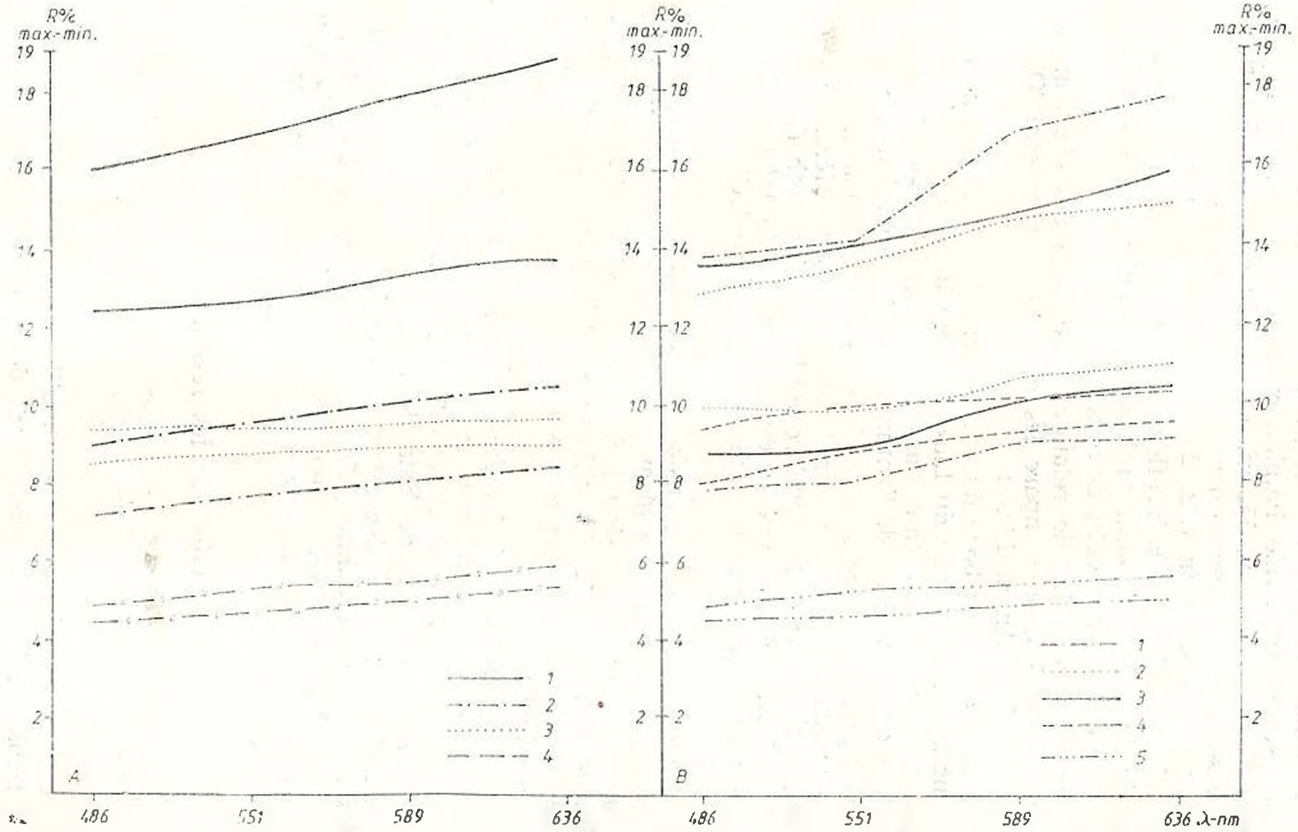

Fig. 2 — Les courbes de dispersion des mesurages R pour la vitrinite.
A : 1, graphite (Baia de Fier) ; 2, houille anthraciteuse (Lupac) ; 3, houille anthraciteuse (Baia Nouă-Cucuiova ; 4, houille (Petroşani) : B : 1, anthracite (Schela) ; 2, anthracite (Valea Izvorului) ; 3, anthracite (Jieţ) ; 4, houille anthraciteuse (Pregheda) ; 5, houille (Anina).

de ces charbons démontrent quelque fois seulement des inflexions très larges et de petite intensité pour les plans basaux (002) et (004) ;

— les charbons du Lias — houilles et anthracite — présentent des caractéristiques différentes tant au point de vue du R et △ R que de la diffraction Rx. Ainsi l'houille d'Anina présente les plus basses valeurs du R vis-à-vis de tous les charbons analysés. Par la diffraction Rx, on n'a pas remarqué, dans son cas, l'individualisation d'une certaine inflexion dans les plans basaux (002) et (004). En échange, l'houille anthraciteuse de Pregheda relève des valeurs sensiblement plus élevées par rapport à l'houille d'Anina et l'anisotropie est elle aussi un peu plus élevée.

Fig. 3 — Métagrès à l'anthracitoclaste aux „ombres de pression" de quartz fibreux. Formation de Schela — Vallée de Jieț, lame mince, N +, × 60.

En corroborant ces caractéristiques avec l'inflexion diffuse, mais bien individualisée, du plan basal (002), on constate l'individualisation incipiente de certaines structures prégraphitiques. Ces observations sont en corrélation aussi avec celles optiques qualitatives qui ont révélé à Pregheda le caractère anisotrope de la vitrinite. Le caractère plus avancé de carbonification de la vitrinite de ce gisement a été déterminé par l'action des solutions épigénétiques qui ont cristallisé la pyrite et le quartz (fig. 3). Il en résulte donc que les charbons liasiques et carbonifères de Banat ont un degré de carbonification intermédiaire entre les phases à contenu réduit en carbone (lignite, charbon brun) et les phases à contenu riche en carbone (anthracite, métaanthracite).

Par rapport aux caractéristiques pétrographiques des roches hôte, on constate une corrélation avec leur caractère intensément diagénétique, argumenté par la présence de l'illite, de la muscovite et du quartz secondaire de surcroissance sur des granoclastes initiaux (Fig. 4).

Fig. 4 — Schiste à pyrophyllite et chloritoïde prismatique radiaire. Formation de Schela — Gorj, lame mince, N +, × 60.

L'anthracite de la formation de Schela (Schela-Gorj, Izvorul-Jiu, Jieț), la troisième catégorie de charbons analysés, se place par ses caractères roentgénostructuraux et optiques quantitatives (R) dans la catégorie des phytoclastes (phase de transition) caractéristiques à l'anchimétamorphisme (Popescu et al., 1982). Qualitativement, la vitrinite qui représente le microlithotype majoritaire est visiblement anisotrope et à Schela-Gorj on observe même des bandes à caractère graphitique, fait justifié par les valeurs relativement élevées du R (14-18%) et du $\triangle$ R (fig. 2). Dans ce cas, les roches hôte, les schistes à pyrophyllite et à chloritoïde, les métagrès et les schistes à paragonite-muscovite, ont des caractéristiques minéralogiques qui, au point de vue du degré de métamorphisme, se trouvent en corrélation parfaite avec le degré de carbonification de la vitrinite (tabl. 2) (fig. 5, 6).

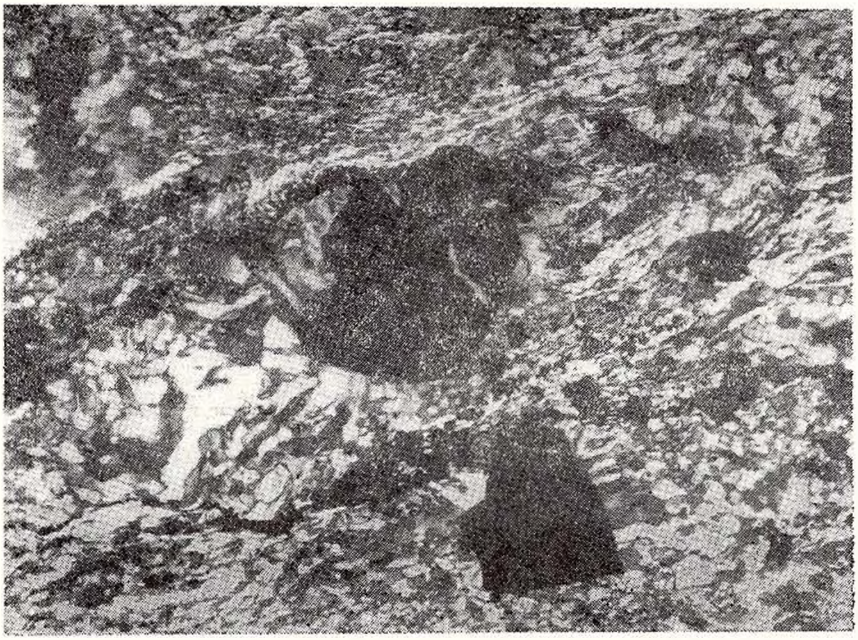

Fig. 5 — Vitrinite fragmentée et cimentée du quartz épigénétique et de la pyrite. Pregheda-Banat, section polie, N II, × 80.

Fig. 6 — Grès quartzeux à l'illite/muscovite sur des fissures et en nids. Pregheda-Banat, lame mince, N +, × 100.

TABLEAU 2

La relation entre la reflétivité (Rmax en air), l'espace interréticulaire(d_{002}) des phases de transition de charbons à graphite-minéraux silicatés, indicateurs des zones, stades pétrogénétiques (Carpathes Méridionales)

Rang des charbons	Rmax air 552 m %	d_{002} (R)	Minéraux silicatés indicateurs	Zones: stades pétrogénétiques
Houille Houille anthraciteuse	5,3—5,5 8,8—10,1	>3,70	illite/séricite, quartz épigénétique	Diagenèse
Anthracite	13,8—14,2	3,43—3,59	pyrophyllite, paragonite/muscovite chloritoïde	Anchimétamorphisme
Graphite	>16*	3,36	muscovite, biotite	Métamorphisme

* Valeurs déterminées en huile

Conclusions

Les charbons des Carpathes Méridionales apparaissent dans des formations sédimentaires afférentes au domaine danubien et au domaine gétique et dans des bassins post-tectoniques. Les périodes où les charbons se sont accumulés en importantes quantités ont été le Carbonifère, le Lias et l'Oligo-Miocène.

Les caractéristiques pétrographiques et optiques quantitatives (R) et roentgénostructurales différencient d'une part les charbons d'âge liasique, en fonction de la position de la formation hôte dans l'autochtone danubien, et relèvent d'autre part le caractère similaire en ce qui concerne les charbons du Carbonifère tant du domaine gétique autant que du domaine danubien.

Cette comparaison est déduite des conditions géologiques essentiellement différentes où se sont formés les charbons du Carbonifère — bassins limniques —, par rapport à ceux liasiques — bassins paraliques.

L'évolution géologique ultérieure des formations liasiques à charbons a conduit à leur placement dans des sous-unités structurales différentes du domaine danubien, fait qui a déterminé une différentiation des conditions thermiques qui ont conduit à de différents stades de carbonification de la masse végétale. La formation de Schela, à l'anthracite, a été surmontée, à la suite des phases d'orogenèse autrichienne et laramienne, tant par les formations de la nappe de Severin autant que par celles de la nappe gétique, à l'épaisseur cumulée qu'on estime à plus de 6 km (fig. 7). Ainsi, la température d'approximative-ment 200°C caractéristique au moment de la formation de l'anthracite et du pyrophyllite (Popescu, Constantinescu, 1982) a été atteinte dans les conditions d'une marche géothermique normale (33m/1°C).

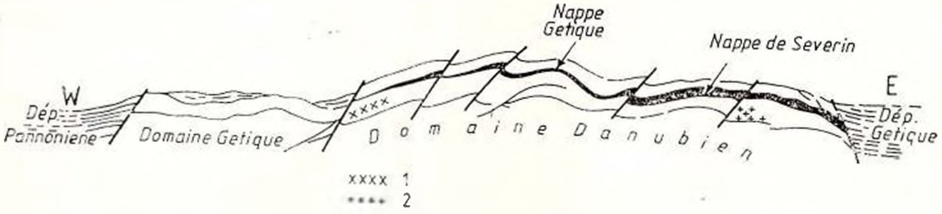

Fig. 7 — Coupe schématique montrant l'évolution tectonique des Carpathes Méridionales pendant le Sénonien-Paléogène (d'après Codarcea, 1940).
1, zone de Svinița-Svinecea avec les gisements de Cucuiova-Baia Nouă et Biger-Pregheda ; 2, formation de Schela.

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CONTINUITY, PERIODICITY AND EPISODICITY IN MAGMA GENESIS PROCESSES ASSOCIATED TO THE CLOSING OF THE ALPINE OCEAN IN THE CARPATHIAN AREA

BY

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The incompatibility between Stille's concept regarding the evolution of orogeny and associated magmatic processes — which assumes the alternation of short periods of orogeny and magmatic activity with long quiet periods — and the model of global tectonics — which implies the continuous movement of lithospheric plates and the continuity of associated processes —have been noticed and commented upon as early as the seventies (Evernden, Kistler, 1970). As regards, for example, the North American Cordillera it was shown that the periodicity of tectogenesis and of magmatic activity was apparent ; these were characterized by episodicity only in places, but they belonged to phenomena essentially marked by continuity along the continental margin (Gilluly, 1973). In order to account for the discontinuity of magmatic activity in time and space (along the plate margin) the following hypotheses are stated : (a) the occurrence of several plates subduction-related to the North American Plate, which led to (b) variations of compositions of subducted material and (c) variations of subduction velocity in different sectors of continental margin ; it is to note that at least the former category of variations may occur within a single plate too ; (d) variations of features of subducted lithosphere or of the upper mantle in which subduction takes place ; (e) variations of depth of low velocity zone of seismic wave propagation (LVZ), which could have stimulated or attenuated the magma genesis process.

In the Carpathian area — especially in the East and South Carpathians and Apuseni Mts — the tectogenesis induced, at the end of Mesozoic and beginning of Tertiary, by the collision of Euroasiatic plate with the plates in the south-west and south is systematized in two major deformation phases : Dacidic and Moldavian phases. The magmatic activity brought about by collision-related subduction has been delimited, for a long time, into "Upper Cretaceous-Paleocene

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magmatic activity" (corresponding to the former "banatites") and "Neogene volcanic activity". Later on, the former was divided into a Turonian-Senonian stage, characterized by andesitic associations and a Paleocene one, characterized by granodioritic associations ; the latter was divided into a Lower Badenian stage and an Upper Badenian-Pliocene stage, characterized by slightly different features in different regions ; the Neogene volcanism exhibited activity "phases", with no evidence of their length nor of the quiet periods between.

The dating of these stages was exclusively based on stratigraphic criteria, because of which the continuity of magmatic activity was never possible to be discussed in detail. The main idea, unanimously accepted and clearly discussed, pointed to an important discontinuity — corresponding to the Eocene, Oligocene and lowermost Miocene, that is ca 15-ca 18 m.y. B.P. — between Upper Creteaceous-Palogene magmatic activity and Neogene volcanic activity. Although the stratigraphic image could account for the continuity of magmatic activity within each stage, it is quite obvious that as far as the tectonic activity was thought to occur during short stages delimited by quiet periods, the associated magmatic activity was characterized by periodicity. Thus, the major discontinuity was implicitly accompanied by two other ones, less important, between Upper Cretaceous and Paleocene and Lower Badenian and Upper Badenian respectively.

The idea of periodicity of magmatic activity was generated mainly by the insufficient information about the age of rocks ; by the lack of radiometric determinations, some other facts added.

The increasing number of mining and drilling works carried out in the last 20-30 years has revealed the ignoring of the products of subduction related magmatic activity ; in several Neogene volcanic areas, there are completely covered subvolcanic bodies the age of which was proved — by recent radiometric determination — to be different from the age of volcanic rocks, generally greater than the latter. Thus, the lower limit of Neogene volcanic activity is still a problem as far as it has been stated by taking into account the age of rocks emplaced under subaerial conditions only ; this problem arose many years ago and remained unsolved in the Ţibleş and Rodna Mts in connection with the age of some subvolcanic bodies related, in outcrops, to Paleogene rocks only. Another cause was the ignoring of the true relations between magmatic rocks and magma generating subduction processes and especially the ignoring of the fact that during some periods and in some sectors the compression generated oceanic lithosphere-oceanic lithosphere subduction ; the lack of magmatic products on the continent, during some periods and in some sectors, was attributed to the complete cessation of magmatic activity, by omitting the traces of this activity present in basin deposits of appropriate age and location. It is also to note the rather late demonstration of the fact that the oceanic basin corresponding to North and East Carpathians closed gradually from NW to SE — and therefore, the subduction, the magma generation processes and the emplacement of rocks had a similar evolution — and thus favoured the interpretation of age differences among rocks from different sectors as products of independent activity "phases" or "stages". Finally, one should mention that due to the presence in the Carpathian area of two oceanic basins — one corresponding to the East and South Carpathians and one to the West Carpathians (Fig. 1) —

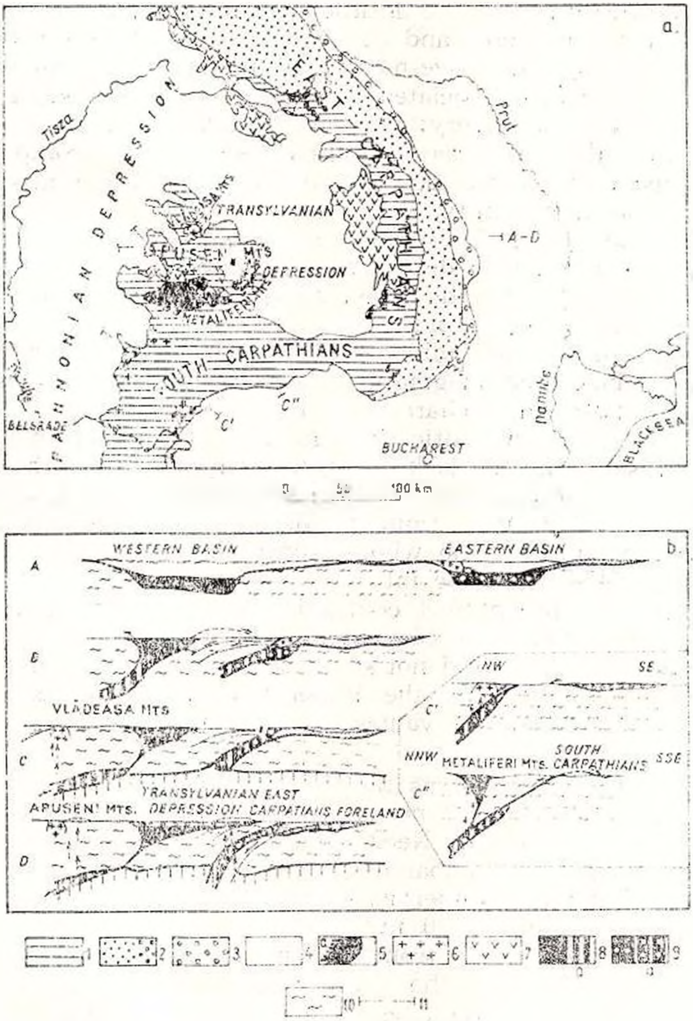

Fig. 1 — Sketch of geological structure of Carpathian area and its evolution during the Alpine time (acc. to Rădulescu, Săndulescu, 1973, and Săndulescu, 1980).

1, inner zones ; 2, outer zones (flysch) : a, outer Dacidic flysch ; 3, molasse ; 4, posttectonic formations and foreland areas ; 5, mafic and ultramafic rocks : a, Transylvanian Nappes ; 6, Upper Cretaceous-Paleocene igneous rocks ; 7, Tertiary volcanic rocks ; 8, oceanic lithosphere : a, in consumption area ; 9, oceanized lithosphere : a, in consumption areas ; 10, continental lithosphere ; 11, location of sections. A, Trias-Neocomian ; B, Mesocretaceous ; C, C′, C″, end-Senonian ; D, Actual.

with different evolution, was created the impression of lack of magmatic activity during certain periods and in some sectors ; in fact, the differences which influenced the evolution of magmatic processes concerned but the compression and the location of subduction areas. The occurrence of the two basins and of the two subduction areas does not impede on treating the above-mentioned problem in common for all the magmatic processes associated to subduction during Alpine time in the Carpathian area, as everything was generated and controlled by a unique phenomenon all over the area, namely the compression of oceanic basins between the Euroasiatic Plate and the plates adjacent to its south-western extremity.

Recent radiometric determinations have allowed the specification of the age of rocks under discussion and have led mainly to stating some new stages of magmatic activity associated to subduction during Alpine time (Lemne et al., in press).

By using all age determinations at hand (Fig. 2)[2] one may show that, on the whole, the magmatic processes associated to subduction during Alpine time were characterized by continuity and the stages identified represented intensifications as part of a "permanence".

First, it is to note the lack of any discontinuity in both the Upper Cretaceous-Paleocene and the Neogene magmatic activity. Around the moment 65 m.y. B.P. the frequency of age values is identical with the one corresponding to the whole Upper Cretaceous-Paleocene magmatic activity, while the time interval of about 3-18 m.y.B.P. included no gap, supporting the perfect continuity of rock formation processes in both cases ; if this was easy to foresee in the case of Neogene volcanic activity, which could not show but very reduced discontinuity, it is significant to note that the Upper Cretaceous-Paleocene time is remarkably covered by age values of a time interval considered to correspond to a discontinuity.

A second remark concerns the time interval ca 18-ca 55 m.y.B.P. corresponding to the "quiet" period between Upper Cretaceous-Paleocene magmatic activity and Neogene volcanic activity. This interval includes less values as compared to those prior or posterior to it ; however, the values are numerous enough and homogeneously spread, fact which shows that the 40 m.y. interval considered to correspond to the "quiet" period is far from marking the cessation of magmatic activity. It is probable that the number of values will increase in the future, on the one hand by discovering new rock bodies of this age and on the other hand by finding out that the rocks assigned either to Neogene volcanic activity or to Upper Cretaceous-Paleocene magmatic activity exhibit, in fact, intermediate radiometric ages.

In the case a longer interval of 5 m.y. is considered to represent a "break" of magmatic activity, then in the Carpathian area we find only the situation of this kind between 32.5 and 38.8 m.y.B.P. The distribution of values between ca 18 and ca 55 m.y.B.P. shows that this is not a simple decrease of the discontinuity interval length from ca 40 m.y. to 6-7 m.y. but it is due to the inaccurate study of reality. One may thus infer that between Upper Cretaceous-Paleocene magmatic activity and Neogene volcanic activity there is no discontinuity at all, but only an important decrease of intensity.

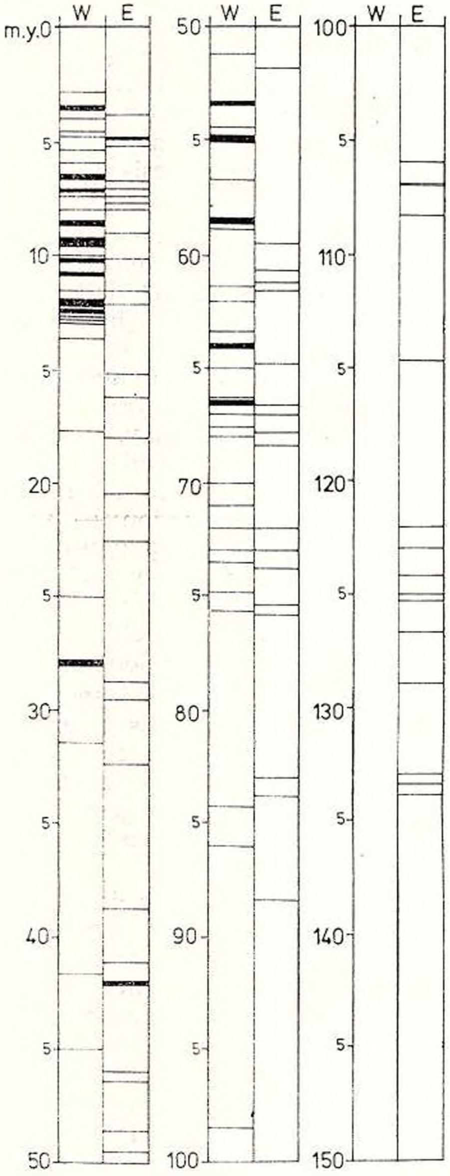

Fig. 2 — Distribution of age of rocks associated to subduction in the Carpathian area, between 0 and 150 m.y.B.P.
W, E, subduction associated rocks in the western and eastern basins.

The third remark to be made as a result of radiometric determination concerns the existence of intrusive calc-alkaline rocks of acid-intermediate chemical nature which belong obviously to subduction magmatic activity, older than the lower accepted limit of Upper Cretaceous-Paleocene magmatic activity ; however, it is to note a "discontinuity" of ca 10 m.y. between ca 88 and ca 98 m.y.B.P. The measured values regard, with a sole exception, the products associated to subduction in the eastern basin and reach ca 130 m.y., being homogeneously distributed between ca 100 and ca 134 m.y.B.P. Thus, it is possible that the magmatic activity started, episodically and with low intensity, just from the beginning of compression, corresponding to the Austroalpine phase (intra-Barremian) which was not completely lacking in magmatic activity as it had been considered so far (Rădulescu, Săndulescu, 1980) ; moreover, the values exceeding 120 m.y. arise the question whether magma genesis and implicitly the compression started as far back as the end of Neocomian.

The settling of compression regime as far as the end of Neocomian is also accounted for by other two remarks. On the one hand, it is to note the occurrence of "island arc" products within the Mureș ophiolitic zone (Savu, 1976) ; their radiometric ages, although of ca 66 m.y. — corresponding to the lower part of the interval when igneous rocks generated by subduction were massively emplaced in the continental lithosphere of central and northern parts of the Apuseni Mts — extend to about 122 m.y. and show that at the upper boundary of Neocomian compression had started in the Carpathian area and generated in the western basin an oceanic lithosphere-oceanic lithosphere collision and subduction — with an island arc formation — before the appearance of a subduction area at the margin of continental lithosphere and the emplacement of magmas in it. On the other hand, radiometric ages of 120-125 m.y. reported for some granitoids in the South Carpathians (Bîrsa Fierului) and considered to represent deformation stages, indicate a tension and compression regime, too.

The special attention paid separately to the products associated to each of the two basins leads to a fourth remark : most of time, the magmatic activity progressed concomitantly in the two areas ; this demonstrates the simultaneous occurrence of two subduction areas at the margin of continental lithosphere of two microplates and implicitly the presence of the same process of compression all over the Carpathian area.

There are few instances in which the magmatic activity took place in only one of the two zones. The most important example of this kind is given by the already mentioned magmatic activity earlier than ca 100 m.y.B.P. ; the other ones corresponding to the intervals of ca 18-ca 25 m.y.B.P., ca 32-ca 42 m.y.B.P., ca 23-ca 29 m.y.B.P. and ca 52-ca 59 m.y.B.P. could stand for an insufficient information or could be accounted for by the statement below.

Although there existed two oceanic basins and several microplates, the lithospheric compression in this area of the Earth, from Mesozoic to Tertiary, was obviously a unitary phenomenon ; formerly, this is to be considered as a whole and secondly independently for each of the two basins. It is absolutely possible that the two basins reacted at various moments either similarly or differently to the general tension. Thus, in one basin could be active a subduction zone situated at the margin of the continental block, while in the other one compression could generate oceanic lithosphere-oceanic lithosphere collision and subduction inside it ; or even at certain times tension could produce the compression of one of the two basins, while the other one was inactive. In case this is true, the relationships between the age of igneous rocks emplaced in and on the continental lithosphere and the continuity of subduction magma genesis and emplacement — on the whole and in each basin — should be considered from a different point of view : the absence of rocks emplaced in and on the continental lithosphere at a certain time and in a certain area does not necessarily mean the cessation of magmatic activity, of subduction or of compression ; this situation should be viewed for both basins by taking into account the possibility of subduction processes inside them and not only at their margins.

By examining the timing of subduction related magmatism during the Mesozoic and the Tertiary in the Carpathian area the following main ideas arose :

1. The lithospheric compression of present-day Carpathian area was continuous, starting from the end of Neocomian to the complete consolidation of the area. It occurred differently in different zones and at different times : simultaneously in both basins or predominantly in one of them, at certain times ; this was due either to the same process — subduction at the margin of continental lithosphere — or to different ones — coexistence of the previous one with oceanic lithosphere-oceanic lithosphere subduction. Consequently, compression is not always similarly reflected by deformation of sedimentary formations.
2. Subduction associated magmatic processes were continuous, on the whole, but (a) obvious intensity variations and (b) variations regarding their location and the location of their products are to be noted ; the variations are due to the different ways compression proceeded.

The note presented above is based on both published papers and several unpublished age determinations made by E. Călinescu (1976, 1977, 1978, 1979), N. Iosipenco (1980, 1981), M. Lemne (1975, 1976, 1977, 1978, 1979, 1980, 1983), S. Mînzatu (1977, 1980), O. Romanescu (1977, 1978, 1979, 1980, 1983), M. Soroiu (1974), A. Tănăsescu (1975, 1977, 1978, 1979, 1980, 1983), E. Vîjdea (1975, 1976, 1977, 1978, 1979, 1980, 1983) during investigations carried out together with A. Ştefan (1983), T. Berza (1980), M. Borcoş (1977, 1978, 1979, 1980, 1983), G. Istrate (1977), H. Savu (1983), D. Russo-Săndulescu (1980, 1983), I. Tiepac (1977), G. Udubaşa (1977, 1983) ; most of these reports are to be found in the Archives of the Institute of Geology and Geophysics in Bucharest.

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[2] Figure 2 does not show the frequency of each value, so that the conclusions inferred are exclusively related to the continuity of magmatic processes and not to subordinate ones, such as their intensity.

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NEOCRETACEOUS-PALEOGENE SUBDUCTION IGNEOUS ROCKS IN THE ROMANIAN CARPATHIANS — MUTUAL RELATIONSHIPS, SUCCESSION AND AREAL DISTRIBUTION

BY

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Introduction

These last years important details have been supplied regarding the petrology and the age of the Carpathian Neocretaceous-Paleogene igneous rocks, generally called banatites. It is to note here that the collective term of "banatites" proposed by Cotta (1865) for a group of rocks rather varied from mineralogic and structural view points, most of them granodiorites, was subject to controversed evolution ; some specialists defined them from strictly petrographic point of view, while others regarded them as an Upper Cretaceous-Paleocene petrographic province developed in the western part of Romania, south of the Danube in Yugoslavia and farther to the Balkans. Recently, some scientists have delimited the older extrusive products from the younger intrusions. One often mentions the "Laramian igneous rocks" which stand for a wider time interval ; in our opinion, this term is however inadequate, as in the main it relates the subsequent igneous rocks, placed within a wider time interval, to a stage of tectogenesis (deformation of crust by compression) which corresponds to a relatively short time interval, thus leading to the wrong assignment to syncinematic magmas. Finally, it is to consider Cotta's definition and to adopt the term of banatites as more adequate from petrologic point of view and deprived of restrictions as regards their assignment to the time intervals.

The occurrences of Neocretaceous-Paleogene igneous rocks in outcrops, boreholes or disclosed by geophysical researches (Visarion, Săndulescu, 1979) as well as their location with regard to the major tectonic units of the Romanian Carpathians allow their assignment to two big zones : (1) the South Carpathians, where they are related genetically to the oceanic crust consumption in front of and below the Getic Domain and (2) the Apuseni Mts, where they are related to the oceanic crust

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consumption in the Transylvanides (Rădulescu, Săndulescu, 1973 ; Săndulescu, 1980).

Several granitoid intrusions, which either cross or occur in the immediate neighbourhood of ophiolitic complexes (in the South Apuseni Mts or distinguished on magnetic maps below the molasses of the Pannonian Depression) are uncertain as regards both their location and age (Pl.).

Within each of the two distinct zones (South Carpathians and Apuseni Mts) volcanic activity and intrusive processes took place. The age of intrusions was considered posterior to volcanics, according to the succession relationships in the Rusca Montană basin in the Poiana Ruscă Mts (Giușcă et al., 1966 ; Kräutner, Kräutner, 1972).

In the Apuseni Mts, the age of intrusions was similarly considered, although the partly similar mineralogical and chemical composition of volcanics and plutonics (Istrate, 1978 ; Ștefan, 1980) as well as the volcano-plutonic complex in Vlădeasa (Giușcă et al., 1969) account for the relatively simultaneous presence of extrusive and plutonic processes.

The intrusive processes within Neocretaceous-Paleogene igneous rocks surpassed the volcanic activity, fact which calls for clearing up the problem related to the occurrence of some unconsanguineous magmas all over the investigated time interval. Thus, the study of some composite massifs such as Bocșa and Surduc (Russo-Săndulescu et al., 1978, 1983b) as well as the examination of the K/Ar radiometric ages obtained for these plutons allowed the elaboration of a model (Russo-Săndulescu et al., 1983a), the validity of which is discussed upon in the following pages referring to the banatitic rocks in the Romanian Carpathians.

Neocretaceous-Paleogene Igneous Rocks in the South Carpathians

An outlook on their occurrence (Pl.) points to the prevalence of plutonic rocks in the western, innermost part of Median Dacides, mainly within the Supragetic Nappe area or exceeding a little the front of this nappe, "the area of plutonic banatites" (Russo-Săndulescu, Berza, 1977). However, the volcanic rocks occur in the largest areas of the Getic Nappe, just like "hypabyssal banatites", namely within an outer area as compared to the South Carpathian structure.

The study of time and space relationships as well as the petrochemical composition of banatites from the plutonic area of the South Carpathians point to two distinct magmatic stages :

1. *Coniacian-Maastrichtian stage* (K/Ar ages of 87-68 m.y.), represented by complex intrusions which form big plutons (Bocșa 1 and Bocșa 2, Surduc — Russo-Săndulescu et al., 1978, 1983b) and small dykes in places.

The early pulses of gabbroic magma are supposed to have started their evolution at greater depth, under the conditions of a relative "tectonic quiescence", pointed out by the occurrence of layered gabbronorite and anorthosite nodules, evincing initial cumulate crystallization. The basic magmas which generated these rocks characterized by primitive chemistry (SiO_2 — 40-45% and MgO — 7-14% at Surduc —

Russo-Săndulescu et al., 1983b) could be considered as the least modified meltings. The repeated magma injections, or perhaps only the mixing up of initial magma and some locally fractioned "lots" determine at the present-day level of the intrusion an irregular distribution of anorthosite and gabbronorite nodules or only of cumulate crystals (basic euhedral plagioclase or corroded basic plagioclase nuclei, clino- and orthopyroxene), characterized by a new type of differentiation (?), "schlieren"-like, varied as regards their size and chemical-mineralogical composition (gabbroic, monzodioritic, monzonitic and even sienitic). The alkaline potassic nature of magmas is accounted for by the occurrence of biotite and potash feldspar in the gabbroic schlieren as well as by the occurrence of potassic sienites (which correspond chemically to shoshonites).

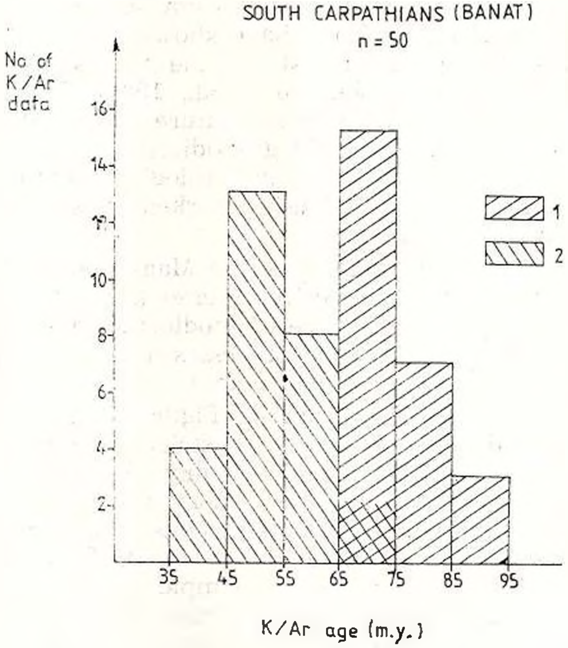

Fig. 1 — Histogram of K/Ar isotopic age analyses of banatitic rocks during the Conician-Maastrichtian (1) and the Maastrichtian-Eocene (2) in the South Carpathians.

As regards these former intrusions, already described (Surduc, Bocşa 1 and Bocşa 2 types), the apparent K/Ar ages corresponding to the Coniacian-Maastrichtian interval and culminating between 75-65 m.y. have been reported (Fig. 1).

The spatial distrubtion of alkaline potassic intrusions from Surduc to Hăuzeşti, generally follows the South Carpathian bend and they occur in the innermost areas as compared to the supposed subduction paleoplane (Rădulescu, Săndulescu, 1973 ; Săndulescu, 1980).

2. *Maastrichtian-Eocene stage* (K/Ar ages of 65-42 m.y.), marked by a different spatial distribution, following a north-southward line, is characterized by big plutonic intrusions in the north and the apophyses of a pluton in the south (eastern Bocşa or B_3, which due to a younger tectonic contact occur to the east of B_1 and B_2 intrusions ;

Ocna de Fier-Dognecea ; Oravița-Moldova Nouă), to the east of the previously described plutons.

Considering that the second stage intrusions occur in the area of former banatitic igneous rocks (Coniacian-Maastrichtian, as for example at Surduc), then the Maastrichtian-Eocene banatites cover a wider area, both in the Supragetic and Getic nappes.

The petrographic and chemical features as well as the field relationships within the plutons Bocșa 3, Ocna de Fier-Dognecea, Oravița-Ciclova-Moldova Nouă have shown that within the typical calc-alkaline Maastrichtian-Eocene stage, the intrusions exhibit multipulse behaviour (Russo-Săndulescu et al., 1984). Thus, were reported a stage of precursory dykes of basic nature (called stage Bocșa 3.1.) and a stage of emplacement of big granodioritic plutons (stage Bocșa 3.2). It is to note the chemical and mineralogical homogeneity within granodioritic plutons, which obviously marked most descriptions of the banatitic province of Romania.

The last products of the Maastrichtian-Eocene stage are represented by late, vein-like differentiates (rhyolites, dacites, andesites in places). These dykes cross the granodioritic plutons thus accounting for the persistence of plutonic processes at greater depths than the present-day levels of intrusion emplacement.

On the histogram of Figure 1, this second stage in the South Carpathians covers the Maastrichtian-Eocene interval, its maximum being located between 55-45 m.y. According to individual apparent age values, some bodies (precursory basic dykes) seem to have been emplaced prior to the end of igneous rocks evolution during the Coniacian-Maastrichtian stage (about 70 m.y.). Therefore the relationships among banatitic plutons are not simple at all, as far as sequences of veins belonging to the preceding stage (characterized by chemistry obviously typical of the last acid differentiates) may occur at the same time with the precursory basic dykes of the following stage.

Neocretaceous-Paleogene Igneous Rocks in the Apuseni Mountains

In the Apuseni Mts one distinguishes a southern area in which occur ophiolite nappes — the "Simic Metaliferi" Mts which belong to the Transylvanides (Săndulescu, 1975, 1980) ; in the west, this zone bends southwards below the Pannonian Depression (Visarion, Săndulescu, 1979). Some of the plutonic bodies, previously considered to stand for banatitic magmatic activity (Săvîrșin, Cerbia), have not been attested by the geochronological study. By taking these into account, as well as the less clearly stated position against the subduction paleoplane, the values obtained for the post-ophiolitic igneous rocks in the Simic Metaliferi were not presented on the histogram regarding the Apuseni Mts.

In the Northern Apuseni Mts the Neocretaceous-Paleogene igneous rocks occur as intrusions of different shapes and sizes in the Gilău and Bihor Mts where they pierce the crystalline schists and the sedimentary covers of several tectonic units : Bihor Unit, Codru nappe system and Biharia nappe system. The greatest part of banatitic rocks crop out in the Vlădeasa massif of the Apuseni Mts.

Most references to the banatitic sequence imply a parallel with the volcanic and plutonic series reported from Vlădeasa (Istrate, 1978 ; Ştefan, 1980). Although the characteristic aspect of the Vlădeasa massif is influenced by the occurrence of acid volcanics in varied facies, an ignimbrite rhyolite formation, earlier andesite and dacite flows, as well as younger intrusions (laccoliths, dykes or apophyses of plutons) of diorites, mainly granodiorites, tonalites and monzogranites do also occur ; microgranitic and aplitic rocks pierce the entire complex as last differentiates.

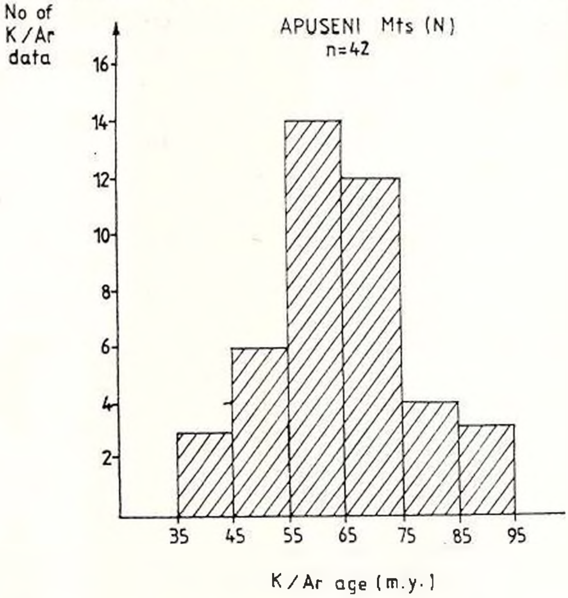

Fig. 2 — Histogram of K/Ar isotopic age analyses of volcanic and intrusive rocks in the Northern Apuseni Mts.

A more general systematization of the above cited authors supposes a former magmatic cycle mainly volcanic and subvolcanic (andesites-dacites-rhyolites) followed by a cycle represented by intrusions (diorites-granodiorites-granites).

The main features of Neocretaceous-Paleogene magmatic activity in the Northern Apuseni Mts are due to the presence of successive magma pulses, pointing to both the differentiation into intermediate magmatic chambers situated below the emplacement level of present-day intrusions and the repetition of some "cycles" of typically calc-alkaline rocks. It is also to note as characteristic feature, the close relationships between volcanic and plutonic formations, of comagmatic nature, temporally and spatially associated, and forming the Vlădeasa volcano-plutonic Complex (Giuşcă et al., 1969), as well as the presence, in this same area of the ignimbritic formation (Istrate, 1978 ; Ştefan, 1980).

By taking into account the above mentioned features, the radiometric ages reported from the Northern Apuseni Mts have been cumulated on the histogram of Figure 2, on which one may note a long time interval during which the Coniacian-Eocene magmatic activity took place (K/Ar age of 90-42 m.y., Lemne et al., 1981) with a wide maximum between 75-55 m.y.

According to the model proposed by Rădulescu and Săndulescu (1973), the entire banatitic magmatic activity in the Northern Apuseni Mts was considered to have resulted from the oceanic crust consumption in the Transylvanides, the remnants of which are still preserved in the major Tethysian suture in the Simic Metaliferi area (Săndulescu, 1980). This area is situated between two big groups of Carpathian tectonic units and delimits the Neocretaceous-Paleogene igneous rocks presented above.

Some of the main features which point to similarities or differences between the two large areas of Neocretaceous-Paleogene magmatic activity in the Romanian Carpathians are :

— All the banatites in the Northern Apuseni Mts cross the Codru and Biharia nappe systems of pre-Gosau age, probably intra-Turonian (Mediterranean). In the Trascău Mts there is an area in which a Neocretaceous (?) volcano-sedimentary formation and probably Laramian nappes are crossed by banatitic dykes, which point to an intrusive activity subsequent to the nappes (Russo-Săndulescu, Berza, 1976).

— South of the Simic Metaliferi, the South Carpathian banatites are located within a more external group of tectonic units, crossing the Supragetic nappes, partly of Mesocretaceous age with Laramian reworking, and the Getic Nappe initiated during the Mesocretaceous and completed during Laramian tectogenesis. From this point of view, in both areas of banatitic magmatism, it is to note that they are subsequent to the nappes which they cross. Another similar feature of banatites from the two zones is represented by the long time interval during which the magmatic activity took place, that is Senonian-Eocene (according to K/Ar age of 90-42 m.y.). It is to mention two maxima in the South Carpathians, and only one maximum value in the Northern Apuseni Mts, that is more homogeneous and on a wider time interval.

Finally, the occurrence of an ignimbrite formation both in the South Carpathians (in the Maastrichtian volcano-sedimentary formation at Poiana Ruscă) and in the Northern Apuseni Mts (Vlădeasa massif), which is prior to some calc-alkaline, generally granodioritic, intrusions, seems to point to an obvious similarity within the Maastrichtian-Eocene stage stated in Banat.

As compared to the model of banatitic magma generation in the South Carpathians, in the Apuseni Mts it does not consist of two stages, lacking in the basic magmas with alkaline potassic tendency and the evidence of initial cumulate crystallization of the first stage. The petrochemical nature of banatites in the Northern Apuseni Mts is more homogeneous resulting in sequences of calc-alkaline rocks (both volcanic and intrusive) typical of subduction magmas. It is possible that this homogeneity is due to the generation of magmas as a result of oceanic crust consumption in the Tethys Ocean (preserved in the present-day major Tethysian suture — the Transylvanides).

Nevertheless, in the South Carpathians, although characterized by oceanic crust consumption, this was generated within an intracontinental rift of Afars-Red Sea type (according to the model proposed by Săndulescu, 1980, 1983). The consumption of this crust started during Mesocretaceous tectogenesis and was perfected during Laramian tectogenesis, concomitantly with the end of Getic and Severin nappes overthrusting. This process has led to the appearance, during the first stage (Coniacian-Maastrichtian), of basic igneous rocks and alkaline-potassic differentiaion products, typical of a distention period, unconsanguineous with the calc-alkaline igneous rocks of the next stage (Maastrichtian-Eocene).

Conclusions

The Carpathian Neocretaceous-Paleogene igneous rocks are related to oceanic crust consumption by means of processes similar to classical subduction, while the igneous rocks resulted, both volcanic and intrusive, are the calc-alkaline types characteristic of subduction areas (andesite-dacite-rhyolite assemblage or their plutonic correspondents). The occurrence in the Banat region of some igneous rocks peculiar from both structural and petrochemical points of view (gabbronorites and anorthosites with layered structure, basic rocks with alkaline potassic tendency, the differentiates of which show even a shoshonite chemistry) points to the existence of some unconsanguineous magmas in the same area. However, by taking into account the two maxima corresponding to the emplacement of plutonic intrusions with peculiar petrochemical features, the magma genesis is difficult to be accounted for by the same simple subduction process.

As far as the recent studies have not solved yet the problem of the time necessary for the genesis and emplacement of magmas, as compared to the initiation of crust shortening and consumption phenomena, the simple reference to radiometricc ages without taking into account the tectogenetic and magmatogenetic conditions characteristic of each area, or even the structure of each pluton, could lead to conclusions inadequate to the Carpathian area.

Thus, it is quite probable that in the South Carpathians, the first magmatic stage (K/Ar age values corresponding to the Coniacian-Maastrichtian) depend on oceanic crust consumption, which started during the Mesocretaceous ; the emplacement of these magmas reached its maximum during the Maastrichtian, in the Getic and Supragetic nappes (in this region the Laramian tectogenesis had started much earlier).

The second maximum, between 55-45 m.y., took place rather late as compared to Laramian movements (just like it is the case with the first stage as compared to Mesocretaceous movements) in the South Carpathians.

The data above lead to a general conclusion on the emplacement stages of Neocretaceous-Paleogene magmas in the Carpathians. By merely taking into account the isotopic K/Ar ages one may infer the apparent continuation of magmatic activity ; however, the study of the two maxima of South Carpathian banatites corroborated with the characteristic petrology and tectogeneses from this area, reveals the peculiar conditions under which subduction, crust consumption, source areas, genesis and differentiation of magmas took place.

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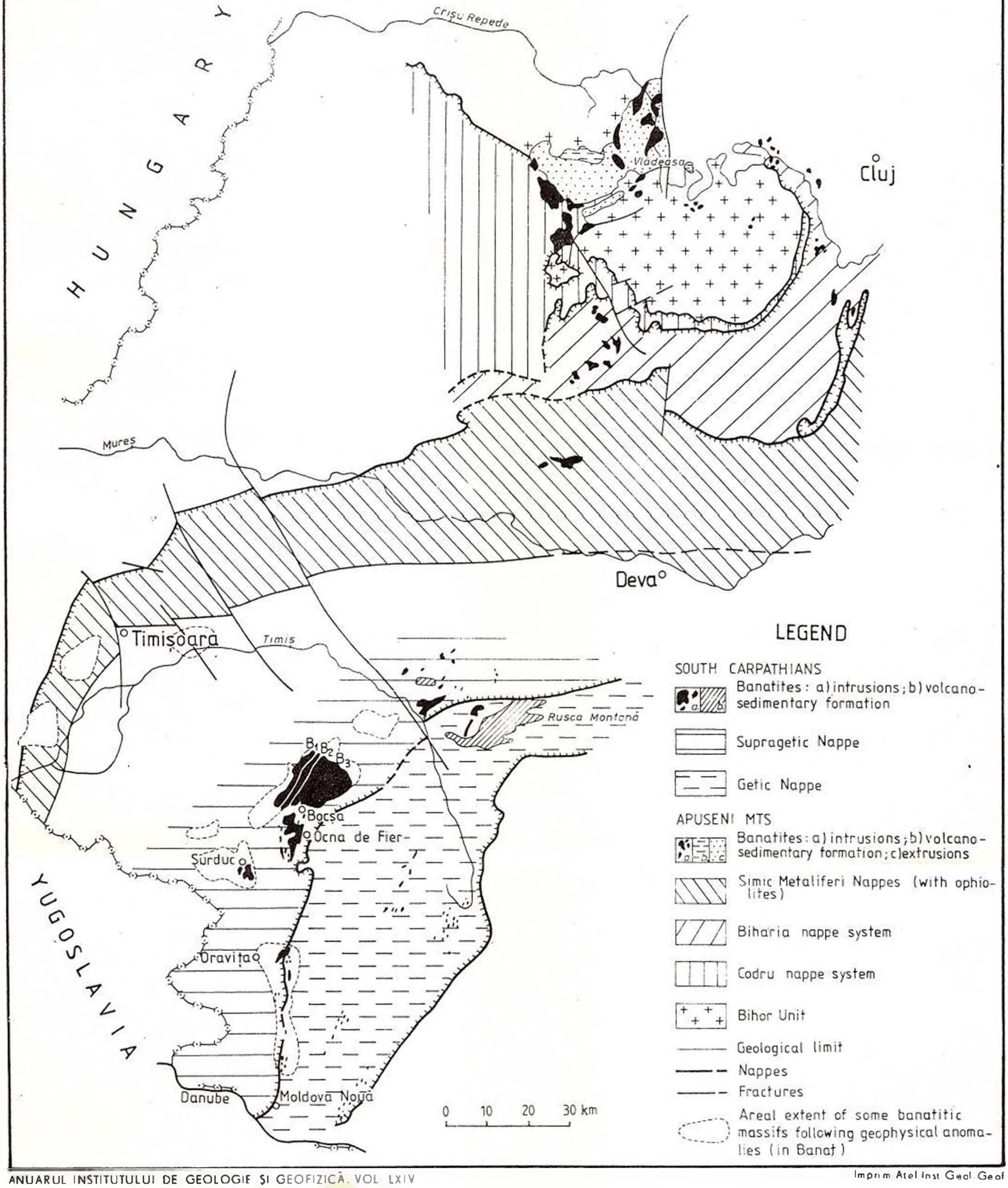
SKETCH OF DISTRIBUTION OF NEOCRETACEOUS - PALEOCENE IGNEOUS ROCKS IN THE ROMANIAN CARPATHIANS

(data acc. to Geological map of Romania, 1 : 1.000.000,
modified, Săndulescu et al., 1978, Visarion, Săndulescu, 1979,
Russo - Săndulescu, Berza, 1977)



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Pétrologie-Minéralogie

TRENDS OF THOLEIITIC MAGMA DIFFERENTIATION IN THE SHEETED DYKE COMPLEX FROM THE MUREŞ ZONE (ROMANIA)

BY

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1. Introduction

The observations concerning the differentiation of the basaltic magmas were made since the beginning of this century. From then on, important progresses were made, as today magmas are characterized according to the tectonic setting and their petro- and geochemical features. This paper presents the differentiation of the tholeiitic magma from which the sheeted dyke complex was formed in the Mureş Zone, its controls and trends.

2. Remarks on the Structure and Petrography of the Sheeted Dyke Complex

The sheeted dyke complex (O_2) is developed in the Drocea Mts (Fig. 1) on a length of 35 km and a maximum width of 10 km (Savu et al., in press ; Savu, 1983, in press a. b). North-westwards it overthrusts a scale of oceanic floor basalts (O_1) and together overlap the flysch (J_3-Cr_1), which encompasses the products of a bimodal volcanism. The 0.5-2 m thick dykes consist of basalts, dolerites and spilites, but in the Dumbrăviţa-Juliţa-Lupeşti region they alternate with dykes of gabbro, ferrogabbro, granophyre and albite-quartzdiorite which are also pierced by thin dykes of dolerite, ferrodolerite, basalt, gabbroporphyrite and dykelets (Rothery, 1983) of hyaloferrobasalt. Dykes and veins of felsite and albitic (trondhjemite) plagiogranite with xhenoliths of basic rocks metamorphosed in amphibole hornfelses are rarely found.

The sheeted dyke complex was formed during the spreading stage of the Mureş Ocean and its presence is a clear proof that ophiolites

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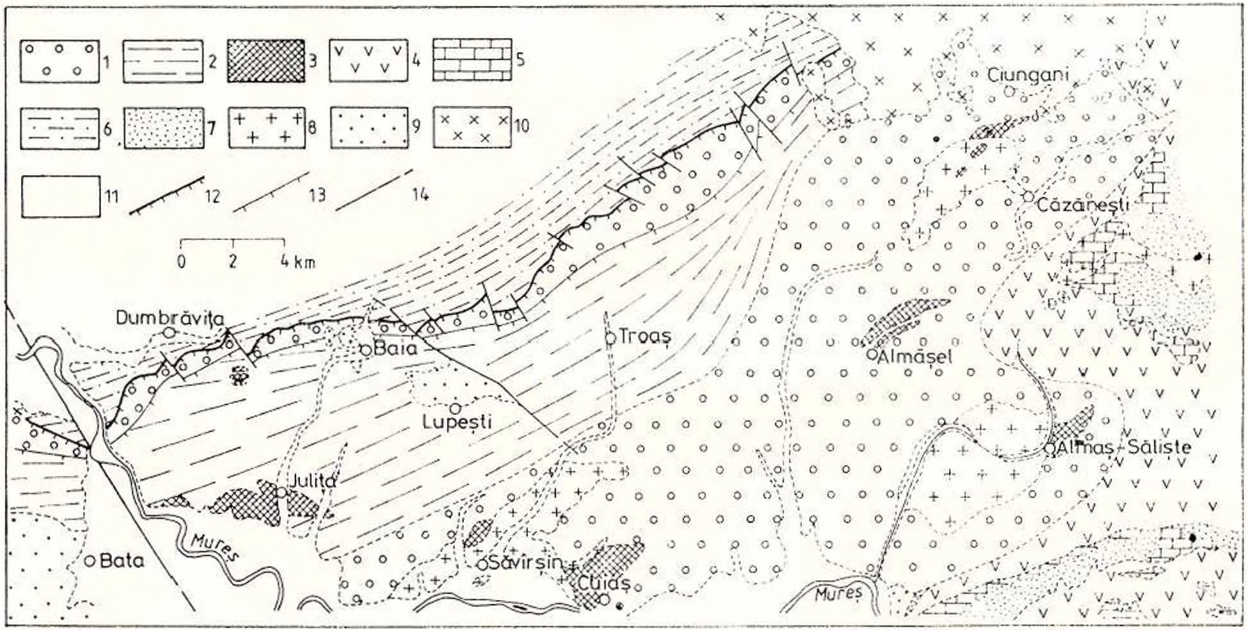

Fig. 1 — Sheeted dykes spreading in the Mureș Zone.

1, basalts complex (O_1) ; 2, sheeted dyke complex (O_2) ; 3, gabbro bodies ; 4, series of island arc volcanics (J_3-Cr_1) ; 5, Stramberg Limestones ; 6, flysch (J_3-Cr_1) ; 7, Upper Cretaceous ; 8, Lower Cretaceous-Paleocene acid intrusions ; 9, Neogene ; 10, Neogene volcanics ; 11, alluvial deposits ; 12, overthrust ; 13, reverse fault ; 14, fault.

from the Mureş Zone basement appeared in oceanic floor conditions (Savu et al., in press). It is similar to other sheeted dyke complexes in the world, which are considered to be formed in spreading conditions (Coleman, 1977).

The rock structures are intergranular, ophitic and hypidiomorphic-granular. The first mineral to crystallize is plagioclase (labrador An 50-70 or albite An 8-10), followed by clinopyroxene, magnetite and ilmenite ; the albitic felsites are an exception, where garlands of elongated crystals of clinopyroxene are formed before the albitic mass which encompasses them. The clinopyroxene is an augite (c Λ Ng = 44°) rarely diopside and in the albitic felsites a Ti-augite (c Λ Ng = 47°). Albitic plagiogranites contain a brown-greenish amphibole (c Λ Ng = 22°). In the granophires, sometimes in the albitic plagiogranites, mirmekitic, rarely micrographic textures are frequent. The presence of the latter ones and of some small interstitial crystals of finely twinned albite of low temperature seem to indicate the initial existence of a subsequently albitized potassium feldspar.

3. Differentiation of the Tholeiitic Magma

The parental magma of the sheeted dyke complex was an oceanic floor tholeiitic magma (Fig. 2). The previous researches on the tholeiitic magma differentiation have shown that this one had only one trend, according to which the residual magma gradually enriched in iron, alkalis and SiO_2 ; thus, a continuous curve is established on a FMA diagram (Wager, Deer, 1939). Our studies on the sheeted dyke complex from the Mureş Zone (Savu et al., in press; 1983 unpublished data) have shown that its rocks can be separated into the following groups[3] : 1) non-differentiated basic rocks, 2) differentiated basic (ferrobasaltic) rocks and 3) differentiated acid-albitic rocks (Tab.). These derived from the same tholeiitic magma, formed in a magmatic chamber (MacDonald, 1982) located in the mantle at about 100 km deep under the spreading centre. Their separation was not produced after a monovariable curve,

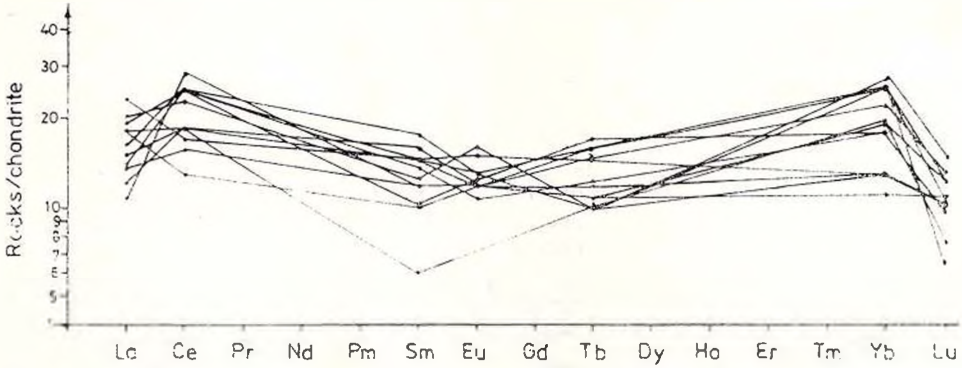

Fig. 2. — Chondrite — normalized REE patterns for non-differentiated basic rocks.

TABLE

Chemical composition of the three groups of sheeted dyke rocks (50 anal.)

Oxides and elements	Non-differentiated tholeiitic rocks	Differentiated basic rocks	Differentiated acid rocks
SiO_2 (%)	47.47–51.75	43.20–52.14	51.76–75.53
Al_2O_3	12.87–15.95	10.87–17.73	11.50–14.14
Fe_2O_3	1.36– 5.30	2.46–11.12	2.25– 6.77
FeO	3.26– 9.32	3.60–14.84	0.64– 9.64
MgO	6.10– 8.65	3.22– 6.21	0.21– 2 99
CaO	7.84–12.64	6.49–15.81	1.58– 7.03
Na_2O	0.86–3.70	2.31– 4.93	4.65– 5.71
K_2O	0.06– 0.88	0.05– 1.30	0.05– 3.39
TiO_2	0.60– 2.35	2.39– 6.95	0.34– 3.32
P_2O_5	0.08– 0.32	0.06– 1.64	0.12– 0.88
Ni (ppm)	30–200	14–127	3.5–18
Co	23–55	4–65	5.5–22
Cr	46–565	1.5–190	1.5–38
V	200–325	190–900	12–210
Sc	28–40	11–42	7.50–26
Zr	37–200	54–280	206–850
Y	14–55	10–87	60–150
Yb	1.7–6.5	4.8–10.5	6.5–18.5
La	<30	<30	<30–38
Ba	12–80	10–110	<10–60
Sr	95–240	18–240	43–155
Pb	<2–11	<2–5	<2–23
Cu	2–52	2.5–53	4–32
Ga	8.5–22	12–51	12–30
Sn	<2–3	<2–7	<2–4.5
FeO (tot)/MgO	0.70–2.0	2.0–5.85	3.0–18.75

but the tholeiitic magma differentiation had two trends : 1) a trend which led to the enrichment of one part of the magma in iron (Fenner, 1929) and 2) a trend which led to the enrichment of the residual magma in SiO_2, alkalis and gas, especially dissociated water (Fig. 3). One can notice that on the first trend (a) the differentiation starts from the non-differentiated basic magma and goes towards the FeO corner of the diagram, following a straight line which leads from the basalt-doleritic rocks to ferrogabbros and hyaloferrobasalts. On the second trend (b) the differentiation starts in the separation zone of differentiated basic (ferrogabbroic) rocks from the non-differentiated ones, following a curve which turns and goes towards the alkalis corner, reaching somewhere lower the FeO-Na_2O+K_2O side. Although the acid-albitic rocks are maintained in the tholeiitic domain, after its appearance and position, this curve remembers rather the intermediary magmas curve of Izu-Hakone type (Miyashiro, 1975, Fig. 4) ; this aspect would result as well from other geochemical peculiarities of the differentiated acid-albitic rocks. The behaviour of Ti (Fig. 4) as well as that of V (fig. 5) indicate the same two trends of tholeiitic magma differentiation.

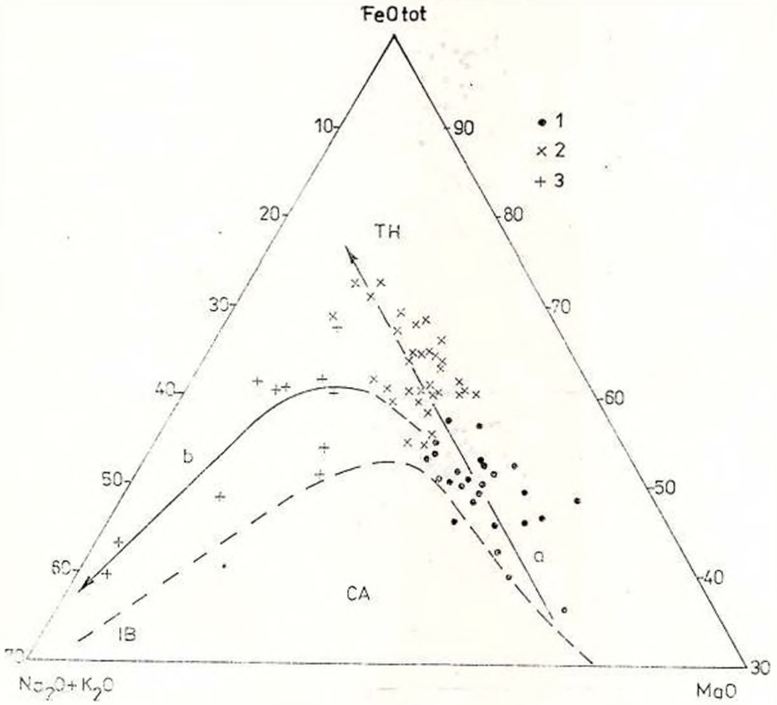

Fig. 3 — FeO tot — MgO — Na$_2$O + K$_2$O diagram with the separation curve (IB) of tholeiitic rocks from the calc-alkali ones, according to Irvine and Baragar (1971).

1, non-differentiated basic rocks ; 2, differentiated basic (ferrobasaltic) rocks ; 3, differentiated acid-albitic rocks ; a, differentiation line of ferrobasaltic magmas ; b, differentiation curve of acid magmas — the legend of this diagram can be used for the others, too.

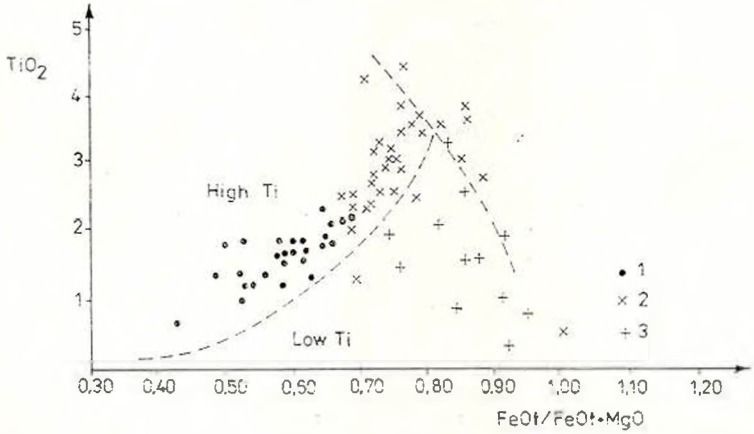

Fig. 4 — TiO$_2$-FeO tot/FeO tot + MgO diagram (after Serri and Saitta, 1980).

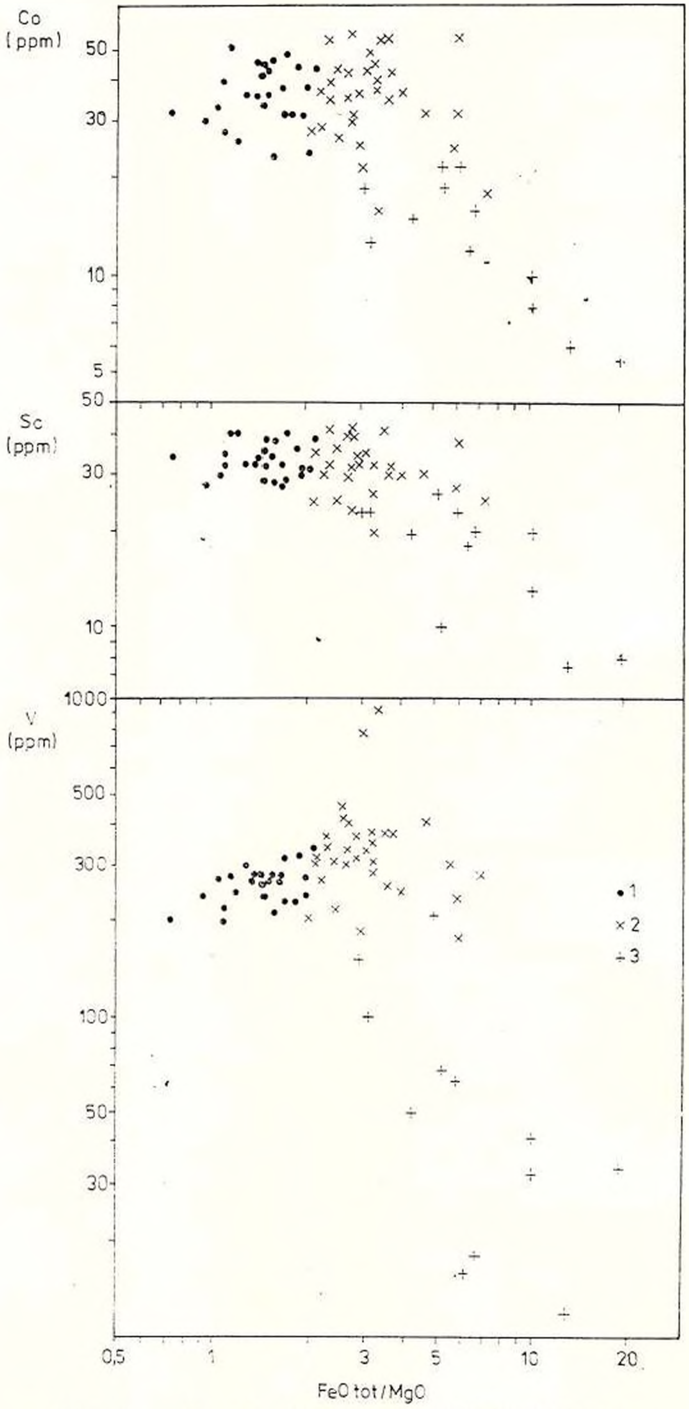

Fig. 5 — Co, Sc, V — FeO tot/MgO diagrams.

3.1. *Characteristics of the Non-Differentiated Tholeiitic Magma.* This magma approximately corresponds to the primary tholeiitic magma (Fig. 3), whose differentiation was possibly very weak. From this one there formed many rocks of the sheeted dyke complex (basalts, dolerites, gabbros) and evidently those of the basaltic complex (O_1). It belongs to the "high-Ti" type of magmas (Fig. 4), which according to Serri and Saitta (1980) shows that it was similar to the magma from median oceanic ridges. Within the rocks formed from this magma, the ratio FeO tot/MgO (Miyashiro, 1975) is lower than 2 ; the Na_2O content is below 3.70% and that of K_2O is seldom higher than 0.60%. Because of that the non-differentiated basic rocks are located in the normal tholeiitic field on the diagram in Figure 3, which is to be noticed as well on diagrams for minor elements (Figs. 5-7). The REE content is generally constant and the patterns of these elements are situated on the diagram in Figure 2 between the values of 6.5-20.8 of the rock/chondrite ratio.

3.2. *Characteristics of Differentiated Basic (Ferrobasaltic) Magmas.* These magmas represent the differentiated magmas enriched in iron (Fe_2O_3 = 2.46-11.12% ; FeO = 3.60-14.84%) and other chemical elements (Kennedy, 1948) from which ferrogabbros, ferrodolerites and hyaloferrobasalts have resulted. The differentiation of these rocks from the tholeiitic magma is clearly illustrated in the diagram of Figure 3, where they are situated in the tholeiitic rocks field ; they are continuously distributed along the line (a). Besides the iron, these magmas enrich in TiO_2 (1.28-6.95%), as it can be noticed in the diagram of Figure 4, as well as in V (Fig. 5). The Co and Sc contents are approximately constant as they show only a weak tendency of decreasing. A very strong decreasing for this group of rocks, as for the other two, is noticed for Ni and especially for Cr (Fig. 6). On the contrary, Zr, Y and Yb show a clear tendency of increase along the whole series of sheeted dykes (Fig. 7). Therefore, the rocks from the ferrobasaltic magmas group occupy on these diagrams as on other diagrams for minor elements, an intermediary position, as they are situated between the non-differentiated basic rocks and that of differentiated acid-albitic rocks.

3.3. *Characteristics of Differentiated Acid Magmas.* These magmas have resulted from the residual liquid from which there previously separated the chemical elements which led to the formation of differentiated basic (ferrobasaltic) magmas. From these magmas there formed some spilitic rocks, albite-quartzdiorites, granophyres, plagiogranites and albitic plagioaplites.

The acid magmas gradually enriched in SiO_2 (51.76-75.53%) and alkalis, especially Na_2O (4.65-5.71%), some minor elements such as Yb, Y and Zr (Fig. 7) as well as gas. They were depleted of Mg, Ca, Fe (Fig. 3), Ti (Fig. 4) and minor elements such as V, Co, Sc (Fig. 5) and Ni, Cr (Fig. 6). The value of FeO tot/MgO ratio is higher than 6. The fact that on all diagrams for minor elements, except that for V,

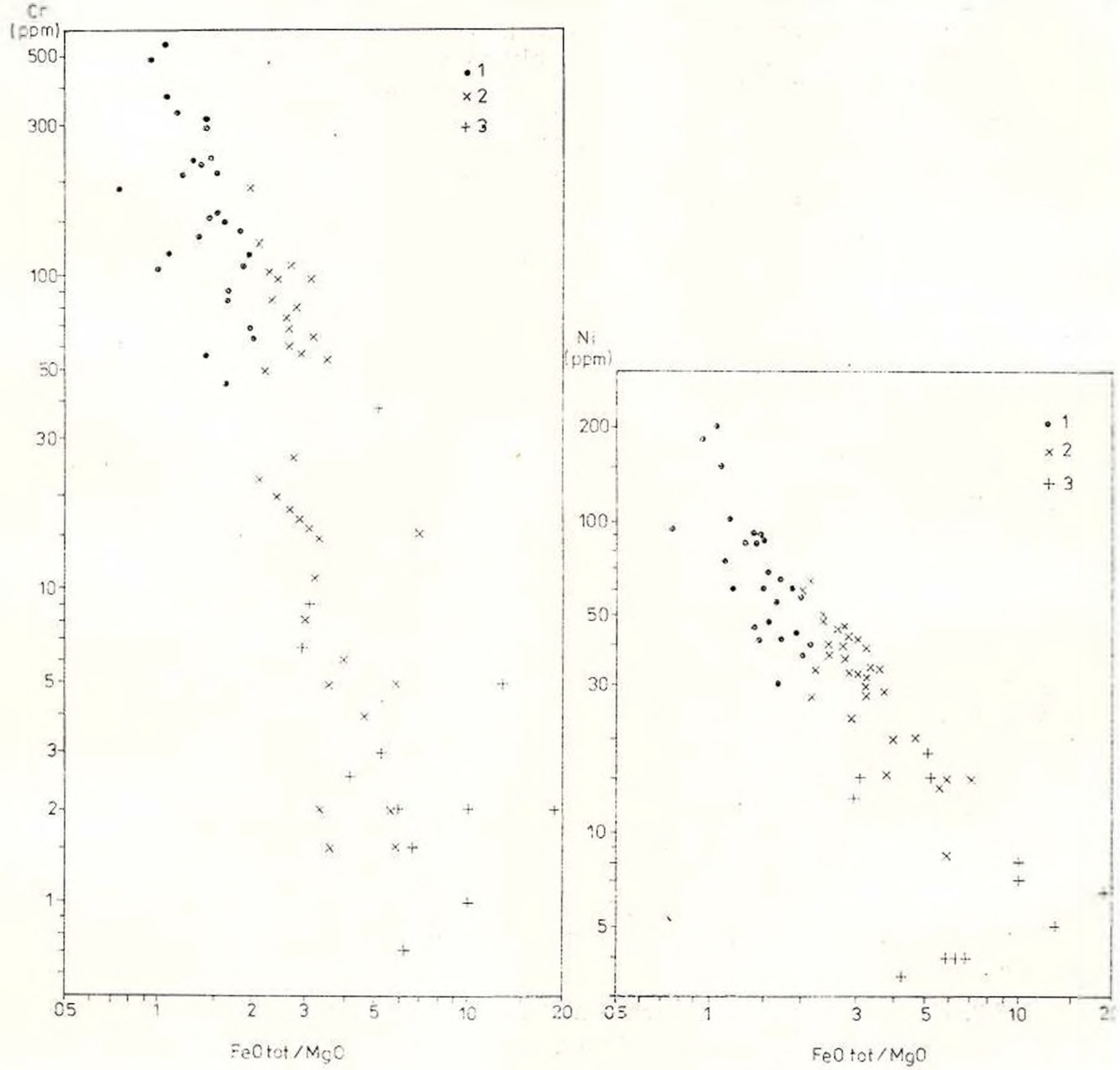

Fig. 6 — Cr, Ni — FeO tot/MgO diagrams.

the acid-albitic rocks are situated in prolongation of rock fields from the sheeted dyke complex, no matter if they indicate positive or negative correlations, clearly demonstrates that differentiated acid magmas were separated from the primary tholeiitic magma as the differentiated basic (ferrobasaltic) ones have done, and so they have no other origin. Another argument is the very low content of K_2O ($< 0.40\%$) of these magmas. They show as well the lowest contents of Sr and Ba as compared to the whole series of the analysed rocks.

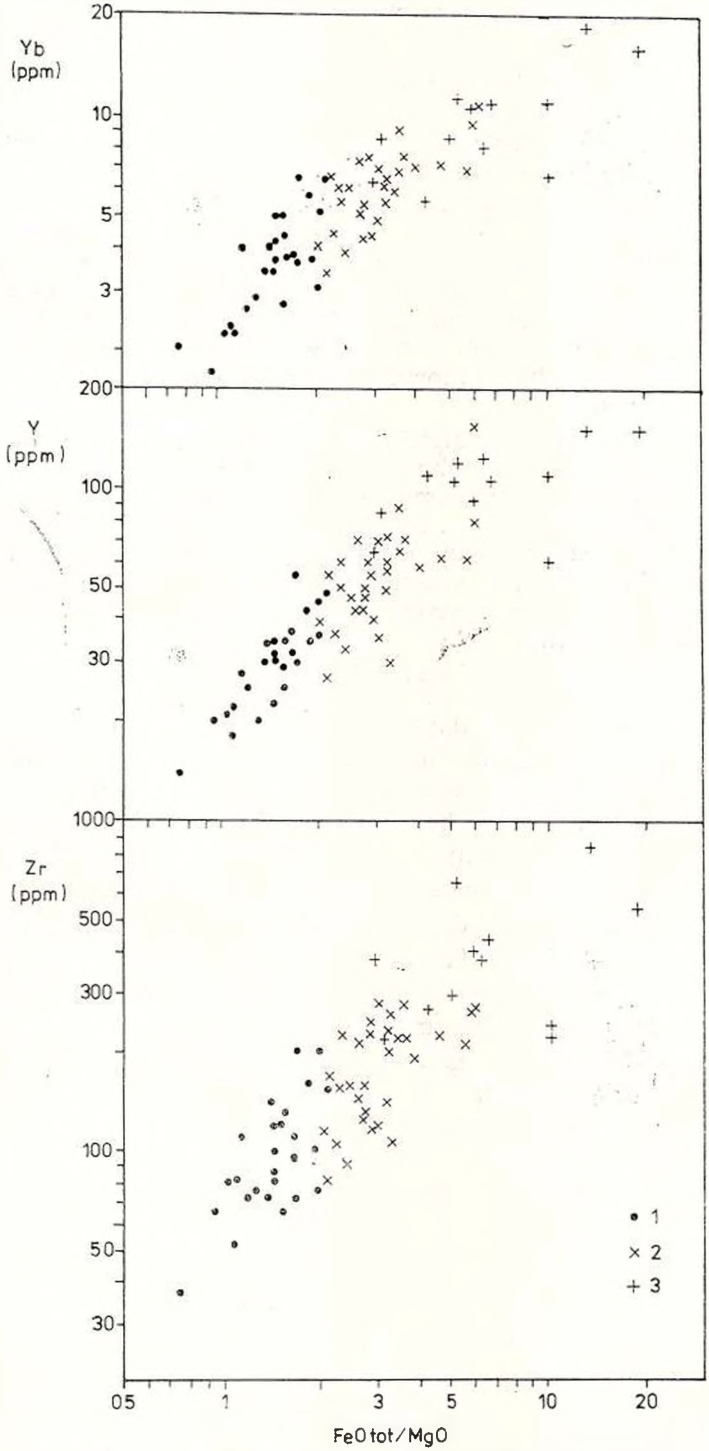

Fig. 7 — Yb, Y, Zr — FeO tot/MgO diagrams.

4. Final Remarks

The tholeiitic magma which was formed from the pyrolith originating in the magmatic chamber located in the mantle (Savu et al., 1981) is differentiated, following the below mentioned line which leads to the formation of the three rock groups.

Pyrolith → Primary tholeiitic → acid magmas
magma ↘
basic (ferrobasaltic) magmas.

It is possible that ultrabasic rocks (magmas) from the gabbro-peridotitic complex (O_3) of the ophiolitic series situated in depth would have been first separated from the initial pyrolith, so that the primary tholeiitic magma got the character of a "high-Ti" magma, SiO_2, Al_2O_3 and CaO saturated, which imprinted from the very beginning its differentiation trends.

As concerns the acid-albitic rocks associated to ophiolites from Corsica, Ohnenstetter and Ohnenstetter (1980) think that they were formed by imiscibility, a phenomenon that appeared after the beginning of ferrogabbros crystallization. In the case of the sheeted dyke complex from the Mureș Zone, we can see on the diagram of Figure 3 that the acid-albitic rocks curve (b) is separated from the differentiation line (a) in the same zone where ferrobasaltic magmas start separating from the normal tholeiitic one.

The cause of this differentiation consists in the chemical peculiarities of the primary tholeiitic magma, which allow the crystallization first of plagioclase, which determines the enrichment of the residual magma in Fe, Ti, V, Si, Na, lanthanides and other minor elements, as well as gas, especially dissociated water. These substances, by their character and quantities to be found in the residual magma, are incompatible with the formation of a natural magma. Therefore, the imiscibility phenomenon appears, due to which a ferrobasaltic magma is separated on the one hand, which will be the source of ferrogabbros with concentration of Fe, Ti and V and on the other hand an acid magma, rich in Si, Na and dissociated water which develops as an undercooled magma. In this magma there originated spilites, granophyres, albitic plagiogranites and sulphide mineralizations from the ophiolitic series (Savu, 1973).

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[3] Rocks (magmas) discrimination method: first there are separated non-differentiated rocks, with FeO tot/MgO < 2, then the basic (ferrobasaltic) differentiated rocks with FeO tot/MgO $= 2$-6, $SiO_2 < 52\%$, which are enriched in iron, Ti and V on the one hand, and the differentiated acid-albitic rocks with FeO tot/MgO > 3. $SiO_2 > 52\%$, $Na_2O > 4\%$, rich in La, Y, Yb and Zr on the other hand. The orthospilites with $Na_2O < 4$ are situated in the second group, at the beginning of imiscibility (Fig. 3).

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Pétrologie-Minéralogie

CHLORITIZATION OF BIOTITES AND ITS BEARING ON K-Ar AGES OF SOME ALPINE MAGMATITES FROM POIANA RUSCĂ MASSIF

BY

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Introduction

The present paper deals with 15 K-Ar ages on biotites and their interpretation in view of the fact that in most cases biotites were more or less affected by chloritization. The biotites under consideration were extracted mainly from granodioritic rocks located in the south-western part of the Poiana Ruscă Massif (South Carpathians), where they form several subvolcanic bodies (Fig. 1), which represent the northernmost extremities of the Tincova-Krepoljin and Ruşchiţa-Bor banatitic alignments (Kräutner, Kräutner, 1972), belonging to the Tethyan-Eurasian Metallogenetic Belt (Janković, 1977). Details regarding preparation techniques of the concentrates and determination methods are to be found in Strutinski et al. (in press).

K-Ar Diagrams versus Harper Isochrons

The determined ages (Tab.) pointed out that, despite the petrographic likeness, the igneous rocks of the Tincova and Ruschiţa groups represent two distinct moments of magmatic activity (Strutinski et al., in press). This situation is obviously illustrated by the K-Ar diagrams (Fig. 2), where it can be seen that the samples are plotted on two straight lines, corresponding to the two groups of intrusions. On the upper line the granodiorites of the Tincova intrusion (samples 27PR, 29PR, and 31PR), as well as their porphyritic varieties (9PR, 25PR), occurring as dykes (apophyses ?), are figured. On this line a granodiorite porphyry dyke from Ruschiţa (sample 5PR) is also plotted, a possible evidence of the first phase of emplacement being active also

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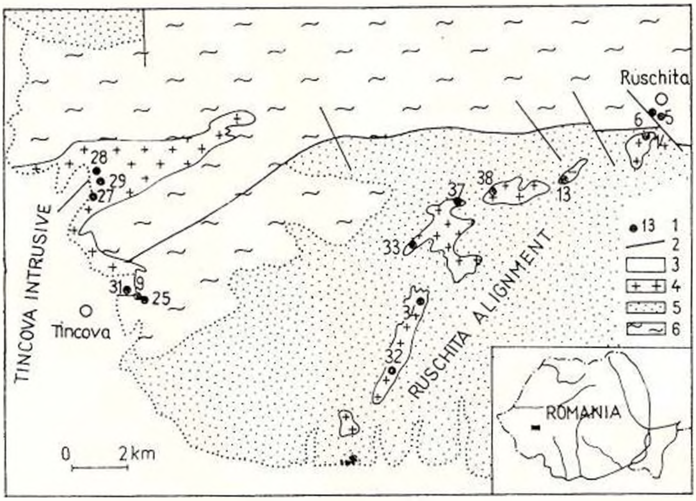

Fig. 1 — Geologic sketch of the south-western part of the Poiana Ruscă Massif showing location of samples radiochronologically dated.

1, metamorphic basement; 2, Mesozoic sedimentary rocks; 3, subvolcanic, mainly granodioritic, intrusive bodies; 4, Neozoic sediments and alluvial deposits; 5, faults; 6, samples.

TABLE

*Analytical data of K-Ar ages on biotites of the intrusive and dyke rocks from the south-western part of the Poiana Ruscă Massif (South Carpathians)**

Sample No.	Rock type	% K	%C**	^{40}Ar rad 10^{-9} g/g	Age (m.y.) ($\pm 1\sigma$)
5 PR	Granodiorite porphyry	6.61	22.3	37.61	80.3±2.4
6 PR	Quartz diorite porphyry	4.69	38.8	26.63	80.1±2.4
9 PR	Granodiorite porphyry	6.09	23.6	35.10	81.3±2.4
13 PR	Granodiorite	5.76	30.4	30.77	75.5±2.3
14 PR	Granodiorite	6.97	18.6	34.88	70.8±2.1
25 PR	Granodiorite porphyry	5.68	29.6	32.92	81.7±2.5
27 PR	Granodiorite	7.035	17.3	39.83	80.0±2.4
28 PR	Granodiorite	5.40	20.0	28.26	74.0±2.2
29 PR	Granodiorite	6.54	14.6	36.59	79.0±2.4
31 PR	Granodiorite	6.46	15.2	35.94	78.5±2.4
32 PR	Granodiorite	5.88	13.3	31.29	75.2±2.3
33 PR	Granodiorite	2.54	27.5	13.42	74.6±2.2
34 PR	Granite porphyry	7.19	20.4	37.45	73.6±2.2
37 PR	Dacite	4.48	34.3	22.05	69.7±2.1
38 PR	Granodiorite porphyry	5.33	21.3	27.76	73.6±2.2

* Analyst M. Soroiu; ** Contamination.

in the Ruschiţa region, yet on a much smaller scale. On the lower line the samples of several bodies of the Ruschiţa alignment are plotted (13PR, 14PR. 32PR, 33PR, 34PR. 38PR). Sample 6PR from Ruschiţa is situated right between the two lines, so that it could eventually represent an independent phase of less importance. Sample 28PR should actually lie on the upper line. The reason why it does not fit in rests on the fact that the biotite concentrate contains a greater percentage of impurities, which influenced the K and Ar determinations negatively. The same explanation is valid for samples 37PR and 38PR, which belong to the Ruschiţa Group.

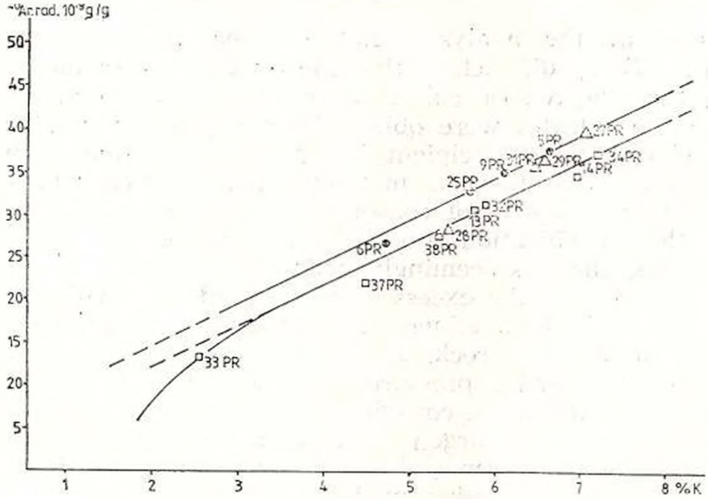

Fig. 2 — K-Ar diagrams for biotite concentrates of the Alpine igneous rocks (banatites) from the south-western part of the Poiana Rusca Massif.

□, intrusive bodies of the Ruschiţa alignment ; △, intrusive body of Tincova ; ●, dykes.

A careful examination of the correlation between the K-contents of biotites and the calculated ages emphasizes the general tendency of increasing ages with decreasing K-contents. Such inverse relations are mentioned from several other places (e.g. Hopkins, Silberman, 1978), where they were attributed to a gain of extraneous argon by the rocks or minerals subject to age determinations. In this view, the K-Ar diagrams of Figure 2 would be nothing but Harper isochrons, both showing small amounts of excess argon in the biotites under consideration. However, this can hardly be accepted. According to Harper (1970), excess argon in coeval rocks is constant, irrespective of their K-contents. In our case that would mean either that biotites had from the very beginning nearly the same K-contents as those actually determined, or that by potassium loss an equivalent loss of radiogenic argon occurred. The first alternative can be ruled out, because the biotites in question are almost all partially chloritized. Their K-contents, generally less than 7% (Tab.), are good evidence in this respect (Wilson, 1972). The second alternative may be unacceptable, as well, for the reasons displayed further on.

Discussion and Conclusions

In case of the analysed biotites it has been proved, optically, as well as by X-ray diffraction, that the potassium decrease is generally reflecting their degree of chloritization. It should be mentioned that the biotite concentrates were obtained from generally unaltered rocks, and that chloritization, incipient but frequent, is due to weathering rather than to hydrothermal metasomatism, so that thermal events younger than the moment of consolidation may be excluded. Admitting now that the chloritization process is responsible for the increase of the K-Ar ages, there is seemingly no need any longer to resort to an extraneous source for the excess argon. According to Giletti (1971), the distribution of ^{40}Ar in a mineral is a function of the partial pressure of this gas in the host rock. If, for any reason, during the history of the rock, the partial pressure increases up to a certain degree, incorporation of ^{40}Ar in the considered mineral would occur, and in this case "extraneous" excess argon would accumulate. Otherwise, different diffusion rates of potassium and argon during the chloritization process should be regarded as another mode of argon accumulation. Supposing the diffusion rate of potassium to be higher, there will be a relative enrichment of argon in the chloritized biotite, leading to the accumulaion of what should be called "native" excess argon. A reason for the relative inertia of argon in the crystal lattice of biotite subject to chloritization would be expanding tendency of the lattice, as a consequence of potassium substitution by brucite layers (Tröger, 1967). As a response, an increase of the partial pressure of argon in the rock must be inferred, not as much as to permit incorporation of argon by the biotite lattice, but sufficient to lower the diffusion rate from within it. In practice this process seems to operate down to contents of about 4% potassium. Beyond this threshold a rapid expulsion of argon is noticed, indicated by the downward bending of the straight lines on the K-Ar diagrams (Figs. 2, 3, 4). This process is triggered most probably by the generation of fissures and cracks in the rock, at a moment when the resistance of mineral aggregates fails against the increasing pressures developed by the chloritization process. Such fissures, transecting the whole rock in from of a tiny veined network, have been observed microscopically in sample 33PR. The fissures are often filled with chlorite migrated from pseudomorphs after hornblende and biotite.

From our hypothesis it clearly evolves that the credibility of a K-Ar biotite age rises with the increase of the K-contents, that is as chloritization approaches zero. Within a series of determinations on the same geological body, whose age, obtained on the purest biotite, should be the oldest, represent only an acceptable upper limit of the real age. Nevertheless, it should be stressed that such age interpretations apply only to rocks sufficiently young, or, at least, which prove that they did not undergo any other thermal event after the time of their formation.

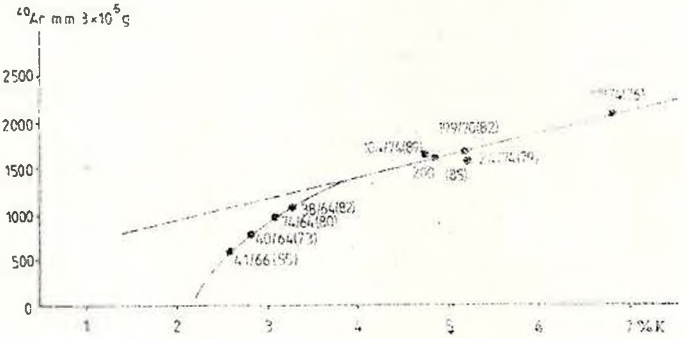

Fig. 3. — K-Ar diagram for biotite concentrates of the Central Srednogorié, Bulgaria (acc. to Boyadjiev and Lilov, 1981). Calculated ages (m.y.) in brackets.

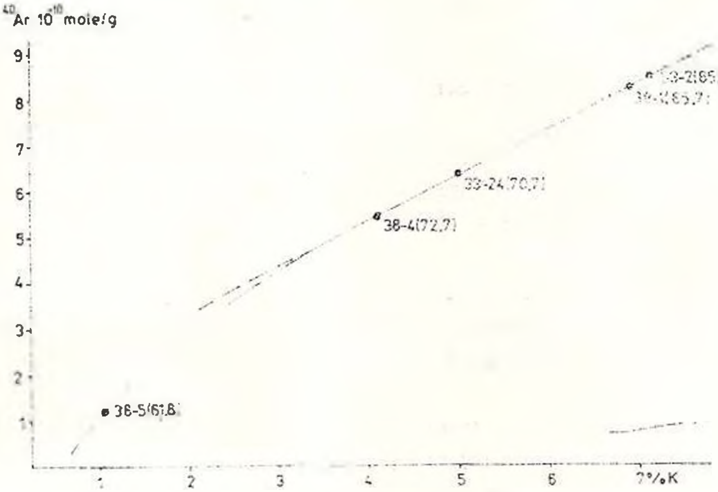

Fig. 4 — K-Ar diagram for biotite concentrates of the Utuado and San Lorenzo batholiths, Puerto Rico (acc. to Cox et al., 1977). Calculated ages (m.y.) in brackets.

Similar interpretations as for the Poiana Ruscă Massif can be performed, for instance, for the Upper Cretaceous intrusions from Central Srednogorie-Bulgaria (Boyadjiev, Lilov, 1981) or Puerto Rico (Cox et al., 1977), or for the Tertiary ones from South Park-Breckenridge (USA) (Bryant et al., 1981). The K-Ar diagram for the Alpine intrusions from Central Srednogorie (Fig. 3) points out that they are coeval and do not exceed 76 m.y. Similarly, as concerns the Utuado and San Lorenzo batholiths from Puerto Rico, the K-Ar diagram (Fig. 4) shows their cogenetic character and sets the upper limit of their age at 67,5 m.y. There is no evidence of two magmatic phases in case of the San Lorenzo Batholith, in spite of K-Ar ages on hornblende concentrates. It has to be mentioned that hornblende ages are subject to greater errors than biotite ages, because of the very small K and Ar contents implicated, in connection with chloritization, biotitization or other alteration processes, which may produce serious and uncontrollable perturbations in the activity of hornblendes as "geological clocks". Some other cases in which the credibility of hornblende ages is doubted are cited by Rice et al. (1982). As for the South Park-Breckenridge region (Bryant et al., 1981), the discrepancy between biotite ages of 44 to 50 m.y. and zircon, sphene and apatite ages of 35-42 m.y. may of course be due to excess argon, but not derived from Precambrian basement, as the authors presume, but by internal accumulation of "native" argon, as emphasized in this paper.

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TYPOMORPHISM OF SOME ORE MINERALS AND A P_vT CLASSIFICATION OF CERTAIN ORE DEPOSITS

BY

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The following considerations include data on some Alpine ore deposits genetically related to Laramian and Neogene magmatites in Romania. The two calc-alkaline magmatic provinces show different distribution areas (Fig. 1) and the main magmatic events are productive. Mineralizations are known both in volcanic and subvolcanic or hypabissal environments. In some parts of the distribution areas of the Neogene magmatites there occur together typical lava flows and subvolcanic bodies. Shallow depth hypabyssal or subvolcanic andesites may hardly be distinguished from the volcanic rocks of the same petrographic type. The subvolcanic mineralizations are characterized by a greater vertical extent as compared to those genetically related to the volcanic structures. This is why an attempt was made to find out some typomorphic features of the mineralizations formed in two major settings : volcanic and subvolcanic or hypabyssal (Udubaşa et al., in press). This paper represents a further research on this topics.

Typomorphic Assemblages of Oxide Minerals in Igneous Rocks

This discussion includes mainly igneous rocks of intermediate composition. In the study areas the rocks exhibit a nearly continuous variation of texture, from holocrystalline to porphyritic (with glassy matrix) rocks. The accessory opaque minerals are magnetite, ilmenite, maghemite, sphene and hematite, occurring in nearly fresh or slightly transformed rocks. Rutile becomes important as the alteration proceeds (Udubaşa, 1982). Pseudobrookite appears only in highly oxidized volcanics. The gradual oxidation of the primary oxide minerals in igneous rocks is largely described by Haggerty (1976).

Magnetite *(mgt)* is more homogeneous in volcanic rocks (especially in lava flows) but sandwich and trellis type ilmenite *(ilm)* lamellae

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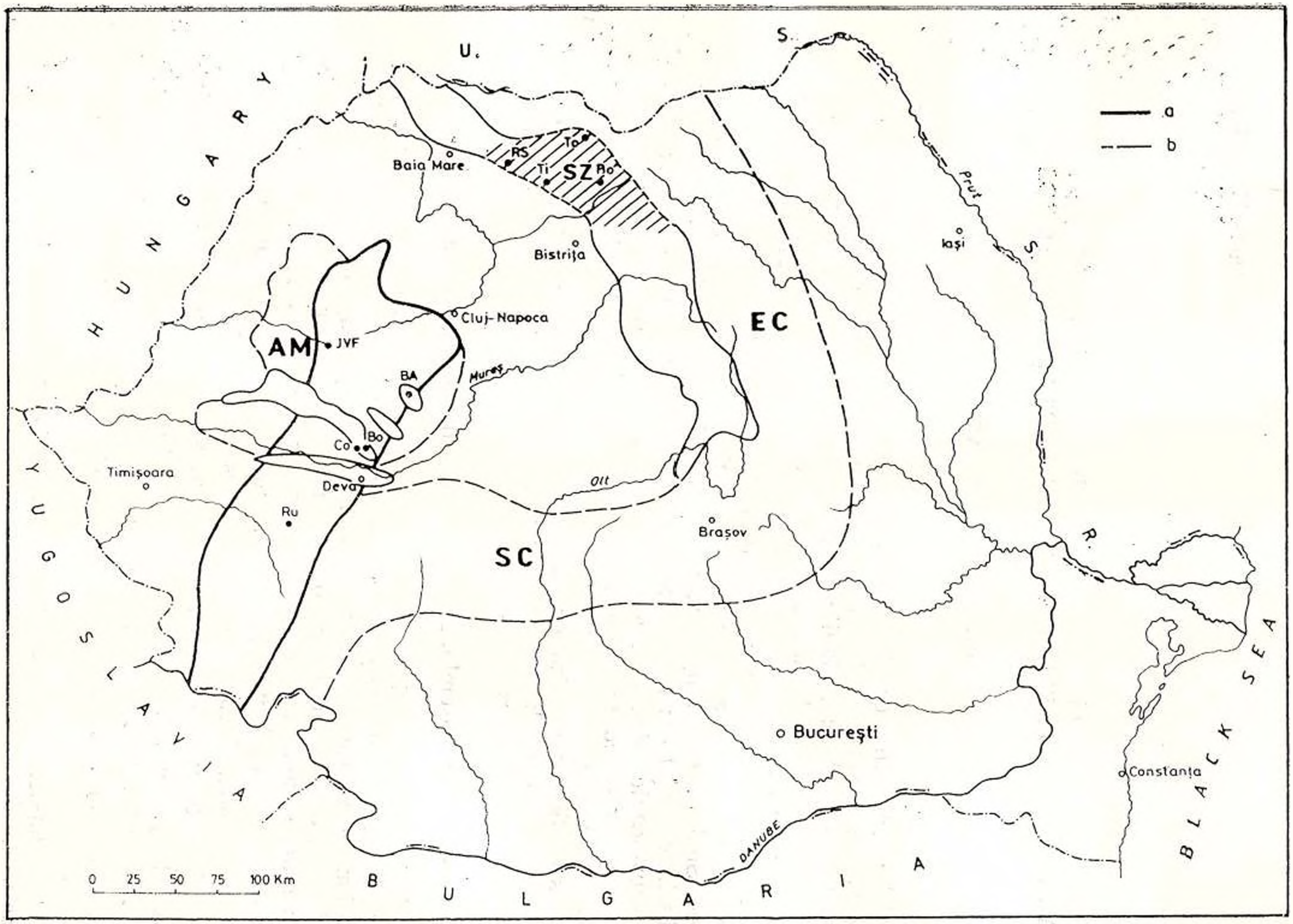

Fig. 1 — Distribution areas of Laramian (a) and Neogene (b) magmatites in Romania.

AM, Apuseni Mountains, EC, East Carpathians, SC, South Carpathians, SZ, subvolcanic zone of the EC Neogene volcanic chain. Subvolcanic and hypabyssal rock areas and the discussed ore deposits : RS, Rotunda-Strîmbu, Ţi, Ţibleş, To, Toroiaga, or separate grains are not rare. In such rocks it is the maghemite *(mgh)* which seems to be a very frequent and typical alteration product of the *mgt* in nearly fresh or slightly altered rocks. In most cases *mgh* forms coronas around *mgt* grains or irregular veinlets along the joints ; it may be sometimes overlooked if no oil immersion is used. More complex features shows the *mgt* richer in Ti : nearly concentric hands of varying colour shades (from blue to grey) completely replace the *mgt* grains (Fig. 2, II b and c). Maghemite has been often described in volcanic rocks (Katsura, Kushiro, 1961 ; Buddington, Lindsley, 1964 ; Ramdohr, 1975 ; Dankers, 1978 ; Freer, O'Reilly, 1980 etc.). However, its apperance in some intrusive rocks is also mentioned (Davidson, Wyllie, 1968 ; Czamanske, Mihalik, 1972 ; Gasparini, Naldrett, 1972 etc.).

Magnetite grains in subvolcanic or hypabyssal rocks are more complicatedly featured. The sandwich and trellis type *ilm* lamellae are abundantly developed. In addition, very fine, ulvospinel *(usp)*-derived *ilm* lamellae may be observed in places. The most typical feature of such (nearly fresh or slightly autometamorphicaly altered) rocks is the development of the secondary sphene. The sphenitization is also occurring on the fine, *usp*-derived *ilm* lamellae and the replacement is sometimes zonally evoluted (Fig. 2, I a). The secondary sphene has not been observed in samples of lava flows. In turn, *mgt* in some hypabyssal rocks may contain *mgh* as regular spots obeying the crystallographic planes in *mgt*. Elsewhere sphene was often observed in intrusive rocks (Davidson, Wyllie, 1968 ; Czamanske, Mihalik, 1972 ; Gasparini, Naldrett, 1972 ; Tsusue, Ishihara, 1975, etc.) ; it has been very rarely mentioned to occur in some basalts (Basta, Shaalan, 1974).

Breccia Formation

The best developed mineralized breccias are known in some porphyry copper (especially of Neogene age) and in some base metal ore deposits (Laramian and Neogene). At Rodna the breccia formation represents an intramineralizaion process and it seems to have appeared after the main ore phase was formed. The breccia is andesitic in character and bears both fragments of metamorphic limestones (partly replaced by ore) as well as fragments of ores belonging to the first phase and metamorphosed ores of Blazna-Gușet type. The Baia de Arieș ore deposit also exhibits andesitic breccia with metamorphic limestones, whereas the breccia occurring at Bocșa-Săcărîmb is nearly purely andesitic in lithology ; it developed in an older lava flow situated some tens of metres over the hypabyssal body of quartz microdiorites. The subvolcanic body at Coranda-Hondol is enveloped at the top by brecciated Cretaceous black shales and sandstones. A similar constitution has the mineralized breccia at Rotunda-Strîmbu. At Rușchița the breccia seems to represent a mixed formation of both eruptive and tectonic origin. Less developed breccias are present at Julești-Valea Fagului. No breccia at all is known at Toroiaga (Borcoș et al., 1982). For more details see Udubașa et al. (in press).

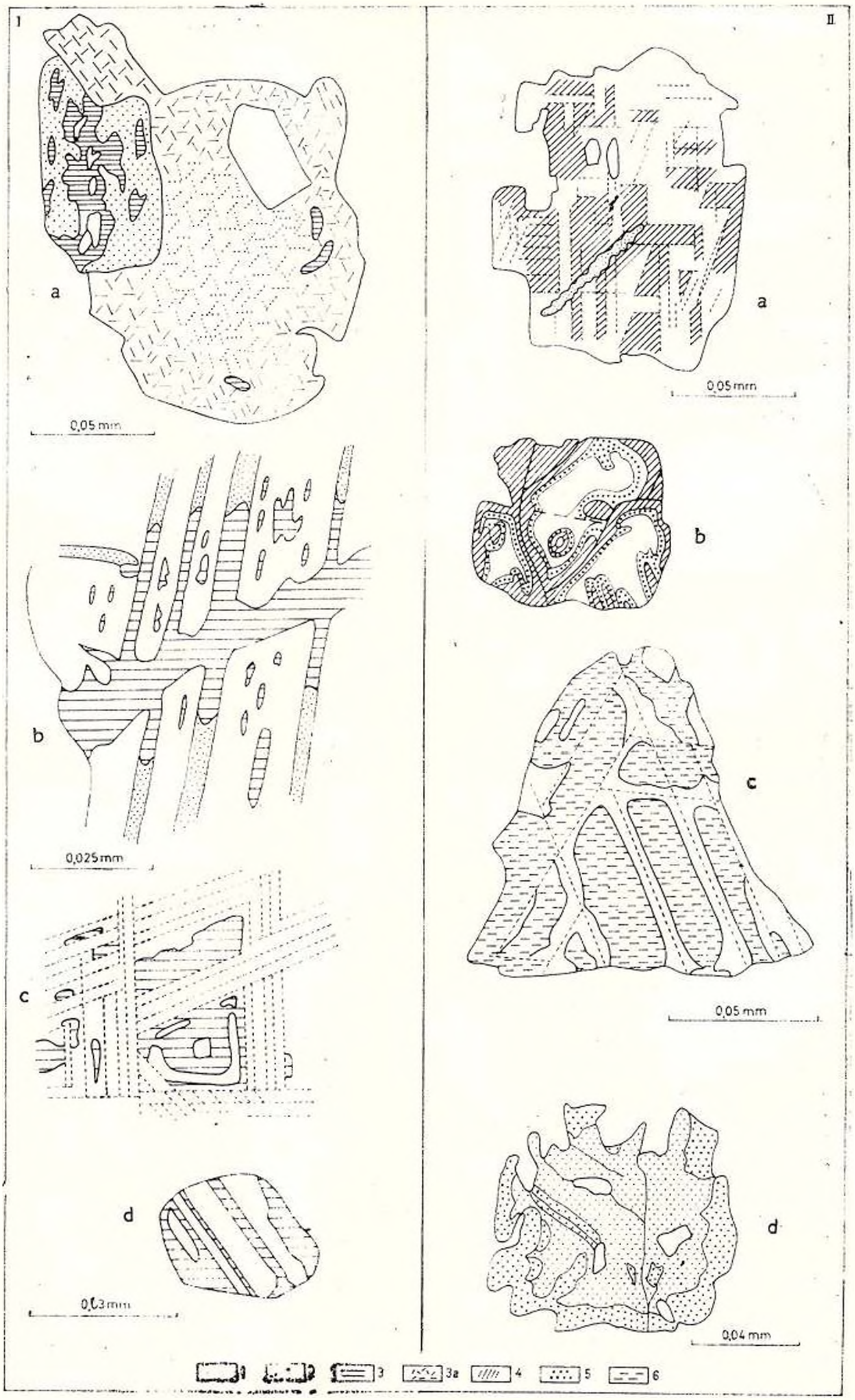

Fig. 2 — Types of oxide mineral intergrowths in subvolcanic (I) and volcanic (II) rocks.

Ore Mineral Assemblages in Ores

The ore deposits under discussion have a lead-zinc character ; copper-rich ores or copper enrichment show those at Toroiaga, Juleşti-Valea Fagului and Ţibleş (southern part). Sphalerite and galena together with pyrite (and chalcopyrite) are the dominant ore minerals. There occur also some sulphosalts (among them bournonite predominates) as late differentiates or late phases. Some porphyry copper systems are accompanied by base metal ore veins as late formations ; the ores in such veins always contain iron poor sphalerite, e.g. Bolcana-Troiţa, Metaliferi Mts ; South Ţibleş etc.

Pyrrhotite *(po)* as separate aggregates is present only at Ţibleş, Rodna and Toroiaga ; commonly it occurs within the multiphase inclusions in sphalerite *(sph)*, too. Pyrite *(py)* is always present but it is less frequently occurring in the copper-rich, bornite bearing ores at Juleşti-Valea Fagului as well as in the nearly pure lead-zinc ores from Ruşchiţa. The same is true for arsenopyrite. Galena is an ubiquitous mineral and bears some sulphosalts at Rodna, Toroiaga, Coranda-Hondol and Bocşa-Săcărîmb. Chalcopyrite is abundant at Toroiaga (as an early mineral associated with *po* and *py)* and at Juleşti-Valea Fagului (associated with bornite). At Bocşa-Săcărîmb there were observed some chalcopyrite crystals in vugs.

Sphalerite exhibits the most complicated features. It contains varying iron contents (from 1 up to 15 wt.-% Fe), and, accordingly, inclusions of different compositions. The iron rich (10-15wt.-% Fe) *sph* bears multiphase inclusions containing chalcopyrite *(cp)*, pyrrhotite *(po)*, chalcopyrrhotite or the intermediate solid solution *(iss)*[2] of the Cu-Fe-S system, cubanite *(cb)* and mackinawite *(mck)*. These minor phases are differently textured and are of varying dimensions (Fig. 3). The complexity of inclusions is decreasing as the iron content in *sph* decreases : the *sph* with 3-6 wt.-% Fe contains only *cp* (Baia de Arieş ; the ores contain also alabandite) or *cp*+*iss* (Bocşa-Săcărîmb) and the iron poor

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I a, a grain of *mgt* with *usp*-derived fine *ilm* lamellae, zonally transformed into sphene ; granodiorite, Ţibleş Neogene igneous complex ; I b, sandwich type *ilm* lamellae in *mgt* partly altered to sphene ; lamprophyre, dyke swarm of Liassic age, East Făgăraş Mts ; I c, trellis type *ilm* lamellae ; the *mgt* "islands" are sphenitized ; microdiorite, Rotunda-Strîmbu Baia Mare area ; I d, sandwich type *ilm* lamellae completely transformed into sphene ; monzodiorite, Ţibleş igneous complex ; II a, *mgh* developed in a *mgt* grain with fine *ilm* lamellae ; Neogene pyroxene andesite, Baia Mare Area, N Romania ; II b, banded *mgh-timgh* developed on a homogeneous *mgt* grain ; Neogene pyroxene andesite, Baia Mare area ; II c, "scaly" hematite formed on *mgt* via *mgh* with preservation of *ilm* lamellae ; lava flow of Neogene quartz andesite, Bocşa-Săcărîmb ; II d, *timgh* formed at the borders of a *ilm* grain ; Neogene pyroxene andesite, Baia Mare area. 1, magnetite ; 2, ilmenite ; 3, sphene ; 3a, sphene (in I a only) ; 4, maghemite ; 5, titanomaghemite ; 6, hematite.

sph at Coranda-Hondol, Ruşchiţa (with 1-2.5 wt.-$^0/_0$ Fe) bears only *cp* and occasionally *mck*. Two-phase inclusions consisting of *cp*+*bn* were observed in the *sph* from the Juleşti-Valea Fagului ores.

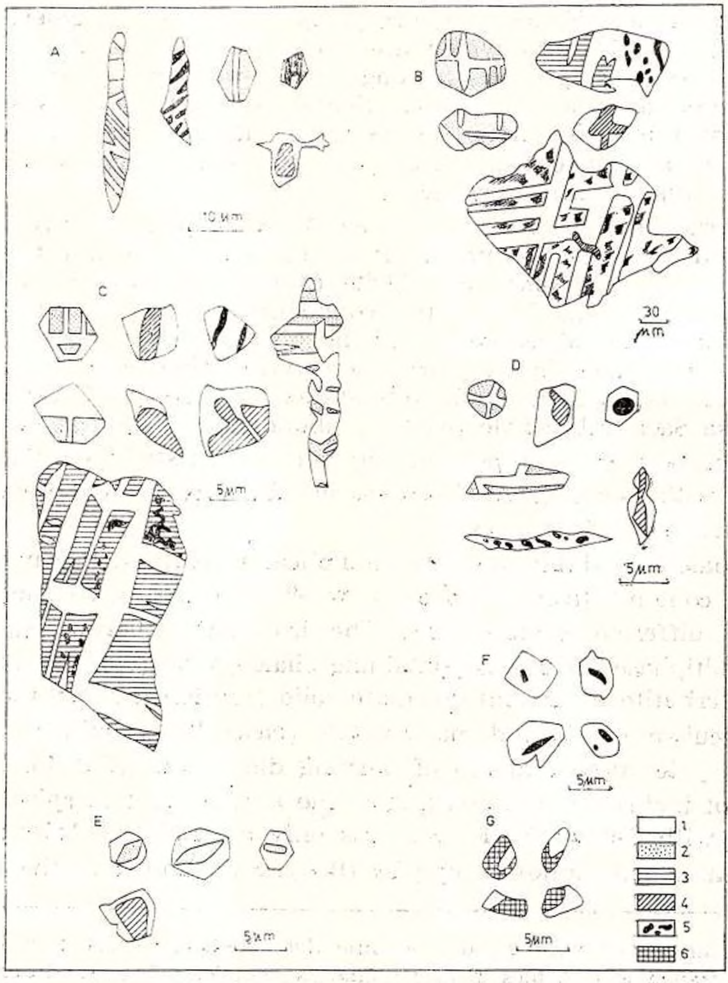

Fig. 3. — Types of inclusions in sphalerites : A-D, iron rich *sph* (10-15 wt.-$^0/_0$ Fe) ; E, *sph* with 6 wt.-$^0/_0$ Fe ; F-G, iron poor *sph* (1.0-2.7 wt.-$^0/_0$ Fe). The greater inclusions in B and C reflect the derivation of the coexisting minerals from the high T *iss* ; the *po* is grown in a phase resembling *cp*, not distinguishable from the *cp* of the laths (B) ; coexisting *cb* and *mck* developed on a sol-sol matrix with lanceolate *cp* laths suggesting a paramorphic transformation of a high *T* phase. The smaller inclusions contain only frozen assemblages, in which *mck* does not occur together with *iss (cpo)* as *cb* sometimes does (in B).

A, Rodna ; B, Toroiaga ; C, Ţibleş ; D, Rotunda-Strîmbu ; E, Bocşa-Săcărîmb ; F, Coranda-Hondol ; G, Juleşti-Valea Fagului. 1, chalcopyrite *(cp)* ; 2, "chalcopyrrhotite" or *iss* ; 3, cubanite *(cb)* ; 4, pyrrhotite *(po)* ; 5, mackinawite *(mck)* ; 6, bornite *(bn)*.

Some Typochemical Features

In order to distinguish the volcanic from the subvolcanic mineral assemblages, the diagrammatic representation of some minor elements contained by the main sulphides was used (Udubaşa et al., in press). The most interesting diagram is given in Figure 4. The subvolcanic *sph* concentrates into a relatively well delimited field, no matter that the sphalerite is iron rich or iron poor. However, the iron rich *sph* plots mostly towards the Mn corner whereas the iron poor *sph* plots towards the Cd corner. The Mn : Cd ratio, is thus fairly constant, reflecting a typical feature of the subvolcanic *sph,* i.e. an "equilibrium concentration" of these elements.

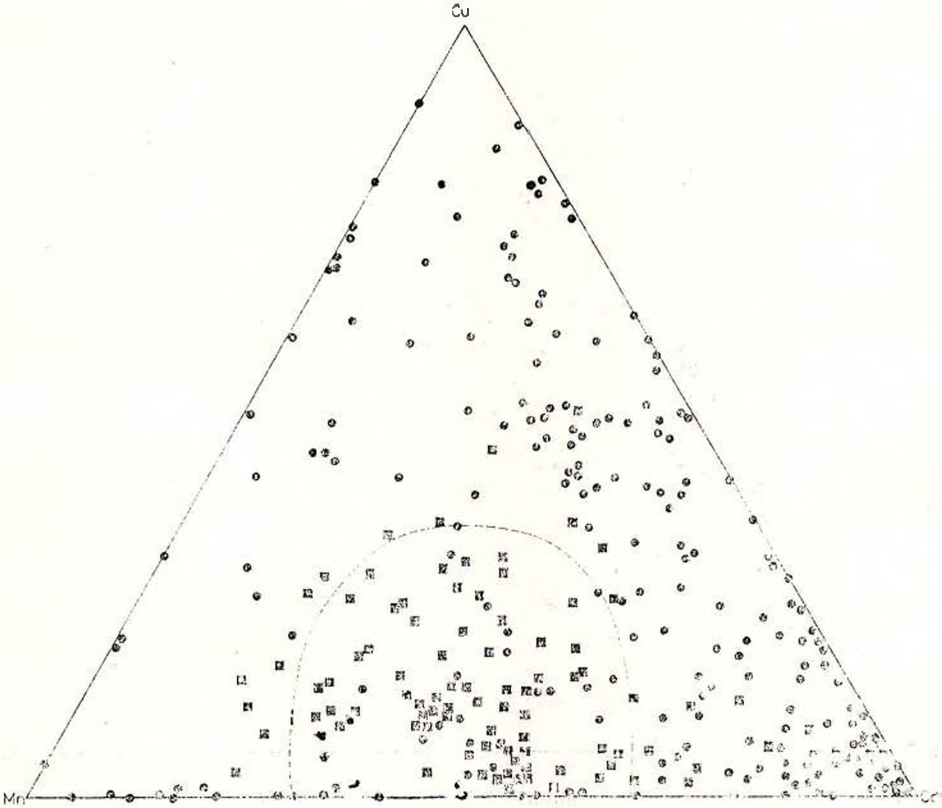

Fig. 4. — Cu-Mn-Cd diagram for sphalerite.
1, subvolcanic sphalerites ; 2, sphalerite from other ore deposits (mainly of volcanic setting) ; 3, field of subvolcanic sphalerites.

The Ag-Bi-Sb diagram does not discriminate the volcanic and subvolcanic galenas. Most of the plottings concentrate at corresponding values of the compounds $AgBiS_2$ and $AgSbS_2$ (Fig. 5). However, the field of plottings is much larger, showing the extent of solubility of Ag, Bi and Sb in natural galenas. The space towards the Bi-Sb edge is practically free (less than 4% of total plottings) suggesting a very limited solubility of the BiSb pair, as inferred from the experimental data of Amcoff (1976) too. The Laramian galenas concentrate towards the AgBi edge whereas the Neogene galenas towards the AgSb edge; however, galena from high-T assemblages of Neogene age tends to migrate to higher Bi contents too. The trend is not only temperature dependent but indicates also that the galena seems to be geochemically much more linked to the age and regional characteristics of the ore deposits than the *sph* is.

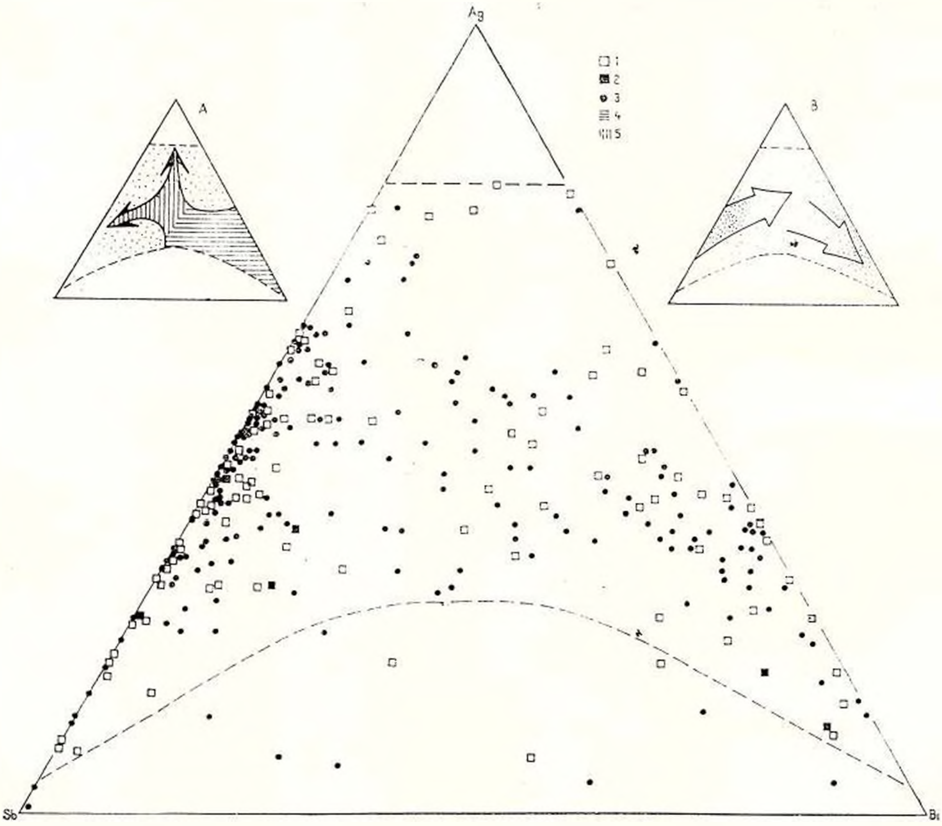

Fig. 5. — Ag-Sb-Bi diagram for galena: A, the evolution trend from Laramian (4) to Neogene (5); B, the behaviour of galenas from low T to high T assemblages (arrows) in Laramian and Neogene ore deposits.
1, subvolcanic galenas; 2, average values of subvolcanic galenas; 3, other galenas (mainly of volcanic setting); 4, (in A) field of Laramian galenas; 5, field of Neogene galenas.

Discussion

The ore deposits under discussion are genetically related to Laramian and Neogene subvolcanic and hypabyssal rocks of andesitic composition consolidated at various depths. In such cases one can admit that the total volatile pressure (P_v) of the postmagmatic systems was roughly equilibrated by the lithostatic pressure (P_l).

The primary oxide minerals in the related igneous rocks indicate for all cases a relatively low f_{O_2}. The assemblage *mgt*+sphene seems to be a typomorphic feature of the non-erupted rocks (generally slightly altered) as against the typical lava flows constantly exhibiting the assemblage *mgt*+*(ti)mgh*. This fact points to a low P_{CO_2} as shown by the sphene stability diagram of Schuiling and Vink (1967), within the consolidating and decompressing magmatic systems. In some cases, e.g. the Toroiaga igneous complex, *mgh* has been observed forming regular spots in the inner parts of the *mgt* grains (high T *mgh* ?) ; sphene is, howver, rather frequently occurring too. The greater f_{O_2} in this case correlates with the high T massive deposition of *cp* (bearing *sph* stars) prior to the *sph* and *gn* deposition. The “excess” copper in these ores might have originated also in the remobilisation (at least partly) of the pre-existing metamorphosed ores of Lower Paleozoic age intruded by the Neogene andesitic rocks and cut by the hydrothermal vein ores.

If the *sph* geobarometer (Scott, 1973 ; Scott, Barnes, 1971 ; Lusk, Ford, 1978 ; Hutcheon, 1980 ; Shimizu, Shimazaki, 1981 etc.) is applied to such ore deposits, it is interesting to note an inverse relationship between T and P_v. High T assemblages containing *po, apy*, iron-rich *sph* with multiple inclusions of *cp, iss, po, cb* and *mck* seem to associate with low P_v postmagmatic environment, i.e. ores at Toroiaga, Ţibleş and Rodna (if skarn minerals are present then the ores are subsequently formed). Low T assemblages with iron poor *sph* bearing inclusions of *cp*±*mck* may further be ascribed to high P_v conditions, i.e. ores at Ruşchiţa and Coranda-Hondol ; such ores associate with large scale breccia formation suggesting high volatile pressure in the postmagmatic fluids. The other deposits containing *sph* with moderate iron content, i.e. Baia de Arieş and Bocşa-Săcărîmb, may be regarded as formed at intermediate T and P_v; they show also mineralized breccia.

TABLE 1

Iron content in sphalerite and related mineral assemblages as function of T and f_{S_2}

Mukaiyama, Izawa 1971)			This study	
Mole %FeS	Assemblages		Mole % FeS	Assemblages*
>23	*cp, cb, hpo*	high T low f_{S_2}		
14–23	*cp, hpo*	↑ ↓	14–25	*cp, iss, cb, po, mck*
9–14	*cp, py, mpo*		10–12	*cp, po, iss*
0–12	*cp, py*		1.5–2.7	*cp, mck*
< 1	*cp, bn, py*	low T high f_{S_2}	1–2	*cp, bn'*

* as inclusions in *sph*; all assemblages contain *py* too.

As Mukaiyama and Izawa (1971) showed for some Japanese ore deposits the *sph*-bearing assemblages are temperature dependant (Tab. 1). If such a T arrangement is correlated with the P_v regime of mineral deposition, as shown by the iron content in *sph* and by the presence of breccia in certain ore deposits, one can obtain a classification of mineral assemblages in terms of T and P_v (Tab. 2). Such a treatment of facts allows to attenuate the apparent discrepancy between the experimental data regarding the relation of the iron content in *sph* and its T of formation. This is not an attempt to revive the *sph* geothermometer, but to find out an explanation for the iron rich *sph* occurring in high T and low P assemblages.

TABLE 2

The P_vT grouping of the ore deposits presented in this study

P_v, T estimated	Ore deposits	Main characteristics
High T, low P_v	Toroiaga Rodna Tibleş Rotunda-Strîmbu	No breccia formation 17–25 mole % FeS in *sph* Multiphase inclusions in *sph* (see Table 1) Sphalerite has a nearly constant Mn : Cd ratio, but Mn is statistically slightly enriched Galena is slightly enriched in BiAg
Intermediate T and P_v	Bocşa-Săcărîmb Baia de Arieş	Abundant breccia formation 5–10 mole % FeS in *sph* Two-phase inclusions in *sph* (*cp, iss*)
Low T, high P_v	Coranda-Hondol Ruşchiţa Tibleş (south) Juleşti-Valea Fagului Bolcana-Troiţa*	Abundant breccia formation 1–4.5 mole % FeS in *sph* Inclusions in *sph* consist of *cp, mck* Sphalerite has a nearly constant Mn : Cd ratio, but Cd is statistically slightly enriched Galena is slightly enriched in SbAg

* Pb-Zn ore veins cutting the porphyry copper system.

The model presented above can be applied only to subvolcanic and hypabyssal environments, which are complementary assured by the typomorphic assemblage *mgt*+sphene in the related igneous rocks (as against the stabilized assemblage *mgt+(ti)mgh*, currently occurring in lava flows).

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[2] *cpo* (chalcopyrrhotite) or *iss* as used in this paper is a phase having an intermediate colour between *po* and *cp* ; it is lighter than the *cb*, mostly isotropic, but sometimes becomes slightly anisotropic (gradual transformation into *cb* ?) especially when *mck* is appearing (see Fig. 3 B). The phase is somewhat similar to the talnakhite but it does not tarnish and the optical properties fit well with those of *cpo* described by Ramdohr (1975). However, it is clear that the iron rich *sph* contains inclusions of a high T solid solution which decomposes step by step by decreasing T, giving *cp+cb*, *cp+po* or *cp+cb+mck* (metastable assemblage), or the quenching products — the *iss* itself — occurring in relatively quickly cooled ores.

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ORE TEXTURE AND STRUCTURE OF SULPHIDE-IRON OXIDES OF THE PRECAMBRIAN ALTÎN TEPE DEPOSIT, DOBROGEA (ROMANIA)

BY

ION BERBELEAC[1], AVRAM ŞTEFAN[1]

Introduction

Among the great diversity of stratiform metamorphosed hydrothermal-sedimentary ores in metamorphic schists the sulphide-iron oxide Altîn Tepe deposit represents a particular type. The peculiarities of ore textures and structures observed in polished hand specimens and polished sections help us to solve some problems regarding especially the origin and metamorphic processes of ore.

Geological Setting

The geological characteristics of the ore deposit are given in a number of works (Codarcea, Petrulian, 1948 ; Gurău, 1970 ; Mureşan, 1969, 1972 ; Ianovici et al., 1971 ; Berbeleac et al., 1983).

The investigated deposit is situated in the Upper Precambrian metamorphic schists (1600-850 m.y.) which form a regionally extended belt trending north-westwards. Several geological units are parallel to the belt : northwards there are unmetamorphosed Paleozoic (Carapelit Formation — Permo-Carboniferous) and Mesozoic rocks (Triassic-Upper Cretaceous) ; southwards lies the greenschist series of Upper Precambrian age (800-575 m.y.) which underwent a low-grade metamorphism, while north-eastwards there occur Triassic rhyolitic rocks.

The rocks of the Altîn Tepe Series represent a normal monoclinal flank with north-western strike (N 35-65° W). The B axis dips by 30-45° towards the south-east. The greenschist series overthrusts the Altîn Tepe Series and a great number of faults are present.

According to the age the Altîn Tepe Series may be subdivided in two lithostratigraphic units : 1) the lower unit (Fig. 1 a), over 2 500 m thick, which includes high grade metamorphosed sediments

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(gneisses, amphibolite, quartz biotite, etc.), plutonic (gabbro, diorite) and volcanic (tuffs, basaltic lavas, etc.) rocks and 2) upper unit, probably 150-200 m thick, also high grade metamorphosed but strongly retromorphosed, consisting of quartz-feldspar-chlorite schists (80-100 m, Fig. 1, b, c) interlayered with quartz-muscovite-biotite±garnet schists, quartz-chlorite-sericite schists and amphibolite schists in the lower part

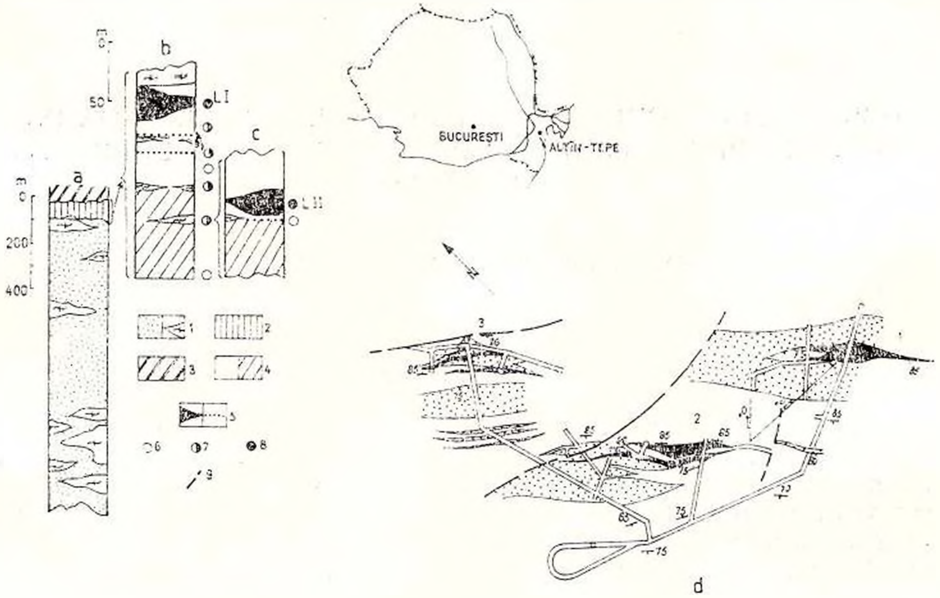


Fig. 1 — Stratigraphic column of the Altîn Tepe Series (a-c) and the geometry of ore lenses at level —750 m(d). a, Baspunar-Altîn Tepe zone ; b, upper unit at level —750 m in the area of lenses nos. 1 and 2 (c).
1, a, gneiss, micaschist, biotite quartzite, etc. ; b, amphibolite ; 2, upper unit ; 3, Greenschist Series ; 4. a, chlorite-quartz schist ; b, chlorite-feldspar-quartz schist ; 5, a, massive ore lenses ; b, disseminated ore ; 6, disseminated pyrite ore ; 7, disseminated sulphide ore : 8, sulphide-iron oxide ore ; 9, fault. L I, L II, lense number.

and sericite-chlorite quartz schists and chlorite-albite schists with minor intercalations of quartzite, quartz-feldspathic schists and amphibole schists in the upper part (70-100 m, Fig. 1, b, c). The ore deposit lies in the upper part of the upper unit.

The presence of staurolite and garnet in the rocks of the Altîn Tepe Series points to the metamorphic amphibolite facies.

Ore Deposit

At present the mining works reach the level of -750 m (580 m below sea level) and a recent borehole has reached the ore at the depth of 530 m below this level. It means that on the plane of plunge (30-45° SE) the ore deposit was controlled along more than 3 km.

Two types of mineralization occur in the Altîn Tepe deposit: 1) sulphide-iron oxides massive ore lenses, obviously related to the chlorite-sericite quartz schists, quartz-albite-epidote±barite schists, quartz-sericite schists and other types of rocks and 2) disseminated pyrite-chalcopyrite±sphalerite, galena ore, which lies both in quartz-sericite schists and chlorite-quartz±albite schists from inside massive ore lenses or from a halo in their hanging wall and footwall (Fig. 1d).

The massive ore forms four small lenses (nos. 1, 2, 3, 4) in the upper levels of the deposit and three small lenses (nos. 1, 2, 3) in the lower part of the deposit. The dimensions of the massive ore lenses are small and on the strike the length ranges between 30-80 m, while the thickness varies between 2-18 m.

Ore bodies form either alternations of banded highly impregnated to compact pyrite-chalcopyrite±sphalerite, galena with magnetite±sulphides, hematite or with quartzitic, sericitic or chloritic albite schists which are poorly mineralized.

The dissemination bodies (1A, 1B, 2A, 2B, 3A, 3B) are related especially to quartz-rich sericite schists and appear at the exterior of the massive ore or at different levels of the upper part associated with chlorite-quartz±albite schist and amphibolite schist. This type of ore is larger than the massive ore and consists mainly of pyrite, minor chalcopyrite, very small amounts of sphalerite and galena and accidentally magnetite.

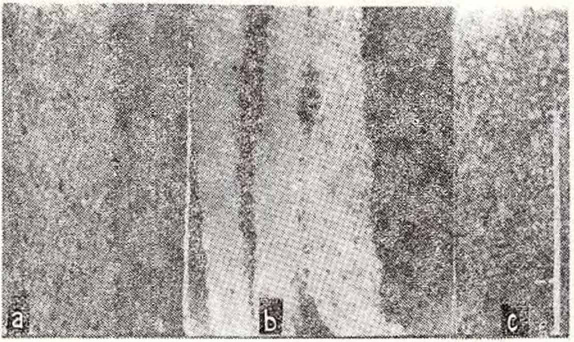
The authors think that the original ore deposit has a hydrothermal-sedimentary origin, being later modified by prograde and retrograde metamorphism.

Ore Texture and Structure

Many prominent features of the Altîn Tepe ore are, in the authors' opinion, compatible with the pro- and retrograde metamorphism. These features suggest many similarities to the ore bodies described by Vokes (1969), Suffel et al. (1971) and Lawrence (1973). Some of these characteristics of ore metamorphism are generally commented upon by other authors (Stanton, 1972; Shadlum, 1982). Before describing the ore polymetamorphism from Altîn Tepe, we remind and point out the fact that the ore bodies are situated within a retrogressive sequence of rocks.

Prograde Ore

Banded texture. The most striking primary textural feature of the ore is usually the fine compositional banding ranging from 0.1 to 10-15 cm in thickness (Fig. 2 b). However, many compositional bandings between 15-150 cm in thickness are known in massive ore lenses, while others, ranging between 3-20 cm in thickness are found in disseminated ore (Fig. 2 a). The layering is structurally concordant with the ore host-rock contacts and with the bedding primary origin. The most typical peculiarity of textures is the presence of numerous alternating rhythms or microrhythms with the same or different structures



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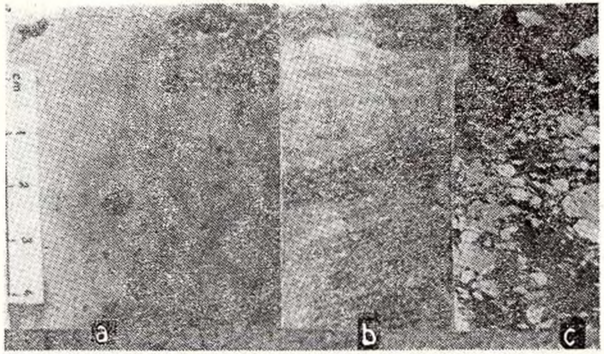

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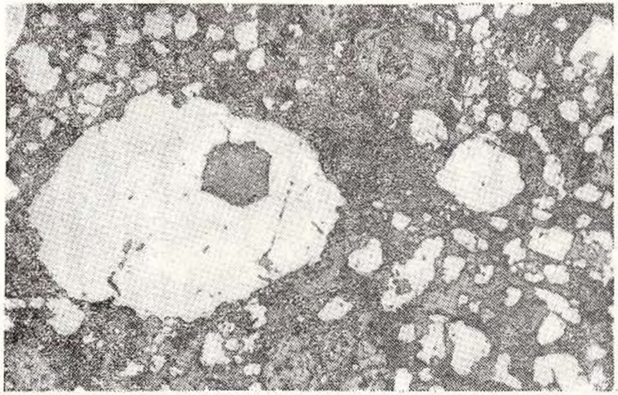

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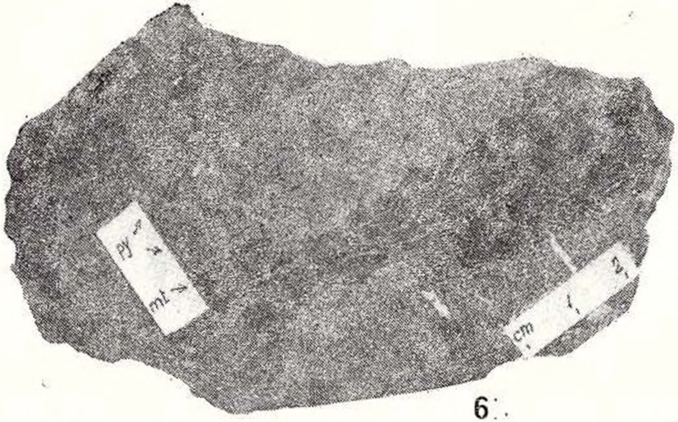
Fig. 2 — Banded texture of prograde ore : a, disseminated ore ; b, massive pyrite-iron oxide ore ; c, retrograde texture. Polished specimens : a, body 2B, level —450 ; b, lense no. 2, level —750 ; c, lense no. 1, level —750.

Fig. 3 — Retrograde textures : a, the breccia texture with fine-grained ore and large fragments of quartz ; b, the "bull structure" with slight banded texture ; c, the porphyroblastic structure of disseminated ore (prograde ore). Polished specimens : lense no. 1, level —750 m.

Fig. 4 — Poikiloblastic structure of pyrite and magnetite, // Nic 60 ×, lense no. 2, level —750.

Fig. 5 — Pyrite in chalcopyrite cement (prograde ore, // Nic, 60 ×, lense no. 2, level —750.



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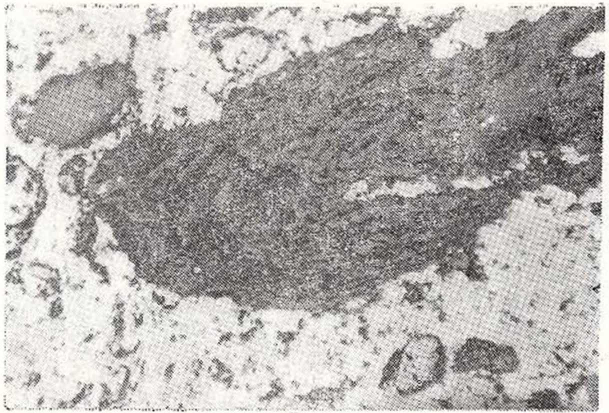

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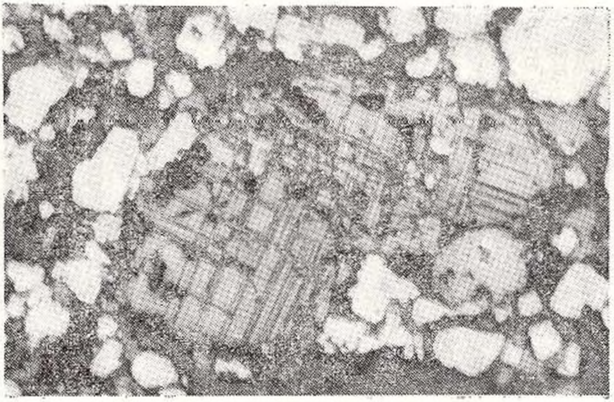

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Fig. 6 — Retrograde texture : "Mixed ore" with slipping folding. The fissures are filled by barite. Polished specimens : lense no. 1, level —750.

Fig. 7 — Retrograde ore. Breccia texture with fragments of quartz crenulated and folded schist and relics of magnetite (grey) and pyrite (white). The cement consists of chalcopyrite, sphalerite and galena. // Nic 120 ×.

Figs. 8, 9 — Superimposed twinning in magnetite (3) and chalcopyrite (4). Magnetite is grey ; pyrite is white ; sphalerite is grey-white ; chalcopyrite is white-grey. Etched. // Nic 120 ×, lense no. 1 level —750.

and mineralogical composition. The layered sulphide and iron oxide bands in massive ore lenses and sulphide bands in disseminated ore, are interlayered with schist bands (Fig. 2 c).

In the almandine-amphibolite facies the competent minerals, such as quartz, pyrite and magnetite from banded massive ore, are frequently stretched, become discontinuous and form boudins. Sometimes even schistose rock fragments or muscovite, sericite and chlorite are detached, contorted and rounded up by tectonic rolling (Fig. 3 a, b). Even thin bands of quartz-pyrite and quartz-magnetite$\pm$barite ore are laminated or completely detached. In connection with this, pyrite and magnetite “bulls”, which are rounded or nearly rounded, are present. In our opinion this type of texture may be assumed as a retrogressive texture. The insignificant presence of these textures, typically composed of “mixed” ore, in lense no. 2, may be due to the decrease of the retromorphism degree towards the lower parts of the upper unit — sequence of the Altîn Tepe Series.

In prograde ores the average structure ore is medium- and coarse-grained (Fig. 3 c). It is not uncommon to see pyrite and magnetite crystals as much as 2-3 mm in size. Chalcopyrite and sphalerite are minor constituents, being comparatively medium-grained. It is important to underline that in the above-mentioned structure persist all the main assemblages : pyrite-quartz ; magnetite (hematite)-quartz-barite and pyrite-chalcopyrite-sphalerite (galena)-quartz. Some heterogeneous structures which have been observed in the ore may be due to the variation in the initial mineralogical-chemical compositions and metamorphic processes. For instance the poikiloblastic structures for magnetite, sulphide, quartz or any silicate are present (Fig. 4). Resorption, recrystallization and the structures involving co-recrystallization of sulphides and gangue minerals have been noticed. The co-recrystallization of chalcopyrite with quartz, muscovite, marked by smoothly curved or curvilinear interphase boundaries are sometimes remarkable. Also, chalcopyrite shows broad twin lamellae (Fig. 9). This mineral as well as sphalerite serve frequently as cement for magnetite and pyrite aggregates (Fig. 5).

Retrograde Ore

Mesoscopic characteristics of the ore from the retrograde zones are best seen on polished slabs under a binocular microscope. The ore is practically brecciated and intersected by numerous stress zones ; the fissures are filled by plastic injections or “filter pressed” chalcopyrite. Like the retrograde ore from the Broken Hill sulphide ore bodies (Lawrence, 1973) the fragments of wall rocks and prograde ore are prominent in the retrograde ore of Altîn Tepe.

Breccia texture. A very common and characteristic type of ore from lenses 1 and 3 are breccia textures. These textures consist of two principal types : 1) “bull texture” formed by medium and coarse pyrite, magnetite, quartz and silicate fragments, rather closely packed in a matrix of chalcopyrite, sphalerite and galena (Fig. 3 b) and 2) breccia texture, which consists of very fine-grained minerals (Fig. 3 a, Fig. 6). The latter type appears locally, generally along the longitudinal faults or shearing zones (Fig. 10).

The breccia textures represent especially more than 80% from the total volume of the lenses 1 and 3. The main features of these textures are the following : 1) the banded aspect which consists of rhythms of mixed ore composed of pyrite and magnetite "bulls" and fragments of rocks or other minerals in a matrix of sulphides, quartz and barite ; 2) the disappearance in general of the bed primary origin of ore, initially marked by well-defined bands of uniform mineralogical-chemical compositions and the appearance of local enrichments ; 3) the presence of intrusive deformations which are responsible for cataclastic ore and slipping microfolds (Fig. 6) and 4) the existence of a great homogeneity of the ore structures.

The breccia texture, which consists of fine-grained ore, shows a slight orientation and bandings (Fig. 3 b). This ore has a remarkable number of inclusions, which differ in composition : quartz-magnetite ± ±barite ; quartz-sericite-chlorite schists, milky quartz, relict pyrite and magnetite porphyroblasts. The inclusions vary from large blocks, especially of milky quartz and quartz-sericite-chlorite schists ; some of these are severely folded and crenulated, others are rounded or angular in shape (quartz) (Fig. 7). Chalcopyrite±galena and rarely sphalerite show injections, especially into quartz-magnetite+barite ore and milky quartz. The sources of the ore fragments are the quartz veins from the country rocks or from massive ore and the barren rocks. The breccia cement of fine-grained ore is very rich in chalcopyrite and sphalerite. Although galena is a minor constituent, here it shows the highest content.

We have previously mentioned the relict prograde texture and structure of the ore within the lenses 1 and 3. (Fig. 10 a, c). Between the relics of prograde ore and retrograde ore there is a transition zone marked by homogeneous, pyritic or iron-oxide ore, but with breccia texture. A common and characteristic feature of this ore is the existence of the slight signs of flowage and particularly of shearing.

As a result of retrograde metamorphism the initial structure of the ore was partly subsequently modified.

The principal modifications of massive ore consist in : 1) the decrease of the grain-sizes (0.1-0.5 mm) ; 2) the substructures of chalcopyrite and galena ; 3) complex superimposed twinning of magnetite and chalcopyrite (Figs. 3, 9) ; 4) the presence of chalcopyrite inclusions (exsolution) within sphalerite ; 5) the fine dispersed sulphides in gangue minerals ; 6) the increase of the degree of martitization in magnetite and 7) the appearance of small euhedral pyrite crystals in massive breccia ore (0.1-0.5 mm).

As regards the disseminated ore we point out that the pyrite aggregates or isolated grains have been subject to fracturing and crushing. Texture and structure peculiarities of these aggregates show quite distinctly their formation in connection with mylonitization processes.

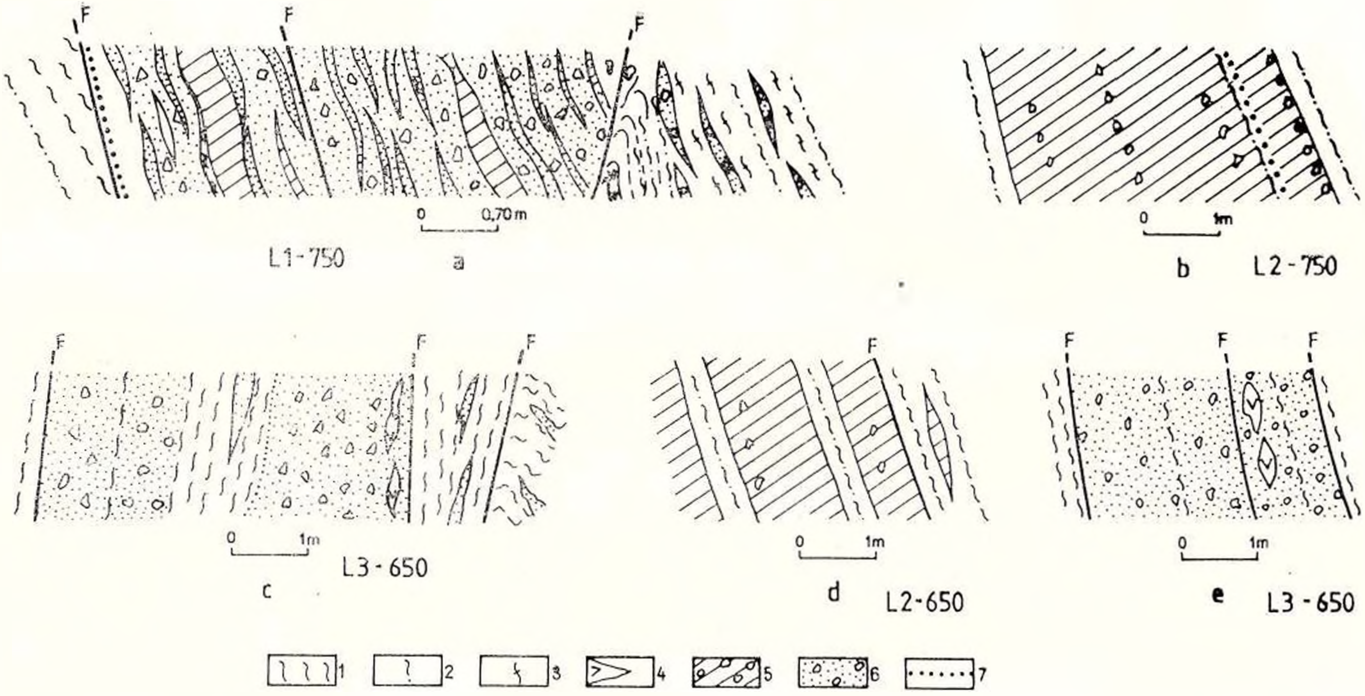

Fig. 10 — Details of the ore lenses.
1, chlorite-quartz ± feldspar schist ; 2, chlorite-quartz schist and sericite-quartz-schist ; 3, amphibolite and chlorite-quartz feldspar-epidote schist ; 4, quartz blocks ; 5, prograde ore with relics of retrograde ore ; 6, retrograde ore ; 7, magnetite ore : L 1, 2, 3 — lense number ; mining levels —650 and —750 m.

In some cases the growth of pyrite has been disturbed, showing elongated shapes or "en échelon" distribution.

The pyrite and sphalerite deposition, like in massive ore, was obviously accompanied by brecciation and partly corroded by abrasive solutions of later pyrite, quartz and barite. This pyrite was immediately redeposited as fine skeletal grains.

Conclusions

1. The Altîn Tepe ore deposit lies within a retromorphic sequence in the neighbourhood of an important overthrust and is composed of predominantly quartz-sericite-chlorite mylonitized schists. Initially the rocks of this sequence of the upper unit of the Altîn Tepe Series have been metamorphosed in amphibolite facies.

2. Both types of massive and disseminated ore preserve the peculiarities of pro- and retrograde metamorphism. The effects of the retrograde metamorphism are generally more visible in the massive ore of lenses 1 and 3 and the surrounding disseminated ore.

3. The most essential texture and structure ore peculiarities are the following : a) the rhythmically stratified textures, with fine to coarse interstratification of sulphide (especially pyrite) with layers of quartz-magnetite±barite and quartz-sericite-chlorite schists ; b) they preserve in general the initial limits between the ore layers and schists and the ore homogeneity and c) the presence of the ore medium-coarse granoblastic structure.

4. The typical peculiarities of retrograde ore consist in : a) the breccia texture with slight aspect of banded texture, which shows two principal aspects : the "bull texture" which consists more often of rounded grains of pyrite, magnetite and quartz in a matrix of plastic sulphide and gangue minerals ; b) the breccia texture with very fine-grained sulphide in cement and different fragments of quartz is associated with faults and shear zones and c) the fine-grained granoblastic structure and the superimposed twinning of the magnetite, chalcopyrite and sphalerite.

5. The textural and structural features of the Altîn Tepe ore deposit show many similarities to other ores from Broken Hill, Australia (Lawrence, 1973), Norway (Vokes, 1969), Kholodninskoe, U.S.S.R. (Shadlum, 1982) etc.

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PORPHYRY COPPER SYSTEMS IN THE SOUTH APUSENI MOUNTAINS — ROMANIA

BY

SERGIU BOŞTINESCU[1]

Introduction

The Badenian-Sarmatian calc-alkaline magmatism in the South Apuseni Mts (Metaliferi Mts) evolved in an area with complex tectonic structure, consisting of Precambrian and Paleozoic metamorphic rocks, Mesozoic island arc ophiolites, Mesozoic deposits, Upper Cretaceous-Paleogene and pre-Badenian magmatites and Miocene deposits.

The magma access ways, partly corresponding to the deep fractures bordering Miocene basins are most of all NW-SE oriented. Thus, several well individualized zones are delimited : Baia de Arieş, Roşia Montană-Bucium, Stănija-Zlatna, Căraciu-Brad-Săcărîmb and the Mureş Valley.

The prevailing andesitic volcanism resulted in the emplacement of some explosive products and in the formation of a lot of simple or very complex apparata, partly preserved in certain places ; moderately sized subvolcanic bodies have been emplaced at depth, typically by the end of the volcanic activity.

Significant metallogenetic processes are related to the Sarmatian phase of this activity, resulting in some impregnations within andesitic bodies, breccias and host rocks, more or less complex vein systems, metasomatic replacements with metallic minerals ; gold-silver mineralizations prevail at the upper levels, grading at depth to a dominantly lead-zinc character.

The subvolcanic realm is characterized by mainly porphyry copper deposits, partly turned to account.

Among these, the mineralizations from Deva (Borcoş et al., 1972) and Poieni (Ionescu, 1974) have been initially ascribed to the porphyry copper type. The other occurrences, mainly identified during the seventies, some by geophysical methods, have been the subject of local studies, among which those concerning the structures : Poieni (Ionescu et al., 1975 ; Petrulian et al., 1978), Bolcana (Udubaşa et al., 1978 ;

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Boştinescu et al., 1980), Valea Morii (Borcoş et al., 1980) ; Musariu (Borcoş et al., 1980), Voia (Berbeleac et al., 1982), or with a regional synthesis character (Ianovici et al., 1977 ; Boştinescu et al., 1981 ; Vlad, 1981 ; Andrei, 1981 ; Boştinescu et al., 1983).

A general view of the porphyry copper mineralizations in the Metaliferi Mts may be outlined based on the present available geological data which are summarized here below.

Main Features : Deva, Tarniţa

The porphyry copper structures in the South Apuseni Mts generally have common features, but differ in some details, some of which suggesting two evolutionary trends — Deva and Tarniţa ; the summary of their geological features is shown in Table 1.

Distribution

Excepting the northernmost structure — Baia de Arieş — the genetic conditions required for the generation of more or less evolved porphyry copper systems have been attained everywhere in the area where (Fig.) the Neogene volcanism occurs.

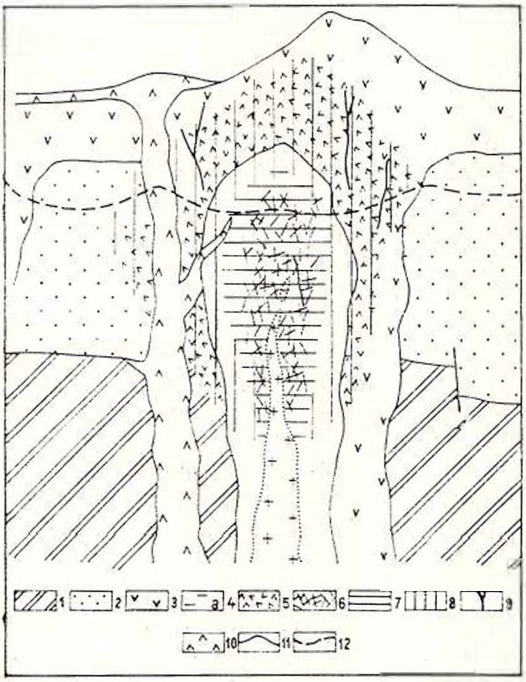

Fig. — Synthetical model of a porphyry copper structure.
1, metamorphites or ophiolites ; 2, Mesozoic and Neogene deposits ; 3, early volcanites ; 4, subvolcanic body : a, axial zone ; 5, breccias, brecciations ; 6, mineralized zone; 7, inner alteration zone ; 8, outer alteration zone ; 9, veins ; 10, late andesitic flows ; 11, topographic surface during the emplacement time ; 12, present-day topographic surface.

Thus, in the Roşia Montană-Bucium Zone, the Poieni and Tarniţa structures are known ; in the Stănija-Zlatna Zone : Valea Tisei, Măgura Poienii, Trimpoele ; in the Brad-Săcărîmb Zone : Rovina, Colnic, Musariu, Valea Morii, Bolcana, Voia ; in the Mureş Valley Zone : Deva.

As a whole, the distribution of the porphyry copper systems in the South Apuseni Mts is confined to a NW-SE trending area, which extends towards the NW up to a line that joins Roşia Montană and Brad ; towards the SE, its limit passes near Zlatna and Deva ; the domain of porphyry copper occurrences is thus larger in the Mureş Valley and closes in the Baia de Arieş Zone.

Geological Setting

The upper levels of the basement of the zone with copper mineralizations consists of crystalline schists or ophiolitic formations, apparently the only important element differentiating them in what concerns the predating geological setting.

This basement usually underlies Cretaceous sediments often in flysch facies, with ages ranging between Aptian and Senonian and Miocene detrital rocks.

The above mentioned formations are crossed and overlain by the products of the Tertiary volcanism, represented by various intrusions and extrusions.

It is frequently obvious that the porphyry copper structures are confined to the vicinity of faults, usually in their intersection points.

In each case, the existence of at least one volcanic or subvolcanic episode, previous to the porphyry copper emplacement suggests the same ways of emplacement and the rather late genesis of the latter.

The regional faulting sometimes favours the isometric character of the areal distribution of rootings. The mineralized intrusion occupies therefore a position near the axis of the whole zone. Thus, a central type structure appears, the development of some vein systems emphasizing the above mentioned symmetry, as it is specially the case in the Poieni, Bolcana and Deva zones.

The mineralized subvolcanoes seem to have finished their evolution which is strongly similar at comparable depths under the topographic surface at the emplacement time.

This evolution implies the appearance of some breccia formations.

In the deep zones of some structures, the surrounding rocks have been modified by the thermal contact metamorphic processes.

The alteration-mineralization processes favoured by characteristical cracks and brecciations are largely developed.

The evolution of the magmatic phenomena in some porphyry copper structures is ended with the emplacement of some small sized intrusions.

Petrography

The petrographical features of rocks implied in the formation of porphyry copper structures vary between close limits.

All these rocks are andesites or quartz bearing andesites with hornblende while all the other mafics may be biotite and/or hypersthene, sometimes augite.

The three main petrographic types — hornblende andesites, hornblende andesites with biotite, hornblende andesites with pyroxene —

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Main characters of porphyry copper

	Neogene pre-volcanism formations	Structural type	Petrographic types		
			pre-ore volcanics	mineralized porphyry)	post-ore volcanics
1 Poieni	Maestrichtian flysch Campanian wildflysch Precambrian metamorphism	central	α am	α am	α am bi α am px
2 Tarniţa	Albian flysch Baric magmatites?		α am	α am	unknown
3 Valea Tisei	Albian flysch Ophiolites		αam αam bi αam px	αam	unknown
4 Măgura Poienii	Albian flysch Ophiolites		αam	αam	unknown
5 Trîmpoele	Miocene sediments Cretaceous deposits Ophiolites?		αam	α am	unknown
6 Rovina	Albian flysch Ophiolites		α am α am px	α am	unknown
7 Colnic	Albain flysch Ophiolites	central	α am α am px	α am	unknown
8 Musariu	Badenian sediments Albian flysch Ophiolites		α am α am bi α am px	α am bi	α am px
9 Valea Morii	Miocene sediments Albian flysch Ophiolites		α am α am bi α am px	α am px	unknown
10 Bolcana	Miocene sediments Ophiolites Metamorphites	central	α am α am bi	α am	α am
11 Voia	Miocene sediments Ophiolites?		α am bi	α am	α am
12 Deva	Miocene sediments Turonian-Senonian sediments Paleozoic metamorphism	central	α am α am bi	α am bi	α am α am px

Abbreviations: α-andesite; fr-fresh: am-amphibole; bi-biotite; px-pyroxene; kf-potash feldspar; cl-chlorite; act-actinote; ep-epidote; cm-clay minerals; ser-sericite; alu-alunite; anh-anhydrite; ze-zeolites; cp-chalcopyrite; py-pyrite; bn-bornite; mgt-magnetite; cc-chalcosine; mo-molybdenite; po-pyrrhotine; ttr-tetrahedrite.

BLE 1

systems in the South Apuseni Mts

Breccias, brecciations	Contact phenomena	Alteration — mineralization						Base metal veins
		Axial zone	Inner zone			Outer zone		
important	present	frequent	kf, bi, py, cp, mgt, mo, bn, ttr (anh, ze)			cm, ser, py	cm, py, alu	minor
important	present	frequent	kf, cl, act, py, cp, mgt, (anh)	kf, cl, bi py, cp, mgt (anh)	kf, cl, cp, py, cp, (anh)	cl, cm, py	cm, py	present
abundant	unknown		kf, cl, bi, ep, act, py; cp			cm, cl, ser, py		abundant
present	insufficiently known							present
resent	present	?	kf, cl, ep, act, bi, py, cp			cm, cl, py		present
important	unknown	frequent	kf, cl, act, py, mgt, cp, po, (anh)	kf, cl, bi py, cp, mgt, po (anh)	kf, cl, ep, py, cp, mgt, po, (anh)	cl, cm, py	arg, py	present
present	unknown	?	kf, bi, act, ep, py, mgt, cp			cm, cl, ser, py		important
present	present		kf, cl, ep, py, cp, act, bi			cl, cm, py	cm, py	abundant
present	unknown		kf, cl, act, ep, bi, py, cp, mgt, ru			cl, cm, py	cm, py	abunadnt
present	unknown		kf, bi, cl, py, cp, mgt, bn, ru,(anh)			cm, cl, py	cm, py	important
present	present		kf, cl, ser, py, ep, (anh)			cm, cl, py, alu		important
abundant breccia-pipe	unknown	frequent	kf, bi, bn, ep, mgt, cc, (py, anh)			cm, mgt		minor

represent the early volcanites, the mineralized intrusion as well as the subsequent magmatic emplacement.

Usually, the structures are typically porphyric, sometimes weakly expressed in the deeper zones. The coarse-grained structure of the mineralized subvolcanic bodies is the main feature which distinguishes them, when fresh, from the other intrusions. The mining works and mainly the two relatively deep drillings made in the Musariu and Deva structures which reached the absolute depth of approximately 1200 m. underline the gradual passage towards the porphyry microdiorite facies while in places the structures may be even dioritic.

Fissurations, Brecciations

An advanced fissuration, within a horizontally and vertically limited zone is characteristic for the mineralized intrusions. This zone is not generally found within sterile porphyries. The intensity of this phenomenon may be moderate, i.e. Tarniţa, but can form crack-breccia aspects — Poieni and Bolcana breccias.

The stockwork developed within the body on the Băilor Brook from Deva contains microfissures with orientations statistically conformable to those of the two main faults in the area (NW-SE and WSW-ENE) which intersect each other in the ore zone, remembering the situation from Chaucha, Ecuador (Goosens, 1973).

The subvolcanic intrusion border with host rocks mainly at upper levels is the site of various brecciations, sometimes at microscopic scale. Fragments are angular to very rounded, consisting of andesites from the subvolcanic body, from the host rocks or both of the former and of the latter. The breccia matrix is only a small part of the rock body and is in fact represented by a microbreccia.

At Rovina, some of the structural aspects suggest brecciation in a partly plastical stage of some local magmatic injections.

At Poieni, Bolcana and Deva, some penetrations of fluid tuffitic material resemble those described in the Venice Alps (De Vecchi, De Zanche, 1974).

Some of the features, mainly of the Bolcana structure, suggest a quite incipient breccia pipe type column. But at Deva the mineralized andesitic-microdioritic body has resulted into a breccia pipe consisting of variously sized andesite-microdiorite and host rocks fragments as well as an earlier breccia fragment.

Several processes have been claimed for the genesis of such formations ; at Deva, the breccia pipe is the result of tectonic crushing described by Butler (1913) and Kuhn (1941), but mainly of the repeated collapse imagined by Perry (1961).

Contact Phenomena

In the neighbourhood of the subvolcanic intrusions, the deeper parts of the sedimentary host rocks may undergo the effects of contact metamorphism.

The Albian formations at Tarniţa and the Miocene rocks at Musariu and Voia (mainly their argillitic sequences) are affected in this way ; thus, some biotite hornfelses with a restricted areal development appear. The alternations consist of recrystallization of the primary minerals and the blastesis of minute biotite crystals, uniformly dispersed within the rock.

Some more advanced stages of the metamorphic process have been revealed at Tarniţa, resulting in andalusite and sillimanite bearing biotite hornfelses (Valentina Răduţă, oral communication).

Alteration — Mineralization

The alteration processes display very well expressed zonality, mainly controlled by the distance from the intrusion axis and less by structural elements ; the lithologic control is less important.

In the South Apuseni Mts, these processes, more than any other features, tend to discriminate between the Deva and Tarniţa porphyry copper structures.

The subvolcanic body axis corresponds to a discontinuous zone where the coarsely crystallized rock is practically unaltered. At Deva and Poieni andesites contain biotite with a deuteric aspect and at Tarniţa the axial zone has suffered a weak propilitization, mainly represented by chloritization.

Anyhow, for most of the mineralized intrusions belonging to the Tarniţa type, the hydrothermal alterations seem to be superimposed over a previous late magmatic propilitization.

Towards the outer zone we can outline two alteration areas equivalent to those of feldspar stability and destruction respectively, previously described by several authors.

Within the inner zone corresponding to the potash silicate facies of the main genetic models (Lowell, Guilbert, 1970 ; Hollister, 1975 ; Hollister, 1978), the paragenetic assemblages of the Deva type contain potash feldspar and biotite, together with chalcopyrite, pyrite, bornite, magnetite, molybdenite, sometimes rutile, while at depth anhydrite and locally zeolites appear.

The widely developed outer zone is characterized by the presence of clay minerals and pyrite.

Sericite, which frequently appears in the argillization zone, sometimes tends to be better represented towards its limit with the inner zone, which in the Poieni structure is more obvious ; this concentration corresponds to the phyllitic facies of the reference regions (Lowell, Guilbert, 1970), although it has not the same areal development and individualization.

Alunite is formed within the argillization zone from Poieni and Voia.

Quartz is abundant in all alteration zones both as a metasomatic product and as veins.

Pyrite has a characteristical behaviour in the Deva ore, being practically absent at middle levels of the mineralized column.

For the Tarniţa type, in the inner alteration zone the potash feldspar and neoformation biotite are associated with chlorite, actinote, albite and epidote. Sometimes a zonal succession is obvious within this zone ; from the inner to the outer zone successively follow assemblages with actinote, biotite and epidote, the albite being more important towards the borders. Anhydrite appears at depth, sometimes in large amounts.

Metallic minerals are represented by pyrite (sometimes pyrrhotite), chalcopyrite, magnetite ; bornite and molybdenite are very subordinate or sporadic.

In the outer zone, the main neoformations consist of clay minerals while pyrite is constantly present.

Typical for the Tarniţa type structures is the presence of chlorite in the argillization zone. Its concentration in the inner zone marks a subzone which could be as well an equivalent of the above mentioned phyllitic facies of alteration, when the iron excess characteristically modifies neoformation parageneses. Chloritization is sometimes very intense, and selectively emphasizes the primary structure of metasomatized rocks.

Quartz is also an ubiquitous hydrothermal product in this type of mineralization.

Even in the inner alteration zone the argillization and sometimes sericitization processes generally affect tectonized rocks.

Vein Mineralizations

The mineralized intrusion is crossed mainly at Tarniţa and Bolcana by fascicles of veins or small veins with a predominantly lead-zinc character. They cross the copper mineralization and represent a final stage of the metallogenetic activity associated with the subvolcanic body.

Otherwise, as for example at Musariu and Valea Morii, gold-silver veins are widely developed, belonging to complex systems, important at the metallogenetic district scale.

In some structures of central type (Colnic and Bolcana), the vein systems have less clear relationships with the subvolcanic body. For the Bolcana veins, a zonality of gold-silver and base metal characters was delimited (Cioflica et al., 1966), whose inner term is identified in the copper rich nature of mineralizations associated with the subvolcanic intrusion.

Geochemical Data

Some geochemical features are typical for the porphyry copper mineralizations from the Metaliferi Mts. For example, the Sr/Ba ratio is constantly near 1, but concentrations of the two elements are higher for the Deva type. The Zn/Pb ratio is approximately 14, contents in the Deva type being also higher ; the Deva ore is an exception, as the Pb concentration is rather high (Zn/Pb = 3).

In ores of the Deva Group the Mo/Cu ratio is high, while in those belonging to the Tarniţa Group the values of the Au/Cu ratio increase. Thus, Kesler's (1973) two classes of porphyry copper (CuMo and CuAu) are present ; the author attributes them to continental setting and to island arc setting respectively, which apparently correspond to the Deva and Tarniţa types of the Metaliferi Mts.

According to Sillitoe (1979), gold accumulation in porphyry copper deposits was favoured in the proximity of their present crustal setting.

Although there are not proofs, the gold source of the Tarniţa type porphyry copper could be related to the ophiolitic formations. As Sillitoe (1979), suggests, the concentration of gold in these ores is connected to the geochemical processes which determine iron enrichment within the feldspar stability field.

Concluding Remarks

The porphyry copper occurrences in the South Apuseni Mts, controlled by NW-SE fractures, are confined to a north-east trending zone, which parallels the main structural elements in the area.

These occurrences show very similar essential features, but some aspects evidently separate them into two distinct trends, corresponding to the Deva and Tarniţa types.

The Poieni, Deva and to a lesser degree Bolcana structures belong to the first type, while Tarniţa, Valea Tisei, Trîmpoele, Rovina, Colnic, Musariu, Valea Morii and Voia belong to the second ; Măgura Poienii is not completely known.

These two main types correspond to an ophiolite basement (Tarniţa) or to an area where ophiolites are absent (Deva).

The structural, petrologic and alteration-mineralization features of the porphyry copper systems in the Metaliferi Mts, mainly those of the Tarniţa Group, are similar to several examples of island arc setting.

They may hardly be matched in the known genetic models (Tab. 2). The Deva type resembles Lowell and Guilbert's model, while the Tarniţa type is ascribed to the dioritic model, with the difference that large areas of argillitic alteration occur in the Metaliferi Mts.

TABLE 2

A comparison between porphyry copper systems of the Metaliferi Mts and the main "porphyry copper" models

Features	Lowell and Guilbert model	Dioritic model	Metaliferi Mts
	(after Hollister 1975)		
Intrusion origin			
Mineralized intrusion	quartz-granodiorite monzonite	sienite-monzonite	diorite
Other present intrusions	quartz-diorite	diorite	quartz diorite
Alteration			
Central	potash (orthoclase-biotite and/or orthoclase-chlorite)	potash (orthoclase-biotite and/or orthoclase-chlorite)	potash (orthoclase-biotite; orthoclase-chlorite-actinote-biotite-epidote)
Outside the central zone	phyllitic (quartz-sericite-pyrite)	propilitic (chlorite-sericite or chlorite-epidote)	argillitic (sometimes clay minerals-sericite or clay minerals-chlorite)
Outside the phyllitic zone	argillitic		
Outside the argillitic zone	propilitic (chlorite-epidote)		
Diffuse pyrite	in potash and phyllitic zones	in potash zone or in both	in potash zone (except Deva) and in argillitic zone
Mineralization			
Quartz in fissures	common	sporadic	common
Orthoclase in fissures	common	sporadic	rare
Albite in fissures	traces	common	rare
Magnetite	scarce	common	common (sometimes abundant)
Pyrite in fissures	common	common	common (except Deva)
Molybdenite	common	rare	sporadic
Chalcopyrite/bornite ratio	3 or more	3 or less	much more than 3 (subunitary at Deva)
Chalcopyrite dissemination	present	important	present, sometimes important
Gold	rare	important	important
Structure			
Breccia-pipe type	can appear	rare	column only at Deva
Stockwork type	important	important	common

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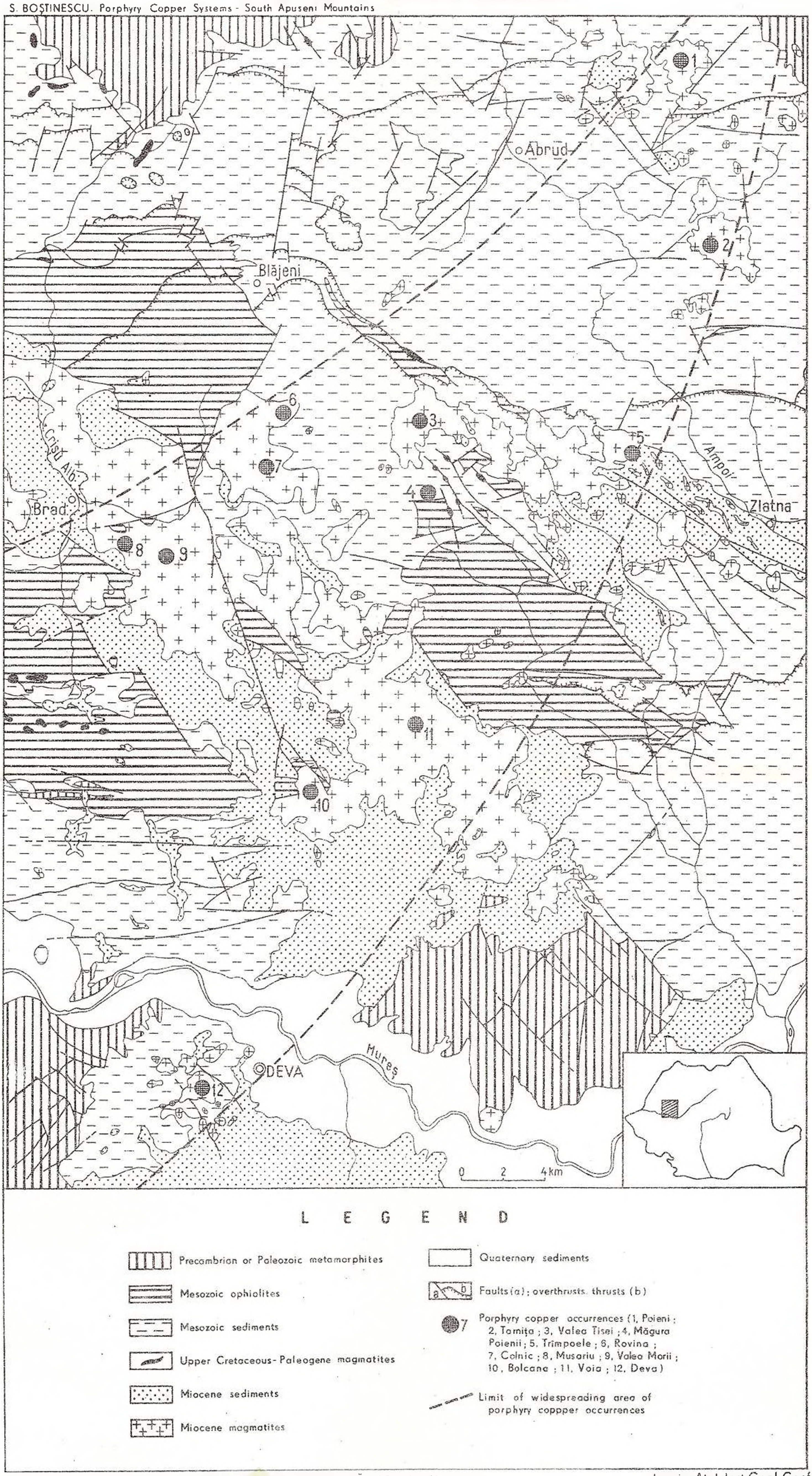
DISTRIBUTION OF PORPHYRY COPPER OCCURRENCES IN THE SOUTH APUSENI MOUNTAINS

S. BOŞTINESCU. Porphyry Copper Systems - South Apuseni Mountains



Imprim.Atel.Inst.Geol.Geof.

ALPINE METALLOGENY IN ROMANIA

BY

GRAŢIAN CIOFLICA [1], ŞERBAN VLAD [2]

Introduction

The purpose of this paper is to discuss metallogeny in association with the evolution of tectonics and magmatism during Alpine time on the Romanian territory. It provides therefore an outlook on the metallogenetic development during a complete cycle by taking into account the ore types associated with successive intracontinental rifting (North Dobrogea, Ditrău), spreading areas (South Carpathians), subduction related settings (South Apuseni Mts, East and South Carpathians), collision and post-collision related settings (East and South Carpathians, South Apuseni Mts) (Fig. 1).

Metallogenesis Related to Intracontinental Rifting

The Ditrău alkaline massif (East Carpathians) with related carbonatites and associated Mo ores was emplaced in a Mesozoic intracontinental rift. The alkaline massif is a quasi-ring-like intrusive (Anastasiu, Constantinescu, 1980) and exhibits a complex structure : the innermost part is built up of foid rocks, surrounded discontinuously by syenite and monzonite rocks ; hornblendite and diorite rocks are confined to the north-western margin of the composite pluton, and granites and alkali granites occur especially at the contact with the basement (Fig. 2). These rocks belonging to the main magmatic event are associated with lamprophyre, microsyenite, even foid, alkali granite and aplite dykes. Albitite segregations and carbonatic veins with sulphide minerals are found in places. Mo-bearing carbonatites are developed commonly as veins, but also as nests, bands, veinlets and disseminations. Carbonatites and related ores formed discontinuously from the late ortho-magmatic to the hydrothermal stages. In the north (Constantinescu et al., 1983) ilmenorutile, ilmenite, monazite, tapiolite, columbite and sulphide minerals are the characteristic association within diorite and hornblendite rocks, whereas xenothime, sulphides and niobotantalates are common within alkali syenite rocks found in the eastern and southern parts of the massif.

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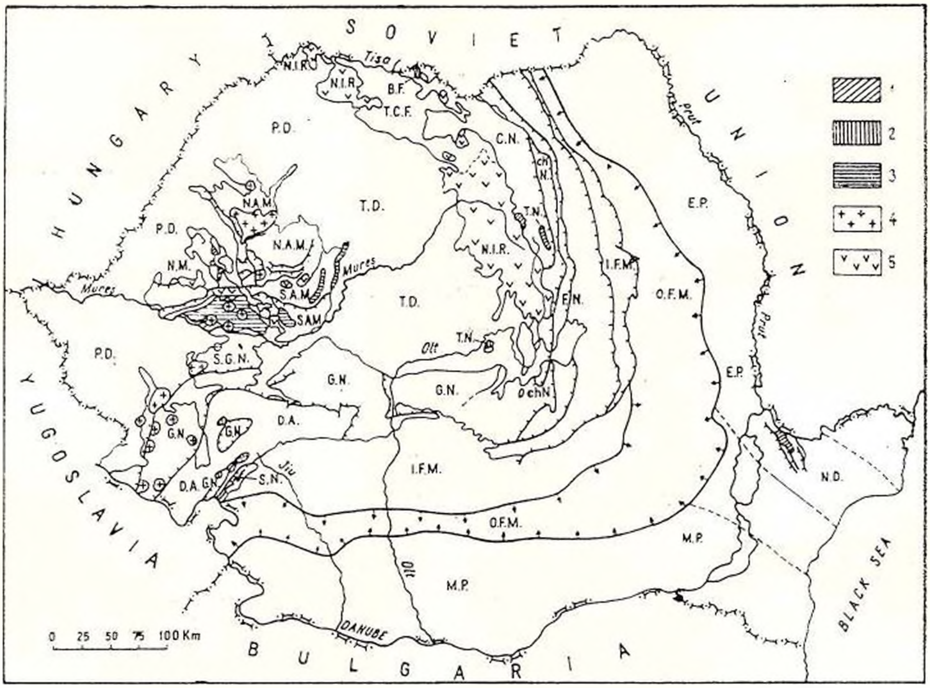


Fig. 1 — Distribution of Alpine magmatism in Romania (modified from Cioflica et al., 1980).

1, short-lived rift-related igneous rocks ; 2, ocean floor spreading-related igneous rocks ; 3, subduction-related Mesozoic (Lower Jurassic-Neocomian) magmatism ; 4, subduction-related Senonian-Paleocene magmatism (banatites) ; 5, subduction-related Neogene volcanics. TCF, Transcarpathian flysch ; CN, Central East Carpathian Nappes ; BF, Black Flysch nappes ; Tn, Transylvanian Nappe ; Ch N, Ceahlău Nappe ; FN, Flysch Nappe ; IFM, inner flank of molasse ; OFM, outer flank of molasse ; SGN, Supragetic Nappe ; GN, Getic Nappe ; SN, Severin Nappe ; DA, Danubian Autochthon ; NAM, North Apuseni Mountains ; SAM, South Apuseni Mountains ; TD, Transylvanian Depression ; ND, North Dobrogea orogenic system ; EP, East European Platform ; MP, Moesian Platform.

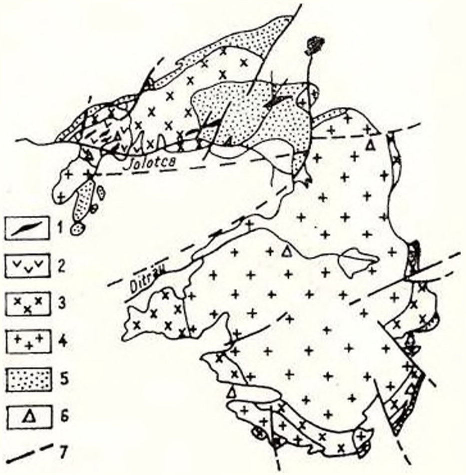


Fig. 2 — Geology and structure of the Ditrău alkaline massif (acc. to Anastasiu, Constantinescu, 1980).

1, hornblendite ; 2, diorite ; 3, syenite, alkali syenite, monzonite, monzodiorite ; 4, foid rocks ; 5, granite, alkali granite ; 6, carbonatites with related ores ; 7, faults.

The North Dobrogea area evolved during Alpine times as an aulacogen-like failed arm in relation with active rifting developed into the Tethys Ocean (Vlad, 1978). It became inactive before reaching the stage of ocean-floor spreading and was filled by clastics and, especially, carbonate prevailing sediments. The lower part of the sedimentary succession was penetrated by bimodal (basaltic-rhyolitic) magmatism. The mineralization is stratabound within Spathian calcareous-terrigenous turbidites penetrated by volcanics. It consists of carbonate-hosted and, of lesser extent, rhyolite-associated ores as follows : stratiform sedimentary barite bodies formed within minor submarine depressions ; small Ba ± Pb-Zn veins in steeply dipping fractures mostly along the crests of anticlines ; restricted Fe-bearing infiltration skarns in the vicinity of the Luncavița-Consul Line ; Pb-Zn ± Ba, F stockworks in K-altered rhyolites.

Metallogenesis Related to Ocean Floor Spreading Areas

Ophiolites of the South Carpathians are related to the Severin Nappe. They were considered of ocean floor type according to geodynamic interpretations (Rădulescu, Săndulescu, 1973). An elongated basin with oceanic crust acted between the Getic and the Danubian realms. It promoted basaltic flows and pyroclastics associated with Lower Sinaia Beds during Lower Cretaceous times. The Severin Nappe which contains them is tectonically emplaced between the Getic Nappe and the Danubian Autochthon. It migrated completely from the roots during Upper Cretaceous compressions, when collision between Getic and Danubian realms was reached. The metallogenesis related to these obducted ophiolites yielded Cu-pyrite ores at Baia de Aramă. Cioflica et al. (1981) provided geological and geochemical evidence to characterize these ophiolites as tholeiitic ocean floor basalts formed in a small ocean setting. Accordingly the related ore deposition was ascribed to the Joma type of Pearce and Gale (1977). The Cu-pyrite ores are located in a lava unit of prevalent basaltic character. The basalts that exhibit in places pillow structure contain small size stratiform pods of massive chalcopyrite, with subordinate amounts of early pyrite and subsequent sphalerite in quartzose gangue. The massive ore is commonly underlain by pyrite ± chalcopyrite, sphalerite stockworks (Fig. 3).

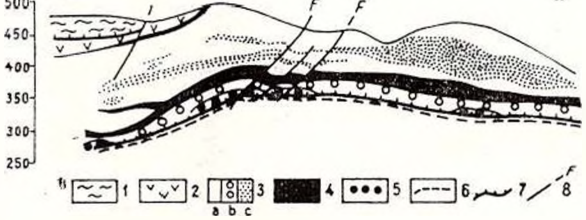


Fig. 3 — Section through mineralized ophiolites from Baia de Aramă (acc. to Cioflica et al., 1980).

1, crystalline schists of the Getic Nappe ; 2, serpentinite protrusion ; 3, ophiolite complex (a, upper ophiolites ; b, lower ophiolites ; c, mineralization) ; 4, Cretaceous black argillite ; 5, Sinaia Beds ; 6, Upper Cretaceous flysch of the Danubian Unit ; 7, thrust ; 8, fault.

Metallogenesis of Subduction-Related Settings

The Mesozoic magmatic rocks of the South Apuseni Mts are located between crystalline schists of the North Apuseni Mts and of the South Carpathians. They derived from a tri-stadial magmatism, that is first stage-tholeiite series (Lower Jurassic-Callovian) and second stage-calc-alkaline series (Upper Callovian-Neocomian) representing island arc magmatism, and third stage-spilitic complex (Barremian-Lower Aptian ?) representing active marginal basin magmatism (Cioflica et al., 1980).

The related metallogenesis consists of Fe-Ti-V and Ni late magmatic segregations in gabbroic intrusions and Cu-pyrite veins and stockworks in basaltic lavas of the tholeiite series, whereas the calc-alkaline series comprises Mn volcano-sedimentary ores.

The Fe-Ti-V segregations occur in layered gabbroic bodies (e.g. Căzăneşti-Ciungani) as nests, lenses and disseminations consisting of titanomagnetite and ilmenite. The Ni ores are found at Ciungani where a small size metallic pod contains pyrrhotite, pentlandite and sporadic chalcopyrite and magnetite.

The Cu-pyrite volcanogenic ores associate with basalts and fall in the Gjersvik type of Pearce and Gale (1977) (Cioflica et al., 1981). At Pătîrş (Fig 4) the mineralization consists of pyrite + chalcopyrite veinlets and a massive pyrite pod, surrounded by a disseminated pyrite halo. At Căzăneşti-Ciungani, Almăşel, Roşia Nouă, Corbeşti and Pietriş similar Cu-pyrite ores (veins, stockworks) are controlled by various fractures and brecciated basalts. Mn volcano-sedimentary ores are associated with jaspers at Şoimuş-Buceava-Pîrneşti and Godineşti.

The Senonian-Paleocene calc-alkaline magmatic belt (banatites) related to distinct subduction-related settings runs from the Apuseni Mts to the South Carpathians. The magmatic emplacement was tectonically controlled and exhibits a specific polystadial character : 1) volcanics consisting of andesite, dacite, rhyolite rocks and associated pyroclastics ; 2) subsequent intrusive stage divided into three phases : a) early minor diorite bodies ; b) plutons and subvolcanic bodies of monzodiorite, diorite → granodiorite and granodiorite → granite composition ; c) final acidic dykes and concurrent basic and lamprophyre dykes.

The intrusions of the main intrusive event were accompanied by recrystallization and metasomatism. The latter produced significant replacement zones with widespread mineralization inside and around

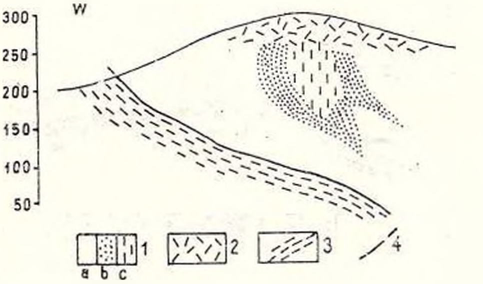


Fig. 4 — Section through mineralized ophiolites from Pătîrş (acc. to Cioflica et al., 1980). 1, ophiolites (a, unmineralized ; b, pyrite impregnations ; c, pyrite + chalcopyrite stockwork) ; 2, Gossan ; 3, Cretaceous flysch ; 4, reversed fault.

bodies of the main magmatic event of subsequent dykes. Porphyry copper of Lowell and Guilbert type and skarn deposits prevail, whereas vein deposits occur scarcely (Cioflica, Vlad, 1980) (Fig. 5).

The North Apuseni Mts sub-belt is characterized by major granodiorite → granite magmatism and widespread base metal deposition.

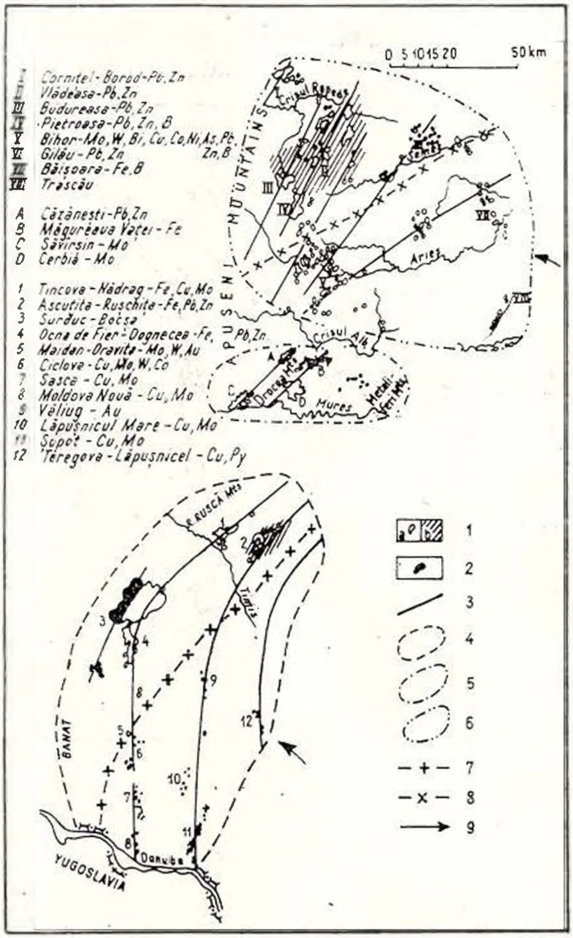


Fig. 5. — Distribution of the Senonian-Paleocene magmatism and ore deposits in Romania (acc. to Cioflica, Vlad, 1980). 1, igneous bodies belonging to the granodiorite → granite evolution line: a, plutons and subvolcanic bodies; b, volcano-plutonic complex; 2, igneous bodies (plutons and subvolcanic bodies) belonging to the monzodioritic, dioritic → granodioritic evolution line; 3, petrogenetic alignment; 4, Banat-Poiana Ruscă Mts sub-belt; 5, Drocea-Metaliferi Mts sub-belt; 6, North Apuseni sub-belt; 7, Cu-Pb line; 8, complex-Pb, Zn line; 9, direction of subduction.

The well-expressed zoning is represented by the complex zone of the Bihor-Gilău (Mo, Bi, W, Cu, Co, Ni, Pb, Zn, B and Fe ores associated with skarns), followed landwards by the base metal zone (hydrothermal Pb-Zn ores in the Vlădeasa Massif and Cornițel-Borod Depression).

The South Apuseni Mts sub-belt with transverse position with regard to the adjacent Banat-Poiana Ruscă and North Apuseni Mts sub-belts is represented by monzodiorite, diorite → granodiorite magmatism with Cu-impregnated and Fe skarn deposits and granodiorite → granite magmatism with Pb-Zn and Mo vein deposits.

The Banat-Poiana Ruscă Mts sub-belt is represented by the inner zone (South Banat) with monzodiorite, diorite → granodiorite magmatism and Cu-Mo porphyry and skarn deposits (e.g. Moldova Nouă), and landward, by the outer zone (North Banat-Poiana Ruscă) with granodiorite → granite magmatism and Fe, Pb-Zn skarn deposits (e.g. Dognecea, Ocna de Fier) or restricted Mo-porphyry occurrences (e.g. Oravița). The well expressed Cu(Mo) → Mo → Fe, Pb-Zn transverse zoning is characteristic of the Andean-type subduction related setting (Vlad, 1979).

The Neogene magmatism consists of calc-alkaline products of mainly andesitic type related to westward subduction of the eastern basin (Rădulescu, Săndulescu, 1973). It developed within the inner part of the Carpathians (Fig. 1) and consists of the East Carpathians volcanic arc and volcanic occurrences along NW-SE and E-W alignments within the mature island arc structure of the South Apuseni Mts. Subvolcanic bodies which penetrated the volcanics promoted recrystallization of surrounding rocks as well as metasomatism.

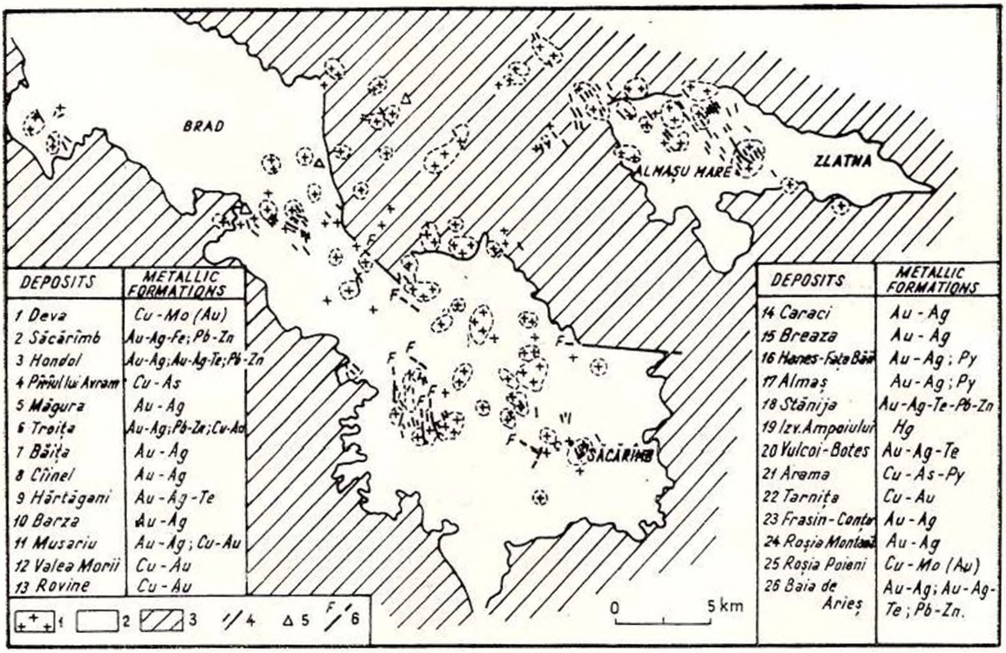
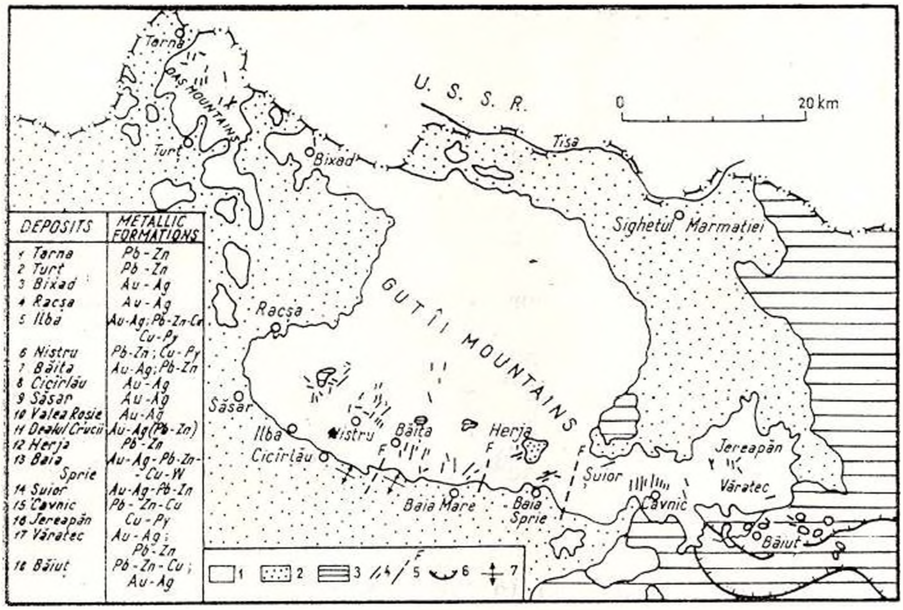


Fig. 6 — Deposits in the South Apuseni Mountains (modified from Giuscă et al., 1969).

1, Neogene magmatism (necks, dykes ; lavas not represented) ; 2. Miocene molasse ; 3, pre-Tertiary basement ; 4, veins ; 5. porphyry copper ; 6. faults.

The related metallogenesis is of hydrothermal type and yielded commonly Au-Ag, base metal and Cu ores with subordinate amounts of Hg, exhalative S and Fe (siderite) (Giuşcă et al. 1969 ; Borcoş et al., 1980). The ores occur commonly as veins and stockworks which are intimately controlled by fractures and breccia pipes.

The South Apuseni Mts (Fig. 6) are characterized by Au-Ag veins and stockworks ; they are found within volcanics and associate with base metal veins. porphyry copper systems of stockwork or breccia-pipe type and peripheral Hg occurrences. The host volcanics filled tectonically controlled Tertiary sedimentary basin which crosses the mature Mesozoic island arc. Porphyry copper deposits are of peculiar character for the Tertiary metallogeny and were assigned to the diorite model (Ianovici et al., 1977). They contrast with the Senonian-Paleocene systems by lack of skarnization halo replaced by peripheral Au-Ag and Pb-Zn. The porphyry systems show a Cu-Au character (Tarniţa, Rovina, Valea Morii, Musariu) and a Cu-Mo(Au) character (Deva, Roşia Poieni).



DEPOSITS	METALLIC FORMATIONS
1 Tarna	Pb-Zn
2 Turt	Pb-Zn
3 Bixad	Au-Ag
4 Racşa	Au-Ag
5 Ilba	Au-Ag; Pb-Zn-Cu Cu-Py
6 Nistru	Pb-Zn; Cu-Py
7 Băiţa	Au-Ag; Pb-Zn
8 Cicîrlău	Au-Ag
9 Săsar	Au-Ag
10 Valea Rosie	Au-Ag
11 Dealul Crucii	Au-Ag(Pb-Zn)
12 Herja	Pb-Zn
13 Baia Sprie	Au-Ag-Pb-Zn-Cu-W
14 Suior	Au-Ag-Pb-Zn
15 Cavnic	Pb-Zn-Cu
16 Jereapăn	Cu-Py
17 Văratec	Au-Ag; Pb-Zn
18 Băiuţ	Pb-Zn-Cu; Au-Ag

Fig. 7 — Deposits in the Oaş-Gutîi Mountains (modified from Giuşcă et al., 1969). 1, Neogene volcanics ; 2, Neogene molasse ; 3, Paleogene flysch ; 4, veins, stockworks ; 5, fault ; 6, overthrust ; 7, anticline axis.

The southern margin of the Gutîi Mts contains significant ore deposits (Fig. 7). In the northern part of the Oaş Mts mineralizations occur between Tarna and Bicsad. In these regions base metal veins prevail ; it is noteworthy that the deeper parts of this setting may promote porphyry copper metallogeny. In the Ţibleş-Toroiaga Mts sulphide mineralization is related to subvolcanic bodies, whereas in the Rodna Mts they connect with breccia pipes or replace crystalline limestones.

In the Harghita-Gurghiu-Călimani Mts only Hg, S, Fe (siderite) and restricted Au-Ag and base metal ores are known up to the present. It is however likely that porphyry copper systems may develop at depth (Peltz et al., 1982).

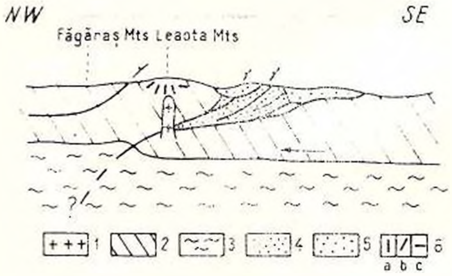
Metallogenesis of the Passive Margin-Related Settings

The sedimentary Fe deposit from Căpuş occurs within epicontinental sediments found north of the Gilău crystalline massif, as limonitic and glauconitic ores (Vinogradov et al., 1963).

Metallogenesis of the Continental Collision-Related Settings

In the eastern part of the South Carpathians, Supragetic (East Făgăraş Mts) and Getic crystalline schists (Leaota Mts) are cut in specific areas by numerous minor veins that contain Bi minerals, Co-Ni sulphides and arsenides, base metal minerals and gold. The spatial distribution of the ores strongly suggests a periplutonic arrangement within the contact aureole of two presumed deep-seated plutons. Regionally, the ore occurs along a NNE-SSW alignment which is to be related to the above mentioned plutons ; accordingly, it is however likely that a deep-seated continent-continent collision belt of Erzgebirge or Cornwall type runs in the east South Carpathians (Vlad, Dinică, 1984) (Fig. 8).

Fig. 8 — Continent-continent collision-related ore deposition in the eastern South Carphathians (acc. to Vlad, Dinică, 1984).
1, continent-continent collision-related granite-alkaline granite pluton ; 2, continental crust ; 3, upper mantle ; 4, Lower Cretaceous flysch ± obducted ophiolites ; 5, molasse ; 6, ores (a, Co-Ni ; b, base metal ; c, gold).



Metallogenesis Related to Post-Collision-Related Settings

At Jitia (East Carpathians) base metal ores of diagenetic and partly epigenetic nature are found within Miocene sediments.

Conclusions

The Alpine ore deposition in Romania was intimately connected with magmatic and other controlling factors during the Mesozoic-Cainozoic cycle of early intracontinental rifting, ocean floor spreading, subduction, continental collision and post-collision rifting.

Unevolved settings were assigned to the early intracontinental rifting. The North Dobrogea aulacogen-like Trough exhibits a characteristic Triassic bimodal magmatism and Ba-Pb-Zn and Fe metallogeny ; the metals resulted by means of concentration from the sialic crust during rifting and hot brines formation. The Ditrău alkaline pluton and associated carbonatites with Mo deposits, occurred above a thermal dome in continental environment ; it is likely that Mo and associated metals were subtracted from the sialic crust by the alkaline melts.

Various evolved settings promoted significant metallogenesis during Alpine times. The deposits formed in oceanic setting are represented by Cu-pyrite volcanogenic ores from Baia de Aramă (South Carpathians). Subsequent compressions yielded various deposits related to pulsative subduction events. Thus, the closing of the South Apuseni Mts basin gave rise to island arc volcanics with associated deposits : Fe-Ti-V and Ni late magmatic segregations, Cu-pyrite volcanogenic ores. Following the suture of the island arc to the North Apuseni Mts sialic block, sedimentary Fe deposits formed on the north-western passive continental margin of the paleo-basin.

A rather complex subduction which promoted also an Andean type magmatic arc in Banat-Poiana Ruscă gave rise to banatitic igneous rocks and associated metallogenesis. During the main magmatic event two evolution lines developed differentially. The monzodiorite, diorite → granodiorite acted as Cu carrier from the deep-seated source, that is upper mantle-subducted oceanic crust. The granodiorite → granite magmatism mobilised various metals from the sialic crust by palyngenesis ; it is situated landwards and yielded Mo, base metal deposits.

The Tertiary subduction inferred from specific evidence gave rise to external andesite arc (East Carpathians) with Au-Ag and base metal metallogenesis and to internal andesite occurrences (South Apuseni Mts) wherein Au-Ag and porphyry copper metallogenesis prevails.

It is to be mentioned that Cu metallogenesis sequences derived from deep-seated sources (upper mantle, subducted oceanic crust) taking into account both scale metallogeny and ore formation within restricted deposits (e.g. porphyry systems and the Baia Sprie xenothermal deposits). Pb-Zn, Mo, Au-Ag sequences derived simultaneously from sialic environments by palyngenesis. The higher Au contents of South Apuseni Mts mineralizations are presumably due to remobilisation from island arc tholeiites, too.

Co, Ni, Bi, Ag, Au, Cu mineralizations and related magmatic-tectonic features from the Leaota-Făgăraş Mts provide evidence for interpreting such settings as continental collision magmatism and related metallogenesis. The post-collisional event of the southern East Carpathians promoted sedimentary Pb-Zn metallogenesis at Jitia, in connection with the Lower Miocene molasse. Finally, intracontinental basalts (7,3 ± 0,6 m.y, according to Rădulescu et al., 1981) are characteristic of late post-collisional rifting and lack in metallogenetic importance.

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THE TECTOSTRUCTURAL FACTOR, A FUNDAMENTAL CRITERIUM TO OUTLINE THE METALLOGENETIC (PETROMETALLOGENETIC) PROVINCES — EXEMPLIFICATION ON THE ROMANIAN TERRITORY

BY

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The first observations and studies on provinces and metallogenetic epochs were made at the beginning of our century ; this idea was developed by de Launay (1913) in Europe, Emmons (1913) and Spurr (1923) in America. Subsequently, many other researchers dealt with this question which became not only interesting, but also useful. Among them, we can speak of Stanciu (1930), Lindgren (1933), Blondel (1938), Hills (1947), Turneaure (1955), Bateman (1956), Routhier (1963), Petraschek (1965), Janković (1967), Vokes (1971), etc. The Soviet geologists had an important contribution to this question, starting by Obrucev, Fersman, and continuing by Smirnov, Bilibin (1955), Smirnov (1959), Magakian (1959, 1967), Tatarinov (1967) and others.

The notions of province, its subunits (sub-province, district, metalliferous field, deposit, ore body), and metallogenetic epoch are widely discussed in literature, but without finding a common reference point for all cases, in order to define their content.

We consider the metallogenetic provinces as geostructural spatial units, remarked by a structogene evolution, magmatism and specifical metallogeny, and the metalogenetic epochs as geochronological sequences during which metallization took place.

A new stage in the study and interpretation of metallogenetic provinces is connected to the idea of general tectonics, to which many researches had important contributions (Noble, 1970, 1974 ; Turneaure, 1971 ; Sillitoe, 1972 ; Closets, 1972, etc.).

According to this theory, the metallogenetic provinces are some spatial units which reflect the geostructural domain where petrometallogeny manifests itself, by outlining for the magmatic domain : provinces of lithosphere compression ; provinces of compression zones (subduction), namely the Alpine type and the island arc type ; provinces of the hot regions, the oceanic type (weaklier expressed) and the con-

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tinental type. Here, the metallogeny of the tectono-magmatically activated zones can be added (Şceglov, 1979).

According to the geological time of formation, we can separate pangeic (petrometallogenetic) metallogenetic provinces, whose genesis is previous to the detachment of the metallized surface from the originating continent, and post-pangeic ones, formed after the detachment of the continental plates from the primitive block. The result is the migration of the metallogenetic provinces, what de Closets (1972) has called "la dérive des gisements" (the deposit drift).

Provinces and Magmato-Metallogene Epochs on the Romanian Territory

Long time after the first study of professor Stanciu (1930), entitled "The mineral provinces of Romania", the problems of the relationship between the metallization time and space of the Romanian territory were a theme requested by many geologists (Socolescu, 1961 ; Dimitrescu, 1961, 1972 ; Ianovici et al., 1966 ; Rădulescu, 1966 ; Rădulescu et al., 1970 ; Lucca, 1967 ; Codarcea-Dessila, 1968 ; Savu et al., 1970 ; Giuşcă, 1974 ; Mârza, 1981, 1982).

The tectonic-magmatism-metallization causality relationship imposes the substantiation of the concept of magmato-metallogenetic (or metallogenetic) province on the tectostructural criterium ; in this case, the activated mobile or stable tectonic structures are the origin of magma formation and concentration. The largest tectostructural zones of the earth crust correspond to the magmato-metallogene belts ; provinces and sub-provinces, etc. are encompassed by their segments (subdivisions). From the functional type point of view, the geostructures activate as spreading zones, compression zones and stable zones, each with characteristical magmatism and associated metallogeny.

According to the geostructural criterium, which was substantiated on general tectonics concept (Rădulescu, Săndulescu, 1973 ; Bleahu et al., 1973) the following provinces associated with the geostructural units (petrometallogene) of the Romanian territory are to be mentioned (Mârza, 1982) : provinces of stable units ; provinces of mobile units ; provinces of rift zones ; provinces of subduction zones.

Taking into account the common geological evolution of the boundary territories between Romania and the neighbouring countries, the formation of metallogenetic (petrometallogenetic) provinces was regarded in a unitary framework (Fig.).

Provinces of Stable Zones

1. The province of the East European Platform : the Moldavian Platform sub-province ; 2. the Moesian Platform province : the Wallachian sub-province ; the South Dobrogea province ; the Central Dobrogea sub-province.

Provinces of Mobile Zones

3. The Carphathian Crystalline province : the East Carphathians sub-province ; the South Carpathians sub-province ; the Apuseni Mts province ; 4. the North Dobrogea Orogene province.

Provinces of Rift Zones (Riftogene)

5. The Balkan province of gabbro-peridotitic complexes: the North Danubian sub-province of Paleozoic ultrabasic magmatites, belonging to the Balkan-Anatolian-Iranian (Tethys) magmato-metallogene belt. corresponding to the Paleozoic and old Alpine ultrabasic magmatites (over 4000 m) ; 6. The metallogenetic province associated to the Mesozoic ophiolitic magmatism from the Apuseni Mts : the Highiş sub-province, sulphide mineralizations ; the Drocea-Mureş sub-province ; the Trascău sub-province.

Provinces of Subduction Zones (Subductogene)

7. The metallogenetic province associated to the banatitic (Laramian) magmatism ; it belongs to the Dacian-Balkan-Transcaucasian-Iranian Belt, associated to the Laramian magmatism and developed on about 4500 km long : the Banat sub-province ; the Poiana Ruscă sub-province ; the Apuseni Mts sub-province ; 8. The metallogenetic province of the Carpathian neoeruptive chain : the Börzsöny-Zemplény (Tokaj) sub-province in Hungary ; the Vihorlat (Czechoslovakia)-Gutin sub-province ; the Toroiaga-Căliman-Harghita sub-province ; 9. The metallogenetic province associated to the neo-magmatism from the Apuseni Mts, characterized by native gold deposits (telluriums), porphyry copper and less polymetallic ones : the Arad-Săcărîmb district ; the Stănija-Zlatna district ; the Roşia Montană-Bucium district. The Baia de Arieş metallogenetic field.

Main Metallogenetic Epochs on the Romanian Territory

The geological evolution of the Romanian territory has numerous phases or epochs with magmato-metallogene activity, which took place from the Precambrian to the Pliocene (Tab.). The Precambrian epochs belong to the Prebaikalian epoch (Cadomian).

The basic idea with important metallogenetic significances presented in this paper is considered to be the connection of the endogene (petrometallogene) metallogenetic provinces with the geostructural units which from a unitary type of developing the geological processes ; according to their genotype, both the petrogenesis and the associated metallogeny are pointed out by distinct notes (chemism, mineralogy, metallogenetic phase, deposit form, etc.). The geostructural individuality of metallogenetic provinces is marked as well by metallization (or petrometallogene) epochs. Therefore, there are monocyclic and polycyclic regions, the latter being tectono-magmatically reactivated during various phases. In a geostructure, the epochs alternate in time and superpose in space.

TABLE

Main endogene metallogenetic epochs corresponding to provinces of the Romanian territory and generated mineral resources

Province	Sub-province	Epoch	Geostructural type	Metallogenetic type	Useful resources and examples	Observations
1	2	3	4	5	6	7
1. East European Platform province	Moldavian Platform	Prebaikalian	Stable	Metamorphic	Granites, pegmatites, migmatites	
2. Moesian Platform province	Wallachian	,,	,,	,,	,,	
	South Dobrogea, Central Dobrogea	Karelian	,,	Sedimentary metamorphosed, volcanogene-sedimentary	Fe (Palazu Mare) Cu (Altîn Tepe)	
3. Carpathian Crystalline province	East Carpathians	Prebaikalian (Upper Proterozoic)	Orogene	metamorphic	pegmatites (Rodna, Preluca)	Rebra Series
		Prebaikalian (Upper Precambrian-Lower Paleozoic)	Island arc	Volcanogene-sedimentary	Sulphides (Pb-Zn) Valea Blaznei-Gușet (Rodna)	,,
			,,	,,	Mn (Iacobeni-Vatra Dornei)	Tulgheș Series
		Lower Cambrian	,,	,,	Sulphides (Cu, Pb-Zn) Baia Borșa-Fundul Moldovei-Bălan	,,
		Jurassic	Tectonomagmatic activation	Orthomagmatic	Cr (Breaza, Suceava district)	
		Jurassic	Hot spot fractures control (?)	,, Pneumat.-hydrothermal	Sienites with Ti, TR, etc. (Ditrău) Sulphides (Mo, Cu) and TR (Ditrău)	
	South Carpathians	Prebaikalian	Orogene	Metamorphic	Granitoids, pegmatites (Semenic-Sebeș-Lotru)	Sebeș-Lotru Series,
		,,	,,	,,	Kyanitic schists (Făgăraș-Lotru-Sebeș-Semenic, etc.)	Făgăraș, Sebeș-Lotru Series Sebeș-Lotru Series
		,,	,,	,,	Mn silica rocks (Sebeș)	

		?	,,	Hydrothermal (metamorphosed)	Au-Valea lui Stan (Lotru)	
		Devonian	Island arc(?)	volcanogene-sedimentary	Fe-(oxide-carbonatic) Poiana Ruscă	Bătrîna and volcanogene-sedimentary series
		?	,,		Sulphides (Pb-Zn), Muncelul Mic-Muncelul Mare-Veţel (Poiana Ruscă)	
		Sudete ?	Orogene Tectonomagmatic activation	Hydrothermal-metamorphic Orthomagmatic	Talc dolomites (Poiana Ruscă) ultrabasites (Cr, Ni, Pt), Sebeş Mts and Almăj	Carboniferous
	Apuseni Mts	Prebaikalian	Orogene	Metamorphic	Granites, pegmatites, migmatites (Gilău Mts)	Someş Series
		,,	Island arc	Volcanogene-sedimentary	Sulphides (Pb−Zn±Cu), Scrind-Răchiţele (Vlădeasa)	,,
		Rifean	Spreading zone	Orthomagmatic	Metabasites with magnetite, ilmenite, chalcopyrite (disseminations)	Biharia Series
		Laramian	Tectonomagmatic	Hydrothermal	Mainly hydrothermal mineralizations associated with meditic magmatites located in crystalline (Trascău-Băişoara, Ciucea-Remeţi) Complex sulphides in the sedimentary border (Băişoara sector)	Reflected metallogeny
4. North Dobrogea Orogene province		Prebaikalian Paleokimmerian	Orogene Tectonomagmatic activation (?)	Metamorphic Hydrothermal skarn Hydrothermal	Pegmatites, migmatites Magnetite-hematite, subordinately sulphides (Iulia), sulphides (Pb-Zn), barite (Somova-Minerii)	
5. Gabbro peridotitic complexes Balkan province	North Danubian	Paleozoic	Spreading zone	Orthomagmatic Hydrothermal-metamorphic	Cr (Banat) Asbestos (Banat)	

1	2	3	4	5	6	7
6. Metallogenetic province associated with the ophiolitic magmatism from the Apuseni Mts	Highiş	Paleoalpine ($T_3-J_1?-C_1$)	,,	Hydrothermal	Basites with sulphides (Highiş)	
				Orthomagmatic	Vanadiferous titanomagnetite (Ciungani)	
	Drocea-Mureş	,,	,	Hydrothermal	Sulphides (Pb-Zn) Vorţa?; sulphides (Cu), Valea Lungă?; Ciungani	
	Trascău	,,	,,	Volcanogene-sedimentary	Mn-Fe (Troaş-Pirneşti)	
				,, (?)	Mn (Pădurea Turzii, Buru)	
7. Metallogenetic province associated with the banatitic magmatism (Laramian)	Banat	Laramian	Spreading zone (island arc)	Hydrothermal skarn	Fe, Mo, Cu, Pb-Zn (Ocna de Fier, Dognecea, Oraviţa, Sasca Montană, Moldova Nouă)	Continues in Yugoslavia (Bor, Majdanpek etc.)
	Poiana Ruscă	,,	,,	,,	Pb-Zn (Cu), Poiana Ruscă	
	Apuseni Mts	,,	,,	,,	Cu-Mo-Bi (Băiţa), Fe (Băişoara) and Cu-Pb-Zn (Vlădeasa)	
8. Metallogenetic province of the Carpathian Neoeruptive chain	Börzsöny Zemplény	Neogene	Subduction zone	Hydrothermal	Complex sulphides deposits	Hungary
	Vihorlat-Gutin	,,	,,	,,	Complex sulphides deposits Au(Oaş-Baia Mare), Hg in Soviet Union	
	Toroiaga-Căliman-Harghita	,,	,,	Hydrothermal (locally skarn)	Sulphides (Pb-Zn±Cu), Toroiaga, Ţibleş, Rodna; Hg-Sintimbru (Harghita); S-Negoiul Românesc (Căliman) Volcanites with sulphides (Birgău), magmatites (Ciceu, Dej) and zeolitic volcanic tuffs	

9. Metallogenetic province associated with the neomagmatism of the Apuseni Mts		,,	Island arc	Hydrothermal	Cu (porphyry copper), Deva, Roșia Poieni, Bucium-Tarnița, etc. Pb-Zn (vein), Băița-Crăciunești, hydrothermal metasomatic at Baia de Arieș, Vein, hydrothermal-metasomatic at Haneș-Larga; Au-Fe (Săcărîmb, Hondol Baia de Arieș, Brad, Roșia Montană etc. Hydrothermalized volcanites in the western part of the country	

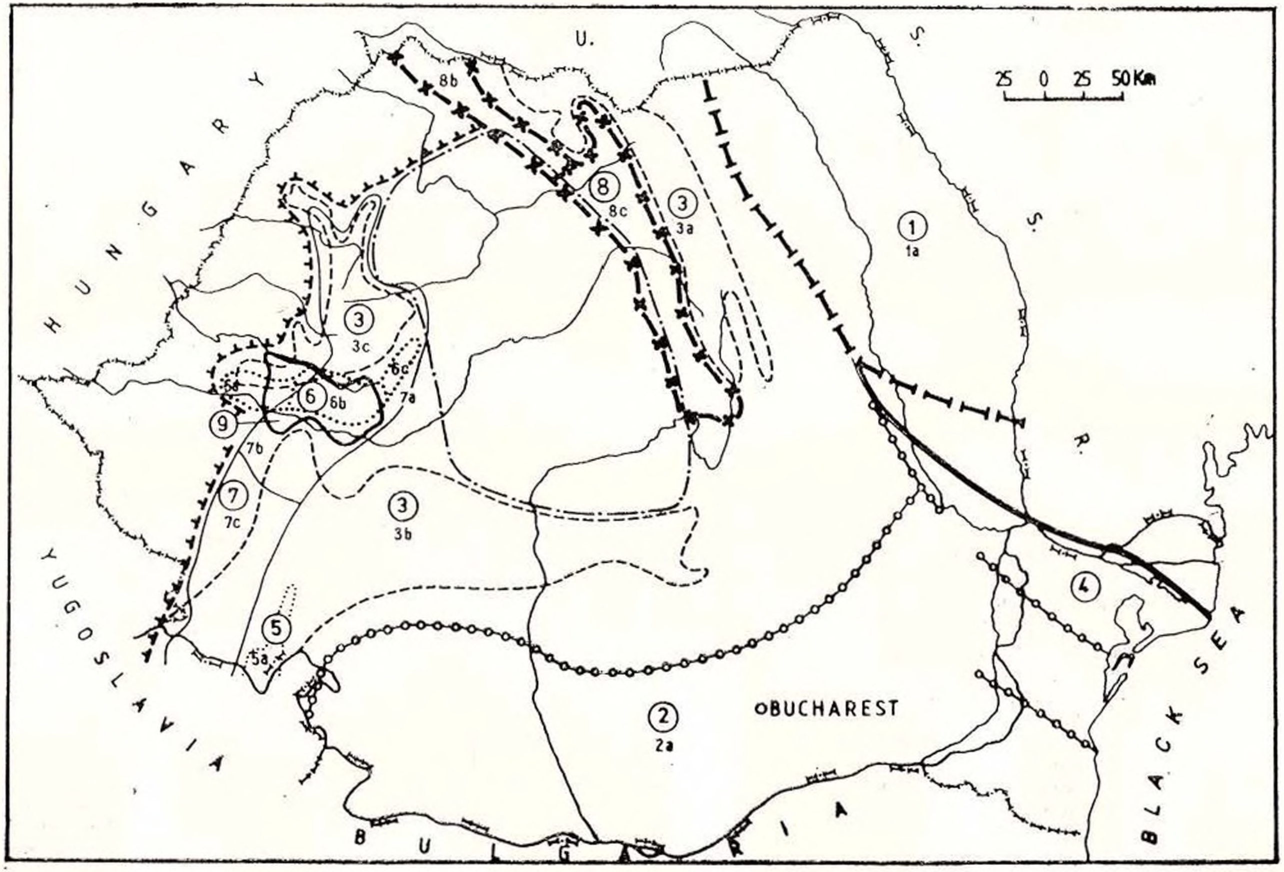
U.
S.
S.
R.
25 0 25 50Km
HUNGARY
YUGOSLAVIA
BULGARIA
BLACK SEA
BUCHAREST
8b
8
8c
3
3a
1
1a
3
3c
6c
7a
6a
6
6b
9
7b
7
7c
3
3b
5
5a
2
2a
4

Fig.

Fig. — Endogene metallogenetic (petrometallogenetic) provinces on the S. R. Romania territory, as compared to structural units.

1, East European Platform province, at the Moldavian Platform sub-province ; 2, Moesian Platform province ; 2 a, Wallachian sub-province ; 2 b, South Dobrogea sub-province ; 2 c, Central Dobrogea sub-province ; 3, Carpathian Crystalline province ; 3 a, East Carpathians sub-province ; 3 b, South Carpathians sub-province ; 3 c, Apuseni Mts sub-province ; 4, North Dobrogea Orogene province ; 5, Balkan province of gabbro-peridotitic complexes ; 5 a, North Danubian sub-province of Paleozoic ultrabasic magmatites ; 6, metallogenetic province associated with the Mesozoic ophiolitic magmatism of the Apuseni Mts ; 6 a, Highiş sub-province ; 6 b, Drocea-Mureş sub-province ; 6 c, Trascău sub-province ; 7, metallogenetic province associated with the banatitic magmatism (Laramian) : 7 a, Apuseni Mts sub-province ; 7 b, Poiana Ruscă sub-province ; 7 c, Banat sub-province ; 8, metallogenetic province of the Carphathian neoeruptive chain ; 8 a, Börzsöny-Zempléný sub-province (in Hungary, continued on the Soviet Union territory) ; 8 b, Vihorlat-Gutin sub-province ; 8 c, Toroiaga-Căliman-Harghita sub-province ; 9, metallogenetic province associated with neomagmatism in the Apuseni Mts.

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Gisements

GEOLOGY OF THE MAIN COAL BASINS IN ROMANIA

BY

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In the Romanian territory, in the Hercynian and Alpine structural setting, there are coal basins of geosynclinal, intermediate, inter- and intramountainous types. The size of deposits and the quality of coal have been determined by the different geological conditions under which developed the humito-genetic provinces that succeeded during five time intervals : Upper Carboniferous, Lower Jurassic, Upper Cretaceous, Oligocene-Lower Miocene and Middle Miocene-Lower Pleistocene (Pl.). One notes the following petrographic types of coal : anthracite, bituminous coal, subbituminous coal (''Hartbraunkohle") and lignite (''Weichbraunkohle").

Geosynclinal Basins

In different geosynclinal units belonging to the southern segment of the Carpathian Orogen, accumulated coal deposits preserved in five basins (Pl.) as follows : *Reşiţa* (1), *Sirinia* (2), *Mehedinţi* (3), *Codlea-Vulcan* (4) and *Rusca Montană* (5). These basins have been extensively investigated (Antonescu, Năstăseanu, 1977 ; Biţoianu, 1973 ; Năstăseanu, 1964, 1978, 1979 ; Năstăseanu et al., 1973, 1981 ; Răileanu, 1953 ; Semaka, 1970, etc.) and it has been concluded that only the first two ones are more important.

Reşiţa Basin

Inside the South Carpathians there are partly superposed coal basins located in the basement of the Hercynian molasse Westphalian-Autunian) and in the Early Lower Jurassic deposits.

Westphalian-Autunian

The Hercynian molasse consists of continental-lacustrian deposits (2000—2500 m thick) and lies unconformably over the pre-Sudetian metamorphic formations. It is partly detached off the metamorphic basement and involved, from west eastwards, in some overthrust units : Reşiţa (upper unit), Dealu Vremii, Lupac and Semenic (lower unit).

The molasse deposits of the first three units include several distinct conformable lithostratigraphic members.

The Doman Beds consist of breccias which contain coarse-grained cobbles of crystalline rocks and are probably of Westphalian C age. They

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represent the base of the molasse deposits in the Reşiţa (Fig. 1 a) and Lupac (Fig. 1 b) units.

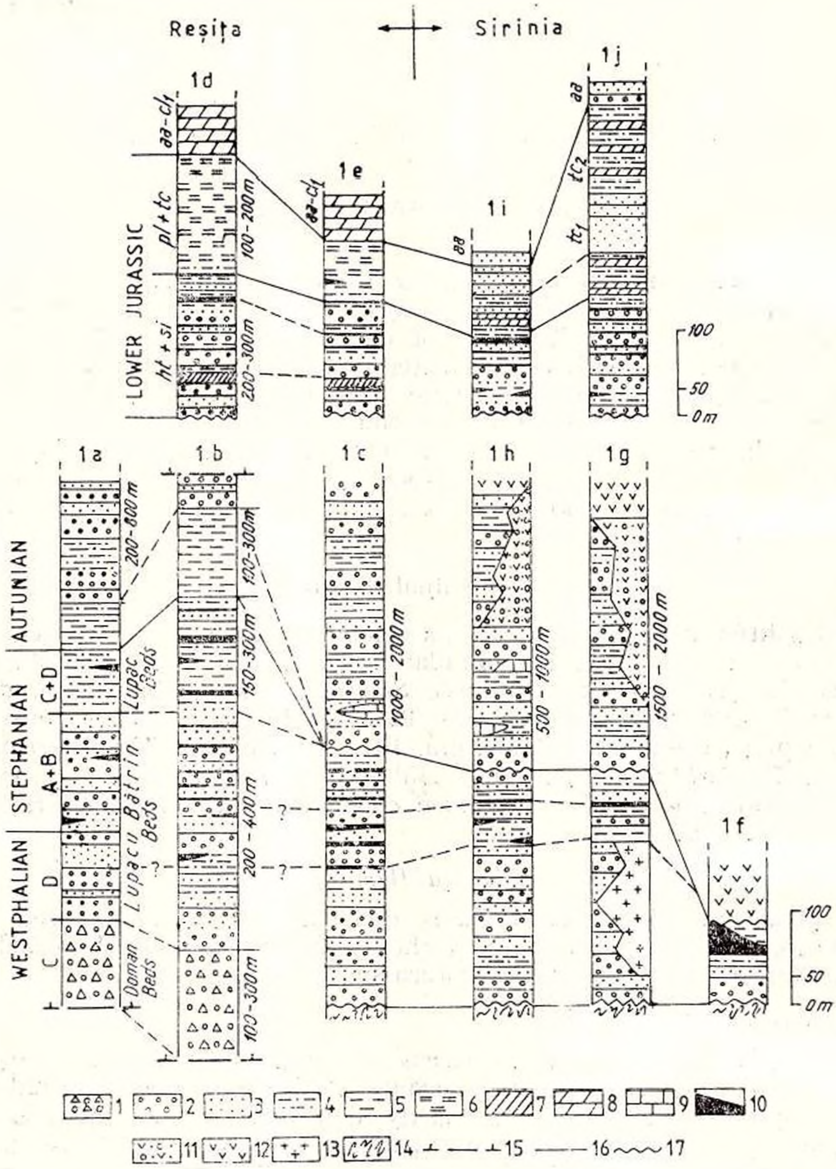

Fig. 1. — Correlation of coal-bearing formations in the Reşiţa and Sirinia basins. 1, breccias ; 2, conglomerates ; 3, sandstones ; 4, siltites ; 5, clays ; 6, bituminous clays ; 7, refractory clays ; 8, marly limestones ; 9, limestones ; 10, coal ; 11, volcano-detrital rocks ; 12, rhyolites and ignimbrites ; 13, basalts and andesites ; 14, metamorphic rocks ; 15, overthrust plane ; 16, geological boundary ; 17, unconformity.

The Lupacu Bătrîn Beds are mainly of sandy-conglomerate nature and consist of at least four energetic bituminous coal layers, 0.30—3 m thick, in the Lupac Unit (Fig. 1 b), as well as of unimportant lenses in the Reşiţa Unit (Fig. 1 a). The paleoflora association (*Neuropteris gigantea, N. ovata, Pecopteris feminaeformis*, etc.) reported from both units points to the Westphalian D and Stephanian A—B ages of these beds.

The Lupac Beds consist of black siltites with frequent ferruginous spherolites and include three energetic bituminous coal layers, 0.30—3 m thick, also reported from the Lupac Unit (Fig. 1 b). The abundant fossil flora (*Annularia stelata, Calamites suckowi, Pecopteris arborescens, Sphenophyllum longifolium*, etc.) accounts for the Stephanian C—D age.

The black shales horizon contains usually *Walchia* and *Callipteris conferta* remains, species of *Florinites* and *Potoneisporites* genera, macro- and microflora typical of the Autunian.

The sandy-conglomerate complex, with shales and red siltite interlayerings, occurs at the top of the molasse deposits. It includes the same macro- and microflora associations as the preceding horizon.

As regards the Semenic Unit, the lithostratigraphic members mentioned above are no longer recognized, while the conformity of the molasse deposits is interrupted by a sedimentation gap corresponding to the Stephanian B—D. In the abandoned deposit at Secu (Fig. 1 c) were exploited four layers of coking bituminous coal, 0.30—2 m thick, belonging to Westphalian D and Stephanian A. The coal layers overlie Lower Westphalian deposits 250 m thick. At Secu, the metamorphic rocks of the coal layers are unconformably overlain by the red sandy-conglomerate complex which contains *Walchia* and *Callipteris conferta*, typical of the Autunian.

Lower Jurassic

The Lower Jurassic rocks (300—450 m) overlie unconformably the Hercynian molasse and the metamorphic formations of the Semenic Unit.

The Hettangian (ht) and the Sinemurian (si) are represented by a psammo-psephitic complex with rare interlayered refractory clays, clays, coaly shales and coal layers. The Anina deposit (Fig. 1 d) consists of eight layers of coking bituminous coal, while the abandoned Doman deposit (Fig. 1 e) contained only three energetic bituminous coal layers ; the thickness of the layers varies between 0.30—3 m. The macro- and microflora associations (*Pterophyllum rigidum, Podozamites mucronatus, Aletopteris dentata, Nilssonia orientalis, Chasmatosporites major, Monosulcites minimus, Cerebropollenites mesozoicum*, etc.) point to the age of the coal complex.

The Pliensbachian (pl) and the Toarcian (tc) are represented by a complex of argillaceous oil shales with thin sferosiderite interlayerings and coal lenses in places. The age of the oil shales has been stated based on a rich palynological association (*Auritulinosporites scanicus, Todisporites major, Foraminisporites jurassicus, Cyathidites australis*, etc) as well as on ammonite remnants (*Grammoceras fallaciosum*). Aalenian (aa)-Lower Callovian (cl_1) marly limestones occur conformably.

The coal-bearing formations in the Reşiţa Basin are characterized by complex structure, as a result of several tectogenetic phases, out of which the Laramian one was the most important ; it is then that the nappes were emplaced and numerous N—S or NNE—SSW trending reverse faults were formed. Post-Laramian tectonics is mainly characterized by vertical and/or subvertical faults.

Sirinia Basin

Outward the South Carpathians, overlying the Inner Danubian Unit, there are the same coal-bearing formations, characterized by slightly different lithofacies as compared to those typical of the Reşiţa Basin.

Westphalian-Autunian

The Hercynian molasse (2000—2500 m) consists of two superposed rock complexes delimited by an unconformity : a lower discontinuous terrigenous complex underlying an upper volcano-sedimentary complex.

The terrigenous complex overlies unconformably the pre-Sudetian metamorphic formations. It is built up of conglomerates, sandstones, siltites, coal shales and coal (energetic bituminous coal). To sedimentary rocks igneous rocks (basalts and andesites) associate in places.

At Baia Nouă (Fig. 1 f), in a small tectonized syncline, was mined a coal layer (1—40 m thick) at the base of the terrigenous complex and directly overlain by Autunian rhyolites. The paleoflora association (*Neuropteris gigantea, Mariopteris sauveuri, Sphenophyllum cuneifolium,* etc.) points to the Westphalian C and D age of the coal layers in the Baia Nouă exhausted deposit.

In the Cameniţa (Fig. 1 g) and Dragosela (Fig. 1 h) valleys there is another lens-like coal layer (0.30—3 m thick), located at the top of the complex and including a macroflora association, similar to that one reported from Secu, typical of Westphalian D and Stephanian A.

The volcano-sedimentary complex exhibits varied lithological features : it consists, at the base, of conglomerates and sandstones with siltite and red clay interbeddings and fresh water limestone lenses in places ; in the middle, terrigenous rocks are less numerous and soon replaced by igneous rocks (tuffs and agglomerates), while at the top, there are ignimbrite and rhyolite flows. Isolated flora debris *(Walchia)* point to the Autunian age of this complex.

Lower Jurassic

Older formations are unconformably overlain by Lower Jurassic formations represented by two facies : a well represented detrital facies and a partly carbonate facies, more limited and deprived of coal.

The detrital facies (500—700 m) includes three lithostratigraphic complexes assigned to Hettangian-Sinemurian, Pliensbachian and Toarcian.

The Hettangian and the Sinemurian consist of sandy-conglomerates coarse grained at the base and fine grained at the top, with frequent interbedded coal shales and coal layers.

The Cozla deposit (Fig. 1 i) includes three layers of coking bituminous coal, 0.30—5 m thick, while the Bigăr deposit (Fig. 1 j) includes five layers of energetic bituminous coal, 0.30—2 m thick.

The age of these coal deposits has been stated according to the paleoflora associations similar to the ones reported from the Reşiţa Basin.

The Pliensbachian is represented mainly by argillaceous-siltic rocks which include marine fauna *(Gryphaea cymbrium, Aequipecten aequivalvis*, etc.).

The Toarcian is sandy at the base (Lespezi Beds) including ammonites *(Dactylioceras semicelatum)* and argillaceous-siltic at the top (Zamoniţa Beds). It is conformably overlain by sandy rocks (Moşnic Beds) of Aalenian age.

Its structure consolidated during the same main tectogenetic phases (Pfaltzic, Austrian and Laramian) as the Reşiţa Basin. This is characterized by numerous reverse faults, along which the western compartments overthrust the eastern ones. The coal deposits are also affected by post-Laramian vertical and/or sub-vertical faults.

Intermediate Basins

Outward the Carpathian Orogen (Pl.) it is to note the Dacic (6) and Moldavian (7) basins. Their basements include both outer orogen units and inner sectors of neighbouring platforms (Moesian, Scythian, East European).

The Neogene formations in these basins include lignite deposits, but only the Dacic Basin is worth studying.

Dacic Basin

The area bordered by the Carpathians, the Balkans and the Măcin Mts is the Neogene sedimentation area of the Dacic Basin. There are numerous studies regarding this basin, but in view of a concise account only syntheses have been used (Andreescu, 1972 ; Marinescu et al., 1981 ; Motaş et al., 1974 ; Pană et al., 1981, etc.).

The Neogene molasse (5000—15000 m) consists of marine deposits with evaporites, at the base, and limnic deposits with coal, at the top.

In the Romanian territory this basin covers two major units : 1. the Carpathian Foredeep, including the Wallachian Subunit, between Trotuş and Argeşel valleys, and the Getic Subunit, west of Argeşel Valley ; 2. the Moesian Platform, including the northern Danubian sector and the southern Dobrogea.

The coal-bearing formations are of Meotian, Pontian, Dacian, Romanian and Lower Pleistocene age. They show great lithofacies variations, depending on the underlying tectonic unit. The thickness of coal deposits is reduced (200—500 m) in the Moesian Platform and increases gradually, from south northwards, to thousands of meters (5000—8000 m) in some areas of the foredeep.

The Meotian starts with marls or sandstones including *Congeria*, *Psilunio*, *Teodoxus*, etc., then a *Dosinia* layer overlain by sands con-

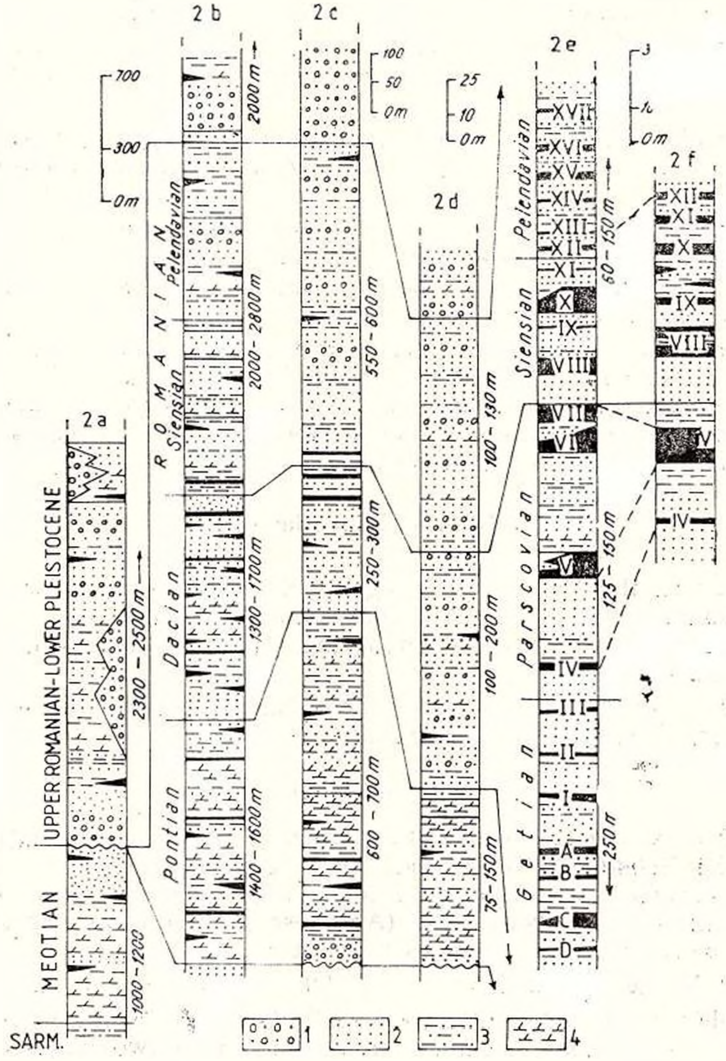

Fig. 2. — Correlation of coal-bearing formations in the Dacic Basin.
1, gravels ; 2, sands and sandstones ; 3, sandy clays ; 4, marls.

taining different species of the genera *Psilunio* and *Congeria*. In the Trotuș, Valley Basin (Fig. 2 a) the Meotian rocks contain unimportant coal lenses.

The Pontian is mainly represented by marls abunding in limnocardiid, dreissenid and viviparid faunas which point to the Odessian, Portaferrian and Bosphorian. The Pontian formations include several thin and lens-like coal layers ; 1—3 interlayerings, 0.10—2 m thick, have been reported at Rîmnicul Sărat (Fig. 2 b), Valea Dîmbovița-Valea Argeșel (Fig. 2 c) in the Wallachian Subunit and at Schitu Golești (Fig. 2 d) in the Getic Unit.

The Dacian consists of mainly sandy rocks abunding in mollusca (lymnocardiids, dreissenids, unionids, viviparids, etc.). Its base (Getian) is defined by the *Pachydacna* beds, while its top (Parscovian) is represented by the *Psilodon* beds.

The Romanian is represented by clays, sands and gravels. Its base (Siensian) includes the beds with *Unio sturdzae, Potomida saratae* and *Viviparus bifarcinatus ;* the middle part abunds in sculpted unionid genera and subgenera *(Rugunio, Rytia, Cuneopsidea, Pristinunio, Psilunio,* etc.), while its top is represented by the Cîndești Beds (gravels) also present at the Lower Pleistocene level.

The Dacian and Romanian formations include ca 21 coal layers, 0.10—9 m thick, the most important belonging to the Parscovian (layers IV—VII) and the Siensian (layers VIII—X) of the Getic Subunit (e.g. Motru, Fig. 2 e ; Rovinari, Fig. 2 f). In the Wallachian Subunit, the coal layers abund in places, less than 0.50 m thick (e.g. Rîmnicu Sărat, Fig. 2 b), or are rather scarce, 0.50—1.50 m thick (e.g. Valea Dîmbovița-Valea Argeșel, Fig. 2 c). On the Moesian Platform the coal layers are few, 0.10—3 m thick and occur mainly in the Dacian formations.

The coal deposits in the Wallachian Subunit are obviously influenced by salt tectonics. Salt pierces the whole cover of Neogene rocks, in places, and generates diapir folds including almost vertical coal layers. These folds are normalized westwards, in the Getic Subunit, exhibiting slightly dipping flanks, or disappear completely in the Moesian Platform, where the Dacian formations include two big elevation areas and two sinking areas due to the positive and negative movements of the pre-Neogene tectonic compartments.

Inter- and Intramountainous Basins

The post-Laramian tectonic depressions which divided the Carpathian belt included several basins (Pl.) : *Pannonian* (8), *Petroșani* (9), *Almaș-Agrij* (10), *Transylvanian* (11), *Hațeg* (12), *Țara Bîrsei* (13), *Borsec-Bilbor* (14) *and Comănești* (15) basins. The first two are most important and will be described in the following lines by considering the latest studies (Marinescu et al., 1981 ; Moisescu et al., 1979).

Pannonian Basin

The Pannonian Basin is located among the Carpathians, the Alps and the Dinarids and only a reduced area lies in the Romanian territory.

The coal-bearing formations are Badenian, Sarmatian, Pannonian s. str. and Pontian in age. Their thickness is of 2000—2500 m in the lowered tectonic compartment and much less in the elevated ones.

The Badenian starts with coarse-grained continental formations (breccias, gravels, sands) ; then follows a paralic facies (marls, clays, limestones, tuffites) which abunds in fossil specimens (*Pycnodonta, Corbula, Cardita, Arca, Cyprinide*, etc.). Extended coal interlayerings characterize the Langhian, such as : the Ţebea-Brad deposit includes three subbituminous coal layers, 0.20—3.60 m thick ; the Bozovici deposit consists of 11 layers, 0.20—3 m thick and the Caransebeş deposit (Fig. 3 a) contains three lignite layers, 1—4 m thick.

The Sarmatian is frequently represented by detrital and carbonate rocks, rich in fossils (*Ervilia, Abra, Pirenella*, etc.), as well as a clayey-marly facies with coal, such as in the Borod gulf. The Borod deposit (Fig. 3 b) consists of two lens-like lignite interlayerings, 1—2 m thick.

The Pannonian s. str. (Malvensian) consists of clayey-sandy rocks with fossils (*Origoceras fuchsi, Congeria banatica*, etc.), is well developed all over the basin and includes only minor lignite lenses.

The Pontian formation stands out as the most important as regards both its great thickness (100—1500 m) and the large number of lignite interlayerings, belonging to the Portaferrian especially. It consists of gravels, sands (bituminous in places), sandstones, tuffites and clays abunding in congerias, lymnocardiids and ostracods. The best known coal accumulations occur in the following regions : Şimleu Silvaniei (Fig. 3 c), containing 27 coal layers, out of which 19 are 0.50—3 m thick ; Beiuş, containing two lens-like layers, 1.5—2.50 m thick ; and Lugoj (Fig. 3 d), with ten layers, 0.50—3.60 m thick.

The coal deposits exhibit a simple structure, the Neogene rocks forming slightly dipping wide anticlines and synclines.

Petroşani Basin

This basin lies in the South Carpathians and consists of molasse deposits in continental-lacustrian facies, accumulated during the Oligocene-Lower Miocene interval. It contains the most important bituminous coal (partly coking) deposits in the country.

The coal-bearing formations are 1800—2000 m thick and belong to the Chattian and Aquitanian which succeed conformably (Fig. 3 e).

The Chattian consists of : 1) basal conglomerate horizon, unconformably overlying the older formations ; 2) middle sandy-argillaceous horizon, including 18—21 coal layers, 0.50—30 m thick, characterized by a rich paleontological content (*Corbula giba, C. carinata, Gobraeus protractus, Glyptostrobus europaeus*, etc.) and 3) upper sandy horizon deprived of coal.

The Aquitanian is represented by a lower argillaceous-sandy horizon including *Crassostrea, Tellina, Mytilus aquitanicus*, etc., with five to eight lens-like coal layers, 0.20—0.80 m thick, and an upper sandy horizon with marine fauna (*Chlamys gigas*, etc.).

The Petroşani deposit exhibits a simple structure, characterized by several ENE—WSW trending anticlines and synclines, affected by vertical or subvertical faults.

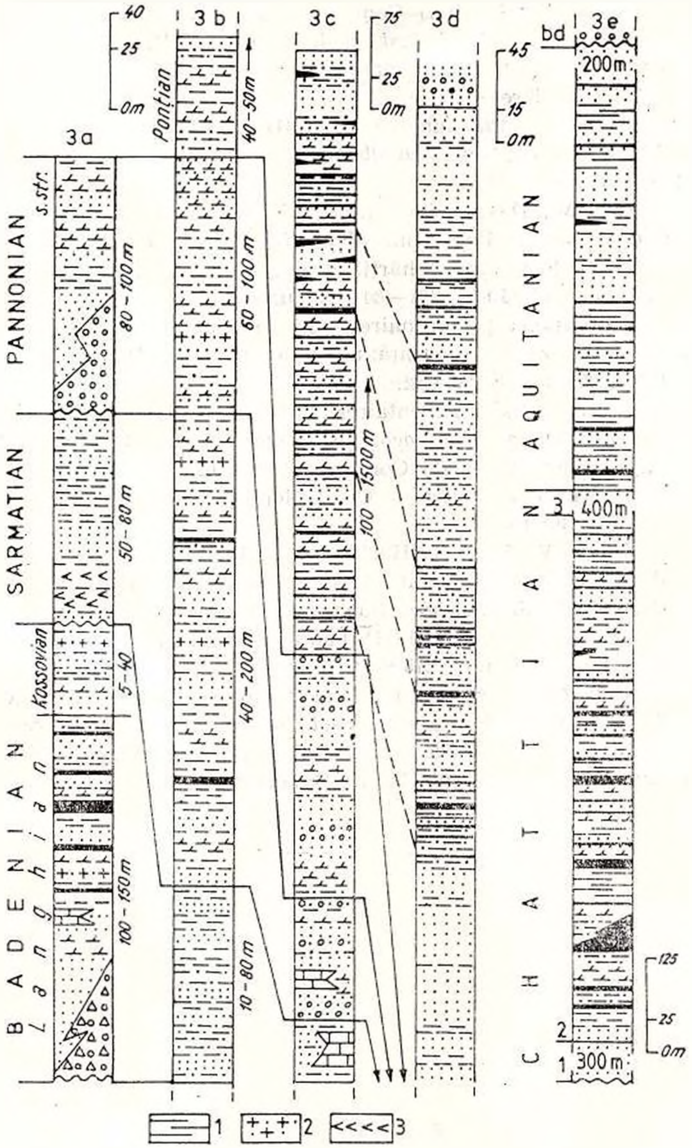

Fig. 3. — Correlation of coal-bearing formations in the Pannonian Basin and synthetic lithostratigraphic column of the Petroşani deposits.
1, oil shales ; 2, tuffite ; 3, gypsum.

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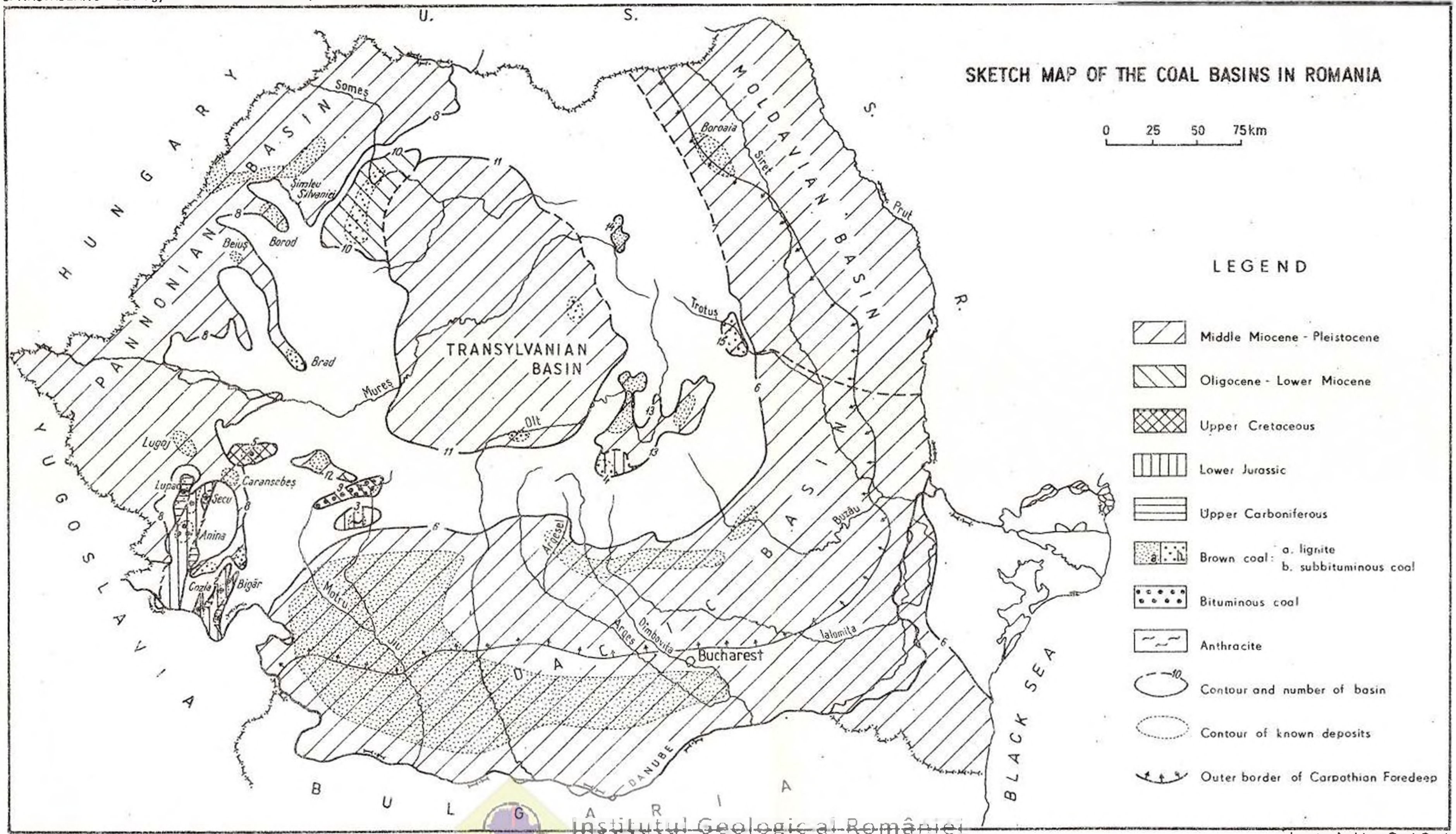

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THE EVOLUTION OF THE HYDROCARBON FIELD DISTRIBUTION IN THE MOESIAN PLATFORM

BY

DUMITRU PARASCHIV[1]

Stratigraphically, the Moesian Platform basement, consolidated at the end of Proterozoic, underlies a more or less complete sedimentary sequence, the age of which ranges from the Cambrian (?) to the Quaternary. The sedimentary cover of the platform may reach thicknesses of 10 000 m, consisting of an alternation of terrigenous and carbonatic formations, most of them of marine origin.

The prospection and exploration activity, performed up to now within the platform, allowed the discovery of about 140 oil and gas structures. Hydrocarbons are stored in the Devonian, Permo-Triassic, Dogger, Malm-Lower Cretaceous, Senonian, Badenian-Sarmatian, Meotian, Pontian and even Dacian formations.

The major idea which backed this hydrocarbon prospection and exploration activity was that the Moesian Domain represents a quiet and almost continuous accumulating area, beginning with the Cambrian. The sedimentary character changed over long periods, namely the large alternance of terrigenous and carbonatic sequences, the formation or reactivation of some ruptural accidents and the gradual fading of the positive and negative forms inherited from the basement level, were the main events which occurred in the quiet and unitary evolution of this domain. The geological and geophysical data gathered during a first stage of researches pointed out a pattern according to which the general sedimentary deposit arrangement is of a faulted monocline falling in steps in front of the Carpathians. Some geologists have extended this image to the whole sedimentary sequence, while others have fully accepted it, at least from the Jurassic. As far as the hydrocarbon pool formation and distribution is concerned, it was generally (Paraschiv, 1979) and even exclusively (Pătruţ et al., 1983) agreed that the oil and gas were generated and entraped during the Sarmatian-

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Pliocene subsidence and as such the huge monocline sectioned by ruptural accidents controlled the accumulation distribution (Fig. 1).

In time, the economic potential of this geological conception began to decrease. At the same time, a vast volume of data accumulated and, together with the progress of fundamental researches and the "nonconformist" distribution of some pools, made possible new ideas and interpretations, to be applied in prospection and exploration designing.

Within the new conception, the history of the Moesian Platform comes out in a much more dynamic development, very complex and non-unitary. When outlining this image, some essential elements to be referred to later on, have been taken into account.

As regards the profound structural aspect, the Moesian area is not a unitary one. At least two big crust segments make it up : the western part of the Black Sea Microplate, between the Ianca-Palazu and Belciugate faults, and the Moesian Microplate west of Belciugate Fault. Constituted in their turn of several blocks, those microplates manifested themselves through dynamical and geothermal differentiated regimes which brought about variations in the deposit thickness and the facies distribution and even in the rock diagenesis (the incorporated organic matter included). One may say that this deep structure, proper to the Moesian area, is responsible for the detail variations of the sedimentary cover constituion and of the diagenesis of the organic matter.

The major features, very much as the whole ensemble evolution of the territory under discussion, have to be connected with the relations established and which followed between the two main geodynamic factors of a total importance : the Euro-Asian continental Block — which acted from the east, and the African continental Block which acted from the south. In rhythmical expansion, those geodynamic giants exerted alternatively their influence on the area between the Carpathians and the Balkans. The action of the global factors determined the displaying of the sedimentary basins. the facies zones, the major structural lines, the fluid dynamics and the preferential zones of hydrocarbon accumulations, elements which were lain perpendicularly on the pushing direction of the blocks. At the same time, the displacement of the respective plates started and varied connections between crust segments of all sizes took place within the Moesian area.

The alternative predominance of the pressure exerted by the two big plates makes necessary the separation of several evolution periods in the Moesian Platform history, since during each stage the surfaces of oil interest were oriented and distributed in a different way. During the periods when the forces from the eastern part imposed themselves, the preferential accumulating zones were generally oriented N—S, and when the pressure from the south was predominant, the pools aligned themselves to some E—W tendencies.

Another essential element considered is that the hydrocarbon could be formed and accumulated not only during the Neogene but also during other stages of the platform history, beginning with the Paleozoic and up to the Pliocene. This hypothesis is backed among others also by the fach that the pools discovered on the Bulgarian territory are situated in regions in which the Neogene is absent or is very

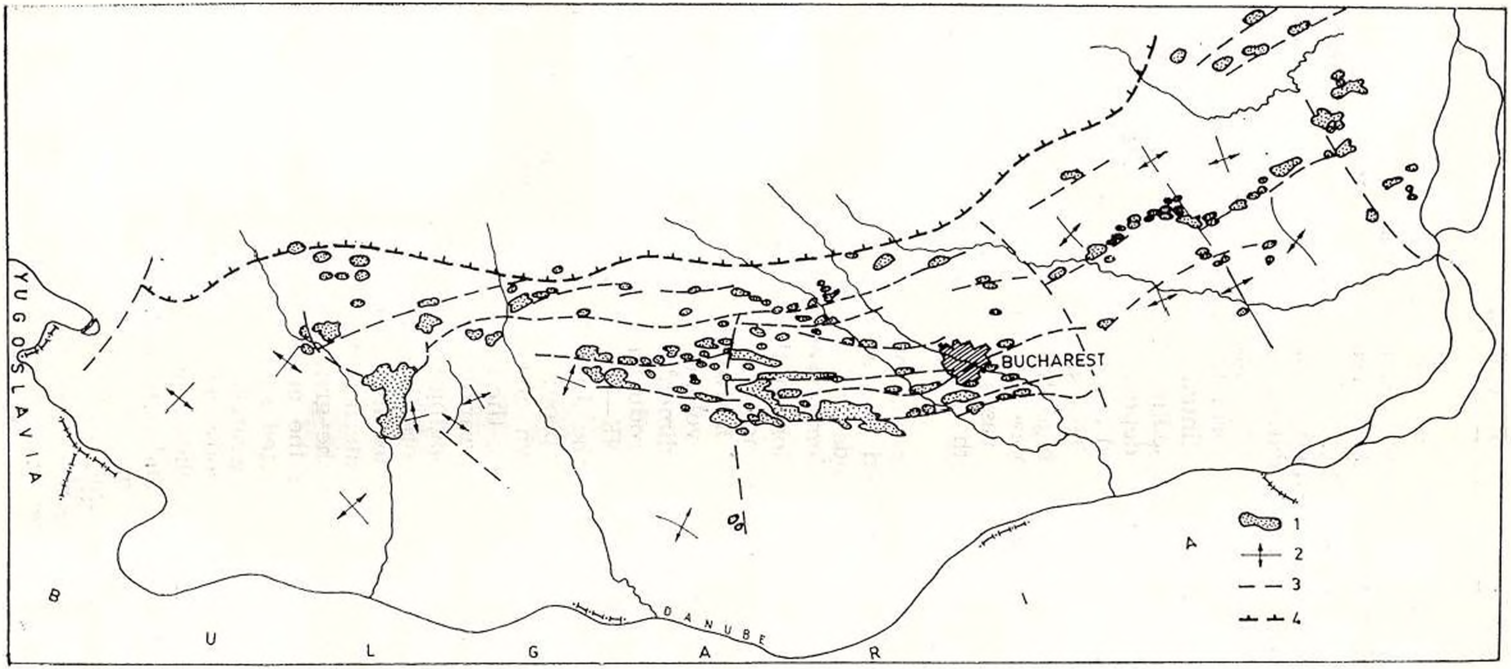

Fig. 1 — Present-day distribution of hydrocarbon pools of the Moesian Platform.
1, hydrocarbon pool ; 2, dome structure ; 3, fault ; 4, pericarpathian line.

weakly developed. Doubtless, within each stage of evolution, fluid remobilizations and partial or total pool redistributions previously formed. were produced. Owing to these reasons, the more recent evolution stages and more particularly the last one, appear more prolifical, but this fact has also to be related to the older formation phases.

The actual structure dating back from the Badenian if not earlier, was completed under the pressure of forces coming from the south which started the platform underpushing process. Consequently, its northern margin sunk continually in front of the Carpathians in a fracture. Concomitantly, the sedimentation was resumed in hyper-subsidient regime. The Neogene deposit isopachs, the facies zones and the major strctural lines (fractures) are displayed E—W, perpendicularly on the pushing force direction. The stratigraphic terms follow from the south to the north very much as the facies evolution. As a result, the organic matter diagenesis more prominent on the sunken northern margin of the platform decreases southwards. The lateral migration sense was the same from north to south, the pools displaying on E—W oriented alignments (Fig. 1).

As it will be seen later on, the Neogene evolution stage of the Moesian Platform determined the re-dynamization of the previously accumulated and formed fluids and their partial re-distribution along the newly created structural elements east-westward.

Another stage of evolution in the Moesian Platform history corresponds to the Cretaceous and Jurassic. During this time the Euro-Asian Plate was much more active in comparison with the African one ; thus the Moesian area developed under the influence of pressures exerted from the east. The sedimentary basins, the isopachs, the facies zones, as well as the main productive alignments seem to have been oriented N—S, NW—SE or NE—SW. Possible to have been noticed from the very beginning of the Jurassic, such tendencies were locally extended up to the Eocene ; proof of it is that the Lom Basin goes on northward close to the town of Craiova, and the Varna Depression is passing beyond the parallel of the locality of Slobozia.

A better studied stratigraphic sequence — seen its economic importance — is that one belonging to the Albian. On close examination of Figure 2, one can see that both the lithofacies evolution and the denudation were controlled by the displacement of the Euro-Asian continental Block. Thus in the western sector of the platform, pelitic pelagic deposits were accumulated ; the gritty-calcareous facies is predominant in the central sector ; while the gritty-siltic rocks are predominant in the eastern part. The denudated surfaces are also oriented N—S. Probably, the hydrocarbons generated by the pelagic sequence migrated laterally in the gritty-calcareous rocks. In fact, the surface of economic interest of the Albian is related to the gritty-calcareous facies zone. In the geological context under review, it is difficult to admit that the productive alignments of the Albian were subject to other initial tendencies than those trending N—S.

It is true that some pools and, first and foremost, those on the Corbii Mari-Petrești structure appear as stretched east-westward, but they are hydrocarbons re-distributed during the Neogene ; proof of

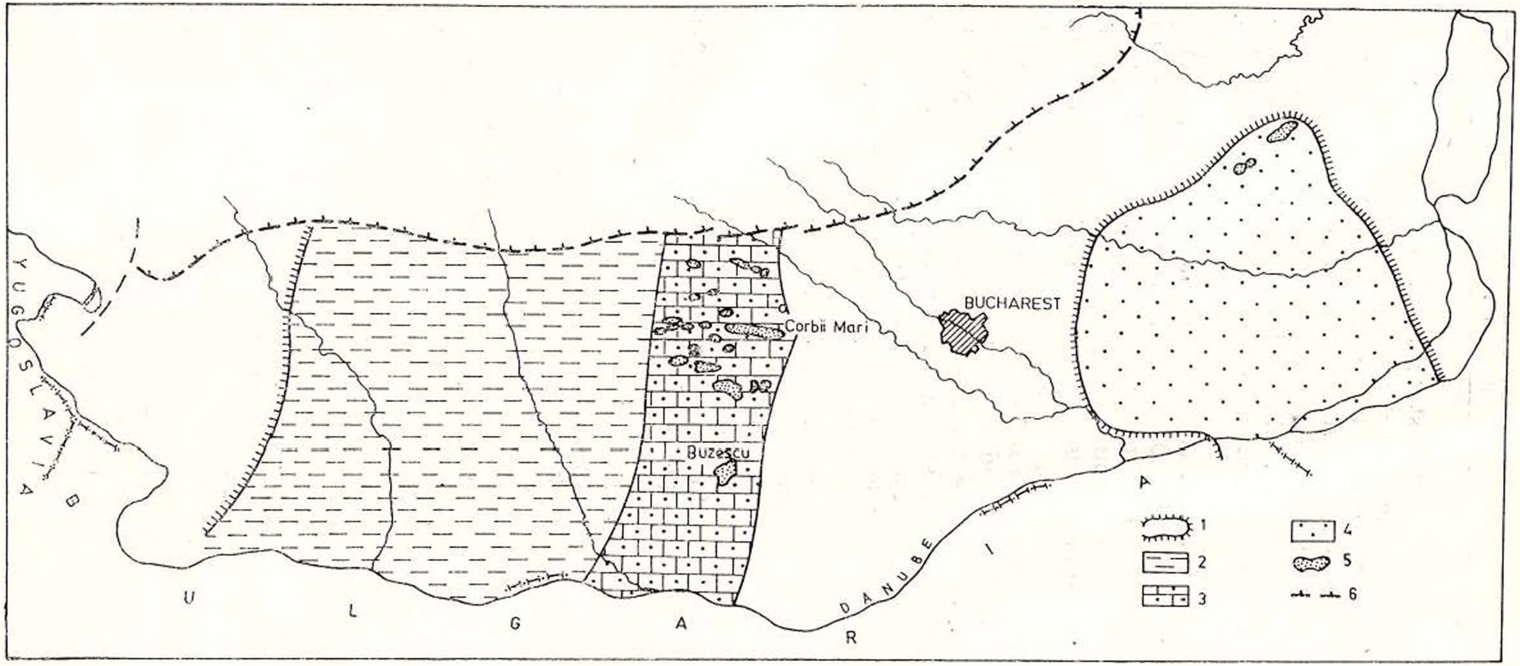

Fig. 2. — Albian lithofacies and hydrocarbon pools located in this formation.
1, line marking the disappearance of the Albian ; 2, Upper Albian in pelagic facies ; 3, Albian in calcareous-gritty facies ; 4, Lower-Middle Albian in gritty-siltic facies ; 5, hydrocarbon pool ; 6, pericarpathian line.

it is that the above-mentioned structure is produced by the Sarmatian as well. But on other structural elements in the gritty-calcareous facies zone, hydrocarbons are to be found only in the Albian. More than that, in the Buzescu-Nenciulești sector, the unproductive Sarmatian is not involved in the tectonics proper to the Albian gas-bearing deposits, a fact suggesting that the first phase of the hydrocarbon accumulations in the respective zone took place before the Miocene, maybe, the end of the Cretaceous.

Similarly, the Cretaceous deposits, older than the Albian were studied ; the same goes for those belonging to the Malm, which, together, make up a plate almost exclusively carbonatic. Figure 3 shows the Barremian-Aptian distribution and facial differentiations of these terms. The respective picture is very much similar to that of the Albian, namely N—S orientation of surfaces covered by Barremian-Aptian deposits as well as the zones affected by denudation. The western sub-basin — more developed and made up of micritic limestones — is divided by a limestone zone with intraclasts and reefs NE—SW oriented. The eastern sub-basin strongly fragmented by erosion is aligned NW—SE together with its appended areas. Hydrocarbon pools, obviously controlled by NW—SE tendencies in the eastern sub-basin, also underwent Neogene ''corrections" more particularly in the Ciurești perimetre.

According to the reference material (Costea et al., 1978, Vinogradov et al., 1978) the rock distribution and the facies evolution of the Lowermost Cretaceous and Malm deposits are very much alike with those of the Mesozoic stratigraphic sequences discussed before.

At the level of the Jurassic terrigeneous sequence (the Balş Formation), the isopachs map (enclosure 4) shows the NW—SE disposition of the main lines and implicitly of the basin depocentre. Hydrocarbon pools are more often encountered on the margin of the basin, corresponding, however, to thicknesses over 100 m. Obviously, subject to some directions parallel to the basin's axis (NW—SE), some of the pools (Ciurești, Oporelu) also bear the mark of the Neogene re-dynamization.

The elements under discussion may state that most of the pools located in the Cretaceous (Senonian, Albian, Neocomian) and Jurassic (Malm, Dogger) are displayed along some zones or alignments oriented N—S, NW—SE or NE—SW in accordance with the sedimentary basin dynamics of the respective periods. One can also see some exceptions, namely E—W orientations suggesting older accumulation re-distribution during the Neogene. There are also some other discrepancies, such as subtle traps (lithological variations of porosity and permiability).

The Eokimmerian diastrophism marks an important change in the evolution of the Moesian Platform, namely posterior to the respective event, the area under discussion developed unitarily, while prior to the Jurassic, the eastern sector (the Black Sea Plate) continually underwent the influences of the Euro-Asian continental Block ; the western sector (the Moesian Microplate) being alternatively subject to the pressure of the two plates. In fact, productive formations older than the Jurassic, i.e. Triassic and Devonian, are known up to now only west of Bucharest meridian.

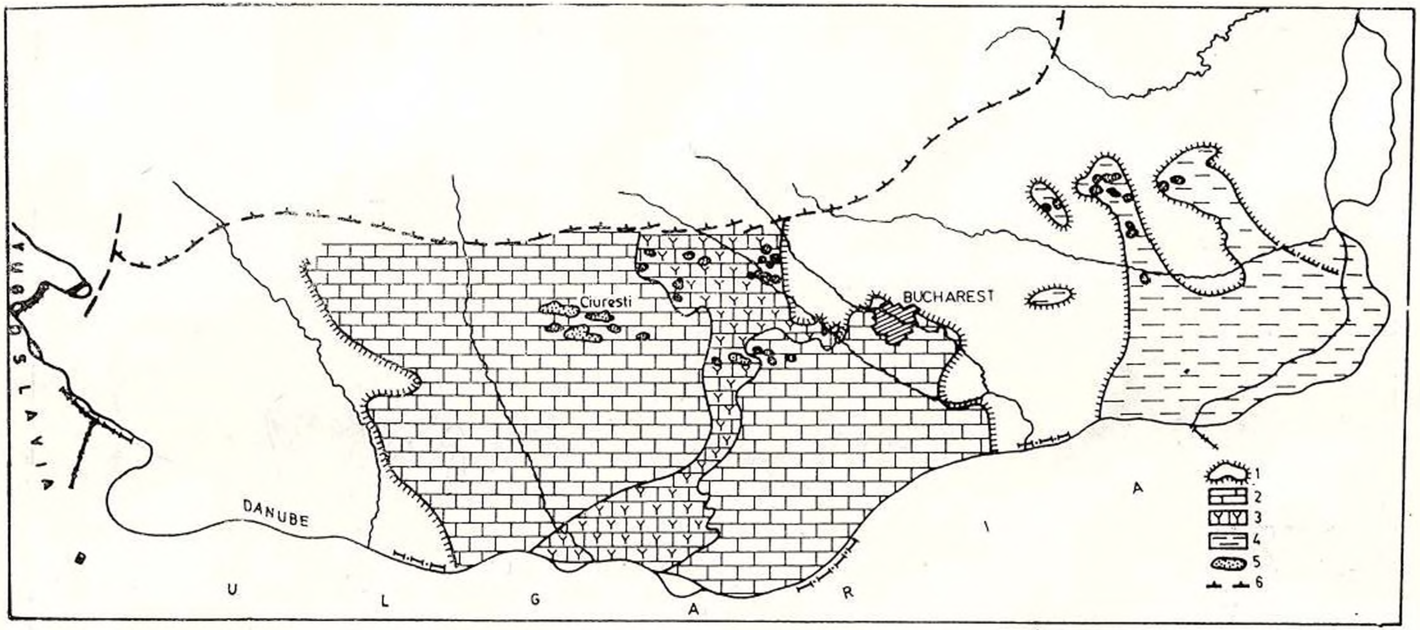

Fig. 3. — Barremian-Aptian lithofacies and hydrocarbon pool distribution located in these formations.
1, line marking the disappearance of Barremian-Aptian deposits ; 2, micritic-calcareous facies ; 3, calcareous facies with intraclasts and reefs ; 4, marine littoral facies with continental sequences ; 5, hydrocarbon pool ; 6, pericarpathian line.

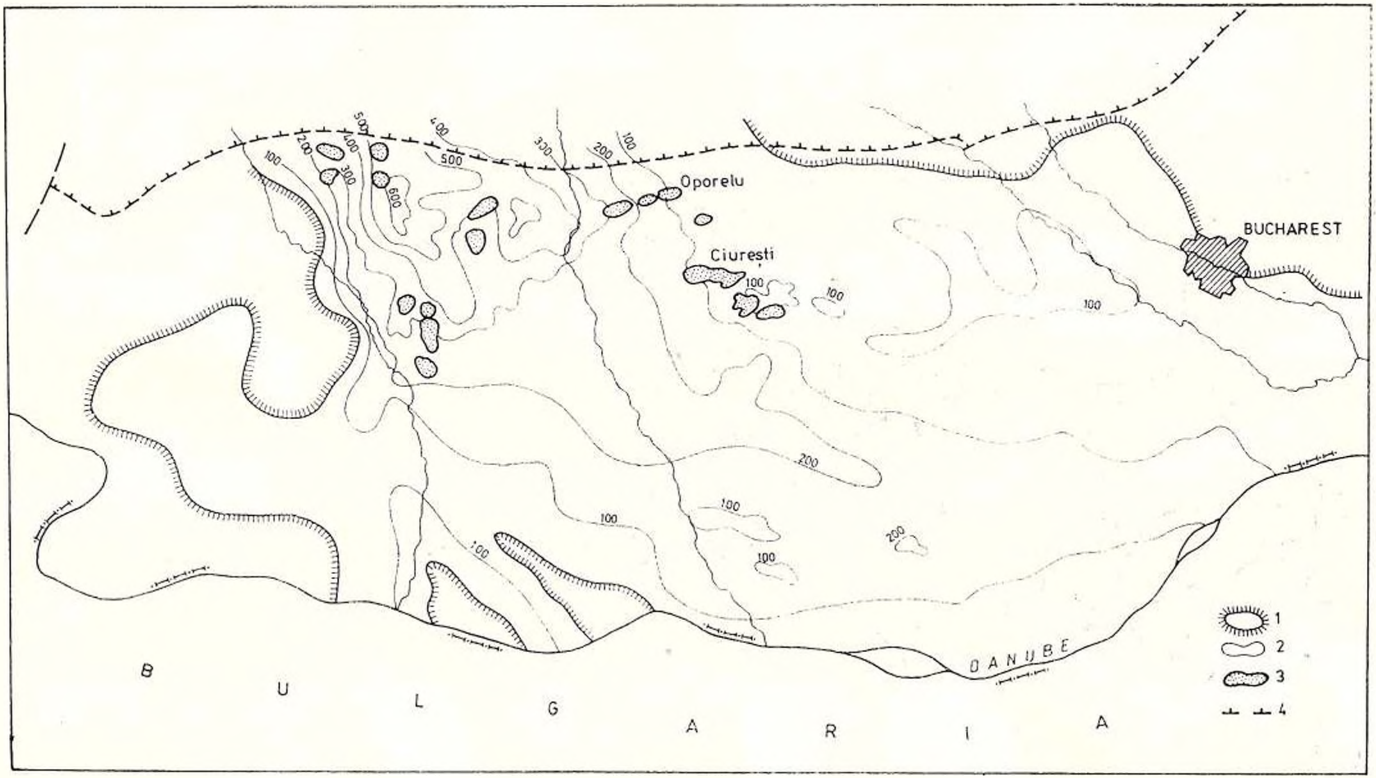

Fig. 4 — Isopach map of the Balș Formation and its hydrocarbon pools.
1, line marking the Balș Formation disappearance ; 2, isopachs ; 3, hydrocarbon pool ; 4, pericarpathian line.

The Triassic deposit distribution, the evolution of these facies and a number of plicative structural elements (E—W oriented) seem to point out that the western half of the Moesian Platform was subject to pressures coming from the south. The hydrocarbon pools, however, have a chaotic spreading. This is probably due to the fluid re-dynamizations and re-distributions following the Triassic. As a matter of fact, most of the accumulations located in the Triassic form common hydrodynamic units with those in the Dogger. On the other hand, the study of the lithostratigraphical terms of the Triassic is still unsufficient and a further action will bring out additional data useful in this respect. However, it is to be expected that the "sealed" pools of the Triassic will follow the E—W tendency.

Only one productive structure was discovered in the Devonian, therefore the reconstruction of the old tendencies cannot benefit by the necessary fundamental elements. One may say, however, that both the Devonian and the Triassic belong to some stages of different evolution. Besides geological arguments, it is to be stressed that the curve indicating the degree of the organic matter transformation shows an important threshold and gradient change at the boundary between the Permo-Triassic and the Devonian. This implies notable changes in the geodynamic and geothermal régime of the platform at the respective moment. The distribution of the hydrocarbon pools seems to be more and more difficult to decipher starting from the present situation to the past. Nevertheless, some tendencies in their distributions are forseeable. It is, however, a fact that the zones and the preferential accumulation alignments have undergone particular changes in the geological past and that the conception regarding their prospection and exploration has to take into account such situations.

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ON THE NATURAL DEGASIFICATION OF THE HYDROCARBON-BEARING DEPOSITS IN ROMANIA

BY

DUMITRU PARASCHIV[1]

The detailed geological mapping (sc. 1 : 25.000) performed in all sedimentary basins of Romania has led to the identification of about 1.000 points with oil and gas natural seepages associated with formation waters, H_2S, CO_2, N_2, etc. (Tonescu, 1953). This large number of occurrences is on the one hand an indication of the petrologenous and petroliferous potential of the respective sedimentary basins, and on the other hand an image of the proportions and stage reached by the natural depletion process of pools.

The recently drawn synthesis map (Pl. I) shows that hydrocarbon occurrence distribution at the surface is ununiform, both spatially and in point of rock age, where they are to be found.

First and foremost, it would be worth stressing that out of the total of 1,000 points with seepages, 73% represent oil springs associated or not with gas or formation water, and only 27% are gas emanations, mud volcanoes included. Most of the gas occurrences are grouped in the Neogene basin of Transylvania, proved up to now to be exclusive gas-bearing, as well as in the Getic Depression, together with the Pliocene zone of the Carpathian Bend. This means that the gas seepages are associated — with small exceptions — to the deposits of Sarmatian-Pliocene age, more rarely Badenian, while the oil is present in zones where deposits older than the Sarmatian or Badenian, crop out. This remark is apt to bring under discussion the stage of the organic matter diagenesis and hydrocarbon forming, namely that the oil window begins as a rule from the basis of the Sarmatian-Pliocene or even of the Badenian downwards. In other words, both in the Carpathian Flysch and Foredeep, in the internal depressions, the Lower Miocene, Paleogene and Cretaceous deposits have reached the maturation corresponding to the oil generation.

The above-mentioned remark comes to prove generally the geochemical, biostratigraphic and petrographic investigation results, but to

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a certain extent, in disagreement with the hydrocarbon physical state of the undepleted pools, discovered at depth in the afore-mentioned formations. It refers to the fact that in the pre-Badenian reservoirs, there are also free gas accumulations, but particularly associated gas (gas cap). Under these circumstances, the gaseous hydrocarbon penury at surface would be also explained by the advanced stage of the pool depletion in the sense that the gas diffusion phase is almost consumed, the large majority of the accumulations being in the advanced depletion stage, namely the oil leakage or oxidation and even the associated water drainage.

In accordance with what we stated above, it might be admitted that a first major remark, allowed by the analysis of the hydrocarbon occurrences at surface, is the hydrocarbon generating potential and the physical state belonging to some formations, as well as to the pool depletion stage. This is due to the fact that the degasification should not be understood and treated separately, it actually representing a first depletion stage, beginning with the gas diffusion, going on with the oil and associated water drainage, and ending with the oil reservoir and source rock denudation.

Out of the 1,000 seepages, 68% occur at the Neogene level, 26% in the Paleogene and 6% in the Cretaceous. In the deposits older than the Cretaceous, oil springs or gas emanations were not evinced, although numerous pools were discovered at depth in such formations. This situation is due to the denudation which has affected to a greater extent the deposits of recent age, upper situated, and ever less, the sedimentary sequences of older age.

The third and the most important ascertained fact is held out by the occurrence distribution, according to the tectonic peculiarities of facial-structural units of oil interest. If out of the total number of productive structures discovered up to now (about 560), 55% belong to the Carpathian Domain and 45% to the foreland, the seepage average at surface is completely different, namely 98.5% in the folded zones and only 1.5% in the outer Carpathian zone. Therefore, within the Moesian Platform, characterized by a strong disjunctive tectonics, through the existence of about 130 pools (out of which 42 of gas) located in the Devonian-Upper Pliocene stratigraphic interval, only four hydrocarbon seepages were pointed out, and in the North-Dobrogea Promontory, where 11 pools were discovered, hydrocarbons do not occur at surface.

This sharp differentiation has its origin in the tectonic movement complexity and proportion, the crust segment seismicity within the studied areas included ; these movements have led to the deformation, fragmentation, raising and then the denudation of the hydrocarbon prospective sedimentary formations. Such processes took place in the past, but they are still continuing. A proof of it is the recent crust movement map (Cornea et al., 1979). Comparing the hydrocarbon seepage distribution at surface to the neotectonic movements (Pl. I), one may notice that almost all oil and gas occurrences at surface correspond to the zones affected by positive crust movements. Most of such points are registered between lines +1 and +4 mm each year, which means that the main agent of the pool depletion is the denudation, which, in its turn, is stimulated and controlled by the crust movements. Besides, the large majority of oil springs and gas emanations occur in valleys, a fact stressing the distructive role of erosion. But within the valleys, the gravity, namely the land sliding, occurs actively enough. The reference material (Tonescu, 1953) mentions numerous cases of appearance or disappearance of occurrences as a consequence of strata sliding. For instance, at Suslănești (Argeș) gas emanations due to the slope gravity processes were noticed. At Gura Drăgănesii, Păltinișu-Nehoiu and Badila, the mud volcanoes, previously known, have vanished as a result of land sliding.

The depletion of oil and gas accumulations has been proved more active where the region raising was associated with the strata fracturation, which facilitated and accelerated the gas diffusion and oil and formation water drainage. In keeping with the mapping, more than half of hydrocarbon occurrences at surface lie along some disjunctive accidents. Out of them mention should be made of the oil springs and gas emanations along the tectonic line, putting into contact the flysch zone with the Miocene one (Tg. Ocna, Tg. Trotuș), of the seepages close to the Cașin-Bisoca Fault, of the accidents within the flysch zone (Slănic Moldova, where H_2S, CO_2 emanations are added) and of Zîmbroaia, Podeni, Valea Dulce, Ocnița, Lăculețe-Glodeni, Slănic-Prahova, Sările-Valea Rea, Plopeasa, Berca-Beciu-Arbănași, Govora, Măldărești, Pietrari, Cîlnic, Aiud (in Transylvania), etc. At the same time with the hydrocarbons, the moffete gas (CO_2, N_2, Ne, etc.) emanated both in the flysch zone and in the Transylvanian Depression.

The plicative elements — with all their range of shapes — from domes to overlapping folds — accelerated the pool degasification. The depletion processes were more active in the case of the broken folds, namely in the circumstances of plicative and disjunctive element associations. Thus, a large number of occurrences were noticed along some anticlines, as those of Rotilești, Berca-Beciu-Arbănași, the axis of the Homoriciu Spur (Cătiașu) and of the Văleni Spur, of Ursei (Vișinești Fault), Stîrmini, Ciocadia, Pitic, Voitești, Rusi (Transylvania), Beclean-Someș, as well as along numerous salt diapirs, as for instance : Turda, Apostolache, Vulcănești, Buștenari, Doftana, Cîmpina, Telega, Cîmpinița, Slătioarele, etc.

The most important effects on the gas diffusion, generally, and the screen destruction, especially, seem to have been had by the seismicity, namely the earthquakes. One may reach this conclusion by examining the oil and gas occurrence distribution map (Pl. I) and the isoseismal line map and one may find out that the maximum density of hydrocarbon seepages occurs at the Carpathian Bend (Vrancea), where the maximum seismal values (8.9) are also encountered. It would be possible that the earthquakes reactivate the fault network, cause other new disturbances, and accelerate the slope gravity processes, thus facilitating the oil and gas drainage to surface.

One of the classical forms covered by the pool degasification in Romania is represented by mud volcanoes. Most of them were identified in the Neogene Transylvanian Depression, north-west of the Getic Depression, and at the Eastern Carpathian Bend (Pl. I). In Transylvania, the volcanoes are to be found in the gas-bearing dome zones, subject to denudation. In the Getic Depression the respective phenomenon is usually associated to the folded anticlines. At the Carpathian Bend, the volcanism determined by the gas emanation is particularly controlled by disjunctive accidents, among which the Andreieşu and Berca-Arbănaşi faults are noticed. The crater cone, made up of the overflowed and deposited mud, all-around the occurrences, varies between some centimeters and 6—8 cm high, its maximum radius being of 25 mm (at Soroştin in Transylvania).

Worth mentioning is that the quasitotal of mud volcanoes in Romania cropped up on valleys within the major river beds and the river terraces. This means that the fluids (water and mud) brought about by gas, mainly come out of ground water sheets.

The most intensive present-day processes of natural depletion of hydrocarbons take place, as mentioned above, at the Eastern Carpathian Bend, and, more particularly, in the Berca-Arbănaşi anticlinal zone (Pl. II). There, the detailed maps showed a normal and continuous sequence of deposits, beginning with the Meotian, present in both axial peaks of the structure and ending with the Romanian, filling the adjacent synclines. The Meotian is formed of marl, siltstone, sand, calcareous sandstone and, more rarely, limestone alternation, grouped in 27 complexes cumulating 600—800 m in real thickness. The Pontian, also terrigenous, comes out somewhat more pelitic than the Meotian especially in its third lower part. The Pontian thickness is between, 1,100—1.500 m. The Dacian is noticed by a variation of lateral facies, somewhat stronger in comparaison with that found out at the lower term level. In point of lithography, the marls, siltstones, sands and coaly schist interbeddings and coals are predominant, the respective ensemble totalling 500—600 m. The Romanian is made up of sands with clayey interbeddings or schists and coals, marls and tuffaceous sandstones, gravel, 1,500 m thick. These deposits are characterized by crossed bedding, more especially towards the upper arenitic-psephitic part of the sequence.

Structurally, the Berca-Arbănaşi zone appears as an anticline (Pl. II), 30 km in length. It is axially affected by two longitudinal faults, which, in their turn, are crossed and disconnected by other transversal disjunctive accidents. The longitudinal system of faults has a hesitant peculiarity, which determined the inversion of the relations between the anticlinal flanks from one end to the other of the structure. In this way, at Arbănaşi, the eastern flank somewhat raised overlaps the western one, while at Berca the western flank appears higher showing a slight tendency of overlapping the eastern one. The strata dips are of 35—85° and the whole network of faults seems to be tight.

The hydrocarbon accumulations are stored in the Meotian sands. In the southern end of the structure, gas was identified in the Sarmatian too. The number of productive layers differs from one block to another, both on the eastern flank — more prolific — and the western flank. The gritty reservoirs contain oil in most cases, but they are also saturated with free or associated gas.

The existence of numerous faults and the Meotian outcrop (or its raised position) saturated by hydrocarbons have led to the partial and local deterioration of the protecting screens and, consequently, to the pool depletion. Therefore, in the folded axial zone of the anticline, mud volcanoes occur grouped in four sectors called "La Fierbători", "Pîclele Mari", "Pîclele Mici" and "Valca Arbănaşului" (Pl. II). The gas emanations have led to the formation of some cones and craters the heights of which vary between some centimeters and 2.5 m. The gas erupts rhythmically and with variable force, entailing waters, mud, rock fragments and some oil. The volcanism intensity of this type increases during the rainy periods, whence the conclusion that both pool fluids and fresh waters participate in the respective process. The mud overflows the crater cones and by drying it deposits on their flanks, which are gradually raising. The material such sedimented consists, among others, of rock fragments and even clay blocks, sandstones and gypsum, ranging in age from Burdigalian up to Pontian included (Ciocîrdel, 1959). The diversity of rocks, brought about by gas, and their ages suggest different relationships between the hydrocarbon source (Meotian) and the other stratigraphic terms, along the main disjunctive accidents. Generally, it may be admitted (Ciocîrdel, 1959) that gas coming from the Meotian, erodes and conveys the material from the older formation of the overthrusting raised flank. At Pîclele Mari "the block clays" overlie a surface of 1,000/600 m reaching a maximum thickness of about 20 m.

The study of the existing data leads to the conclusion that the degasification and pool natural depletion processes were more intense and even more extended, not long ago. Thus, the presence of some rock blocks, in comparaison with the fine material deposited at present, suggests the substantial diminution of the underground energy in the Berca-Arbănaşi sector. About 30 years ago, at Tîrcov, the mud jet raised 4 m high above the ground (Tonescu, 1953). Meantime, the eruption stopped. At Ocna Sibiului, the greatest cones in the country, 6—8 m height and 24 m radius, were identified. At present, the respective volcanoes are inactive. A similar situation was also registered at Şincai. Part of volcanoes are extinct at Soroştin, Lopatna and Bădila. At Slătioarele, as well as at Lopatna, the volcanoes are sticking in the mud and fossilizing the vegetation. The "Living Fires" or "Unextinguished Fires" of the Carpathian Bend, as well as the continuous gas emanations on the edge of the Lake Băicoi, used as fuel for cooking, were kept up only in the toponymy and reference materials. The natural depletion paroxysm process of the hydrocarbon pools might correspond to the latest Pliocene — early Quaternary when strong positive crust movements took place particularly at the Eastern Carpathian Bend (over 1,000 m high) together with a denudation recrudescence. Later on, the intensity of the respective process decreased at the same time with the diminution of the affected pool energy. Within the previously raised zones (flysch and inner flank of the Carpathian Foredeep), the depletion is in the final stage in the sense that the erosion reached the source rock. In the latter situation, the only indices are still the hydrocarbon smell given off by pelitic rocks (Spiratella marls, Pucioasa Beds, black schists, Sinaia Beds, etc.).

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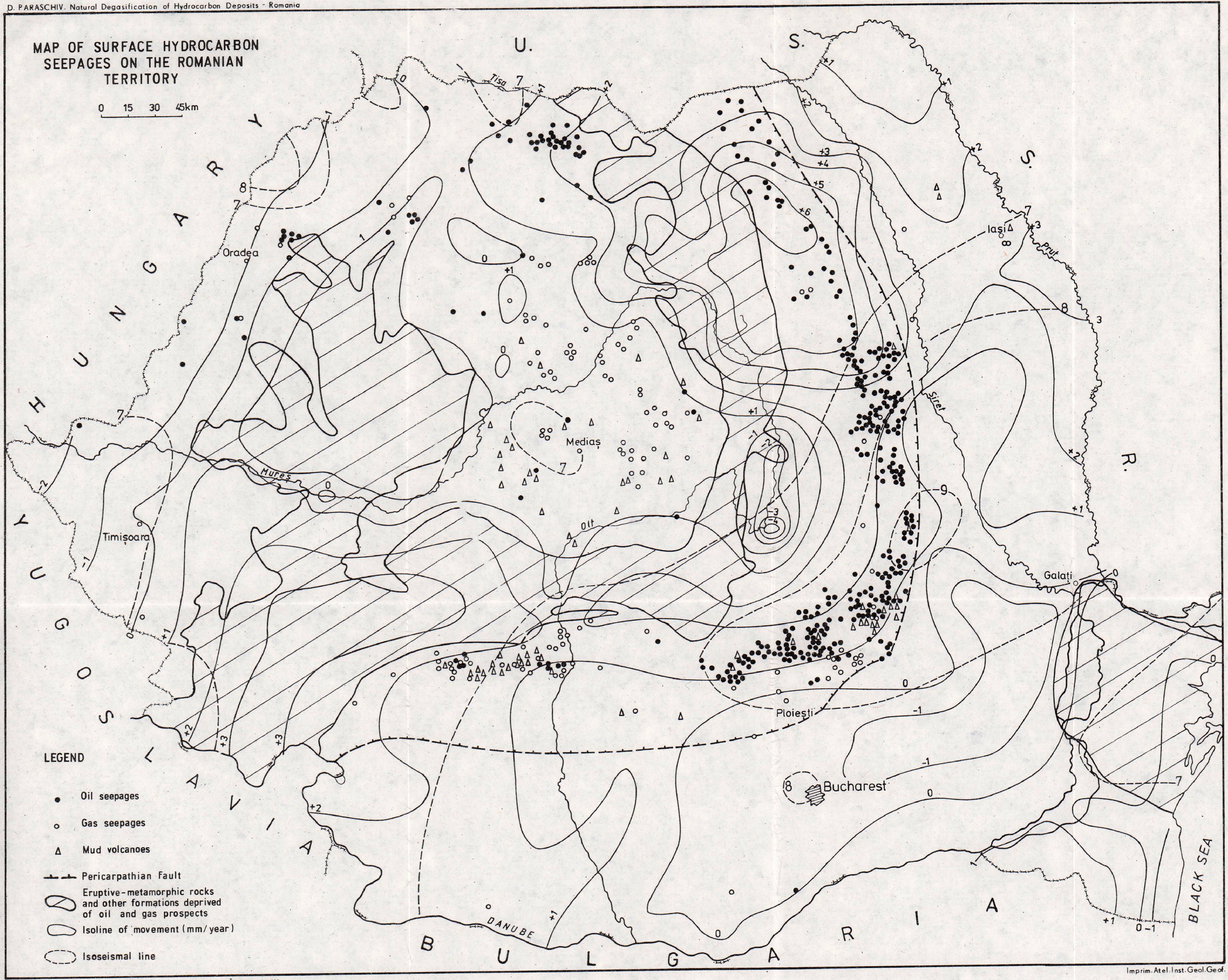
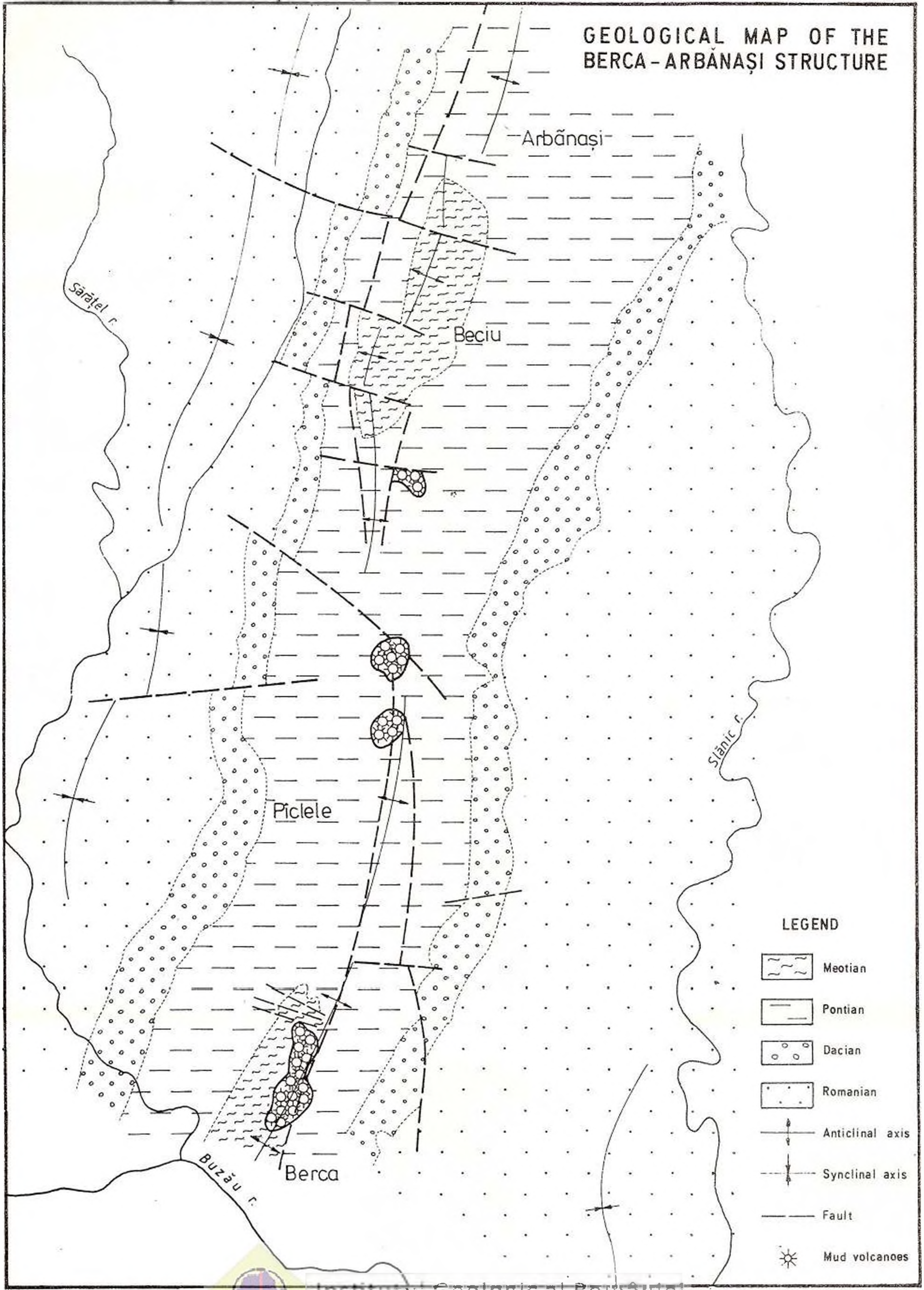


MAP OF SURFACE HYDROCARBON SEEPAGES ON THE ROMANIAN TERRITORY

Imprim. Atel. Inst. Geol. Geof.

GEOLOGICAL MAP OF THE BERCA-ARBĂNAŞI STRUCTURE



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PHYSICAL-GEOLOGICAL MODELS REGARDING THE NEOERUPTIVE ROCKS IN THE BAIA MARE AREA : A CONTRIBUTION TO THE STUDY OF SOME METALLOGENETIC STRUCTURES

BY

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The Oaș-Gutîi segment of the intracarpathian chain consisting of Neogene igneous rocks is characterized by a wide variety of eruptive rocks, generated during a long time interval, starting from Lower Badenian till the end of Pliocene.

The volcanics belong to two formations : andesitic and rhyodacitic, with distinct features and pointing to different magma sources. The andesitic formation consists of a sequence of andesites and dacites, subaerially and subaquatically emplaced, during the Sarmatian-Pliocene interval, which form complex structures including central type volcanoes, arranged on tectono-magmatic lines with mainly mixed activity. The andesitic formation is built up mainly of lava flows, but also of an entire series of pyroclastic rocks, as well as subvolcanic products of intrusive processes. An ever increasing number of subvolcanic bodies has been recently pointed out, more frequently to the east. It is possible that they occur in equally great amount all over the area, but have not been sufficiently marked, so far, below the lava flows. The significant development of intrusive phenomena pointed to the volcano-plutonic nature of the magmatic activity in this area, without any available data about the occurrence of typical plutonics.

The rhyodacite formation consists of tuffs, tuffites, breccias, ignimbrites and subordinately lavas, generated during two main stages : Badenian and Lower Sarmatian. As products of violent explosions, they form piles 20—200 m thick, very widespread, accurately located from stratigraphic point of view and representing real marker horizons.

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The structural images, inferred from the integrate interpretation of geological and geophysical data, prove that the magmatic activity in the Oaş-Gutîi Mts took place on a basement divided into horsts and grabens, which are built up of different rocks : the north-western area, corresponding to the Oaş Mts, which includes the western Dacides and their posttectonic cover in the basement, and the central and eastern area — the eastern Dacides and the Paleogene formations of the transcarpathian flysch. The latter constitute nappes. This division is partly achieved along the E—W trending Bogdan Vodă fracture.

The metallogenetic activity is associated with the andesitic magmatic activity, its products occurring in sedimentary formations and igneous rocks, the latter of Badenian to Pontian age inclusively. At the springs of the Săpînţa Valley, there is a hydrometamorphism aureole also within Upper Pliocene pyroxene andesites (Fig. 1).

The mineralizations are mainly of base metal or gold-silver type and exhibit relatively varied parageneses, usually with frequent occurrences of : galena, blende, pyrite and chalcopyrite associated with small amounts of numerous other minerals containing Pb, Zn, Cu, Ag, Fe, Au, Sb, S, etc. The gangue is represented by quartz, carbonates, clay minerals, gypsum and barytine in places. The characteristic type of ore deposit is the vein ; other types recently reported are the gold stock, the breccia pipe and the impregnations. The mineralization areas are bordered by extended hydrothermal transformation aureoles : argillisations, silicifiations, adularisations, carbonatations, sericitisations, pyritisatios and alunitisations.

Recently, direct and indirect information has been accumulated, pointing to the relationship between the mineralizations and the volcanic structures, to a less extent, and the intrusive processes to a greater extent, revealed only by subvolcanic types so far.

The complexity and great variety of the geological features of the Baia Mare area, due to numerous tectono-magmatic processes, required its study by using geophysical data, mainly gravimetric and magnetometric, correlated with direct geological and petrophysical ones.

The method adopted for the detailed study of some metallogenetic structures was based on the elaboration of mathematical models of simulation of the geological sources, responsible for the registered gravimetric and magnetometric anomalies. Although they refer to local instances, these models have been selected as illustrative of the whole investigated area. The models have been chosen by taking into account the following main criteria : appropriate geological study by means of outcrop data correlated with drilling and mining ones, appropriate study of petrophysical features of rocks and formations — density and magnetic susceptibility —, significant contrasts among the various components of the geological structure and their bi- or tridimensional nature.

Among the sectors which agree with the above mentioned criteria, the following are to be discussed in the present paper : Tarna Mare-Turţ, Herja-Baia Sprie, Cavnic-Roata and Băiuţ. Simplified geological sketches, each accompanied by three sections pointing to the level

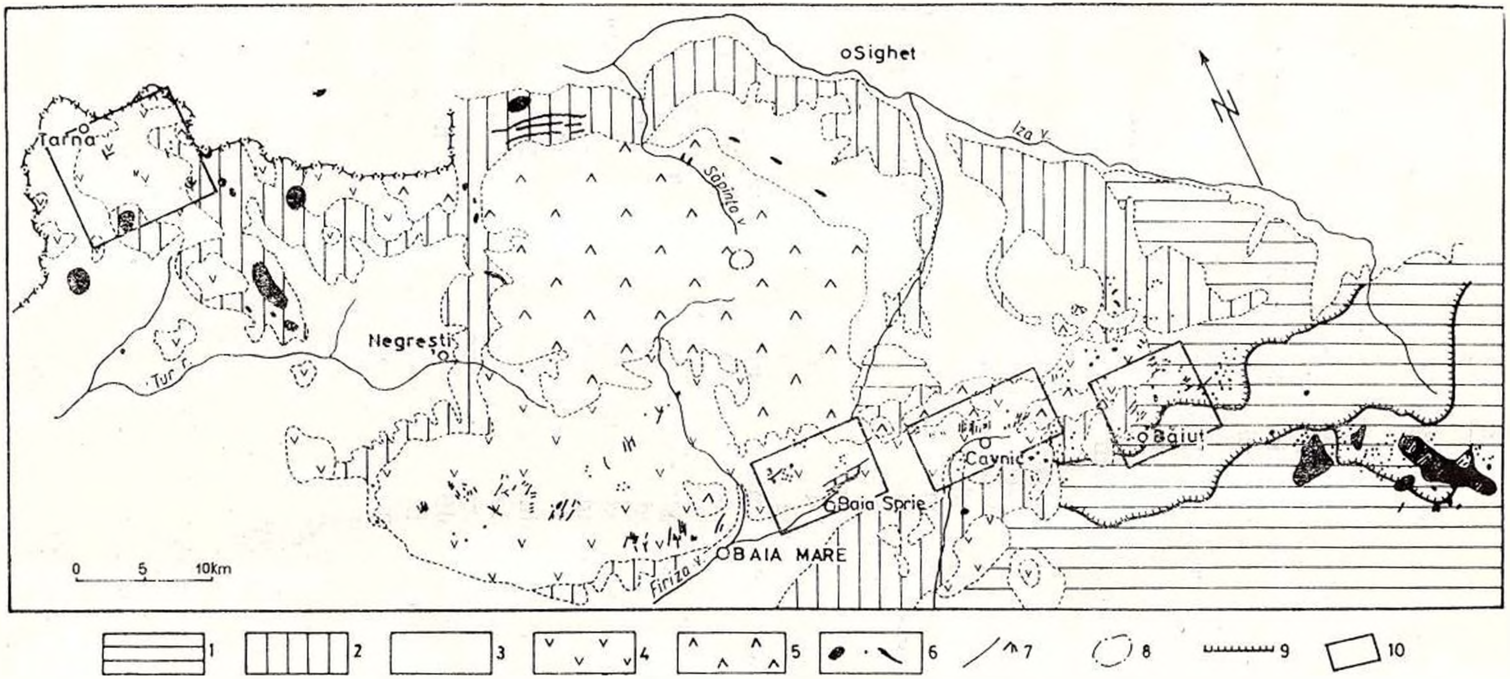


Fig. 1 — Geological sketch of the Oaș-Gutîi Mountains.

1, Paleogene ; 2, Neogene ; 3, Quaternary ; 4, pre-mineralization Neogene volcanics (Badenian-Pontian) ; 5, Pliocene-Upper Pliocene pyroxene andesites ; 6, intrusive bodies ; 7, hydrothermal veins ; 8, hydrometamorphic areas within Upper Pliocene pyroxene andesites ; 9, overthrust ; 10, contour of study areas.

of geological knowledge at different times, have been drawn up for each area [4].

The geological sources plotted on the present sections underwent a physical-mathematical modelling — by means of Talwani's algorithm, the bidimensional variant — in view of achieving an integrative harmony among the geological setting, the morphostructure and the intensity of the measured geophysical anomalies. It is to mention that the anomalies used were taken over from detailed geophysical maps drawn out for these areas. The simulation of geological sources started by according the average density and magnetic susceptibility values with certain domains on the geological sections. These physical domains have been delimited by simplifying the geometry of the considered geological sources.

The physical-mathematical modelling consisted in the computing of the gravimetric and the magnetometric effects, in several iterations, for almost all the analysed sections and have been retained only the variants which showed an adequate concordance between the measured — and/or processed — geophysical anomalies and the calculated effects. Thus, it was necessary, following the case, to conceive other geological sources — physical domains — as well, which correspond to the local structural type.

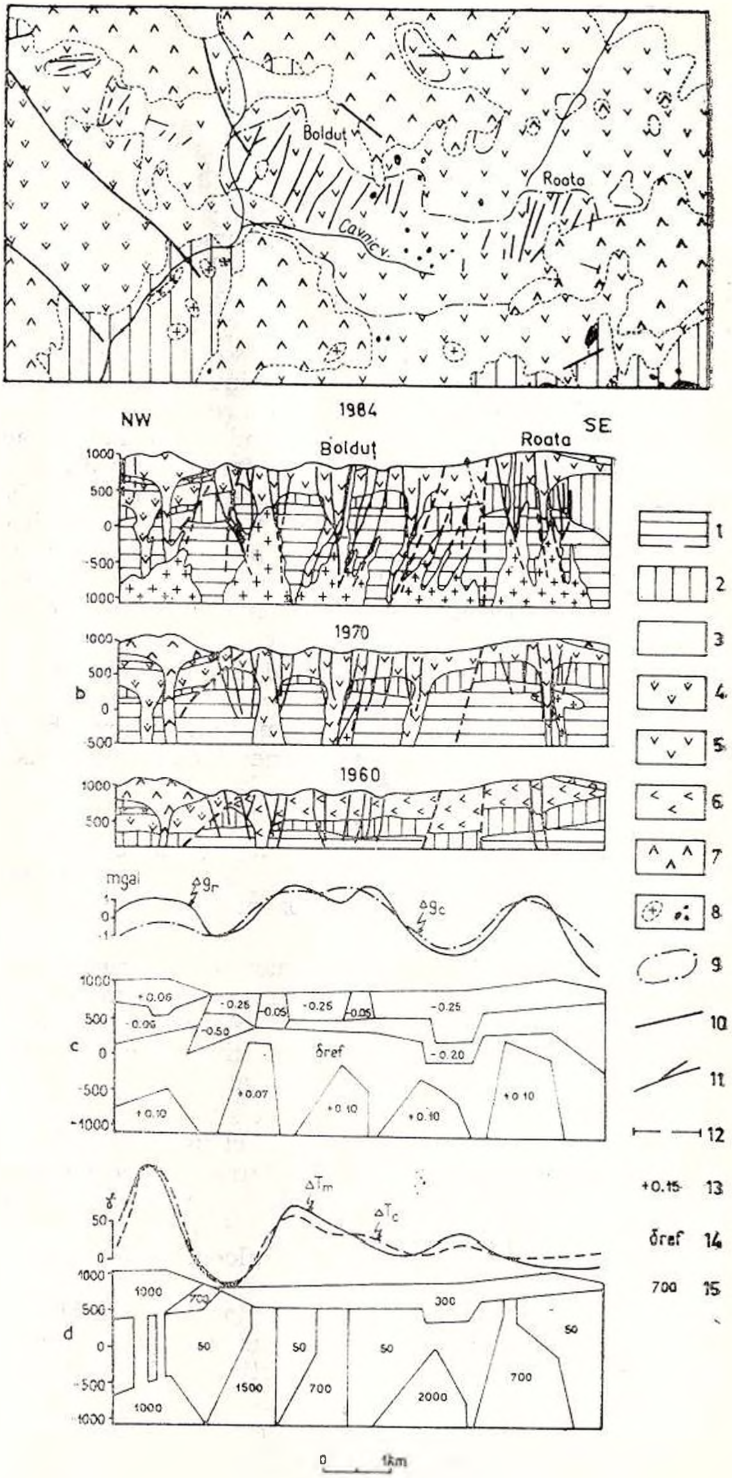
Cavnic sector (Fig. 2 a), situated in the east of the Gutii Mts, exhibits a wide outcropping area of the volcanoclastic complex, which overlies the volcano-sedimentary formation and mainly the sedimentary rocks of Upper Miocene age, Pontian included. The oldest formations which occur at depth — reported by mining works and drillings — belong to the Paleogene transcarpathian flysch. The whole complex mentioned above is pierced by intrusive andesites, microdiorites and microgranodiorites, in one case only. The numerous mineralizations are exclusively of vein type and occur vertically. The vein system is localized both in eruptive and sedimentary rocks of different age. The mineralization consists mainly of lead, zinc, iron and copper sulfides, subordinate sulfosalts, while the gangue consists of quartz, calcite, rhodocrosite, rhodonite, baryte and gypsum ; it is worth mentioning the high frequency of wolfram and molybdenum as minor elements, and even of molybdenite in a recent occurrence. It it to note that the sedimentary formations, mainly those overlain by volcanoclastics, exhibit intense thermic transformations, while the igneous rocks underwent hydrothermal alterations such as argillisation, silicifiation, chloritisation, alunitisation, etc. According to latest data — geological, geophysical, mining and drilling works — the vein-like mineralizations are controlled by subvolcanic processes posterior to the generation of igneous eruptive rocks.

Due to the great amount of detailed geological, petrophysical and geophysical information (Fig. 2 b) on this area, plausible and well founded models of gravimetric and magnetometric anomaly sources have been achieved.

The gravimetric model (Fig. 2 c) accounts adequately for the geological meaning of local gravimetric anomalies and offers additional information on hidden geological structures. The characteristic feature

Fig. 2 — Geological sketch, geological sections and geophysical models in the Cavnic area.

a, geological sketch ; b, geological sections drawn up according to the knowledge of 1960, 1970 and 1984 ; c, gravimetric model : Δ g_r — residual gravimetric anomaly ; Δ g_c — calculated gravimetric effect ; d, magnetometric model : Δ T_m — measured magnetometric anomaly ; Δ T_c — calculated magnetometric effect. 1, Paleogene flysch ; 2, Neogene sedimentary rocks ; 3, Quaternary ; 4, quartz andesites ; 5, pyroxene andesites and pyroxene + amphibole andesites, pyroclastics, volcano-sedimentary sequences (Pannonian-Pontian) ; 6, Sarmatian Seini pyroxene andesites (1960) ; 7, Pliocene-Upper Pliocene pyroxene andesites ; 8, andesite-diorite instrusions ; 9, hydrometamorphic areas ; 10, faults ; 11, hydrothermal mineralizations ; 12, position of geological section ; 13, density contrast ; 14, reference density ; 15, magnetic susceptibility value.



of this model is the occurrence of subvolcanic bodies, at different levels of depth, within a wide aureole of thermal transformations in adjoining sedimentary formations. This is particularly mirrored by maximum gravimetric anomalies, controlled in fact by the achieved conformity with the calculated theoretic effect.

The hypothesis inferred, some other time, from gravimetric data and regarding the presence of a lowered compartment between Bolduț and Roata structures, was also certified by this model. This structural solution, partly accounted for by the recent drilling data, also has a metallogenetic significance — some mineralizations were reported at lower levels, partly without continuing at the surface.

The magnetic model achieved (Fig. 2 d) supplies information on the geometry of structures and physical characteristics of the components of igneous rocks of different nature and defines more accurately the morphology and location of intrusive bodies, as well as the magnetic regime — maximum anomalies of different intensities. The volcanoclastics, characterized by reduced magnetic susceptibilities and occurring next to surface, account for the magnetic quiescence disclosed by the measured magnetic anomaly and proved by the calculated effect.

The predicting value of the adopted structural model, in which the intrusive bodies prevailed, at different levels, consists in the possibility of the presence at depth of an important unitary subvolcanic — plutonic — source of intermediate nature, the incidence of which may include high temperature mineral concentrations, such as : Mo, W, Sn, Co, Ni ; this is also supported by the data reported from the top of the source under discussion.

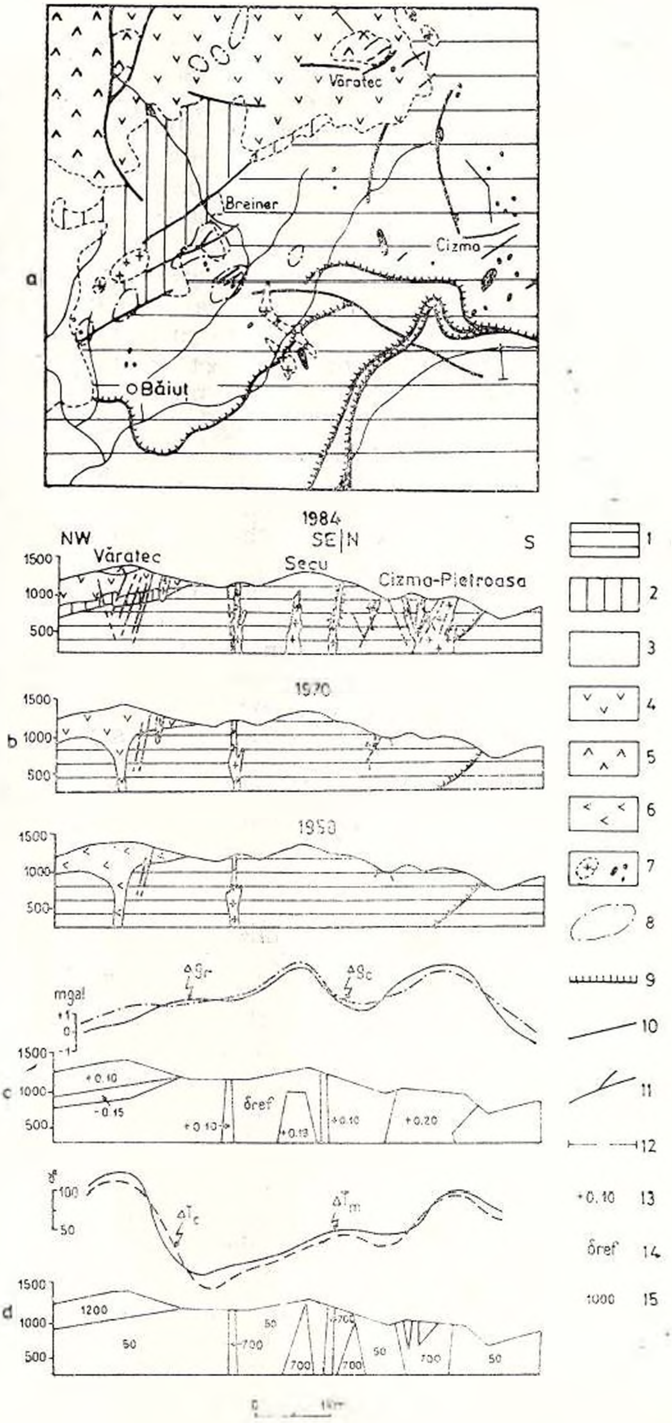
The features of the metallogenetic manifestation of this source may have been determined also by the surroundings of the metamorphic basement, in classical mesozonal facies, rather lifted in neighbouring areas.

Due to the abundant amount of gangue in rhodocrosite and rhodonite, within a mineralization layer of more than 700 m, the Cavnic deposit is an ”exception“ as compared to the other deposits in this region, possibly as a result of the nature of the magmatic source and of specialization of its differentiates.

Băiuț sector (Fig. 3 a) is situated in the easternmost area of the Gutîi Mts, at the boundary of effusive manifestations. Numerous andesite, microdiorite and diorite intrusions do occur. The pyroxene andesite lavas overlie the sedimentary formations belonging to the Neogene molasse (Badenian, Sarmatian, Pannonian and Pontian). Wide areas of this sector are covered by Paleogene sediments which belong to the transcarpathian flysch and exhibit a tectonic style of ”cover nappes“ type (Fig. 3 b). The mineralizations are mainly represented by veins located either in effusive sequences or in sedimentary formations or subvolcanic bodies, the correlation with the latter being more and more obvious. Breccious stocks occur in places.

The gravimetric model (Fig. 3 c) shows that the subsoil of this area is dominated by the positive contrast of density and mass between Paleogene sedimentary formations and intruded eruptive bodies,

Fig. 3 — Geological sketch, geological sections and geophysical models in the Băiuţ area.
a, geological sketch; b, geological sections drawn up according to the knowledge of 1958, 1970 and 1984; c, gravimetric model: Δ g_r — residual gravimetric anomaly; Δ g_c — calculated gravimetric effect; d, magnetometric model: Δ T_m — measured magnetometric anomaly: Δ T_c — calculated magnetometric effect. 1, Paleogene flysch; 2, Neogene sedimentary rocks; 3, Quaternary; 4, pyroxene andesites; pyroxene + amphibole andesites, pyroclastics and volcano-sedimentary sequences; 5, Pliocene pyroxene andesites and pyroxene + amphibole andesites; 6, Sarmatian Seini pyroxene andesites (1958); 7, andesite-diorite intrusions; 8, hydrothermally altered rocks; 9, overthrust and scales; 10, faults; 11, hydrothermal mineralizations; 12, position of geological section; 13, density contrast; 14, reference density; 15, magnetic susceptibility value.



contrast amplified by adjoining thermic contact aureoles. This accounts satisfactorily for the geological meaning of gravimetric maxima, that are in agreement with the calculated effects.

The magnetic model (Fig. 3 d) is apparently easy to discuss, as effusive sequences do not occur in this sector, and only one magnetic susceptibility contrast is present, being related to the contact between the intruded bodies and the adjacent Paleogene sedimentary rocks. Therefore, the maximum magnetic anomalies point accurately to the occurrence known and supposed igneous bodies.

By comparing this structural model to the one of the Cavnic sector, one may infer that in the subsoil of the Băiuţ sector, at depth, there are intrusive bodies characterized by great size and implicitly associated, complex metallogenetic phenomena.

Baia Sprie sector (Fig. 4 a) includes mainly the well known lead, zinc, copper, gold and silver ore deposit, including stibium, arsenic and wolfram minerals, adjacent to the small andesite bodies in the Limpedea Valley in the north. The Baia Sprie deposit, long time considered to be an andesite neck or dyke, the flanks of which hosted the main northern and southern veins, may be described, according to recent geological and geophysical data, as follows : the veins are located on two major fractures longer than 2000 m and between them, to the east, there are some diagonal mineralizations. The two main fractures, approximately trending E—W, mark a compartment including lowered Paleogene formations, filled with Pannonian-Pontian volcanics, as products of a paroxysmal stage prior to this sinking.

The measured gravimetric — residual — effect corresponding to the geological setting of Figure 4 b_2-1984, in agreement with the calculated effect, offers additional information on the significance of the gravimetric minimum (Fig. 4 c_2), due to the deficit of mass, as a result of Paleogene lowering which cannot be entirely compensated by andesite “filling”, independent of the type of hydrothermal alterations. The hydrometamorphic altered andesite ”filling“ is properly outlined by the measured minimum magnetic regime (Fig. 4 d_2) and accordingly by the calculated effect.

Thus, the geological setting presented accounts properly for the significance of gravimetric and magnetometric models.

Given the configuration of gravimetric and magnetometric maxima and the presence of thermic and hydrothermal metamorphism aureoles, the prediction for the geological structure of the Limpedea Valley subsoil may consist in presuming the occurrence at depth of a big igneous mass, which due to its proximity to the Baia Sprie structure may be considered to extend laterally and profoundly and even to represent a unitary geological source ; the metallogenetic implications cannot be neglected under these conditions.

Herja sector (Fig. 4 a) is located in the south of the Gutii Mts. and is characterized by the uplift of the pre-Neogene sedimentary basement, built up of Eocene transcarpathian flysch ; the compartments situated to the north and south of this elevation are step-like following E—W trending fractures, well marked on the map of the gravimetric anomaly by means of gradient belts. The sedimentary rocks

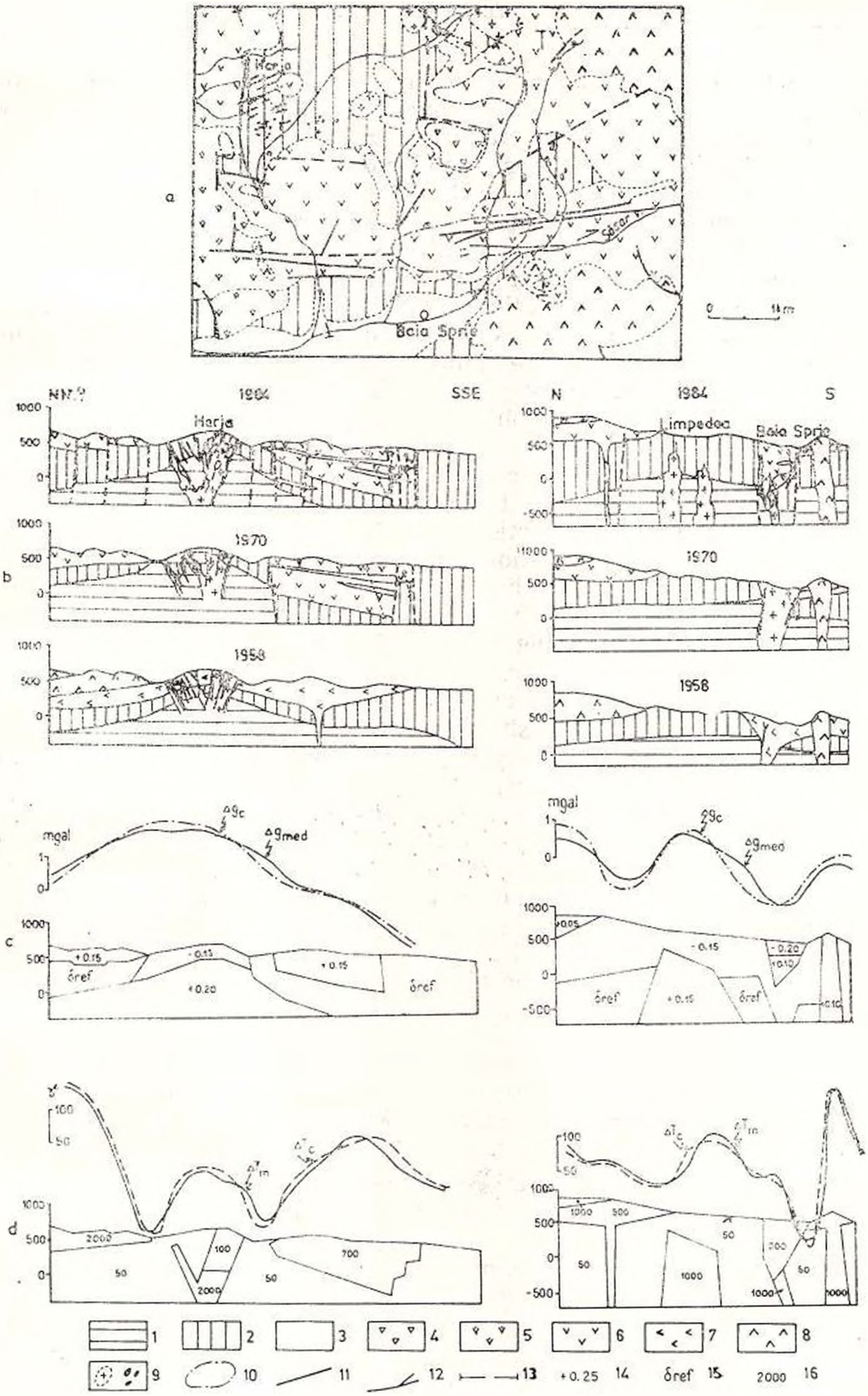


Fig. 4 — Geological sketch, geological sections and geoghysical models in the Herja and Baia Sprie areas.

are crossed by small andesite and microdiorite intrusions or are overlain by quartz andesites and pyroxene andesites. The intrusive bodies are surrounded by thermic metamorphism aureoles.

The Herja deposit consists of more than 100 thin veins which abund in galena and blende. Each vein exhibits a uniform paragenesis· but there are indices that it may be altered at depth. The veins are located both in Paleogene-Pannonian sedimentary rocks and in subvolcanic bodies which pierce the Pannonian.

The physical-mathematical modelling of geological sources represented on Fig. 4 b_1-1984 shows that the measured gravimetric — averaged — anomaly is due almost exclusively to the raising of the Paleogene sedimentary rocks, under the conditions of increasing density, due to thermometamorphic processes produced by intrusive bodies (Fig. 4 c_1).

The measured magnetic anomaly and its accord with the calculated one (Fig. 4 d_1) account for the occurrence of intrusive bodies, deprived of magnetic properties at their top, as well as of some thick, "fresh" lava products to the north and south of them.

The structural model has also a predicting value, as far as the high amplitude and intensity of the gravimetric anomaly suppose the presence, at depth, of a big geological source, overlain by other subvolcanic structures (intrusive), similar to that one located in the known deposit, to which associate adjacent mineralizations as such, but situated at lower levels. One should also consider the possibility of occurrence of some concentrations of other elements than those reported from the Herja deposit, such as copper, wolfram, molybdenum, tin and others, as far as the distance to this supposed unitary geological source decreases.

Tarna Mare-Turţ sector (Fig. 5 a), situated in the Oaş Mts, includes volcanic structures and subordinate subvolcanic bodies built up of pyroxene andesites, pyroxene and amphibole andesites, dacites and microdiorites respectively. They cross or overlie the Neogene molasse, which is very thick and conceals the pre-Neogene basement, on which no direct information has been obtained so far.

The common sulfide ± Au mineralizations, located in Pliocene-Pontian vein-like rocks or, in places, less as breccia bodies, occur in igneous rocks belonging to volcanic structures and subvolcanic bodies, as well as in adjoining sedimentary formations. These are accompanied

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a, geological sketch ; b, geological sections drawn up according to the knowledge of 1958, 1970 and 1984 ; 1, Herja ; 2, Baia Sprie ; c, gravimetric models : $\Delta\ g_{med}$ — mediate gravimetric anomaly ; $\Delta\ g_c$ — calculated gravimetric effect ; 1, Herja ; 2. Baia Sprie ; d, magnetometric models : $\Delta\ T_m$ — measured magnetometric anomaly ; $\Delta\ T_c$ — calculated magnetometric effect ; 1, Herja ; 2, Baia Sprie. 1. Paleogene flysch ; 2. Pannonian ; 3, Quaternary ; 4, rhyodacites ; 5, quartz andesites ; 6, pyroxene andesites ; 7, Sarmatian Seini pyroxene andesites (1958) ; 8, pyroxene andesites and post-mineralization amphibole andesites ; 9, andesite intrusions ; 10, hydrothermally altered rocks ; 11, fractures ; 12, hydrothermal mineralizations ; 13, position of geological section ; 14, density contrast ; 15, reference density ; 16, magnetic susceptibility value.

by relatively large hydrothermal alteration aureoles, represented by argillisations and subordinately, silicifiations and adularisations, which alter the physical properties of rocks.

The physical-mathematical modelling of the geological sources represented on Figure 5 b-1984 lead to the calculated gravimetric and magnetometric effects, which are in proper concordance (Fig. 5 c, d) with the measured gravimetric and magnetometric anomalies. Thus was restricted the ambiguity of interpretation of the measured gravimetric maximum, which is related to the Socea structure and especially to the pyroxene andesite pile, of considerable thickness in this area. From magnetic point of view, it is certified that this andesite pile is highly hydrothermally altered on the whole, accounting thus for the low intensity of the measured magnetic anomaly.

The intrusive bodies, represented on the geological section, which are probably responsible for the generation of the breccia pipes at Socea, may not be detected by gravimetric or magnetometric methods, under the supposed physical-geological conditions.

The dacite intrusive body which constitutes the Ghezuri structure, that exhibits obvious contrasts of density and magnetic susceptibility, is pointed out both by the gravimetric anomaly and the magnetometric one.

The models of the Ghezuri structure and subordinately of the Socea one point to the association of mineralized veins with intrusive sequences, fact also proved by the recent drillings in the Penigher structure (3 km southwards) where the veins are strictly associated with subvolcanic bodies.

One may thus suppose that the physical-geological modelling of other sectors in the Oaş Mts as well may disclose predictive aspects related to the discovering of new subvolcanic — intrusive — structures at different depths and of special metallogenetic importance.

Conclusions. The five sectors envisaged as instances of local geological structures belonging to a regional entity as a metallogenetic subunit, point to an essential feature of each physical-geological model : the provision of global and specific solutions of interpretation by taking into account the geological, petrophysical and geophysical features of the respective sectors.

For all instances has been used a unitary method for the elaboration of models, so that they correspond to the geological, geophysical and petrophysical ”facts“.

It is to emphasize that the structural solutions inferred should be considered as corresponding to a certain stage of knowledge.

As regards the investigation in the future, below the present-day level of depth, the models are to be made actual at least in the following directions :

— the location of igneous bodies, as regards their depth, the shape and extent of hydrometamorphism aureoles as well as their relations to the sedimentary and/or metamorphic formations crossed, possibly older igneous rocks as well ;

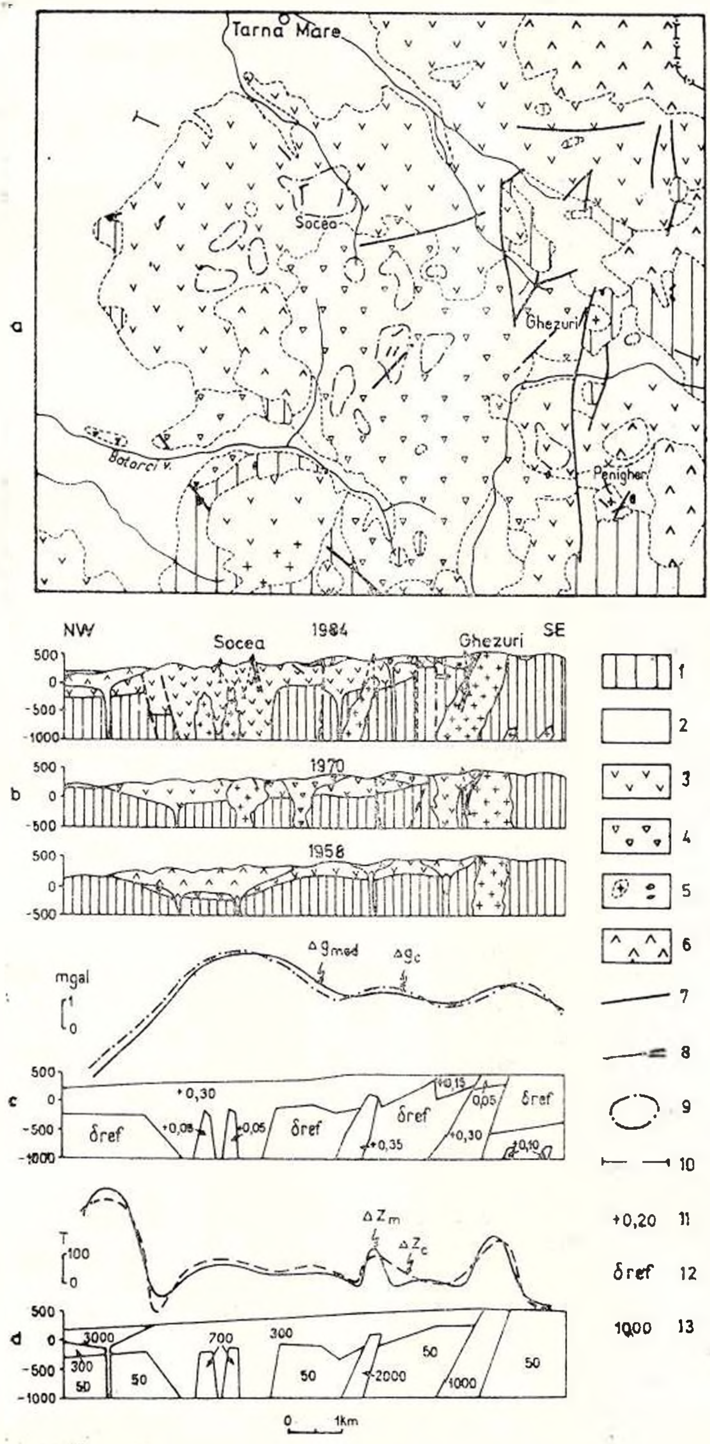

Fig. 5. — Geological sketch, geological sections and geophysical models in the Tarna Mare-Turț area.

a, geological sketch ; b, geological sections drawn up according to the knowledge of 1958, 1970 and 1984 ; c, gravimetric model ; Δ g_{med} — mediate gravimetric anomaly ; Δ g_c — calculated gravimetric effect ; d, magnetometric model : Δ T_m — measured magnetometric anomaly ; Δ T_c — calculated magnetometric effect. 1, Neogene sedimentary rocks ; 2, Quaternary ; 3, andesites and andesite pyroclastics ; 4, dacites, explosion breccias and volcano-sedimentary formations ; 5, andesite-diorite intrusions and dacite-granodiorite intrusions ; 6, Pliocene pyroxene andesites ; 7, faults ; 8, Pb-Zn±Au hydrothermal mineralizations ; 9, hydrothermally altered rocks ; 10, position of geological section ; 11, density contrast ; 12, reference density ; 13, magnetic susceptibility value.

— the features of deep-seated igneous structures and of common regional factors, responsible of their metallogenetic role.

The analysis method used in the present paper represents a means of prediction, both for fundamental and applied scientific investigation and for usual identification and thorough study of the metallogenetic structure in order to achieve a connection of deep-seated spaces of the known deposits and the spaces between the different structures, characterized by common natural factors and the recurrence in the "known" ones close to the surface.

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[4] Horizontal mining works reach a depth of 300 m at Cavnic, 450 m at Băiuţ, 50 m at Herja, —100 m at Baia Sprie and 100 m at Tarna.

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HYPOGENE ALTERATION GENETIC TYPES RELATED TO THE NEOGENE VOLCANISM OF THE EAST CARPATHIANS, ROMANIA

BY

CONSTANTINA STANCIU[1]

The hypogene alteration of the Neogene volcanic formations of the East Carpathians generated several neoformations under different genetic structural conditions, modified in time and space. This paper deals with the classification and succession of the hypogene alteration processes and products, based on the correlation between the petrographic, mineralogic and geochemical data of the neoformations with the results of the volcanogenic and metallogenetic studies. We shall discuss only problems related to the northern sector — Oaş-Gutîi Mts — and the southern one — Călimani-Gurghiu-Harghita Mts — as the central, subvolcanic sector is not thoroughly investigated.

Geological Setting

Contributions brought by numerous researchers (M. Borcoş, O. Edelstein, D. Giuşcă, R. Jude, I. Măldărescu, S. Peltz, D. Rădulescu, N. Stan) led to a better understanding of the volcanism in the Oaş-Gutîi Mts and Călimani-Gurghiu-Harghita Mts.

The Neogene volcanics of the East Carpathians belong to the andesitic volcanic arc of the continental crust of the Transylvanian and Pannonian blocks. In the whole zone it is possible to recognize the products of a calc-alkali volcanism which evolved from acid to basic ; both extrusive (lava prevailing) and intrusive forms are to be found. The fundamental type of activity is a mixed one generating stratovolcanic structures.

In the Gutîi Mts (Giuşcă et al., 1973) the volcanism practically developed uninterruptedly from the Badenian to the Upper Pliocene, on a basement of sedimentary rocks (Senonian-Oligocene) (Pl.). During the development of the process different petrographic types were separated : rhyolites (Badenian) entering into the constitution of a volcano-sedimentary formation, pyroxene andesites (Sarmatian), quartz andesites, subordinately dacites (Pannonian), pyroxene ± hornblende

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andesites (Pontian) ; all this appears in the south and represents the products of a volcanism which migrated from west to east ; the last term is represented by pyroxene andesites (Upper Pliocene) widespread in the north and north-east. The strong erosion destroyed the upper part of certain stratovolcanic structures, exposing supply channels, enrooted zones and subvolcanic bodies.

In the Oaş Mts the volcanic activity is less developed — from the Pontian to the Upper Pliocene — displaying the following succession of products : dacites, pyroxene andesites, quartz andesites, basaltic andesites.

The Călimani-Gurghiu-Harghita Mts (Rădulescu et al., 1981) represent the most recent part of the volcanic arc and are practically constituted only of andesites (two small dacite occurrences have an uncertain position). The basement of the region consists of crystalline schists, Mesozoic and/or Neogene deposits. The activity developed during two main phases. The 1st phase products have been completely destroyed by erosion and gave rise to the lower structural compartment represented by a volcano-sedimentary formation (Pannonian). The superstructures generated by the 2nd phase were built up on a basement constituted of this formation ; they represent the upper structural compartment (Pliocene) made up of stratovolcanos (Fig. 1). The presence of the volcanic edifices with crater and the calderas represent the typical volcanic element, well represented due to the very weak erosion. The petrographic series is reduced : hornblende-biotite±quartz andesites, hornblende andesites, hornblende-pyroxene andesites, pyroxene andesites, basaltic andesites. Subvolcanic andesitic-microdioritic bodies, situated in the central parts of the structures, have been identified by drillings and gravimetric determinations.

Hypogene Alteration

Further on we briefly present the data included in the main papers of the author of this paper (Stanciu, 1973 ; 1976, 1977 ; Stanciu, Medeşan, 1971 ; Stanciu et al., 1984 ; fide Peltz et al., 1982).

The results obtained up to now prove the existence of three types of hypogene alteration processes : the first type is related to the intrusive magmatic processes, partly real porphyry copper systems ; the second type is related to the tectonic fractures, and the third type is generated by postvolcanic processes.

Information on the Călimani-Gurghiu-Harghita chain is quite recent and (in several situations) incomplete due to the fact that the surface observation is highly limited by the poor erosion of the region ; the mine workings, especially the drillings (which only in places reached 1200 m deep), which supplied the most significant data, are included within small programmes of exploration (except the mineable deposits : native sulphur — Călimani, cinnabar — Sîntimbru Băi and argillized rocks — Harghita Băi, located nearby surface). In this region, within which the southern part was more active, all types of Pliocene alteration are to be found ; they appear separately or in association

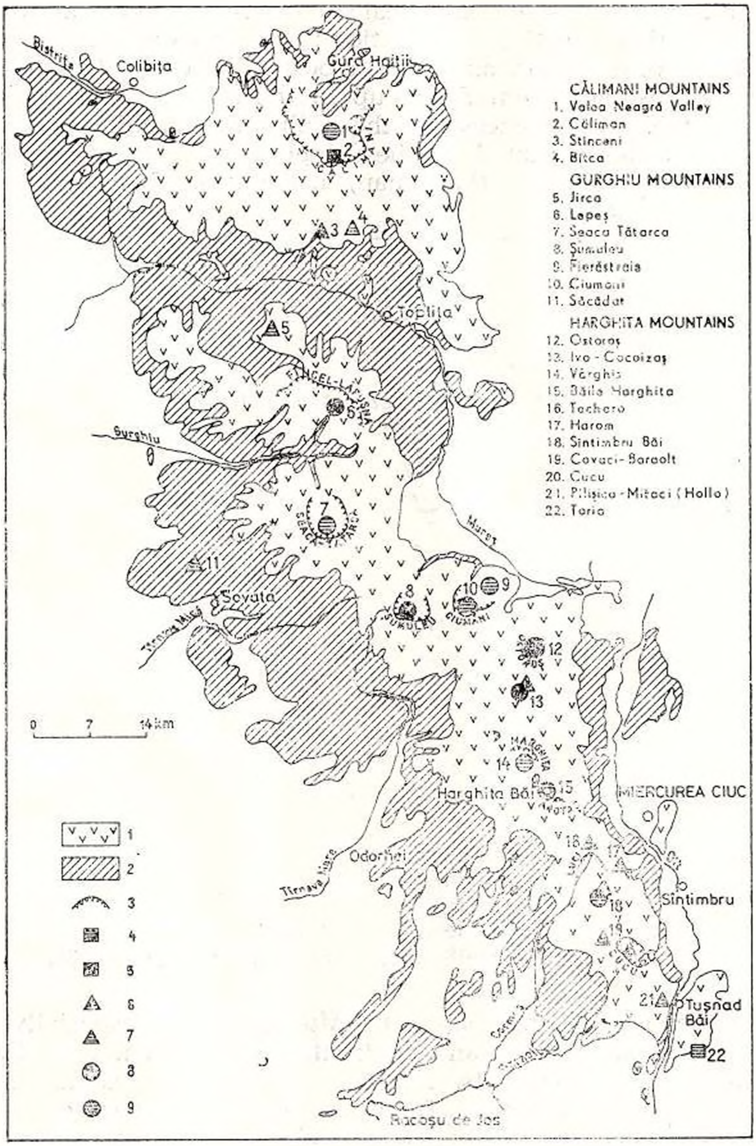

Fig. 1 — Hypogene alteration products of the Călimani-Gurghiu-Harghita volcanic chain (geology after Rădulescu and Peltz, 1981, simplified).

1, stratovolcanic andesitic formation ; 2, volcano-sedimentary formation ; 3, crater and caldera. Postvolcanic alteration ; 4, active solphatarian exhalation ; 5, Recent. Alteration related to tectonized zones : 6, alteration related to mineralizations ; 7, alteration on nonmineralized fractures. Alteration related to intrusive processes : 8, porphyry copper type alteration ; 9, parphyry-like type alteration.

(Fig. 1). Figure 2 presents an idealistic succession of the hypogene products and their spatial distribution ; a porphyry copper system, followed by a vein system and then a postvolcanic one (sensu Sillitoe, 1973) can be observed from depth up to the surface of a crater. The porphyry copper occurrrences in the Călimani-Gurghiu-Harghita are associated with those related to the Neogene volcanism of the Metaliferi Mts and the Banatitic (Laramian) magmatism of the South Banat, well known in Romania.

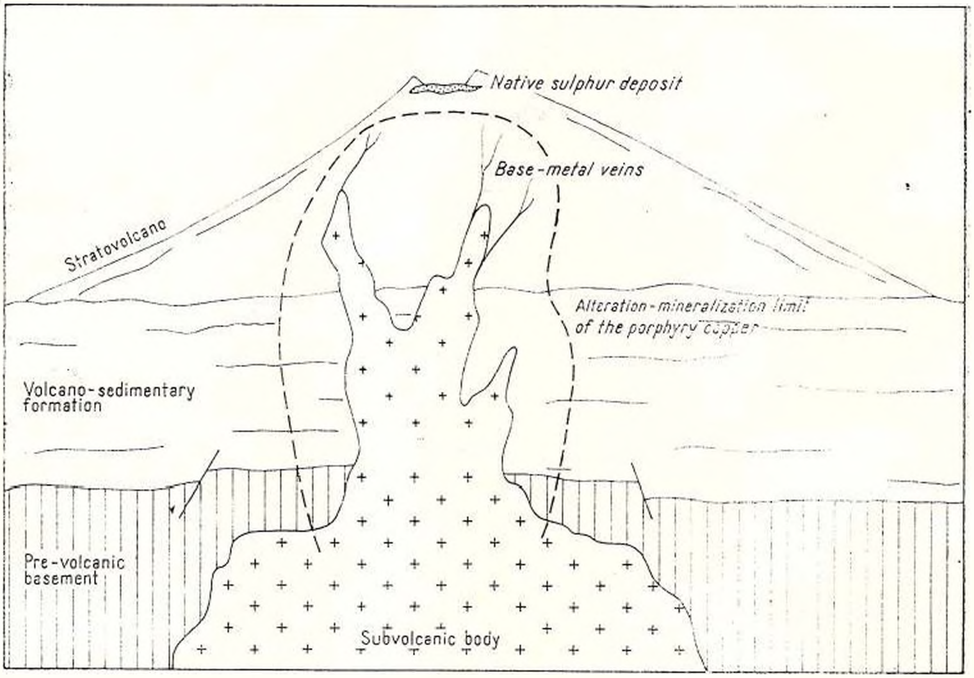

Fig. 2. — Hypothetic sketch on the succession of the hypogene products of the Călimani-Gurghiu-Harghita Mountains (completed after 1976).

The volcanics in the Oaş-Gutîi Mts gave rise especially to an alteration of Sarmatian-Pontian age in the Gutîi Mts and of Pontian-Pliocene age in the Oaş Mts, related to numerous fracture systems which had also metallogenetic functions. Here, the degree of knowledge is much more advanced ; the exposing of vast hydrothermalized areas by erosion but particularly the carrying out of important programmes of exploration and exploitation with galleries (11 mining fields are presently in operation) allowed the obtaining of numerous data as well as of a good correlation with the associated volcanic and metallogenetic elements.

The table shows the main characteristics of the hypogene alteration in zones assigned to the mentioned genetic types associated with mineral resources ; minor alterations were also mentioned beside the active exhalations in order to mention the whole range of hypogene alterations (Tab.).

Alteration Related to Intrusive Processes

The alteration controlled by andesite and/or microdiorite intrusions — representing the apical parts of some small subvolcanoes located at different levels of the stratovolcanic structures in the Călimani-Harghita Mts — are especially situated inside the craters and calderas, where the instrusive processes take part in the building up of the volcanic apparata towards the end of the regional eruptive activity.

Porphyry Copper Type Alteration

The most representative processes developed in the Fîncel-Lăpuşna, Şumuleu, Ostoroş, and Ivo-Cocoizaş apparata, situated in the axial part of the Gurghiu-Harghita Mts. Considering all the paragenetic aspects, very complicated due to the metasomatic interferences in the multi-intensive spaces and the superposition of the alteration processes, a zonal development of the alteration centred on the generating intrusions can be observed ; the succesion is : biotitic-amphibolic-chloritic-argillic-tourmalinic alteration. The biotitic alteration [2] and only subordinately the amphibolitic one — both having an innermost position — formed during a first late magmatic phase, on an initial propylitic (autometamorphic type) background, which in places preserves small amounts of fresh rock. The subsequent alterations are the result of the hydrothermal fluid activity which gives rise to transition chloritic rocks [2] and then, by contamination with the meteoric waters, generates a pervasive argillic alteration (mostly occurring in the stratovolcanic host formation) ; in this way the external part of the porphyry copper system in the Gurghiu Mts is defined. In the Harghita Mts the hydrothermal process continued with an intensely tourmalinic metasomatosis accompanied by tourmaline-quartz depositions in vein breccias ; they develop at the periphery of the intensely tectonized zones of the argillic hanging wall. There is no sufficient evidence but it is possible that such formations might represent the final moments of the breccia pipe structures, which are associated with copper mineralizations in the Andine Cordillera in Chile and Argentina (Sillitoe, Sawkins, 1971).

The metallic minerals occur as disseminations, fissures and blurred veinlets, concomitantly with alteration ; however, a slight metallization was observed in the propylitic, even fresh zones. Pyrite is found in all alteration zones and the magnetite-pyrrhotine-chalcopyrite is conformable with the internal alteration (biotitic and chloritic) ; as the content in Cu is small, no zones of economic interest have been outlined up to now. In the external, argillic zone the paragenesis is much reduced, pyrite being the characteristic mineral. The tourmalinic rocks consist of pyrite ± marcasite, spordically molybdenite, chalcopyrite.

At Şumuleu and Ivo-Cocoizaş late intrusive breccias occurred along certain fracture planes ; at Ivo-Cocoizaş these breccias are accompanied by a xenolithic injection andesite.

In the upper part of the porphyry copper system, small quartz voids and base metal sulphide vein depositions are frequently found ; lateral passage towards fissural depositions and cinnabar±stibnite impregnations are observed at Ivo-Cocoizaş.

The situations studied by us are partly found in Lowell and Guilbert's (1970) and Hollister's (1975) models, several similarities with the Neogene porphyry copper in the Metaliferi Mts being observed (Ianovici et al., 1977).

Hypogene alterations, within which partial aspects of the porphyry type alteration are recognized, also occur in the Călimani caldera, at Stînceni, Seaca-Tătarca, Fierăstraie, Ciumani, Băile Harghita (with pay accumulations of argillic rocks nearby the surface), and Sîntimbru Băi.

Alteration Related to Tectonized Zones

The alteration related to the postvolcanic tectonic activity appears in both volcanic sectors and is well known especially in those zones where it is accompanied by metallogenetic activity, widespread in the Gutîi Mts and subordinately in the Călimani and Harghita Mts.

In the Gutîi Mts, Borcoş and Lang (1973) proved that between the volcanic and metallogenetic activity there are well-defined genetic space relationships, which point out the existence of three hydrothremal phases located on a major W—E tectonic alignment, in the southern part of the massif ; the first hydrothermal phase is connected with Sarmatian pyroxene andesites, the second phase — with Pannonian quartz andesites, and the third phase — with Pontian pyroxene andesites. These phases show an eastward migration, concomitantly with that of the eruptions they are linked to. Each hydrothermal phase has distinct alteration areas (Pl.) ; the main areas are represented by metalliferous accumulations — veins, subordinately impregnations, stocks — dominated by base metal sulphides with local enrichment in Cu in depth and Au$\pm$Ag at the upper part. The Sarmatian areas are associated with base metal sulphide mineralizations $\pm$ Au (Racşa, Ilba, Nistru) ; the Pannonian areas are represented by significant gold-silver deposits (Borzaş, Săsar, Valea Roşie) and subordinately by base metal veins (Tyuzoşa Wilhelm) ; the Pontian areas develop around the base metal veins accompanied by Cu and/or Au$\pm$Ag accumulations (Herja, Baia Sprie, Suior, Cavnic, Văratec).

The rock alteration predates the mineralization and generally has a zonal arrangement, around the access ways of the solutions, which represent also the places of the metalliferous depositions. In fresh or propylitic rocks (autometamorphic type) the following alteration types occur : chloritic-adularic-sericitic-argillic (locally carbonatic and silicic) ; this sequence does not always occur completely. The metasomatosis is achieved by an open system, with a high oxireduction potential. In the beginning the K-rich hydrothermal solutions affected the propylitic rocks and generated nearby them chloritic rocks due to the Fe and Mg redistribution. After that the adularic alteration, typical of the Gutîi Mts (Giuşcă, 1960), occurred as a result of the potassic solutions action ; K introduction is accompanied by the strong leaching of Na, Ca, Mg and Fe. This alteration type was highly intensive during the second phase (Pannonian), representing an excellent environment for the gold-silver depositions. The sericitic-argillic type hydrolitic alterations developed especially towards surface and very late, under low temperature conditions and K^+/H^+ decrease due to the meteoric water contamination.

There is a genetic continuity between alteration and the associated mineralization, proved by certain substitution minerals but particularly by precipitation, in fissures and dissolution voids in rocks, found also in the vein gangue. Several direct relations between alteration and mineralization have been observed : chloritic type — copper mineralization > base metal ; adularic type — gold-silver > base metal ; sericitic and argillic type — base metal > gold-silver ; propylitic rocks are accidentally to be found nearby base metal or cupriferous veins without alteration aureolas.

In the Călimani-Harghita Mts the alteration related to fractures is located in the upper compartment (except the Săcădat area), at the exterior and at the upper part of the porphyry copper system ; the adularic alteration is missing. At Stînceni a common alteration took place around the endogene veins and breccias with base metal sulphides ± Au ; in the Sîntimbru Băi and Ivo-Cocoizaş cinnabar zones the alteration is reduced, and argillizations, accompanied by superposed processes of carbonation, silicifiation ± tourmalinizations, are predominanting. Numerous areas with poor indications of mineralization are controlled by nonmineralized fractures and zones of brecciation surrounded by argillic zones which, in places (nearby the circulation ways) pass to silicic zones (Bîtca, Jirca, Săcădat, Vîrghiş, Techero, Harom, Covaci-Baraolt, Cucu, Pilişca-Mitaci). Such areas represent the terminal part of hydrothermal processes with still unknown continuity at depth. The most frequent cases occur in the Harghita Mts.

Postvolcanic Alteration

This alteration type is specific to the volcanism in the Călimani and Harghita Mts ; it is very significant in the native sulphur deposit in the upper part of the Călimani caldera and is extremely reduced around the present exhalation points.

Alterations, concordant with the stratification of the volcanic products, took place around the sulphur deposition zones. The succession, occurring nearby surface is restricted : chloritic-argillic-silicic ± sulphur $\gg$ iron sulphides > iron hydroxides. The endogene gas solutions underwent essential modifications by reaction with the rocks and by strong contamination with the abundant, meteoric waters. At the beginning small amounts of chloritic rocks occurred, succeeded by argillic rocks ; later on the endogene solutions, in contact with underground waters, become highly acid and, consequently, they gave rise to a strong leaching keeping only Si and partly Al, then reconstituted in silicic rocks ± alunite forming the most significant levels. They accumulate native sulphur — impregnations, subordinately depositions — formed by the H_2S oxidation or its combination with H_2SO_4 ; the leached iron is refixed in FeS_2 and in the most superficial parts it concentrates as hydroxides.

The present volcanic exhalations occur in the Harghita Mts and are represented by several moffetes, in the axial part of the massif, and a solfatara, in the south-westernmost part (at Toria), which intermittently generates small depositions of native sulphur. As the emissions are punctiform, the alterations are not significant. Small amounts of silicified rocks, resembling those in the Călimani caldera, occur around them.

Conclusions

The main characteristics of each of the three types of hypogene alteration are, as follows : 1. the porphyry copper type displays a strong homogeneity (except the amphibolic zones occurring locally and in small amounts) up to the level of the tourmalinic alteration typical of the Harghita Mts. 2. The alteration related to fractures is more diverse, in zones with base metal sulphides ± gold (some terms may be missing), and in the Oaş-Gutîi Mts the adularic metasomatosis is typical and well represented ; in cinnabar zones the alteration is reduced to one main term — the argillic one. 3. The postvolcanic alteration, nearby surface, is also restrictive, generating argillic and silicic neoformations in the native sulphur places and only silicic alteration around the emission points of the present exhalations.

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[2] K-feldspar occurs sporadically and in small amounts in some biotitic and chloritic rocks. The biotitic rocks are penetrated by a granophyre fissure at Ivo-Cocoizaş and by a significant anhydrite front at Lepeş.

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HYDROTHERMAL ALTERATION AREAS OF THE GUTÎI MOUNTAINS

Geology after Borcoş, Lang, Peltz, Stan (1973), simplified
Hydrothermal alteration data compiled by Stanciu

Imprim. Atel. Inst. Geol. Geof.

MAIN CHARACTERISTICS OF THE HYPOGENE ALTERATION

1. PORPHYRY COPPER TYPE ALTERATION

No[1]	STRUCTURE	IGNEOUS HOST ROCK[2]	DEPTH (m)	PROPYLITIC	ALTERATION				
					BIOTITIC	AMPHIBOLIC	CHLORITIC	ARGILLIC	TOURMALINIC
	Gurghiu-Harghita Mts								
6	LEPEŞ	αq - pl	500-1100	▨	cp, mgt		mgt,cp	py(cp, gn, sl)	
8	ŞUMULEU	αam±px μδam,αam-pl	600-1200	mgt ,cp (po)	mgt, cp (po)	▨	mgt, cp (po)	py (cp, sl)	
12	OSTOROŞ	αpxam - pl	300-1200		mgt, po, cp, sl (mo)		cp (mo)	py (cp, sl)	py (cp)
13	IVO - COCOIZAŞ	μδpx, μδam-pl	650	mgt, po (cp)	mgt, po (cp)	mgt	mgt,po,cp	py (mo, po, sl)	py, ms (mo)

2. ALTERATION RELATED TO MINERALIZED FRACTURES

No	DEPOSIT	VOLCANIC COUNTRY ROCK	DEPTH (m)	PROPYLITIC	ALTERATION				MAIN ORE TYPE
					CHLORITIC	ADULARIC	SERICITIC	ARGILLIC	
	Gutîi Mts								
1	RACŞA	αpx-sm	200	▨		▨■○	▨■	▨	Gold-base metal
2	ILBA	αpx+sm vs - bn	200 - 500	▨■△	▨■△	▨■□	▨■□	▨	Base metal - gold
3	NISTRU	αpx - sm	250 - 500	▨	▨	▨■□	▨■□	▨■□	Base metal - gold
8	TYUZOŞA	αq- pn	100	▨		▨■□		▨■□	Base metal+ cupriferous
7	WILHELM	δ -pn	350	▨		▨■□		▨■□	Base metal-gold
6	BORZAŞ	αq - pn	200		▨	▨■○		▨	Gold
4	SĂSAR	αq - pn αpx - sm	200-750	▨	▨	▨■○	▨■○	▨■○	Gold - silver
5	VALEA ROŞIE	αq - pn αpx- sm	500	▨	▨	▨■□○	▨		Gold - silver+ base metal
9	HERJA	αpx-p	400 - 700	▨■□	▨■□		▨■□	▨■□	Base metal
10	BAIA SPRIE	αpx-p	250 - 850		▨■△	▨■□○ +Cb		▨■○	Base metal - gold+ cupriferous
11	ŞUIOR	αpx-p	200 - 500	▨		▨■○		▨	Gold - silver+ base metal
12	CAVNIC	αpx-p αq - pn	500	▨■□	▨■□	▨■□	▨■□	▨■□	Base metal
13	VĂRATEC	αpx-p	200-400	▨		▨■□		▨■□	Base metal
	Oaş Mts								
	TARNA MARE	αpx- p - pl	200 - 600	▨	▨	▨■□	▨	▨■□	Base metal
	Călimani - Harghita Mts								
3	STÎNCENI	αampx±q αampx, μδ-pl	650	▨	▨	absent	▨	▨■	Base metal
18	SÎNTIMBRU BĂI	αq - pl	300					▨■ +Cb, Si, Tm	Cinnabar
13	IVO-COCOIZAS	αpx,am - pl	350	▨				▨■	Cinnabar

3. POSTVOLCANIC ALTERATION

No	DEPOSIT	VOLCANIC COUNTRY ROCK	DEPTH (m)	ALTERATION				ORE TYPE
					CHLORITIC	ARGILLIC	SILICIC	
	Calimani-Harghita Mts							
					RECENT			
2	CĂLIMANI	αpx - pl	350		▨	▨	▨■	Native-sulphur
					ACTIVE EXHALATIONS			
22←	Moffetes and solfatara (Toria) in Harghita Mts						▨	

ABBREVIATIONS

HYDROTHERMAL MINERALS

cp	chalcopyrite
gn	galena
mgt	magnetite
ml	melnicovite
mo	molybdenite
ms	marcasite
py	pyrite
po	pyrrhotite
sl	sphalerite

MAGMATIC MINERALS

am	hornblende
q	quartz
px	pyroxene

ROCKS

α	andesite
δ	dacite
μδ	microdiorite
vs	volcano - sedimentary formation

GEOLOGIC TIME

bn	Badenian
p	Pontian
pl	Pliocene
pn	Pannonian
sm	Sarmatian

ALTERATION

cb	carbonatic
si	silicic
tm	tourmalinic

MISCELLANEOUS

1 -	Number of the occurrence on the plate and fig.1
2 -	Genetically associated with alteration and mineralization
( )	Subordinate or occasional occurrences
■	Alterations in the vein walls or with mineral resources impregnations
○	Gold ± silver
□	Base metal
△	Cupriferous

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PALYNOFACIES. STRATIGRAPHIC-PALEOECOLOGIC CONCEPT AND GEOLOGICAL EXPLORATION TOOL

BY

NICOLAE BALTEȘ[1]

Introduction

The ample geological activity for discovering new reserves of energetical raw materials, particularly hydrocarbons, has brought about a strong development of paleontological sciences, having direct implications in improving prospection and exploration methodologies. In this category, palynological investigations proved to be very useful by the ever wider range of fields tackled, starting with the establishing of the stratigraphic sequence and correlation, up to the characterization of sedimentary environments, the identification of the hydrocarbon source rocks, the more accurate definition of the secondary oil migration sense and stages, a.o. Under these circumstances, it became necessary to use a concept, including all fields of palynological research, which should be conclusive, simple, convenient and allowing its application in all practical geological domains. At the present level of palynology the notion of "palynofacies" may gather — in a satisfactory manner — the requirements of a total utilization, joining at the same time other geological branches employing the term "facies". In this way, the palynofacies represents all aspects covered by the disseminated microvegetal content in all types of sedimentary of metamorphic rocks (Balteș, 1979). These types may be (a) of a structured type, with microspores, pollen, unicellular algae, cuticules, determinable fragments of vegetal tissues, fossil woods, chitinozoans, fungi, bacteria, etc ; (b) of amorphous type resulting from the chemical and decomposition of the former, under several ways : sapropelite (algae), leptobiolite (algae and high plants), ligno-humite (high plants) or as secretion products (waxes, resins, etc.). The first type of palynofacies has a wide application in stratigraphy, sedimentary environment reconstitution and paleogeography. The second one allows the establishing of the thermal maturation degree of the organic matter, leading to the identification of the

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hydrocarbon source rocks, and the determination of the oil potential of a region.

Stratigraphic Palynofacies

Subject to the biological evolution requirements, the vegetal organisms were associated, during the phanerephytic crust evolution, in various, homogeneous systems, ortho- or paragenetic, which characterized different periods according to their life conditions, as well as the paleogeographic evolution of the region. Forming stratigraphic units of values, the palynofacies determined the creation of the ”palynozone“ and, within it, of the ”assemblage zone“ (the totality of present forms or of a group of forms), ”range zone“ (the stratigraphic value of a certain element in the association) and ”acme zone“ (the stratigraphic interval or level of quantitative explosion of some species). Thus, turned into account, the stratigraphic palynofacies becomes an extremely useful tool in geological prospecting and exploration, all the more so as the taxons belong to all sedimentary environments. The whole sedimentary column from the Cambrian to the Pliocene is palynologically determined in Romania. Some stratigraphic diagnoses also refer to anchimetamorphosed Precambrian deposits, more especially, of the foreland units. Most of the assemblage zones define the main lithofacial units, usually at the stage level, while the range zones (parts of them) as o rule, intraformational. Acme zones, although of a local value, characterize limited areas with good biotic and preservation conditions, being able to represent excellent stratigraphic and paleoecological markers.

The palynozone correlation has variable efficiency degrees, which decrease according to distance and strata youth, being considered accurate at the level of the assemblage zone, within the same tectonic unit, and good enough between neighbouring units. Starting from the Neogene deposits, the palynozone correlation becomes more and more difficult, the range zones, and particularly, the acme zones getting an intraregional and even local value.

Environmental Palynofacies

If the paleobotanical reconstitution and more especially the phytosocial assemblage are difficult to be achieved, because of shortcomings in the synonymy between artificial and natural classification, generally, used in biology, some essential aspects of environment may be well characterized by palynofacies. In a lot of continental, marine, shelf, evaporitic etc., sedimentary areas, where the facies transition determines many and subtle variations of environment, and generally the pelagic organisms, the best zone fossils are episodic or very rare, the palynofacies, through its nature, abundance, physical and chemical preserving state, remains the most efficient tool of knowing them, with direct implications in appraising some potential reserves of raw material.

Taking into consideration : (a) the origin and main physical features of the sedimented microvegetal material ; (b) the manner of its transport and depositing in the basin, as against the actual sedimentation ; (c) the preservation conditions of the organic matter within the sediment, as well as during the lithification, one can inversely reconstitute the nature of the deposition environments in different points of a basin with ambiguous lithofacies.

In the Romanian oil industry, this way of interpreting the sedimentation environment is used with variable degrees of detail, leading — together with petrographic and microfacies investigations — to a better knowledge of the facies favourable to the hydrocarbon generation and accumulation. The grouping on major and very simplified sedimentologic categories, the main deposition environments may be palynologically characterized (Fig. 1).

— The continental environment which has a very variable global palynofacies according to the sedimentary nature where it develops and the microflora is accumulated and preserved. Three main types of palynofacies may be distinguished : (a) of detrital sub-aerian origin, palynologically poor, with big spore (0.5—1 mm) fragments, cuticules, vegetal stalks and fossil woods. In Romania, it was encountered in deposits of a red, brown, green or white colour in the Permian, Lower and Upper Trias and Lower Cretaceous of the Moesian Platform, Upper Cretaceous in the East Carpathian Flysch and Lower Eocene in the Transylvanian Depression ; (b) of deltaic sedimentation, rather rich in middle (0.05—0.5 mm) and big spores, fragments of saccate pollen and fossil woods. The lithological variety determines remarkable palynological fluctuations, the most abundant being the clayey-siltic levels, sometimes very thick. A good example is the Lower Miocene molasse in the Carpathian Foredeep ; (c) of coaly sedimentation, palynologically rich and varied, both in the coal beds, sandstones, and more especially, in the interbedded clays. Most of the palynological types are present :

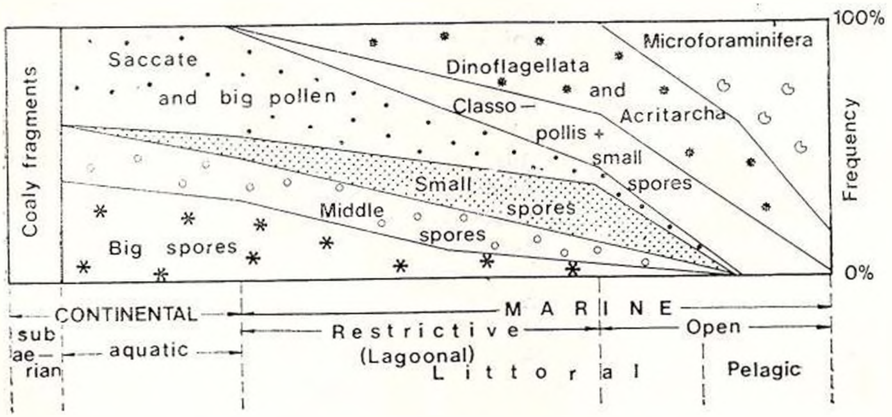

Fig. 1 — Distribution of the main palynologic groups in a hypothetical sedimentary area (after references and own data).

bacteria, spores, all kinds of pollen, fresh-water algae, high plant tissues, waxes, resins, fossil woods, etc. It was identified in the Carboniferous and Lias of the Moesian Platform and the Danube Delta, the Upper Oligocene of the Carpathian Foredeep, Upper Miocene (Sarmatian) and Pliocene with lignite of intramountainous basins.

— The lagoonal environment is characterized by a palynofacies, faithfully representing the interference area of continental and marine environments, rich in nutrients, having a large biotic activity and proper preservation conditions of organisms and organic matter ; as a whole, the palynofacies of this environment can be divided into two types : (a) lagoonal-deltaic with very rich associations in spores and small pollen, saccata pollen and, especially, dinoflagellates and acritarchs. As fragments, more or less determinable, chitinozoans, conodonts, scolecadonts, benthic algae (Codiaceae, Dasycladaceae), bacterian piles occur in old formations. In Romania, such a palynofacies was determined in inner, inter and subtidal platform deposits in the Upper Devonian-Dinantian, Ladinian and Bojacian, which preserves its qualitative features, but the frequences substantially decrease, except for the phytoplankton. Such situations were encountered in some carbonatic-evaporitic sequences in the Upper Devonian-Dinantian, Anisian-Ladinian, and Badenian-Sarmatian of the Moesian Platform (Fig. 2).

— The marine environment determines a very characteristic palynofacies, easily detectable through its constant presence, taxonomic variety and high phytoplanktonic frequences, chitinous microforaminifera too. Subordinately, it can also be associated, according to the sedimentation place, either with amorphous organic matter, or with small spores and pollen, an aspect mirroring the intensity and transport agent of the terrigenous material in the basin. The marine palynofacies occurs in pelagic deposits of outer platform and basinal ones, in limestones of mudstone and wackestone type, spongolites and clays. The most conclusive marine palynofacies were determined at the Dinantian, Callovian-Oxfordian, Albian and Senonian intervals in the Moesian Platform, Upper Eocene in the Transylvanian Depression and East Carpathian Flysch and in the Badenian of he Carpathian Foredeep.

Oil Source Rock Palynofacies

The exploration of some important objectives in new areas, at great depths, or located in complicated tectonic situations, has determined the development methodology of the hydrocarbon source rock identification, the palynofacies being an extremely useful element as well (Balteş, 1981). Determined as a complex of biological, chemical and physical factors, and first and foremost by the quality of organic matter, the palynofacies of the oil source rock represents a special moment of thermal maturation of the kerogene directly depending on the burial depth, geothermal gradient, tectonic evolution and, certainly, the geological time.

In Romania, besides a general characterization of the main generating sequences (Balteş, 1983), detailed oil source rock studies, using the palynofacies, have been achieved. They are dealing with the Wallachian Carpathian Foredeep (unpublished), diapir folds of the same area (Albu, Balteş, 1983), northern (Balteş et al., 1982) and southern (Balteş, 1983 a) flanks of the Getic Depression, eastern part of the Moesian Platform (Balteş, 1982), the Bîrlad (Balteş, 1983 b) and Pannonian (Balteş, Moldoveanu, 1981) depressions, and the Paleozoic of the Moesian Platform (preliminary work).

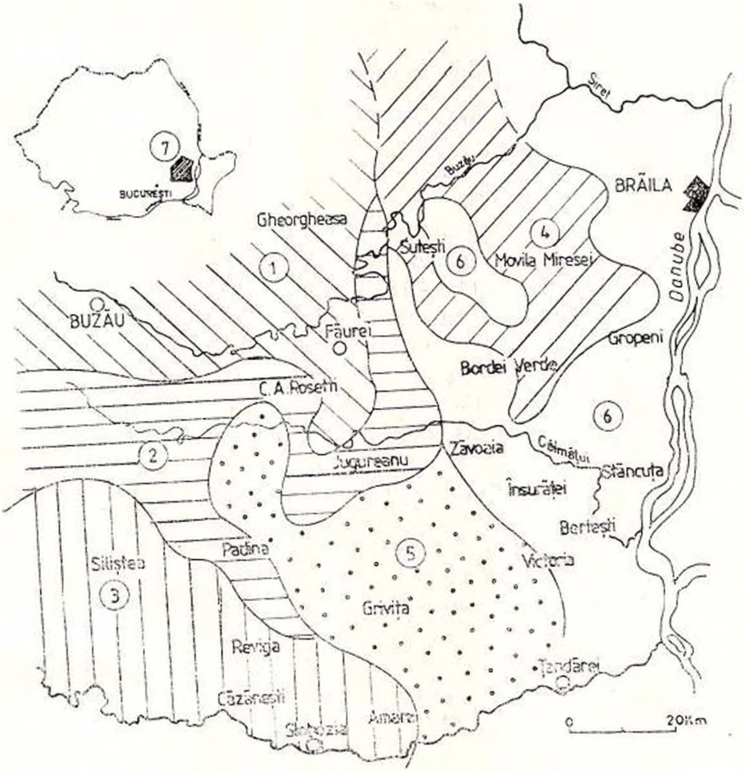

Fig. 2 — Tentative map of the Upper Badenian palynofacies in the eastern part of the Moesian Platform (Romania).
1, marine typical ; 2, lagoonal with frequent marine phases ; 3, lagoonal mainly calcareous ; 4, lagoonal with rare marine phases ; 5, lagoonal-continental mainly detrital ; 6, emersed area ; 7, study area.

As a whole, the successful use of the palynofacies concept marks a superior qualitative stage of research in the Romanian oil geology, thus creating the prerequisites of better information also contributing to the achievement of the programmes of increasing the energetic raw material reserves.

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ETUDE STRATIGRAPHIQUE DU SÉNONIEN INFÉRIEUR DES CARPATHES ORIENTALES (ROUMANIE)

PAR

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Les dépôts sénoniens inférieurs du flysch des Carpathes Orientales roumaines ont fait l'objet de plusieurs études géologiques depuis la fin du dernier siècle. Parmi les chercheurs qui ont étudié la stratigraphie du Sénonien inférieur des Carpathes Orientales, pendant les trois dernières décennies et dont les travaux apportent d'importantes contributions à la connaissance des formations de cet âge, on doit mentionner : Murgeanu et al. (1963), Tocorjescu (1963), Filipescu et al. (1963), Dumitrescu (1963), Mirăuță, Mirăuță (1964), Avram (1967), Neagu (1968), Alexandrescu, Rogge-Țăranu (1981), Szász (1971), Ion (1981, 1983).

Les auteurs de la présente note tentent d'ajouter quelques résultats obtenus par investigation minéralo-pétrographique et micropaléontologique de la séquence sénonienne inférieure des unités d'Audia, médio-marginale et externe.

Origine du matériel d'étude

Le matériel d'étude est représenté par des échantillons récoltés des affleurements et par des carottes. Les affleurements échantillonnés se trouvent dans les vallées de Măgurici, Andrei, Motreanu, Pîrlei, Ciumîrna, Sovîrîta, toutes des environs de Găinești (Suceava), les vallées de Slătioara, Neagra et Pluton, toutes du secteur de Pluton-Pipirig (Neamț), la vallée de Chetag de Covasna (Covasna), la vallée de Putna à Lepșa (Vrancea) : les carottes proviennent des forages d'exploration du Ministère du Pétrole, situés sur les structures géologiques : Straja, anticlinal de Micodina-Gura Boului, Frasin (Suceava), Pluton, Pipirig, Mitocul lui Bălan, Pîngărați, Doamna (Neamț), Geamăna, Gropile lui Zaharache, Zemeș, Moinești, Toporu, Sopoteni, vallée d'Uzului, (Bacău), Lepșa, (Vrancea), Izvoarele Putnei, Ojdula, Oituz, Zăbala (Covasna). Le matériel lithologique a été préparé et étudié du point de vue minéralo-pétrographique, microfaunique et nannoplanctonique.

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Résultats de la recherche

L'analyse de laboratoire a relevé que la séquence stratigraphique faisant l'objet de la présente note débute par des argilites ferrugineuses, tuffacées de l'unité d'Audia et par des grès tuffacés, bréchifiés de la sous-unité médio-marginale. Au point de vue paléontologique, l'intervalle stratigraphique étudié est caractérisé par les associations microfaunique a *Marginotruncana tarfayensis* et nannoplanctonique à *Micula concava concava*, typiques pour le Coniacien basal.

L'association microfaunique comprend des foraminifères agglutinants et calcaires, benthiques, planctoniques, des radiolaires, des débris squelettiques de spongiaires (spicules) et des organismes appartenant au groupe Incertae Sedis. Il faut mentionner particulièrement les foraminifères planctoniques : *Praeglobotruncana delrioensis* (Plummer), *P. stephani* (Gandolfi), *P. marginaculeata* Loeblich et Tappan, *Marginotruncana „renzi“* (Gandolfi), *M. sigali* (Reichel), *M. schneegansi* (Sigal), *M. tarfayensis* (Lehmann), *Dicarinella biconvexa biconvexa* (Samuel et Salaj), *Archaeo globigerina creatacea* (d'Orbigny), *Globigerina caspia* Keller et le groupe Incertae Sedis : *Pithonella ovalis* (Kaufmann), *Stomiosphaera sphaerica* (Kaufmann) et *Calcisphaerula innominata* Bonet.

Le nannoplancton calcaire déterminé à ce niveau stratigraphique est moins convaincant en comparaison des foraminifères accompagnants. Les espèces de l'association ont une évolution lente. Il paraît que la seule espèce qui pourrait suggérer un niveau immédiat supérieur au Turonien serait *Micula concava* (Stradner) Bukry (Bukry, 1969).

La séquence stratigraphique suivante est représentée par un complexe tuffitique carbonaté argilo-ferrugineux se caractérisant par la présence du foraminifère (*Dicarinella concavata concavata* (Brotzen) et de l'espèce de nannoplancton calcaire *Marthasterites furcatus* (Deflandre) Deflandre.

La lithoséquence commence par des tufs vitroclastiques rhyodacitiques à niveaux de silicolites (des couches de radiolarites ou lamines jaspoides) apparaissant presque dans tous les secteurs étudiés de l'unité d'Audia, suivis d'un complexe calcaire, tuffitique (calcaires micritiques, fréquemment argileux, marno-calcaires bioclastiques, biomicrites associés aux tuffites).

Dans l'unité médio-marginale se développe dans les secteurs occidentaux un complexe dominant détritique, grossier (grès feldspathiques à remaniements polygènes, y inclus des roches andésitiques) et dans les secteurs orientaux des grès fins. Dans la demi-fenêtre de Putna-Vrancea aux calcaires à tuffites s'ajoutent des silico-spongolites gréseuses ou des calcaires gréseux à chailles.

Le niveau supérieur comporte des calcaires micritiques argileux, marno-calcaires, biomicrites en couches décimétriques ou en rythmes minces sous forme de laminites, en contenant d'une manière subordonnée deux générations de matériel tuffacé sous forme de lamines discontinues ou de nodules. Même à ce niveau stratigraphique, les secteurs orientaux de l'unité médio-marginale sont différents, du fait qu'ils sont plus détritiques (grès finement calcitiques). L'intervalle respectif a été individualisé dans toutes les zones recherchées.

Parfois, la succession s'achève par des roches carbonatées argileuses fréquemment ferrugineuses (calcaires, marno-calcaires, marnes) généralement bioclastiques et ferrugineuses, argiles tuffitiques, chlorito-sériciteuses ou argilites tuffitiques ferrugineuses.

La microfaune et le nannoplancton calcaires sont représentés par les associations à *Dicarinella concavata concavata* et *Marthasterites furcatus*. Parmi les espèces rencontrées dans les deux associations, il faut mentionner les foraminifères : *Praeglobotruncana aumalensis* (Sigal), *Dicarinella concavata concavata* (Brotzen), *D. biconvexa biconvexa* (Samuel et Salaj), *Marginotruncana angusticarinata* (Gandolfi), *M. pseudolinneiana* (Pessagno), *Globotruncana fornicata* Plummer etc. et les nannofossiles : *Broinsonia signata* (Noël) Noël, *Stephanolithion laffitei* Noël, *S. achylosum* (Stover) Stradner, *Cylindralithus serratuss* Bramlette et Martini, *Helicolithus trabeculatus* (Gorka) Verbeek, *Ahmuellerella octoradiata* (Gorka) Reinhardt, *Marthasterites furcatus* (Deflandre) Deflandre, etc.

La présence de ces espèces confère au complexe lithologique décrit ci-dessus l'âge coniacien.

La succession du Sénonien inférieur finit par un paquet hétérogène de roches. La variation des types de roches est très évidente d'une unité ou sous-unité structurale à l'autre et même dans le cadre de la même unité.

Cependant, le contenu paléontologique (microfaune et nannoplancton) est uniforme et il se caractérise par les associations des foraminifères à *Globotruncana bulloides* et des nannofossiles à *Kamptenerius magnificus*.

La recherche lithologique et minéralo-pétrographique met en évidence le fait que le lithofaciès est :

— biomicritique ou tuffitique pour l'unité d'Audia ;

— calcaire-micritique, argileux-gréseux pour les secteurs occidentaux et micritique argileux et gréseux à ciment basal pour les secteurs orientaux de l'unité médio-marginale.

Selon les données mentionnées ci-dessus il résulte que. sur le fond général calcaire-micritique de ce niveau stratigraphique du Sénonien inférieur, apparaissent aussi par rapport à la source du matériel détritique une grande variété de types de roches, sans remarquer toutefois des changements radicaux dans le géochimisme des eaux du bassin de sédimentation. Ce fait est confirmé par l'uniformité du contenu paléontologique, c'est-à-dire par la présence d'une même association dans tous les échantillons prélevés de ce paquet de roches.

L'association microfaunique comprend également les espèces : *Marginotruncana coronata* (Bolli), *M. marginata* (Reuss), *Globotruncana bulloides* Vogler, *G. minnaeana* (d'Orbigny) attribuées au Santonien (Pessagno, 1967, Ion, 1983) : cette association caractérise le Santonien moyen-supérieur. On remarque l'espèce *Kamptnerius magnificus* Deflandre, fossile-index pour le Santonien, dans l'association nannoplanctonique.

En différentes zones de l'unité médio-marginale on a identifié une séquence de passage au Campanien inférieur, représentée par des marnes à intercalations de siltites, caractérisée par la présence simultanée des espèces *Kamptnerius magnificus* Deflandre et *Ceratolithoides aculeus* (Stradner) Prins et Sissing. On peut souligner que selon la littérature de spécialité, le moment de l'apparition de la nannofossile *Ceratolithoides aculeus* (Stradner) Prins et Sissing est situé à la limite Santonien/Campanien inférieur, l'espèce étant caractéristique pour la partie inférieure du Campanien.

La microfaune, plus pauvre que la précédente, correspond à l'association à Heterohelicidae et *Globotruncanita stuartiformis*, où la participation des foraminifères benthiques (agglutinants et calcaires) est plus évidente.

On peut donc conclure que les dépôts du Sénonien inférieur développés dans les zones étudiées, appartiennent aux unités d'Audia, médio-marginale et externe.

La partie supérieure du niveau basal du Coniacien aussi bien que le segment inférieur du Santonien semblent être le plus souvent absents, soit à cause de la non-déposition, soit à cause de la tectonique très forte qui affecte les dépôts. Les seuls intervalles stratigraphiques constamment rencontrés dans nos recherches appartiennent au Coniacien inférieur (la partie inférieure du paquet caractérisé par l'association à *Dicarinella concavata concavata* et *Marthasterites furcatus*) et au Santonien moyen-supérieur (représenté par l'association à *Globotruncana bulloides* et *Kamptnerius magnificus*).

Considérations sédimentologiques et paléogéographiques

L'étude minéralo-pétrographique des roches du Sénonien inférieur des unités d'Audia, médio-marginale et externe du flysch met en évidence la présence prédominante des dépôts marno-calcaires à intercalation de tufs, tuffites et d'une manière subordonnée des silicolites et localement apparaissent des arénites.

La couleur des dépôts est surtout grise, souvent rouge-brique et plus rarement verte. Le matériel cinéritique très abondant dans la séquence du Coniacien-Santonien provient de deux types d'effusion, à savoir : rhyodacitiques à la partie inférieure et andésites vers la partie supérieure. Des tuffites de la moitié inférieure de l'intervalle comportent deux générations de cendre, une primaire, sédimentée en bassin et une autre secondaire, resédimentée en bassin. Une autre caractéristique des dépôts du Sénonien inférieur c'est que dans la sous-unité médiane, à la base de la séquence, sont logées des arénites lithiques massives, polygènes, contenant aussi des roches effusives andésitiques et des verres.

Du point de vue paléontologique, la succession faisant l'objet de la présente note peut être caractérisée ainsi :

— la faune très rarement signalée, représentée généralement par des Inocerames, Ammonites et Echinodermes, n'a été rencontrée ni dans les échantillons prélevés des affleurements, ni dans les carottes ;

— la microfaune est généralement pauvre et son état de conservation n'est pas toujours satisfaisante. Elle contient des spicules de spongiaires, des foraminifères pélagiques, des radiolaires et des foraminifères benthiques calcaires et arénitiques, les derniers rencontrés particulièrement dans la sous-unité médiane ;

— le nannoplancton se remarque notamment du point de vue de la qualité, par rapport aux séquences du Sénonien supérieur ;

— le groupe Incertae Sedis est mieux représenté sans avoir la grande variété de formes et la fréquence rencontrée dans les dépôts synchrones de la plate-forme moesienne.

Les débris fossiles des dépôts du Sénonien inférieur des trois unités étudiées du flysch des Carpathes Orientales caractérisent des dépôts marins de faciès carbonaté.

Selon Cussey et al. (1977), les foraminifères benthiques imperforés et ceux arénacés se développent dans la zone de la plate-forme interne des domaines intertidal et subtidal, ainsi que dans le domaine interne de barrière de la plate-forme externe. Très différents de ceux-ci, les foraminifères benthiques, perforés se développent de préférence dans la plate-forme externe sur le fond élevé du domaine marin ouvert. Les foraminifères pélagiques, tout comme les radiolaires et les spongiaires, étant cosmopolites, sont rencontrés dans les deux types de plate-forme, mais spécialement dans les limites de la plate-forme externe, dans le domaine marin ouvert, peu profond (100 à 500 m) ou profond (1000 à 5000 m) aussi bien que sur la pente de la plate-forme externe.

Hyman (1940) fait mention que les spicules de spongiraies sont cosmopolites du point de vue de la température, apparaissant autant dans les eaux tropicales que polaires, mais plus fréquents dans les premières.

Enfin, le nannoplancton calcaire souvent improprement dénommé les „coccolithes" se développe dans des dépôts de mer ouverte, allant de petites profondeurs jusqu'aux grandes profondeurs.

Quant à l'énergie cynétique spécifique à l'eau où se développent les groupes d'organismes qu'on vient de mentionner, les auteurs indiquent une faible énergie propre aux milieux de dépôts, de bassin, local, de plate-forme externe ou bien de talus bassinal. Le hydrodynamisme est redevable aux courants de turbidité ou aux glissements, s'agissant d'une turbidité distale et respectivement proximale.

Chilinger (1958) suggère l'origine maréique des courants marins et explique, à leur aide, la forte oxygénation des eaux, le transport et l'accumulation des spicules des spongiaires (le phénomène de tanatocénose). Il montre que les dépôts des spicules sont associés aux eaux relativement froides, troubles, à salinité normale, mais que le grand développement des spongiaires n'est lié que sporadiquement aux régions ayant une activité volcanique intense, c'est-à-dire la concentration en silices de l'eau marine à la suite du volcanisme ne représente pas un facteur de contrôle.

En corroborant les observations de laboratoire avec les données de littérature on peut conclure que les dépôts du Sénonien inférieur des unités d'Audia, médio-marginale et externe du flysch des Carpathes Orientales sont des dépôts marins relativement peu profonds (environ 400 m), développés dans le domaine ouvert du bassin ou éventuellement sur la partie externe de la plate-forme dans des eaux froides, oxygénées, à énergie cynétique faible, à salinité normale. Le bassin a été fortement affecté par une activité volcanique probablement à la suite de la phase d'orogenèse subhercynienne. Le pH des eaux varie en des limites restreintes de 6,6 à 7,2. Le fond du bassin était probablement boueux et les eaux intensivement troublées par des courants de turbidité ou par l'action des vaques. Les conditions bionomiques ont été généralement peu favorables au développement de la vie et surtout de la vie au profondeur (les organismes benthiques). Le plancton montre une variété et une fréquence plus grandes, mais il n'est pas également distribué verticalement, indiquant des connections intermittentes avec le large du bassin par l'intermédiaire des vagues.

Considérations générales sur les dépôts du Sénonien inférieur des Carpathes Orientales

L'orogenèse subhercynienne s'est matérialisée d'une manière accusée dans le géosynclinal des Carpathes Orientales, tant par une activité volcanique spéciale dont les témoins sont rencontrés en grand nombre dans les dépôts du Turonien et du Sénonien inférieur, que par les mouvements épirogénétiques négatifs, affectant ce bassin de sédimentation.

Dans les dépôts crétacés et paléogènes du secteur méridional de l'unité centrale (Ion, 1983) il y a une série de lacunes stratigraphiques affectant la succession du Sénonien (couloir de Vlădeni-Nord, secteur de Sinca). Parfois (couloir de Vlădeni-Sud), l'interruption de la sédimentation à la suite du soulèvement de la zone a affecté presque l'entière séquence (supérieure).

Dans les unités internes de flysch (Neagu, 1968 ; Tocorjescu, 1963 ; Avram, 1967) on a constaté l'existence d'une interruption de sédimentation pendant l'intervalle du Coniacien supérieur-Santonien moyen.

La lacune est située au niveau du Coniacien et partiellement du Santonien dans les secteurs méridionaux de l'unité d'Audia (Bratu, 1966), tandis que dans le secteur septentrional (région de Găinești), les recherches effectuées montrent que le Sénonien inférieur est représenté par le Coniacien basal et inférieur et par le Santonien supérieur, séparés par une lacune.

La situation de la sous-unité médiane (région de Covasna) est similaire. On n'a pas rencontré dans la sous-unité intermédiaire des termes inférieurs du Coniacien.

La stratigraphie de l'unité externe a été étudiée dans les demi-fenêtres de Bistrița (Mirăuță, Mirăuță, 1964) et de Putna-Vrancea (la présente note). Dans la première demi-fenêtre, les recherches effectuées mettent un point d'interrogation portant sur le Sénonien inférieur, tandis que la seconde, elles permettent seulement de contourner le Coniacien inférieur.

En synthétisant les conclusions susmentionnées, il s'ensuit que l'intervalle stratigraphique correspondant au Sénonien inférieur du flysch des Carpathes Orientales a été fortement affecté par l'orogenèse subhercynienne. Les dépôts du Coniacien et du Santonien existent tels quels, pas en succession complète, mais à de multiples interruptions situées aux divers niveaux de la colonne stratigraphique. Les seules séquences lithofaciales, présentes presque constamment, appartiennent au Coniacien inférieur et au Santonien supérieur.

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BIOSTRATIGRAPHY OF THE SILURIAN AND DEVONIAN IN THE MOLDAVIAN AND MOESIAN PLATFORMS (ROMANIA)

BY

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The Moldavian Platform, in the northeast, and the Moesian Platform in the south of Romania, represent Foreland units of the Romanian Carpathians. The deep drillings carried out in the last 10 years through these structural units have supplied a rich and valuable paleontologic and lithologic material, which led to the clearing up of the deep structure geology.

The macro- and microfauna studies as well as the palynological ones regarding the Moldavian Platform were performed by Macarovici (1949, 1956, 1962, 1963, 1965, 1971), Beju, Dăneț (1962), Dăneț (1963), Beju (1971), Patrulius, Iordan (1974), Iliescu (1974), Paraschiv, Muțiu (1974), Iordan (1975).

The macrofauna of the Moesian Platform was studied by Iordan (in Răileanu et al., 1965, 1966, 1967 ; Iordan, 1967, 1971, 1972, 1975, 1977, 1981), Muțiu (in Paraschiv, Muțiu, 1975) ; the phytoplankton and spores by Beju (Venkatachala, Beju, 1961, 1962 ; Beju, 1964, 1967, 1971, 1972) and Iliescu (1971, 1976) ; the microfauna by Dăneț (1964 ; Năstăseanu, 1967).

This paper deals with the present stage of knowledge of the Silurian and Devonian biostratigraphy in the above mentioned units (Fig.).

Silurian

Moldavian Platform

This structural unit is situated in the northeast of Romania and corresponds, from geographic viewpoint, to the Moldavian Plateau. From the geological point of view it represents the southwestern ending of the East-European Platform. It is conventionally delimited by the Pericarpathian line in the west and by the Fălciu-Bogdana-Plopana Fault in the south (Mutihac, 1972), the boundary of this area thus following the line of the Rădăuți-Crasna localities (Barbu et al., 1969).

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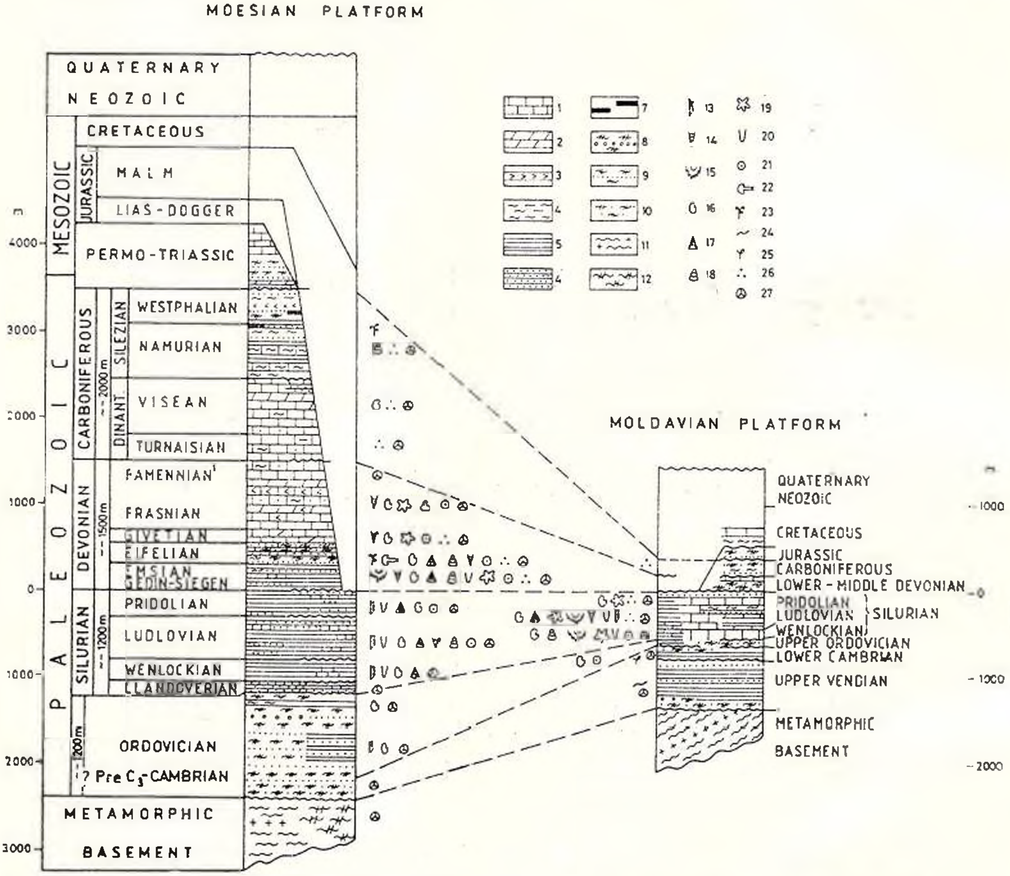

Fig — Correlation of the Paleozoic sequence in the Moesian and Moldavian Platforms.

1, limestone ; 2, dolomite ; 3, anhydrite ; 4, marl ; 5, argillite ; 6, siltite ; 7, coal ; 8, microconglomerate ; 9, sandstone ; 10, quartzite ; 11, crystalline schists ; 12, greenschists ; 13, graptolites ; 14, tentaculites ; 15, trilobites ; 16, brachiopods ; 17, bivalves ; 18, gastropods ; 19, corals ; 20, cephalopods ; 21, crinoids ; 22, placodermi ; 23, plants ; 24, Vendotaenia ; 25, Sabellidites ; 26, microfauna ; 27, palynology.

The Moldavian Platform had a typical cratonic geological evolution and acted like a rigid plate, low dipping to the SW. The pre-Cadomian basement, built up of plagioclase paragneiss and migmatites protruded by pegmatites (Giuşcă et al., 1967) supports a sedimentary cover which includes several sedimentation cycles from Vendian to Neogene.

The Silurian overlies the Vendian-Cambro-Ordovician gritty-shaly lower complex and is overlain by Cretaceous (Patrulius, Iordan, 1974). According to the most recent data, two facies have been identified : 1) the graptolite shales facies in the northwestern part (west of Siret river) and 2) the calcareous shelly-fauna facies in the rest of the platform. The rapid change of facies takes place throughout the important fault (seismically evidenced) which passes east of Fălticeni and Rădăuți localities and is continued by the Rava-Ruska Fault on the territory of Soviet Union (Paraschiv, Paraschiv, 1978).

(1) The graptolite shale facies was intercepted by boreholes 49, 50, 51 Rădăuți and 85 Botoșana and it overlies orthoquartzites and microgritty argillites with bedded graywackes, possibly Cambrian in age by correlation with the Subcarpathian Ukraine. The graptolites reported by Iordan (in Paraschiv, Mutiu, 1974) belong to the species : *Saetograptus colonus* (Barr.), *Bohemograptus bohemicus* (Barr.), *Neodiversograptus nilssoni* (Barr.) which attest a Ludlovian age ; the layers with *Tentaculites ornatus* Sow. were also identified. Chitinozoans reported by Beju and Dăneț (1962) : *Ancyrochitina ancyrea* Eis., *A. moldavica* Beju-Dăneț, *Conochitina elegans* Eis., *Angochitina capillata* Eis., *Lagenochitina baltica* Eis., *L. prussica* Eis., *Sphaerochitina sphaerocephala* Eis., associated with hydrozoan tubes (*Palaeotuba* Eisen) and scolecodonts (*Arabelites* sp., Oenotides sp.) point to Ludlovian and possibly Wenlockian.

(2) The calcareous shelly-fauna facies is the most widespread in the Moldavian Platform. It was identified in the Deleni, 3601, 3602 Todireni-Botoșani, 3501, 3502 Nicolina-Iași, 1 Popești, 23.301 Bătrînești, Hudești, Liteni, 90 Vorona, 91 Lespezi, 80 Preutești, 86 Bosancea boreholes. It is composed and limestones and blackish marly-limestones interbedded with black or bluish-whitish argillaceous marls, black and black-greenish argillites and siltites. According to faunal assemblages we identified the rock sequence of the Wenlockian and Ludlovian.

The Wenlockian, reported from 25.301 Bătrînești borehole is proved by an assemblage consisting predominantly of brachiopods and subordinately of trilobites, gastropods, corals, bryozoans, crinoids, graptolites. The author identified here (Iordan, 1975) : *Eoplectodonta* aff. *transversalis* (Wahl.), *E.* aff. *sowerbiana* (Barr.), *Leptagonia* aff. *joachimiana* Havl., *L. vellerosa* Havl., *Dubioleptina expulsa* (Barr.), *Leptaena rhomboidalis* (Wahl.), *Strophonella* (S.) *euglipha* (Dalm.), *Strophochonetes cingulatus* (Lindstr.), *Atrypa reticularis* aff. *orbicularis* Sow., *Phacops* aff. *fecundus* Barr. All these species were quoted from the Wenlockian in Podolia and Czechoslovakia. In the Bosancea, Lespezi, Preutești and Popești boreholes, the chitinozoans associated with debris of graptolites, trilobites, scolecodonts and *Tentaculites ornatus* ; in the 90 Vorona borehole, the ostracods and conodonts with debris of coral, gastropods and crinoids proved a Middle Silurian age, respectively Wenlockian-Ludlovian (Dăneț, 1963).

The Ludlovian was identified in the Deleni, Todireni-Botoșani, Nicolina-Iași, Bătrînești, Hudești boreholes. The macrofaunal assemblage consists mainly of brachiopods followed in order by ostracods, trilobites, gastropods, corals, tentaculites, bivalves and bryozoans. Among the species identified (Macarovici, 1949—1971 ; Iordan, 1975) it is worth mentioning : *Delthyris elevatus* Dalm., *D. magnus* (Kozl.), *Protochone-*

tes ludloviensis Muir-Wood, *P. dniestrensis* (Kozl.), *Isorthis* aff. *crasa* (Lindstr.), *Mesodouvillina costatula* (Barr.), *Tentaculites ornatus* Sow., *Calymene* aff. *blumenbachii* Brong., *Encrinurus* (E.) *punctatus* (Wahl.), *Leperditia tyraica* Schm., *L. phaseolus* His., *Platiceras* aff. *fecundus* Pern., *Pterinea reticulata* His., *Favosites forbesi* Edw., *Cyathophyllum cyathophylloides* Ryder. This macrofauna is characteristic of the Ludlovian in the Podolia and the Baltic Sea area. The palynological assemblage consists predominantly of chitinozoans and subordinately of spores and acritarchs. Beju (in Macarovici et al., 1963) identified : *Ancyrochitina fragilis* Eis., *A. longicolla* Eis., *Conochitina cornulus* Eis., *Rhabdochitina magna* Eis., *Desmochitina cingulata* Eis., *A. minor* Eis., *Baltisphaeridium* sp., *Leiotriletes* sp., *Ambitisporites* sp. These species were quoted from the Ludlovian of the Baltic Sea area.

The Pridolian was attested in the Hudeşti borehole on the basis of brachiopods, corals, hydrozoans, ostracods and chitinozoans.

Moesian Platform

The Moesian Platform is situated in the south of Romania and corresponds geographically to the Romanian Plain and to the South Dobrogea. From the geological point of view this area corresponds to the northern part of the structural geological unit bordered by the folded Carpathian system to the north and the Balkanides to the south. It is conventionally delimited in the north by a major fault line — Pericarpathian Line ; in the NE by the North Dobrogean Orogen ; in the east by the shore of the Black Sea and in the south by the Danube river. Its basement consists of crystalline schists in the amphibolite-epidote facies, in the west (Optaşi, Oporelu, Balş boreholes) and of the Greenschist Formation in the east, representing an extension of those that outcrop in Central Dobrogea (Silistraru, Bordeiul Verde, Ianca-Berlescu, Tăndărei boreholes). The very thick sedimentary cover (about 7 km) contains a Cambro-Ordovician to Tertiary rock sequence. These deposits exhibit some deformations as a result of the tectogenesis from the surrounding geosynclines. A complex system of faults which delimited some uplifted and sunken compartments and major crustal faults which border the Moesian Platform (e.g. Peceneaga-Camena Fault) or Intra-Moesian faults (Călăraşi-Fierbinţi Fault) had an important role in the evolution and the formation of the present-day aspect of the Carpathian chain (Săndulescu, 1980).

The Silurian lies transgressively over the greenschists and the Ordovician in the east and over the crystalline schists in the west, and supports the Devonian — either conformably or unconformably with sedimentary gap — or it is overlain even by Mesozoic in places. The lithologic facies is represented by the "graptolite shale" one, classical in the lower part and mixed in the upper part of the Silurian. Recently, in the southwestern part of the Moesian Platform, the shelly-fauna facies has been discovered (Iordan et al., 1981). The Wenlockian, Ludlovian and Pridolian series were identified on the basis of graptolites (Iordan, 1981).

The Llandoverian was attested in the Gîrla Mare borehole based on the palynologic assemblage.

The Wenlockian was identified in the Bordeiul Verde, Țăndărei and Ianca-Berlescu boreholes in the east of the platform and in the 3810 Cucueți, 1111 Gîrla Mare boreholes in the west. From borehole 1052 Țăndărei were reported the insectus, centrifugus and murchisoni zones (Lower Wenlockian) on the basis of species : *Monograptus priodon* (Bronn), *M. pseudocultellus* Bouč., *Monoclimacis vomerina vomerina* Nich., *Retiolites geinitzianus* Barr., *Barrandeograptus pulchellus* (Tullb.), *Pristiograptus praedubius* (Bouč.), *Cyrtograptus murchisoni* Carr. (Iordan, 1972, 1981 ; Iordan, Rickards, 1975). From Bordeiul Verde borehole, Murgeanu and Spasov (1968) reported the species : *Monograptus firmus* Bouč., *M. priodon* (Bronn), *Pristiograptus praedubius* (Bouč.) which attest the presence of the firmus Zone from the Lower Wenlockian. The presence of the radians and lundgreni zones (Upper Wenlockian) is attested in the 2803 Ianca-Berlescu borehole by the assemblage : *Plectograptus* cf. *praemacilentus* Bouč. et Münch., *Monograptus flemingii* (Salt.), *Pristiograptus dubius* (Suess), *P. pseudodubius* (Bouč.), *Monoclimacis flumendosae* (Gort.), *Cyrtograptus lundgreni* Tullb., *C. l. gracilis* Bouč., *C. trilleri* Eisel, *Butovicella migrans* (Barr.), *Cardiola* sp. cf. *C. interrupta* Sow., *"Orthoceras'* spp. (Iordan, Rickards, 1971). *Cyrtograptus* sp. is quoted in the western part of the platform in the Cucueți borehole (Paraschiv, 1974). From Gîrla Mare borehole is reported the shelly-fauna facies including brachiopods, spores, chitinozoans, acritarchs and scolecodonts. The following species were identified : *Lissostrophia cooperi* Ams., *Isorthis clivosa* Walms., *Morinorhynchus* cf. *orbignyi* (Dav.), *Leptaena* cf. *rhomboidalis* (Wahl.), *Atrypa* aff. *reticularis* Lin., *Ambitisporites* cf. *ovitus* Hoff., *Veryhachium trispinosum* Eis., *Acanthodiacrodium* sp., *Conochitina gordonensis* Cr., *Eumicites serula* Taug. (Iordan et al., 1981).

The Ludlovian has the largest extension in the Moesian Platform. The Lower Ludlovian was identified in the Tuzla-Costinești, 5083 Mangalia, Călărași and Țăndărei boreholes (Grigoraș, 1956 ; Răileanu et al., 1967 ; Iordan, 1972, 1981 ; Rickards, Iordan, 1975) in the east and in the Optași, Negreni, Cucueți, Iancu Jianu, Făurești, Leu, Balș, Capu Dealului, Gîrla Mare, Oprișor boreholes in the western part of the platform (Paraschiv, 1974 ; Paraschiv, Muțiu, 1974 ; Iordan et al., 1981). The graptolite assemblages, very rich in the east and very poor and sporadic in the west, attest the presence of the nilssoni-scanicus and incipiens zones : *Holoretiolites* (Balticograptus) *balticus* Eis., *Plectograptus macilentus* (Törnq.), *Monograptus uncinatus* Tullb., *"M." incipiens* Wood, *Saetograptus colonus* (Barr.), *S. chimaera* (Barr.), *Bohemograptus bohemicus* (Barr.), *Lobograptus scanicus* (Tullb.), *Neodiversograptus nilssoni* (Barr.), *Euryzone tuboides* Pern., *"Orthoceras"* cf. *primaevum* (Forbes), *Parakionoceras* sp.

The Upper Ludlovian was identified in the Călărași, Zăvoaia and Țăndărei boreholes on the basis of the small juvenile bivalves, flattened orthocone cephalopods, scarce graptolites, small brachiopods, columnalia of crinoids (Iordan, 1981) : *Pristiograptus* cf. *dubius* (Suess), *P. grigoraşii* Iordan, *Bohemograptus bohemicus* (Barr.), *Linograptus posthumus* (Richt.), *Dualina* aff. *fidellis* Barr., *Lunulacardium* cf. *evolvens* Barr., *"Cardiola" insolita* Barr., *"Orthoceras"* cf. *primaevum* (Forb.). In the Gîrla Mare and Oprişor boreholes, situated in the SW of the platform, in shelly-fauna facies were identified : *Fardenia* cf. *wienukovi* (Kozl.), *Howellella* cf. *bragensis* (Wen.), *Triplasma formosum* (Prantl.), *Pisocrinus* sp., *Tentaculites* sp., *Ctenodonta* sp., spores, chitinozoans and acritarchs characteristic of the Ludlovian.

The Pridolian was identified in the Călăraşi, Zăvoaia, Ţăndărei, Gîrla Mare and Oprişor boreholes where continuity of sedimentation from the Ludlovian to the Devonian is established. The fauna assemblage from the first three boreholes consists of graptolites (ultimus-formosus Zone), juvenile bivalves, orthocones, cephalopods, rare brachiopods, trilobites and crinoids : *Monograptus* ex gr. *formosus* Bouč., M. sp. A, M. sp. 1, *Saetograptus* cf. *rarus* (Tel.), *Linograptus posthumus* (Richt.), *Cardiolita* cf. *bohemica* (Barr.), *C.* cf. *fortis* (Barr.), *"Cardiola" insolita* Barr., *Lunulacardium undulatum* Barr., *"Orthoceras" vertebratum* (Barr.), *Geisonoceras* cf. *rivale* (Barr.), *Orbiculoidea* sp. (Iordan, 1977, 1979, 1981). In the Gîrla Mare and Oprişor boreholes the assemblage consists of brachiopods, trilobites, tentaculites, corals, ostracods, spores, chitinozoans, acritarchs and scolecodonts : *Mesodouvillina subinterstrialis* Kozl., *Acaste* cf. *dayiana* Richt., *Tentaculites* cf. *ornatus* Sow., *Favosites gotlandicus* Lam., *Leyotriletes simplex* Naum., *Lagenochitina elegans* Beju, *Micrhystridium stellatum* Deff. (Iordan et al., in press).

Devonian

Moesian Platform

The Devonian rocks overlie the Silurian ones in uninterrupted sequence (Călăraşi, Zăvoaia, Gîrla Mare, Oprişor) or disconformably (5083 Mangalia, Ianca-Berlescu). They are overlain without sedimentary break or facial change by Lower Carboniferous carbonate rocks (Călăraşi, Smirna, Urziceni, Periş, Şoldanu), on swells by unconformable Upper Carboniferous (65 Dobreni) or Permo-Trias or even by Upper Jurassic (Mangalia, Ţăndărei, Zăvoaia, Bordeiul Verde, Ianca-Berlescu). The Devonian sequence includes an argillitic facies at the base, a gritty one in the middle part and a carbonate-lagoonal one in the upper part. On the basis of faunal assemblage of brachiopods, trilobites, tentaculites, bivalves etc. reported from 5082 Mangalia and 2381 Călăraşi boreholes (Iordan, 1981) and of palynological assemblages (Beju, 1971, 1972 ; Iliescu, 1971) the following stages have been pointed out in the Devonian cover of the Moesian Platform : Gedinnian-Siegenian, Siegenian-Emsian, Eifelian, Givetian, Frasnian, Famennian.

The Gedinnian-Siegenian is proved only in the 5082 Mangalia borehole by the assemblage : *Tentaculites gyracanthus* (Eat.), *T. straeleni* Maill., *T. ornatus* Sow., *Nowakia acuaria* Richt., *Multiconus macarovicii*

Iordan, *Comura (Kayserops) kochi* (Kay.), *Schuchertella euzona* (Fuchs), *Chonetes omaliana* (Kon.), *Delthyris dumontianus* (Kon.), *D. infans* Dahm., *Nuculites bisulcatus* (Dahm.), *Sinuitina* sp., *Beyrichia roemeri* Kays., columnalia of crinoids, spores.

The Siegenian-Emsian is attested in the Mangalia, Călăraşi, Zăvoaia, Smirna, Girla Mare and Oprişor boreholes by the assemblage : *Prolationus praelongus* Ljasch., *Volynites velaini* (Mun.-Ch.), *Pilletina asiatica* (Vern.), *P. hammerschmidti* (Roem.), *P. pectinata* (Roem.), *Pseudocryphaeus prostellans* (Richt.), *P. asteriferus* Haas, *Parahomalonotus gervilei rumaniana* Iordan, *Burmeisteria răileanui* Iordan, *Dipleura fornix* Haas, *Leptostrophia index* Havl., *Schellwienella umbraculum* (Schl.), *Fimbrispirifer trigeri* (Vern.), *Grammysia mangalica* Iordan, *Grammysioidea inaequalis calatica* Iordan, *Paracyclas rugosa* (Goldf.), *Orthonota triplicata* Fuchs, *Nuculites elipticus* (Maur.), gastropods, cephalopods, corals, crinoids, spores, acritarchs, chitinozoans, conodonts.

The Eifelian in gritty facies was intercepted in the Mangalia, Călăraşi and Smirna boreholes. Very characteristic here are the land plant — psilophytes — and the placodermi fishes and subordinately brachiopods, tentaculites and crinoids : *Pseudosporochnus krejcii* Pot. et Bern., *Aneurophyton germanicum* Kr. et Wey., *Calamophyton primaevum* Kr. et Wey., *Hyenia* sp., *Mucrospirifer mucronatus* (Conr.), *Nowakia maureri* Zag., *Homoctenus hanusi* (Bouč. et Prantl), *Cupressocrinus crassus* Gold., *?Wijdeaspis* sp. Obr., *Lunaspis broilii* Gross, *Drepanaspis* sp., *Eurypterus* sp., ostracods, spores, conodonts *(Icriodus elevatus, I. nodosus).*

The Givetian marks the beginning of the carbonate-evaporitic facies and is accounted for by : *Tentaculites bellulus potomacensis* Pros., *T. conicus* Roem., *Devonochonetes scitulus* (Hall), *Punctatrypa nalivkini* Havl., *Fimbrispirifer venustus* (Hall), crinoids, corals, ostracods, foraminifers, spores in the Mangalia, Călăraşi and Smirna boreholes.

The Frasnian is proved by the following tentaculite and brachiopod species : *Homoctenus krestovnikovi* Ljasch., *Dicricoconus* sp. A, *Tentaculites* sp. F ex gr. *T. straeleni* Maill., *Chonetes rowei* Cl. et Sw., *Athyris nuculoidea* Coop., *Mucrospirifer bouchardi* (Murch.), *Spinocyrtia martianoffi* (Stock.) identified in the carbonate-anhidritic-bituminous facies in the Mangalia, Călăraşi, Smirna, Comana, Negru Vodă and Viroaga boreholes.

The Famennian is pointed out by microfauna, such as : *Parathurmanina crassa* Lip., *P. suleimanovi stellata* Lip., *Archaeosphaera crassa* Lip., *Orthonella tarnoviensis* Bil. et Gol., *Ellania poyarkovi* Bil. et Gol. (260 Ciureşti, Paraschiv et al., 1977) and recently by macrofauna (brachiopods) and conodonts, too.

The Gîrla Mare and Oprişor boreholes are the only ones situated in the southwestern part of the Moesian Platform, where the Devonian rocks contain a macrofaunal assemblage besides microfaunal and palynoprotistologic assemblages. We mention here brachiopods, tentaculites, trilobites, bivalves, corals, crinoids, eurypterids, plants. Except for the above mentioned ones, from several other boreholes in the west the Devonian was reported only on the basis of microfauna and palynological assemblages ; Beju (1971, 1972) distinguished the D_{1-3} biozones (Chilii, Balş, Iancu Jianu, Bibeşti, Dîrvari, Rîmeşti, Bîrla, Negreni, Poiana, Izvoru, Mitrofani, Făureşti, Strîmbeni, Comana).

Conclusions

The Silurian is widespread both in the Moldavian Platform and the Moesian Platform ; the Devonian was identified in the Moesian Platform and in the Pre-Dobrogean Depression.

The facies of the Silurian is of the "graptolite shales" classical and mixed facies, prevalent in the Moesian Platform, and a calcareous shelly-fauna facies predominant in the Moldavian Platform. The Devonian displays three facies : an argillitic one in the lower part (Gedinnian-Emsian) ; a gritty one in the middle part (Eifelian) and a carbonate-evaporitic facies at the upper part (Givetian-Famennian).

On the basis of the graptolite species, the following biozones and series of the Silurian have been identified : insectus, centrifugus and murchisoni zones (Lower Wenlockian) ; radians and lundgreni zones (Upper Wenlockian) ; nilssoni-scanicus and incipiens zones (Lower Ludlovian) ; bohemicus Zone (Upper Ludlovian) ; ultimus-formosus Zone (Pridolian). Neritic fauna assemblages of Silurian shelly-fauna facies from the Moldavian Platform and from the Devonian of the Moesian Platform led to the identification of the series from the Wenlockian to the Famennian. The trilobites, tentaculites, brachiopods and bivalves have an important role in the Devonian detailed age dating and in the biostratigraphic correlations.

The faunal and microfloristic assemblages in the Moldavian Platform are very similar to and comparable with the ones from the East-European Platform (Podolia) confirming the relations between them. The Devonian trilobite assemblage is very similar to the one from Bithinia (Turkey) ; the tentaculites, brachiopods, bivalves assemblages have a larger distribution being identified in the East-European Platform and in the Rhenan and Hercynian facies in the rest of the world.

As far as the sedimentation conditions are concerned, we should emphasize the reducing environment characteristic of an Euxinic basin for the "graptolite shales" facies and a neritic realm in the shallow water basin for the shelly-fauna facies.

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NANNOPLANKTON ZONES IN THE PALEOGENE AND NEOGENE DEPOSITS OF THE TRANSYLVANIAN BASIN

BY

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Researches on the nannoplankton content of the Tertiary deposits in the Transylvanian Basin were started by Popescu and Gheţa (1972) and were followed by Mészáros and Ianoliu (1973).

In the following years several papers on this subject were published by Mészáros et al. in Cluj-Napoca and Gheţa et al. in Bucharest.

To these one can add Martini and Moisescu's work (1974). All these studies had in view the establishing of the stratigraphic sequence of the Paleogene and Neogene deposits by means of nannoplankton in the north-west of the Transylvanian Basin.

Paleogene Deposits

Within the research work the whole Paleogene sequence of beds was sampled in several areas, such as the town of Cluj-Napoca and its surroundings, Leghia, the eastern border of the Meseş Mts. east of the town of Jibou, etc.

Nannoplankton was yielded by all deposits formed under normal salinity waters or brackish waters.

The Paleogene sequence of deposits in this chronological order and its nannoplankton content will be presented further on (Mészáros et al., 1979).

— The lower variegated complex represents a series of continental sediments devoid of nannoplankton.

— The lower marine series was studied at Luna de Sus, Leghia (west of Cluj-Napoca) and at Agrij (east of Meseş).

— The marly-limestone horizon with *Anomia* and lower gypsum does not contain nannoplankton. The special salinity conditions during the formation of gypsum did not favour the existence of these organisms.

— In the *Gryphaea eszterházyi* horizon (or the lower molluscan marls) the nannoplankton assemblages were found between the *Eupatagus hajnaldy* level and the *Gryphaea eszterházyi* level proper, situated

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towards the upper part of the horizon. These deposits are of Upper Lutetian age, representing the upper part of the NP 15 Zone and the lower part of the NP 16 Zone (after Martini).

— The *Nummilites perforatus* horizon consists of a marly suite where the following species were determined : *Reticulofenestra umbilica*, *R. hampdensis*, *Pemma rotundum*, etc. This assemblage belongs to the Reticulofenestra umbilica Bukry and Bramlette Zone, and NP 16 Zone of Martini, respectively.

— The molluscan marl and limestone horizon (upper molluscan marls) belong to the Reticulofenestra umbilica Bukry and Bramlette, the NP 16 Zone of Martini, respectively.

— The ostreid marl and sandy shale horizon (Mortănuşa Marls) in its lower thirds belongs to the Discoaster saipanensis Subzone of the Reticulofenestra umbilica Bukry and Bramlette Zone, respectively, to NP 17 Zone of Martini, namely the Discoaster saipanensis Zone.

In the median part of these deposits, species of *Chiasmolithus oamaruensis* become more frequent beside the erupting forms of *Reticulofenestra*. This assemblage belongs to the NP 18 — Chiasmolithus oamaruensis Zone.

The last deposits of Mortănuşa Marl thus belong to the NP 19 — Ismolithus recurvus Zone (Mészáros et al., 1979).

— The lower coarse limestone horizon (the Leghia Limestone). One could not prepare substances for the study of nannoplankton in this horizon.

— The upper variegated complex (Turbuta Beds) is a series of continental sediments and consequently devoid of nannoplankton.

— The upper marine series.

— The *Anomia* marly-limestone and upper gypsum horizon and the upper limestone horizon (Cluj Limestone) is very poor in nannoplankton.

— The *Nummulites fabianii* marl horizon, west of Cluj, contains species of : *Sphenolithus pseudoradians*, *Sp. predistentus*, *Istmolithus recurvus*, etc. These deposits belong to the Sphenolithus pseudoradians Zone.

— The bryozoan marl (Brebi Marls) is very rich in nannoplankton. Within the first two thirds, the following species are more frequent : *Reticulofenestra umbilica*, *Lanternitus minutus*, *Istmolithus recurvus*, *Sphenolithus pseudoradians*, *Transversopontis pseudoradians*, *Zygrablithus bijugatus*, etc. These deposits belong to the Sphenolithus pseudoradians Zone.

In the uppermost part of the bryozoan marls (Brebi Marls) at Mera and the Hoia Limestone at Cluj were found : *Ericsonia subdisticha*, *Cyclicargolithus floridanus*, *Helicosphaera reticulata*, *Transversopontis obliquipons*, etc. These deposits of the upper third of the bryozoan marls and the Hoia Limestone belong to NP 21 — the Ericsonia subdisticha Zone and NP 22 — the Helicosphaera reticulata Zone.

Martini and Moisescu (1974) also considered these deposits to belong to NP 22.

— The Mera Beds (Ciocmani Beds). At last, at Cormeniş within the Ciocmani Beds there were identified : *Cyclicargolithus floridanus, Cyclococcolithus formosus, Sphenolithus predistentus*, etc. This assemblage pleads for NP 28 — the Sphenolithus predistentus Zone.

— The Bizuşa marly-limestone beds overlie the Ciocmani Beds in the northern part of the Transylvanian Basin ; it also belongs to NP 23 — the Sphenolithus predistenthus Zone.

— The Ileanda shale Beds, which were studied at Cormeniş, Creaca, contains : *Reticulofenestra lockeri, Cyclicargolithus floridanus, Discolithina latelliptica, Sphenolithus distentus*, etc. These deposits belong to NP 24 — the Sphenolithus distenthus Zone. A similar assemblage was identified in the basal part of the Buzaş Beds, overlying the Ileanda Shales.

— The Cetate Beds, in the Var village, east of Jibou, in the shale from the basal Cetate Sandstone, there was identified an assemblage of : *Sphenolithus ciperoensis, Sph. pacificus, Cyclicargolithus floridanus*, etc. This assemblage belongs to NP 25 — the Sphenolithus ciperoensis Zone and is considered Lower Egerian in age.

— The Zimbor Beds are the last formation in the Paleogene sequence which contains nannoplankton. In the Almaşu Valley, at Cîlgău village, Suraru (1969, 1970) described a marine molluscan fauna at this level. From these deposits the following nannoplankton species were identified : *Discolithina dessueta, D. pygmea, Discoaster broweri, Cyclicargolithus floridanus*, etc. This assemblage also belongs to the NP 25 Zone.

Neogene Deposits

In a sequence at Rohia-Fîntînele (south of Tg. Lăpuş) we succeeded to establish the Oligocene/Miocene boundary by means of nannoplankton within the Vima Beds. Within these beds NP 25, NN 1, and more or less even NN 2 zones were identified. Up to now here is the only place where *Triquetrorhabdulus carinatus* was found as a proof of the NN 1 Zone (Mészáros, Ghergari, 1979). Though the lithologic sequence apparently seems to continue and include even the Chechiş Beds, these ones and partially the Chechiş Beds were erroded before the emplacement of the Hida Beds, as Rusu (1977) also pointed out. The NP 25/NN 1 limit is supposed to be a little earlier than the limit established according to foraminifera.

In order to plot the Oligocene/Miocene boundary by means of nannoplankton, we picked up specimens on a very detailed network to sample the Buzaş Beds, at Poiana Blenchii sequence. Neither in the marl nor in the clay beds was the nannoplankton found out. The absolute lack of it is probably due to the decalcification that altered the remains of mollusca, too.

— In the clay intercalations of the Coruş Beds no specific form of nannoplankton was found.

— The Chechiş Beds were examined in several finding points along the Almaş Valley. In these deposits the presence of *Helico-*

sphaera ampliaperta and *Sphenolithus belemnos* was pointed out. The Chechiş Beds would represent the NN 3 Zone and partially the NN 2 Zone (Mészáros et al., 1977). This association looks very much like that of the Chechiş Beds but it might include the bottom of the NN 4 Zone, too. The presence of the NN 4 Zone (Giurgeşti) was established by Gheţa, Popescu (1975) at the top of the Hida Beds.

— The marls from the bottom of the Dej Tuffs and even the level of the tuffs in the Hoia Hill (Mészáros et al., 1977, 1978), Coasta Marc (Nicorici et al., 1979), Apoldu de Sus (Mészáros et al., 1977), Giurgeşti (Gheţa, Popescu, 1975) were examined. The study of all those sequences pointed out the presence of the NN 5 Zone with *Sphenolithus heteromorphus*.

In Apoldu de Sus at the level of clays with gypsum lenses as well as at those of clays and marls overlying the gypsum we still find the NN 5 Zone. Dumitrică, Gheţa and Popescu (1975) fixed the limit between the NN 5 and NN 6 zones in the middle of the evaporite horizon. Dumtirică, Gheţa and Popescu (1975) separated the evaporite horizon from the Cîmpia Beds, then that with *Radiolaria* and the marls with *Spiratella*, considering all of them as part of the NN 6 Zone. This zone was kept apart in the sequence of the Apoldu de Sus within the marls with *Velapertina* (NN 7 Zone).

— Along the same sequence the NN 8 Zone was separated as the bottom of the Sarmatian age.

As the Miocene sea became sweeter, later nannoplankton associations were not to be found.

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PALYNOLOGY OF THE TERTIARY DEPOSITS IN THE PANNONIAN DEPRESSION, ROMANIA

BY

MARIA MOLDOVANU[1]

Introduction

As there is already a rich documentary material on the Pannonian Depression, the presentation, even in a simple form, of the geology of this unit in this paper would be inopportune, the more so as the present paper refers only to the hydrocarbon exploration.

Located west of Romania and representing only its eastern part, the Pannonian Depression is a geological unit of interest particularly for hydrocarbons, proving completely its insistent geological investigation, as well as the biostratigraphic papers, among which the present paper has to be registered. Its eastern limit is given by a deep fracture which would follow approximately the Carei-Oradea-Timişoara direction (Fig. 1). The Pannonian Depression basement is like that of the Transylvanian Depression, so that it may be similarly interpreted. Thus, one may consider that the crystalline schists encountered by drilling in the Carei region, where they underlie unfolded deposits assigned to the Paleogene, belong to the Pannonian basement.

The sedimentary formations belong to the Triassic and Jurassic in the central section, to the Cretaceous, especially Upper, as relicts of the Transylvanian sedimentation, to the Paleogene in northern and southern sectors, and particularly to the Neogene, which covers the whole depression with its upper terms.

The Neogene forms the proper filling of the depression, and it is represented by the Burdigalian-Helvetian (locally), Badenian, Sarmatian and Pliocene.

On the whole, the Neogene deposits have a monotonous clayey-sandy facies, and at their upper part they represent the Pannonian facies of the Sarmato-Pliocene.

Some details on the Pannonian Depression sedimentation are offered by the adjacent zones (Borod and Beiuş depressions, etc.) where

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the respective deposits present various facies and are very fossiliferous.

In point of tectonics, the Pannonian Depression presents similar features to the Transylvanian Depression. The unit is divided by major faults in three large areas : (1) southern, approximately south of the Mureș River, where most of the wells crossed the Pliocene and the Upper Miocene, isolated the Upper Cretaceous, (2) central comprised approximately between the Mureș River to the south and the parallel of Oradea town to the north, where the sedimentary pile, much raised, is completed by Mesozoic deposits, as a continuation of the Bihor Autochthon, and (3) northern, where Cretaceous and Paleogene deposits are mentioned in flysch facies below the Miocene-Pliocene.

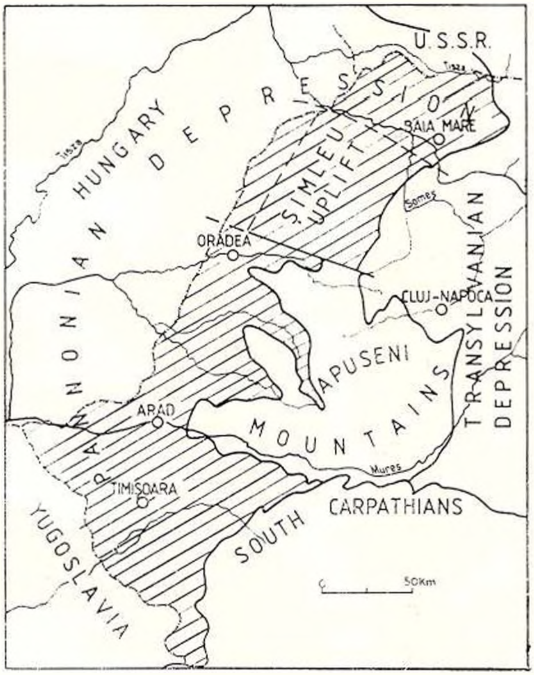

Fig. 1 — Sketch map of the Pannonian Depression (Romanian area).

Palynological Zonation

By synthesizing the previous data (Balteș, 1966, 1971 ; Balteș, Moldovanu, 1981) and those presented in this paper, it results that the distribution of the Neogene microflora allows a detailed biostratigraphic determination and a sharp geochronological dating for the Tertiary deposits, bringing further data on the region stratigraphy (Fig. 2). The global stratigraphic interpretation of the microvegetal content has led to establishing the palyno-stratigraphic zones and subzones.

1. The Wetzeliella clathrata-Wetzeliella ovalis palynological Zone (PP_3), which renders evident the Middle-Upper Eocene (Upper Lute-

tian-Lower Priabonian), present in the Pîşcolt, Curtuişeni, Nisipeni and Turnu structures.

2. The Tytthodiscus-Echinatipollenites palynological Zone (PN_1—PN_2), a comprehensive zone delimiting the Upper Burdigalian-Helvetian.

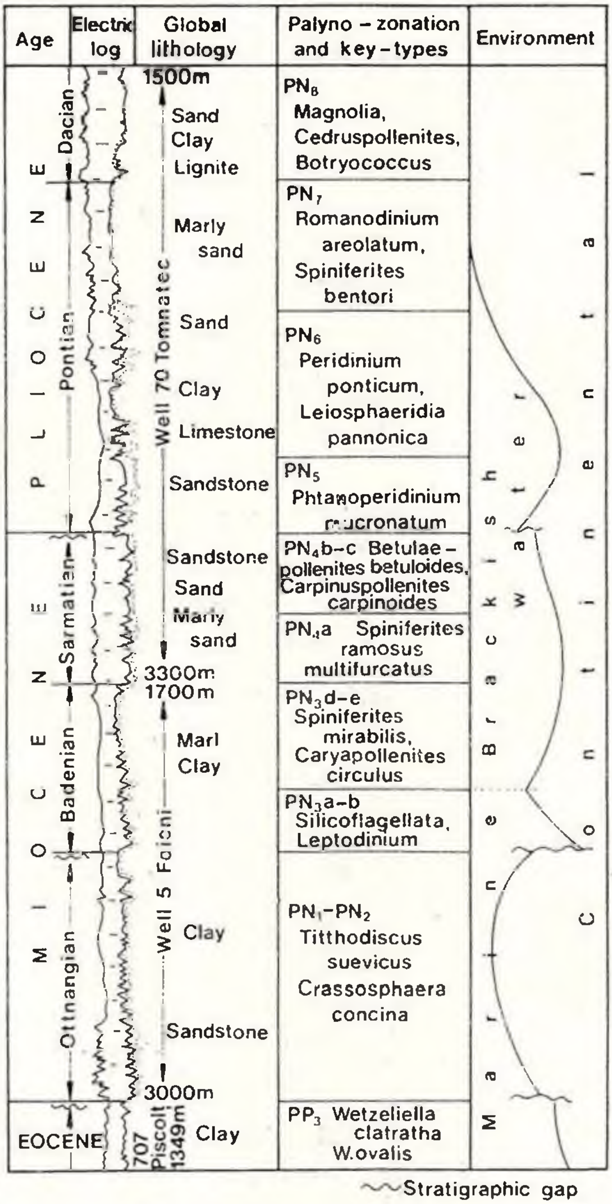

Fig. 2 — Stratigraphic palyno-zonation of the Tertiary deposits in the Pannonian Depression (Romanian area).

3. The Nemathosphaeropsis-Svarbardella palynological Zone (PN_3) is characteristic of the Badenian ; the qualitative and quantitative ratio between the continental and marine elements, leading to the separation of two palynological subzones :

(a) the Silicoflagellata-Leptodinium parviscabratum palynological Subzone (PN_{3a-b}), detected within a gritty-marly-clayey tuffaceous complex of the northern sector structures. This subzone may define the Lower Badenian (Langhian) in the region.

(b) the Spiniferites mirabilis-Caryapollerites circulus palynological Subzone (PN_{3d-e}), largely spread, was identified in a group of argilites, sandstones, argillaceous marls, polygeneous conglomerates and microcrystalline dolomites.

4. The Pterodinium-Neogenisporites neogenicus palynological Zone (PN_4) is characteristic of the Sarmatian.

The study of the chronological development of the phytoplankton has disclosed some palynological discontinuities, determining the separation of two subzones :

(a) the Spiniferites ramosus multifurcatum palynological Subzone (PN_{4a}), situated at the base of the Sarmatian deposits (Volhynian) following — as a rule — the Badenian deposits.

(b) the Betulaepollenites betuloides-Carpinuspollenites carpinoides palynological Subzone (PN_{4b-e}), defined as one of the richest Neogene associations and characterized by the disappearance of the last Badenian elements, while the brackish phytoplankton suddenly decreases up to its disappearance. The lower limit, difficult to trace, was comprehensively diagnosed, as belonging to the Lower-Middle Sarmatian.

5. The Phthanoperidinium mucronatum-Ornatisporites reticulatus palynological Zone (PN_5) marks the beginning of the Pliocene sedimentation in lacustrian-fluvial facies, with a typical palynofacies also pointing out an important palynological gap, which separates it from the Sarmatian.

6. The Peridinium ponticum-Leiosphaeridia pannonica palynological Zone (PN_6) unconformable over the Miocene deposits, and in some situations over the Meotian ones, defines the Lower Pontian spread on a large sedimentary area, and represented by diagenised fossiliferous pelitic deposits.

7. The Romanodinium-Compositopollenites palynological Zone (PN_7) represents the Middle-Upper Pontian. It was reported both in the southern area and in the central-northern one.

8. The Magnoliaceae-Cedruspollenites palynological Zone (PN_8) characterizes the last Pliocene sequence, palynologically identified at the Dacian level (s.l.). The palynological content is almost exclusively continental ; the phytoplankton belonging to the fresh water algae. The coaly facies, sometimes lignite, favours the preservation of a rich palynological association, which begins to outline the vegetal assemblage of the Pliocene and Late Pleistocene.

Paleoecological Considerations

The reconstruction, according to possibilities, of the relations between the way of life of various floristical categories represented palynologically and the sedimentary basin, evinces some particular aspects of the environments, all the more so as some of them are investigated for economic purposes.

Thus, the Eocene association shows a short marine episode, but with deposition in an open basin and a rather intense biotic activity. The Eocene association recalls the calcareous series in the North-West Transylvania.

In the Neogene series of the Pannonian Depression new particular paleofloristic phenomena were noticed, apt to be used in complex geological investigations, in order to characterize the sedimentary basins and their evolution as well as to establish conditions of preserving and accumulating the organic matter. The assimilation of the artificial palynological types, used in the stratigraphic studies, to those similar of the natural system of classification made it possible to interpret ecologically the Neogene palynological assemblages, leading to paleogeographical, paleoclimatic and paleoecologic considerations. The Lower Miocene segment (PN$_1$-PN$_2$ Zone) of some sectors in the investigated area, proves either short sedimentary episodes, or Burdigalian-Helvetian remains. The brackish phytoplankton and the terrestrial material show a restrictive nature of deposition, representing probably the completion of the first Miocene sedimentary cycle resumed concomitantly with the Badenian transgression.

In the Lower Badenian segment (PN$_{3a-b}$ Subzone), the preponderance of the phytoplankton, Silicoflagellates and Dinoflagellates, evinces the marine type of sedimentation. The tropical and sub-tropical continental types diminish considerably (the palm trees disappear) giving way to the temperate types, particularly to the coniferae and falling leaves trees of the piedmont, thus resulting a mixture of temperate semi-arid and sub-tropical elements. The phytoplankton explosion in the Lower Badenian marks a vast transgressive stage to be found in the intracarpathian and extracarpathian basins — an aspect conferring to the respective level the value of a better geological guide mark.

Owing to the presence — though subordinate — of the marine phytoplankton, with specific forms, associated with continental elements, the Upper Badenian deposits (PN$_{3d-e}$ Subzone) suggest a sedimentation of a littoral-inner domain with preponderent transition to the marginal-littoral one (brackish-lagoonal facies). The temperate climate is re-established and continues amplifying. Towards the upper part of the Badenian, as well as in the Lower-Middle Sarmatian, the Miocene basin begins to differ in more and more limited zones, having a brackish-lacustrian feature, with specific fauna and flora : probably these limited basins are going to get closed to the end of the Miocene, the sedimentation getting a lacustrian-fluvial feature in several points, generated by a rich terrigeneous supply and with a paleontological content of a continental nature.

This basin starts its evolution slowly, even at the same time with the Lower Sarmatian sedimentation (the PN_{4a} Subzone) marking a slight change in the ecological coordinates, but which amplifies towards its central part. Even the fact that the uppermost part of the Sarmatian is absent in most of the Pannonian Depression, shows an exondation stage, felt on larger surfaces, following the continuous regression of the Miocene basin. The palynofacies, separated in the Sarmatian interval, are preponderantly temperate-continental ; they tend to become exclusive upperwards, replacing the marine ones, which prove the phenomenon of continual water refreshing and the idea of the presence of a mixed vegetal cover (coniferae and falling leaves trees) having an average altitude (600—1.000 m) in the neighbouring zones of the sedimentary basin. In fact, the presence of the PN_{4b-c} Subzone emphasizes what it was mentioned above, since in its microfloristic composition one may see the rather sudden diminution of the phytoplankton fraction up to its disappearance, as well as the basic change of the rest of the microflora, determined by the installation of a basin with obvious lacustrian-fluvial features. This type of basin was developed with approximately the same features, but more accentuated posterior to the Sarmatian exondation phase. Considered as Meotian (according to the Dacic molasse features) these deposits become practically dominated by brackish-lacustrian algae, and especially of fresh water. In this way the microvegetal content of the PN_7 Zone, rich in continental elements, many of them having a large spreading area, confirm the amplifying of the sedimentation conditions of marginal type, without, however, excluding even the possibility of a deltaic-continental sedimentation at the upper part of the segment defined by this association.

In the Lower Pontian, the explosion of the phytoplankton is proof of a new invasion of the land by the water of an ample brackish basin. In this case, the transgression had a slow and gradual character which explains the PN_6 Zone occurrence on a considerable stratigraphic thickness. Paleoclimatically, the Pontian is characterized by a slight unconformity of some xerophyte elements, but being included in the ensemble of temperate types, does not suggest main changes in latitude, but rather in altitude, having a local feature.

The presence in high percentages of entomophytic pollen of large size, with high specific weight and a reduced spreading area shows the campestrian origin of low altitude of the respective vegetal source.

The last terms of the Pliocene sequence are represented by the Upper Pontian and Dacian deposits having a coaly palynofacies. At this level, the sedimentary areas are quite different, getting the aspect of some small low altitude depressions, progressively filled with terrigenous material, displaced by the hydrographic network (continental facies). The temperate nature with arctic climate accents becomes typical, defined by the domination of the gymnospermes, angyospermes and monocotiledonates flora. Thus, the vegetal cover may be reconstituted with a forested area of a high average altitude (1,100—1,500 m) partially divided by lacustrian-fluvial sedimentary basins, with a great accumulation of vegetal material and lignite formation.

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STRATIGRAPHIE DES DÉPOTS ALBIENS DE LA PLATE-FORME MOESIENNE (SECTEUR ROUMAIN)

PAR

RADU MUȚIU[1]

La présence des gisements de pétrole dans les dépôts albiens de la partie centrale et septentrionale de la plate-forme moesienne a réclamé des recherches géologiques avancées. La première attestation paléontologique de cet étage dans le domaine moesien date de 1962 et revient à Dimitrova, qui dans la zone de la localité de Russe fait connaître la présence de l'Albien moyen ou éventuellement de la partie supérieure de l'Albien inférieur. Les recherches des géologues qui ont étudié cet étage dans le domaine moesien ont un caractère plus ou moins historique étant limitées à une zone restreinte. On doit tenir compte que l'âge des dépôts considérés est restreint seulement à l'Albien moyen. Des études biostratigraphiques systématiques concernant la faune de l'Albien développé dans les limites du secteur roumain de la plate-forme moesienne appartiennent à Muțiu (1969, 1972, 1982 a, b). Le même auteur (1969) a publié une synthèse sur l'évolution stratigraphique et paléontologique des formations albiennes s'appuyant sur les Ammonites couvrant l'Albien entier. Ensuite Muțiu (1972) a complété son étude paléontologique et stratigraphique sur les dépôts albiens de la plate-forme moesienne, en complétant l'inventaire paléontologique de ceux-ci et détaillant l'échelle stratigraphique (Fig. 1). Muțiu (1982) développe

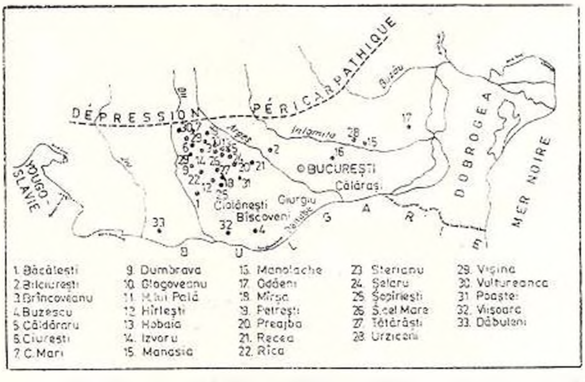

Fig. 1 — Localisation des points fossilifères.

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enfin ses contributions concernant la stratigraphie de l'Albien par l'étude des échantillons prélevés des carottages continus réalisés dans la région de Călăraşi-Chiciu. C'est toujours en 1982 que l'auteur décrit systématiquement des Ammonites hétéromorphes de l'unité moesienne : l'étude fait connaître quatre espèces et une sous-espèce nouvelle. Cet ouvrage enrichit l'inventaire ammonitique des dépôts albiens de la région investiguée (conformément aux données géologiques actuelles). Par l'abondance de la faune et par l'épaisseur des sédiments, cette région tient une importance particulière pour l'étude de l'étage Albien de l'Europe de l'Est.

Inventaire paléontologique

Albien inférieur. On a inventorié la suivante faune dans la région étudiée : *Hypacanthoplites* cf. *milletoides* Casey (région de Tătărăşti), *Hypacanthoplites milletianus* (d'Orb.) (Ciolănesti), *Hypacanthoplites trivialis* Breist. (Bîscoveni), *Nautilus albensis* d'Orb. (Ciolănesti), *Leymeriella (L.) regularis* (Brug.), *Leymeriella tardefurcata tardefurcata* (Leym.), *Leymeriella (L.) tardefurcata intermedia* Spath, *Leymeriella* aff. *revilli* Jacob, *Beudanticeras* aff. *dupinianum* (d'Orb.), *Douvilleiceras Douvilleiceras mammillatum* Schl., *Cleoniceras (C.) morgani* Spath, *Plicatula inflata* Sow., *Grammatodon carinata* (Sow.), *Plicatula gurgitis* Pict., *Aporhais ebrayi* Lor. (faunes de la région de Călăraşi) ainsi que des représentants du genre *Aucellina* (*Aucellina caucasica* Buch.) et *Aucellina aptiensis* Pomp. et *Avellana lacrima* d'Orb., *Solarium monoliferum* Mich. etc.

Leymériellien. Les premiers niveaux de l'Albien sont connus dans les bassins classiques de l'Ouest de l'Europe (Breistroffer, 1947). Ils ont été retrouvés dans la plate-forme moesienne. Ces dépôts sont marqués par l'apparition brusque et l'épanouissement très rapide du genre *Leymeriella* et par la disparition du genre *Acanthoplites*. La plus ancienne association des Ammonites appartenant à l'Albien inférieur identifiée dans la plate-forme moesienne (région de Ciolăneşti-Bîscoveni) et provenant des grès glauconieux durs est formée par : *Hypacanthoplites* cf. *milletoides* Casey, *Hypacanthoplites milletoides* d'Orb. et *Hypacanthoplites trivialis* Breistrof. Elle apparaît habituellement depuis la sous-zone Milletoides jusqu'à la zone Mammillatum (Casey, 1965). On doit souligner que cette dernière forme a été considérée comme espèce index pour la sous-zone Milletianus insérée entre les sous-zones Shrameni et Regularis (Spath, 1923). Les mêmes Ammonites (excepté *H.* cf. *milletoides*) ont été aussi signalés en Dobrogea du Sud étant situés stratigraphiquement au-dessous de la zone *Leymeriella tardefurcata*, éventuellement dans la partie supérieure de la zone Regularis (Chiriac, 1981). La partie supérieure du Leymériellien est caractérisée par la présence de l'espèce *Leymeriella regularis* (Călăraşi), espèce index de sous-zone de la zone Tardefurcata (Spath, 1941 ; Casey, 1961). On a encore identifié dans d'autres régions parmi lesquelles celle de Dumbrava-Ciolăneşti à côté de *L. regularis* des Ammonites appartenant aux genres *Protanisoceras, Tetragonites, Beudanticeras* associés à un riche mélange des faunes d'Aucellinae et d'autres Lamellibranches. On suppose que cette faune a été remaniée dans l'horizon des grès glauconieux à coprolites (type Seimeni) de l'Albien moyen. L'espèce *Leymeriella tardefurcata* a été aussi rencontrée dans les vraies lumachelles à *Leymeriella*, dans des grès glauconieux durs à serpoulidae de la région de Săruleşti.

Douvilléicératien. Les Ammonites qui caractérisent la zone Mammillatum ont été découverts dans la région de Călăraşi et sont représentés par : *Douvilleiceras mammillatum* Sch., *Beudanticeras* aff. *dupinianum* d'Orb. et *Cleoniceras (C.) margani* Spath. Dans les dépôts albiens de la partie sud-est d'Angleterre, espèce omniprésente, *Douvilleiceras mammillatum* fait son apparition dans la sous-zone Kitchini et devenant plus fréquente dans les sous-zones Floridum et Puzosianum (Casey, 1962). Selon le même auteur *Cleoniceras (C.) morgani* Spath a été identifié dans la zone Tardefurcata, sous-zone Regularis et dans la zone Mammillatum, sous-zone Kitchini.

Albien moyen. Les dépôts qui définissent l'Albien moyen de la plate-forme moesienne roumaine se caractérisent par l'apparition des Ammonites des genres : *Anapholites* et *Hoplites* dont dérivent des *Dimorphoplites* et *Euhoplites* etc. ; on a rarement identifié les genres *Protanisoceras* et *Oxytropidoceras*. Dans la région de Giurgiu on a reconnu les fossiles suivantes : *Neohibolites minimus* (List.), *Cymatoceras nekerianus* (Pict.), *Douvilleiceras* sp., *Rhynconella tripartita* Pict., *Terebratula dutempleana* d'Orb., *Plicatula gurgitis* Pict., *Plicatula inflata* Sow., *Ostrea papyracea* Sin., *Inoceramus concentricus* Park., *Puzosia quensteti* (Parona et Bonarelli). *Hamites maximus* Sow., *Prohelicoceras* aff. *subcatenatum* Spath, *Anahoplites intermedius* Spath. *Anahoplites planus fittoni* (d'Arch), *Anahoplites planus discoideus* Spath. Dans les marnes de la région de Cîrleşti on a trouvé : *Anahoplites intermedius* Spath et *Anahoplites praecox* Spath. tandis que des calcaires verdâtres de la région de Corbeanca proviennent *Dimorphoplites niobe* Spath. trans. *D. pinax* Spath, *Hamites* cf. *tenuis* Sow. étant reconnue dans les marnes de Viişoara. Dans l'horizon à coprolites (type Seimeni) on a déterminé : *Hoplites* cf. *dentatus* Sow., et *Hoplites dentatus densicostata* Spath.

Hoplitien. On a prouvé l'existence des dépôts hoplitiens par des espèces des zones : *Hoplites* cf. *dentatus* Sow., Spath (1941) Breistroffer (1947), Owen (1971), en association avec *Hoplites dentatus densicostata* Spath rencontrée dans les grès glauconieux à coprolites (partie nord-centrale de la plate-forme moesienne).

Euhoplitien. La caractéristique de ces dépôts est donnée par les caractères des Ammonites *Anahoplites intermedius* Spath et *Anahoplites praecox* Spath considérés en tant que fossiles index pour la sous-zone Intermedius + Praecox (Breistroffer, 1947) et pour la zone Loricatus (Owen, 1971). L'Ammonite *Dimorphoplites niobe* (calcaires de la région de Corbeanca) est considéré par les mêmes auteurs et par Spath (1941) à titre d'espèce index de sous-zone de la zone Loricatus. L'espèce *Anahoplites intermedius* a été également identifiée dans les grès verts de Călăraşi, en association avec Inoceramus, Plicatula, Cuculaea etc. d'où on a prélevé l'Ammonite *Anahoplites planus discoideus. Dimorphoplites cloris* Spath (Giurgiu) est connu dans le bassin anglo-parisien, notamment dans la sous-zone Daviesi (Owen, 1971).

Albien supérieur. La faune albienne supérieure de la plate-forme moesienne diffère de celle de l'Albien moyen par l'apparition des genres : *Stomohamites, Lechites, Idiohamites, Callihoplites, Hysteroceras, Mortoniceras, Stoliczkaia, Mariella, Scaphites ;* les genres *Beudanticeras, Kosmatella et Oxytropidoceras* que nous avons rencontrés dans la base de l'Albien supérieur montent de l'Albien moyen. L'épaisseur de l'Albien supérieur marneux atteind plus de 500 m dans la partie centrale et ne dépasse pas 50 m dans le Nord. On a inventorié les suivantes faunes : *Hamites* cf. *tenuicostatus* Sow., *Hamites* cf. *compresus* Sow., *Hamites maximus rectus* Brow., *Hamites intermedius* Sow. v. *opalina* Sapth, *Hamites intermedius* v. *distincta* Spath, *Hamites (H.) virgulatus* Pict. et Camp., *Hamites subvirgulatus* Spath, ? *Hamites (P.) boucardianus* d'Orb., *Hemiptychoceras gaultinum* (Pict.), *Lechites gaudini* (Pict. et Camp), *Lechites communis* Spath, *Anisoceras (A.) perarmatum* Pict. et Camp., *Idiohamites tuberculatus* (Sow.), *Idiohamites spiniger* (Sow.), *Idiohamites favrinus* (Pict.), *Idiohamites favrinus* (Pict.) var. *robustus* Muţiu, *Hamites (S.) scharpentieri* Pict., *Idiohamites latus* Muţiu, *Idiohamites tenuis* Muţiu, *Idiohamites bicostatus* Muţiu, *Ostligoceras (O.) puzosianum* (d'Orb.), *Hemiptychoceras* sp., *Mariella* (B.) *bergeri* (Brogn.), *Mariella bergeri* var. *crassituberculata* Spath, *Mariella (B.) miliaris* (Pict. et Camp.), *Spaphites (S.) meriani* Pict. et Camp., *Scaphites* cf. *hugardianus* (d'Orb.), *Scaphites simplex* Juk.-Brown, *Beudanticeras beudanti* Brogn., *Mortoniceras inflatum* Sow., *Mortoniceras (P.) pricei* Spath, *Euhoplites alphalautus* Spath, *Kossmatella agassiziana* Pict., *Puzosia mayoriana* d'Orb., *Puzosia subplanulata* (Schl.), *Stoliczkaia dispar* d'Orb. et les Lamellibranches *Inoceramus sulcatus* Park., *Inoceramus subsulcatus* Wilt., *Inoceramus anglicus* Woods, *Inoceramus concentricus* Park., *Plicatula gurgitis* Pict., *Aucellina grypheoides* Sow.

Hystérocératien. Les dépôts hystérocératiens de la plate-forme moesienne (région centrale : Dumbrava, Hîrleşti, Humele, Glogoveanu etc.) sont représentés par les marnes à *Mortoniceras inflatum* Sow., espèce caractéristique pour la partie basale de l'Albien supérieur (fig. 2).

La sous-zone Dipoloceras cristatum a été déterminée dans la plate-forme moesienne dans les marnes à *Hamites maximus rectus* Brow.

La sous-zone Hysteroceras orbignyi est représentée dans la région étudiée par des marnes grisâtres et des calcaires marneux gris, contenant l'espèce index *Hysteroceras orbignyi*, en association avec de nombreuses espèces, telles *Hamites intermedius* Sow. et *Beudanticeras beudanti* Brogn. et une riche faune d'*Inoceramus sulcatus* Park. (région de Rîca-Dumbrava-Urziceni etc.).

Les dépôts correspondant à la sous-zone Hysteroceras varicosum sont représentés par des marnes et calcaires grisâtres à *Hysteroceras varicosum* Sow., espèce index en association avec de nombreux Ammonites caractéristiques : *Mortoniceras (P.) pricei* (Spath) (espèce index de zone — Breistroffer. 1947), *Euhoplites alphalautus* Spath (Dumbrava), *Kosmatela agassiziana* Pict. (Urziceni), *Scaphites simplex* Juk. (Buzescu), *Idiohamites tuberculatus* Sow., *Idiohamites spiniger* Sow. (Glogoveanu), *Hamites* cf. *gaultium* Pict., *Idiohamites scharpentieri* Pict. etc.

		ZONES			SOUS-ZONES	AMMONITES	EPAIS LITHOLOG
ALBIEN	Supérieur	*Stoliczkaia*	*dispar*	*Ostlingocératien*	*Stoliczkaia* *dispar* *Mortoniceras* *perinflatum* (Vrac. sup.)	*Lechites gaudini (Pict et Comp) Anisoceras peramatum Pict. et Comp Mariella bergeri (Brogn) Mariella miliaris (Pict) Ostlingoceras (O) puzosianum (d'Orb.) Scaphites meriani Pict. Puzosia subplanulata (Schlt.) Mariella (M) bergeri crassituberculata (Spath) Stoliczkaia dispar (d'Orb.)*	marnes
		Arrhapoceras	*Substuderi*	*Todrillien*	*Arrhapoceras* *substuderi* *Mariella gresslyi* (Vrac. inf.)	*Stomohamites virgulatus (Brug.) Scaphites cf. hugardianus d'Orb. Lechites communis spath.*	marnes 10-600m
		Mortoniceras	*inflatum*	*Hysteroceratien*	*Mortoniceras inflatum*	*Mortoniceras inflatum (Sow.)*	
					Callihoplites auritus	*Idiohamites desorianus Pict. trans. I turgidus Sow.*	
					Hysteroceras *varicosum*	*Hysteroceras varicosum Sow. Mortoniceras (P) pricei (Spath) Idiohamites tuberculatus Sow. Hemiptycoceras gaultinum Pict Kossmatella agassiziana Pict. Euhoplites alphalautus Spath. Scaphites simplex Juk. Idiohamites charpentieri (Pict) Puzosia mayoriana d'Orb.*	
					Hysteroceras orbignyi	*Hysteroceras orbigny (Spt.) Beudanticeras beudanti Brg. Hamites intermedius Sow.*	
					Dipoloceras cristatum	*Hamites maximus rectus Br. Oxytropidoceras sp.*	
	Moyen	*Euhoplites*	*lautus*	*Euhoplitien*	*Anahoplites* *daviesi* *Euhoplites nitidus*	*Dimorphoplites cloris Spath* *Hamites cf. tenuis Sow.* *Dimorpholites sp.*	
		Euhoplites	*loricatus*		*Euhoplites* *meandrinus* *Mojsisovicsia* *subdelaruei*	*Anahoplites planus planus (Mant.)* *Anchoplites planus discoideus Spt.* *Anchoplites planus fittoni (Arch.)*	marnes calcaires vertes
					Dimorphoplites niobe	*Dimorphoplites niobe Spath trans.* *Dimorphoplites pinax Spath.*	10-50m
					Anahoplites *intermedius*	*Anahoplites intermedius Spath* *Anahoplites pinax Spath.*	
		Hoplites	*dentatus*	*Hoplitien*	*Hoplites (H) spathi* *Lyelliceras lyelli* *Hoplites (I)* *eodentatus*	*Hoplites cf. dentatus Sow.* *Hoplites dentatus Sow. var densicostata Spath*	grès glauconieux calcares
	Inférieur	*Douvilleiceras*	*mammillatum*	*Douvilleiceratien*	*Protohoplites (H)* *puzosianus* *Otohoplites* *roulianus* *Cleoniceras* *floridum* *Sonneratia kitchini*	*Douvilleiceras mammillatum (Sch.)* *Pictetia sp.* *Beudanticeras aff. dupinianum (d'Orb.)* *Cleoniceras (c) morgani Spath*	grès glauconieux
		Leymeriella	*tardefurcata*	*Leymeriellien*	*Leymeriella* *regularis*	*Leymeriella (L) regularis (Brug.) L. tardefurcata Ley Leymeriella (L) tardefurcata densicostata Spath. Leymeriella (L) intermedia Spath. Leymeriella cf. revilli JAC*	10-150 marnes argiles
					Hypacanthoplites *milletioides*	*Hypacanthoplites cf. milletoides Casey. Hypacanthoplites trivalis Breist. Hypacanthoplites miletianus (d'Orb.)*	grès glauconieux
					Farnahamia *farnahamensis*		grès ferrugineux

Fig. 2 — Biostratigraphie de l'Albien de la plate-forme moesienne.

L'existence de la sous-zone Auritus est prouvée par l'Ammonite *Idiohamites desorianus* Pict. trans. *I. turgidus* Sow., dans les marnes de la région de Glogoveanu.

Turrilitoidien (Vraconien inférieur). L'existence des dépôts de cet âge est révélée par les marnes à *Stomohamites virgulatus* Pict. (Hîrleşti), *Scaphites* cf. *hugardianus* (d'Orb.) (Dumbrava) et *Lechites communis* Spath (Selaru) ; la dernière espèce est consignée comme caractéristique pour le Vraconien inférieur de Salazac (Breistroffer, 1947).

Ostlingoceratien (Vraconien supérieur). Les dépôts otlingocératiens sont représentés par des marnes et marno-calcaires gris à *Stoliczkaia dispar* d'Orb., fossile de zone en association avec des Ammonites caractéristiques : *Lechites gaudini* (Pict. et Camp.), *Anisoceras perarmatum* (Sow.) (à Petreşti Glogoveanu), *Mariella bergeri* (Brogn.), *Mariella bergeri crassituberculata* Spath (à Corbii Mari-Glogoveanu). *Mariella miliaris* (Pict. et Camp.) (à Studina). *Ostlingoceras (O.) puzosianum* (d'Orb) (à Vultureanca-Mîrşa-Odăieni), *Scaphites meriani* Pict. et Camp. (à Glogoveanu), *Mariella bergeri* (Brog.), *Stoliczkaia dispar* (d'Orb.) et *Ostlingoceras (O.) puzosianum* (d'Orb.) ; ils sont considérés en tant que fossiles caractéristiques de la zone (Breistroffer, 1947).

Considérations biostratigraphiques

Les dépôts albiens du secteur roumain de la plate-forme moesienne sont constitués en général par un faciès de grès glauconieux (Albien inférieur), par des marno-calcaires et des grès (Albien moyen) et par des marnes (Albien supérieur). La succession complète de cet étage a été prouvée par cinq Ammonites (espèces index des zones) : *Leymeriella tardefurcata* Ley., *Douvilleiceras mammillatum* (Sch.), *Hoplites* cf. *dentatus* Sow., *Mortoniceras inflatum* (Sow.) et *Stoliczkaia dispar* (d'Orb.) et par huit Ammonites des sous-zones index : *Hypacanthoplites* cf. *milletoides* Casey, *Leymeriella regularis* (Brug.), *Anahoplites intermedius* Spath, *Dimorphoplites niobe* Spath, *Hysteroceras orbignyi* (Spath), *Hysteroceras varicosum* (Sow.), *Mortoniceras inflatum* (Sow.), *Stolicskaia dispar* (d'Orb.).

Du fait de l'affinité de la faune de la plate-forme moesienne avec celle des bassins anglo-parisiens on a choisi les sous-zones de Casey (1961) pour l'Albien inférieur, d'Owen (1971) pour l'Albien moyen et de Spath (1941) et Breistroffer (1947) pour l'Albien supérieur, sous-zones utilisées aussi par Chiriac (1981) pour le territoire de la Dobrogea du Sud.

Les dépôts albiens présentent un fort caractère bassinal dans le secteur situé entre le méridien de Gliganu et la ligne de Titu-Croitori-Preajba-Talpa, où la profondeur dépasse parfois 500 m ; à l'Ouest de la ligne mentionnée les dépôts marneux albiens n'excèdent pas 40 m d'épaisseur. Parmi les régions les plus importantes pour l'étude des sous-étages albiens doivent être considérées celle de Călăraşi (sous-étage inférieur), de Giurgiu (sous-étage moyen) et de Dumbrava (sous-étage supérieur). Les grès glauconieux à coprolites de l'Albien moyen (type Seimeni) ont des caractères pétrographiques variés ; à ce niveau la zone de shelf a été séparé en deux sous-zones : l'une externe (Talpa-Preajba-Corbii Mari) représentée lithologiquement par des calcsiltites et des calcaires micritiques recristallisés où on peut remarquer la resédimentation de nombreux bloques de calcaires récifaux barrémiens-aptiens (à Talpa) marquant la proximité de la rive et l'autre interne (Hîrleşti-Glavacioc-Dumbrava) à grès et calcaires glauconieux. A l'Ouest de la zone de shelf (à Ciolăneşti-Recea) apparaît une zone de transition vers le faciès pélagique de l'Albien.

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PALYNOSTRATIGRAPHIE DE QUELQUES FORMATIONS CRISTALLOPHYLLIENNES DES CARPATHES ORIENTALES ROUMAINES

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Les études palynologiques portant sur les formations cristallophylliennes des Carpathes Orientales roumaines sont à présent des préoccupations prioritaires des géologues roumains du fait qu'elles offrent un matériel concret pour comprendre et expliquer la structure et la tectonique compliquées de ces anciennes séquences lithologiques. Nous voulons y faire une synthèse des résultats de nos recherches palynologiques concernant les plus anciennes formations cristallophylliennes des Carpathes Orientales roumaines.

Parmi les plus importants ouvrages palynostratigraphiques rappelons : Iliescu, Mureșan (1972), Onicescu, et al. (1974), Iliescu, Kräutner (1976), Oniceanu, Olaru (1977), Olaru, Oniceanu (1979), 1983).

On essaie actuellement de réaliser des recherches de détail afin d'obtenir des complexes palynologiques ou des associations strictement caractéristiques des unités lithologiques. De cette manière, on envisage d'étudier palynostratigraphiquement chaque niveau lithologique de la succession des formations des Carpathes Orientales.

Nos recherches palynostratigraphiques ont été réalisée surtout dans la zone des monts Rodna et des monts de Bistrița, là où la zone cristalline des Carpathes Orientales est mieux représentée.

Structuralement, le cristallin des Carpathes Orientales comporte un ensemble de nappes de charriage, en constituant le système de nappes bucoviniennes et sub-bucoviniennes qui repose sur un soubassement métamorphique rétromorphe (Săndulescu, 1967, 1972 ; Bercia et al., 1971 ; Bercia et al., 1976 ; Balintoni, 1981). Les nappes de charriage de ce système comportent également des formations paléozoïques épimétamorphiques représentées par des séries lithologiques dissemblables.

Les formations cristallophylliennes en question ont été générées au cours de trois cycles de sédimentation et métamorphisme, à savoir : un cycle précambrien moyen (Protérozoïque supérieur) accompagné d'un métamorphisme régional allant de 650 à 850 m. a., correspondant à l'orogenèse dalslandienne et reconnu dans la série lithologique de

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Bretila-Rarău et dans la partie inférieure de la série de Rebra-Barnar ; un cycle précambrien supérieur (Infracambrien, Vendien, Protérozoïque terminal) jusqu'au Cambrien inférieur ou possible Cambrien moyen, achevé par l'orogenèse baïkalienne (assyntique tardive) accompagné d'un métamorphisme régional de 472 à 550 m.a. A ce cycle correspond les parties moyenne et supérieure de la série de Rebra-Barnar et la série de Tulgheş. Le troisième cycle est celui paléozoïque s'achevant durant l'orogenèse hercynienne par un métamorphisme correspondant problement à la phase sudète, propre aux séries de Repedea et de Tibău (Bercia et al., 1971 ; Iliescu, Kräutner, 1975 ; Bercia et al., 1976).

C'est dans ce cadre géologique et structural des Carpathes Orientales que nos recherches palynostratigraphiques se sont déroulées et à la suite desquelles nous avons essayé de séparer quelques complexes palynologiques caractéristiques.

Complexe palynologique à Kildinella

Ce complexe palynologique a été reconnu dans des roches métamorphiques, représentées, par des gneiss et micaschistes à amphiboles. Ils appartiennent à la série de Bretila-Rarău d'une puissance de 3000 m, étant métamorphisées dans le faciès des amphibolites à almandin, affectés intensément d'un rétromorphisme hercynien. Le complexe palynologique a été obtenu des micaschistes à grenats et des micaschistes quartzeux orienté parallèlement à l'axe de l'anticlinal de Bretila, les échantillons étant prélevés de la vallée de Rusaia (affleurements et travaux miniers). Le contenu palynologique est pauvre (tabl. 1), étant formé principalement de *Kildinella hyperboreica, K. sinica, Protosphaeridium scabridum, P. vermium, P. flexuosum, Synsphaeridium horediforme, S. conglutinatum, Nucellosphaeridium deminatum, N. bellum.* Ce complexe palynologique qui est caractéristique pour le Riphéen comprend aussi d'autres éléments, tel qu'on relève dans le tableau 1, à une distribution stratigraphique plus grande, éléments qui ne peuvent pas caractériser le niveau lithologique dont ils ont été déterminés. A remarquer que les éléments palynologiques prédominants dans ce complexe palynostratigraphique sont les Acritarches appartenant au groupe Sphaeromorphytae. Quelques représentants de ce groupe apparaissent en chaînes *(Archaeomassulina atava, Archaeomassulina pussila)*, particularité qui les rapprochent aux algues marines. Outre Sphaeromorphytae se développent les premiers représentants de Tasmanaceae *(Nucellosphaeridium deminatum* et *N. bellum)* dont l'apparition était remontée au Cambrien. Leur présence dans ce complexe palynologique à côté des éléments typiques riphéens pourrait représenter un argument pour reconsidérer l'âge des premières Tasmanaceae, en envisageant comme vrai l'âge riphéen et non celui cambrien. Ce fait est d'un intérêt particulier pour la palynostratigraphie des dépôts où ont été rencontrés ces représentants.

Suivant les données présentées dans le tableau 1, le complexe palynologique définie l'âge riphéen pour les micaschistes à grenats et les micaschistes quartzitiques analysés, et de ce fait il devient un complexe palynostratigraphique caractérisant les formations cristallines de la série de Bretila-Rarău d'âge ripheen. Là-dessus, on considère qu'elle est la plus ancienne série lithologique des Carpathes Orientales roumaines, ancienneté attestée également par les analyses isotopiques qui ont indiqué un âge de 850 ± 50 m.a. (Mînzatu et al., 1975).

TABLEAU 1

Distribution stratigraphique des taxons du complexe palynologique à Kildinella

Dénomination des taxons	Riphéen	Vendien
Kildinella hyperboreica Tim.		
K. sinica Tim.	——	
Protosphaeridium scabridum Tim.	——	- - -
P. flexuosum' Tim.	——	
P. vermium Tim.	——	
P. densum Tim.	—	——
P. giberosum Tim.	——	——
Turuchanica alara Rud.	——	- - -
T. ternata Tim.	——	- - -
Synsphaeridium conglutinatum Tim.	——	
S. sorediforme Tim.	——	
Nucellosphaeridium deminatum Tim.	——	
N. bellum Tim.	——	
Spumosata prima Naum.	——	——
Archaeomassulina atava Pych.	——	
Balvinella foveolata Shep.	——	
Archaeosacculina atava Pych.	—	
Lophominuscula rugosa Naum.	——	
Favosphaeridium favosum Tim.	——	——

Complexe palynologique à Granomarginata notata

Il est plus riche et mieux représenté que le complexe antérieur. Ce complexe a été séparé des roches constituées surtout de calcaires et dolomies cristallines à minces intercalations de micaschistes et paragneiss. Ces calcaires et dolomies appartiennent à la série de Rebra-Barnar, savoir au complexe médian de celle-ci. La série de Rebra-Barnar a une puissance de plus de 6500 m, sa partie médiane, généralement calcaire et dolomitique, atteingnant 1000 m d'épaisseur. Les échantillons analysés ont été prélevés de 25 affleurements. Cette série a un caractère mésométamorphique (faciès des amphibolites à almandin), étant affectée d'un rétromorphisme hercynien. Le complexe palynologique présenté (tabl. 2) est la synthèse de plusieurs associations déterminées et il est marqué par les éléments comme : *Granomarginata notata, Archaeofavosima miranda, Bronhopsophoaphaera simplex, Lophominuscula orbillata, Archaeosacculina salebrosa* et d'autres. L'âge est vendien (infracambrien) et l'association palynologique beaucoup plus riche en éléments que celles mentionnées (tabl. 2).

TABLEAU 2

Distribution stratigraphique des taxons du complexe palynologique à Granomarginata notata

Dénomination des taxons	Riphéen	Vendien	Cambrien inférieur
Granomarginata notata Naum.		———	
Brochopsophosphaera simplex Pych.		———	
Archaeofovosima miranda Tim.		———	– – – –
Margoporata glabella Lop.		———	
Leiopsophosphaera microrugosa Naum.		———	
Leiopsophosphaera rugosa Naum.		———	
Archaeosacculina salebrosa Naum.	———	———	
Kildinella hyperboreica Tim.	———		
Protosphaeridium flexuosum Tim.	———		
P. tuberculiferum Tim.	———	———	– – – –
P. densum Tim.	———	———	
P. rigidulum Tim.		———	
P. torulosum Tim.		———	
P. asaphum Tim.	———	———	
Turuchanica ternata Tim.	———	– – – –	
Stictosphaeridium implexum Tim.	———	———	– – – –
S. pectinale Tim.	———	———	– – – –
Margominuscula pumila Naum.	———	———	
Lophominuscula prima Naum.	———	———	
Leiominuscula minuta Naum.	———	———	
Margominuscula rugosa Naum.	———	———	
Baltisohaeridium cerinum Volk.			———
B. ornatum Volk.			———
Micrhystridium palidum Volk.			———
M. obscurum Volk.			———
M. brevicornum Jank.			———
Granomarginata prima Naum.			———
Gyratosphaerina aspera Lop.			———
Dyctyodium priscum Kirj et Volk.			———
Leiomarginata simplex Naum.			———

A part les formes citées ci-dessus, il y a aussi des éléments représentés par : *Kildinella hyperboreica, Stictosphaeridium implexum, Protosphaeradium flexuosum, P. tuberculiferum.* Leur présence dans le complexe palynologique en question trouve l'explication dans le remaniement du complexe inférieur de la série de Rebra-Barnar constituée des micaschistes à amphiboles et qui peut être supposé comme synchrone à la série de Bretila-Rarău. Il est d'âge riphéen. De même, le complexe palynologique comporte une association d'éléments plus jeunes d'âge cambrien inférieur parmi lesquels nous citons : *Baltisphaeridium ornatum, B. cerinum, Micrhystridium palidum, M. obscurum, M. brevicornum, Granomarginata prima, Gyratosphaerina aspera, Leiomarginata simplex.*

On peut constater donc que dans le complexe palynostratigraphique déterminé des calcaires et dolomies de la série de Rebra-Barnar il y a une grande variation d'associations palynologiques à éléments d'âge riphéen, vendien et cambrien inferieur. La présence de jeunes éléments cambriens inférieurs nous mène à une nouvelle manière d'interpréter l'âge du complexe palynostratigraphique et implicitement des roches dont il a été déterminé. Par conséquent, il faut envisager ces roches vendiennes-cambriennes inférieures. En raison des éléments palynologiques du complexe palynostratigraphique déterminé, la partie inférieure de cette série serait synchrone à la série de Bretila-Rarău, d'âge riphéen, tandis que la partie médiane plus jeune est d'âge vendien-cambrien inférieur ou bien cambrien inférieur, ce qui porte à une reconsidération de l'âge de cette série. L'âge radiométrique de 549 à 650 m.a. (Mînzatu et al., 1975) confirme cette hypothèse. Les éléments palynologiques analysés appartiennent aux Acritarches des groupes Sphaeromorphytae, Acanthomorphytae *(Baltisphaeridium, Micrhystridium)*, Herkomarphytae *(Cymatiosphaera, Dictyodium)*, à côté de Tasmanaceae *(Stictosphaeridium)*.

Complexe palynologique à Baltisphaeridium cerinum

Ce dernier complexe palynologique a été déterminé des roches épimétamorphiques représentées par des schistes chlorito-séricitеux, quartzitiques et graphiteux, roches principalement terrigènes. On rencontre dans cette succession lithologique de nombreuses intercalations volcanogènes-sédimentaires, des calcaires et des dolomies cristallines. La puissance de ce complexe lithologique excède 4500 m et appartient à la série de Tulgheş largement développée dans les Carpathes Orientales. Cette série a été divisée, par divers géologues, en cinq formations lithologiques.

Le complexe palynologique déterminé est très riche et du fait de sa constitution il n'a pas soulevé des problèmes en ce qui concerne son âge cambrien inférieur. Nous sommes d'avis que *Baltisphaeridium cerinum* typique pour le Cambrien inférieur est en même temps caractéristique pour ce complexe palynologique, auquel s'ajoutent d'autres éléments comme : *Baltisphaeridium ornatum, B. dubium, Leiosphaeridia pilomifera, Concentrica miranda, Dictyodium priscum, Cymatiosphaera favosa, Ovullum saccatum, Synsphaeridium switjasium, Micrhystridium brevicornum*. A part ces éléments (tabl. 3) on doit mentionner certaines formes remaniées des formations d'âge riphéen ou vendien. Ce complexe palynologique provient des schistes quartzito-séricitеux, schistes chlorito-séricitеux et quartzites noires, prélevés de 20 affleurements. On n'a pas réussi à analyser palynologiquement toute la succession lithologique de la série de Tulgheş, mais les résultats que nous présentons mènent à la conclusion que l'âge du complexe palynologique et des roches dont il a éte déterminé est cambrien inférieur, voire cambrien moyen. Ce dernier âge est soutenu par une série d'éléments du complexe palynologique comme *Granomarginata squamacea, Baltisphaeridium compressum, Alliumela baltica, Micrhystridium obscurum* qui ont été rencontrés aussi en des formations d'âge cambrien moyen

TABLEAU 3

Distribution stratigraphique des taxons du complexe palynologique à Baltisphaeridium cerinum

Dénomination des taxons	Riphéen	Vendien	Cambr. inf.	Cambr. moyen
Baltisphaeridium cerinum Volk.			—	
B. ornatum Volk.			—	
B. dubium Volk.			—	
B. primarium Volk.			—	
B. papillosum Volk.			—	
B. compressum Volk.			—	—
Micrhystridium brevicornum Jank.			—	
M. palidum Volk.			—	
M. obscurum Volk			—	—
Synsphaeridium switjasium Kirj.			—	
Leiosphaeridia pylomifera Pask.			—	
Dictyodium priscum Kirj et Volk.			—	
D. bivertense Pask.			—	
Ovulum saccatum Jans.			—	
Granomarginata squamacea Volk.			—	—
Alliumela baltica Vanderflit			—	—
Cymatiosphaera favosa Jank.			—	
Comembranecea Kirj. '			—	
Concentrica miranda Naum.			—	
C. solida Naum.			—	
Favosphaeridium favosum Tim.			—	
Protosphaeridium tuberculiferum Tim.	—	—	—	
Spumiosina cineraria Lop.	—	—	—	
Gyratosphaerina aspera Lop.		—	—	
Lophasphaeridium tentatum Volk.			—	
Stictosphaeridium torulosum Tim.			—	
Dictyopsophosphaera exilis Lop.		—	—	
Trachypsophsphaera exilis Lop.			—	
Uniporata papulifera Lop.			—	
Leiovalia tenera Kirj.		—	—	
Leiosphaeridia subgranulata Kirj.			—	
Archaeoshavosina minuta Naum.			—	
Leiosphaeridia dehisca Pask.			—	
Archaeodiscima bicostata Volk.			—	
Tasmanites piritaensis Posti et Jank.			—	

(Menner et al., 1979). L'âge isotopique de la série de Tulgheş est compris entre 397 et 472 m. a. (Mînzatu et al., 1975), la limite inférieure du Cambrien inférieur étant considérée à 550 m.a., ce qui correspond au niveau des argiles bleues du littoral de la Mer Baltique (Kremp, 1982).

Ainsi, à la suite de la détermination de ce dernier complexe palynologique, une partie de la série de Rebra-Barnar (niveau moyen à calcaires et dolomies) peut être supposée synchrone à la série de Tulgheş (complexes lithologiques Tg_1 et Tg_2) d'âge cambrien inférieur.

A partir des résultats des recherches palynologiques on peut conclure que :

— le complexe palynologique à *Kildinella* est d'âge riphéen et caractérise la série de Bretila-Rarău, la plus ancienne des Carpathes Orientales ;

— le complexe palynologique à *Granomarginata notata* est caractéristique au niveau moyen calcaire et dolomitique de la série de Rebra-Barnar dont l'âge est vendien-cambrien inférieur ou bien cambrien inférieur ;

— le complexe palynologique à *Baltisphaeridium cerinum* définie la série de Tulgheş (niveaux Tg_1 et Tg_2) d'âge cambrien inférieur, possible cambrien moyen ;

— la partie inférieure de la série de Rebra-Barnar semble être du même âge que la série de Bretila d'âge riphéen alors que la partie médiane de la série de Rebra-Barnar peut être équivalente à la série de Tulgheş. A cet égard la position stratigraphique de la série de Rebra-Barnar doit reconsidérée.

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BIOSTRATIGRAPHIC CHARACTERIZATION OF SOME BOUNDARIES IN THE CENOMANIAN-CONIACIAN INTERVAL

BY

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During the recent years the authors of the present paper have been intensely preoccupied with the establishment of the boundaries and subdivisions of the Upper Cretaceous stage in Romania, on the basis of planktonic foraminifera (J. Săndulescu, 1966, 1969 ; J. Ion, 1975, 1976, 1978, 1982, 1983) and macrofauna (ammonites, inocerami) Szász, 1981, 1982a, b, 1983a, in press). At the same time they have tried as well to establish some micro- and macropaleontological zones which can be correlated with the regions of the stratotypes of the respective stages or with other classical regions in Europe or elsewhere. Some results can be useful to improve the standard biochronological scales based on ammonites, inocerami and planktonic foraminifera. This paper is the first attempt in Romania to correlate the biochronological scales based on macrofauna and planktonic foraminifera for the Cenomanian-Coniacian interval (Plates I and II).

Bio- and Chronostratigraphy Based on Ammonites and Inocerami

Albian-Cenomanian boundary. The last zone of the Albian from the classical regions of Europe is the Stoliczkaia dispar Zone. In Romania the assemblage of this zone occurs in situ in some areas of the Carpathians (Dîmbovicioara Basin) (Murgeanu, Patrulius, 1957 ; Patrulius, 1969), and in other regions (South Dobrogea) it is reworked in the basal conglomerate with which the Cenomanian deposits begin. The first ammonite zone known in Romania is the Manticelliceras mantelli Zone which, besides the index species, contains as well other species of *Mantelliceras*, species of the *Hypoturrilites, Hyphoplites, Stoliczkaia (Lamnayella)* genera, etc.

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The species belonging to the two last genera, as well as the species *Neostlingoceras carcitanensis* suggest that the basal part of the Cenomanian is present in the South Dobrogea, too ; between this assemblage of ammonites and the first ammonite assemblage from the Western European Cenomanian therefore, there are no important differences.

Besides the Mantelliceras mantelli Zone, we have accepted as well the Mantelliceras orbignyi Zone (sensu Juignet et al., 1978) for the Lower Cenomanian. Correlations of the respective zones with those from the Western Europe were made in other paper (Szász, 1982b).

The boundary between the Lower and Middle Cenomanian is obvious in the whole Europe, being marked by the disappearance of the species of the genera *Mantelliceras, Hyphoplites, Mariella, Sharpeiceras*, etc., and by the appearance and development of the species of *Acanthoceras, Calycoceras, Euomphaloceras*, etc. For the Middle Cenomanian we have adopted the two zones proposed by Amédro et al. (1978a, b), namely the Acanthoceras rhotomagense Zone and the Acanthoceras jukesbrownei Zone. In Romania there are ammonite assemblages characteristic of both zones, but in two far regions (south of the Babadag Basin for the former biozone, the Hațeg Basin for the latter one).

Between the Middle and Upper Cenomanian there is a continuity of sedimentation in the Hațeg Basin, where the beginning of the Upper Cenomanian is marked by the appearance of the *Eucalycoceras pentagonum* species, together with the characteristic assemblage of the Eucalycoceras pentagonum Zone (sensu Juignet, Kennedy, 1976). For the last Cenomanian zone (Metoicoceras geslinianum Zone, sensu Amedro et al., 1978 a, b), characteristic ammonites were not found yet in Romania.

As concerns the inocerami, during the Cenomanian, at least in Romania, they have a very irregular distribution in time and space and were not used yet for zonations.

Cenomanian-Turonian boundary. It is still a very disputed subject in the world, first of all because in different regions the Turonian begins with various ammonite assemblages. There are also differences among authors as concerns the ammonite genera, by whose appearance the beginning of the Turonian could be marked. So, several genera which have been considered as exclusively Turonian *(Vascoceras, Nigericeras, Thomasites, Jeanrogericeras*, etc.) after some authors (Wright, Kennedy, 1981) already appear during the Cenomanian. Anyhow, a whole ammonite zone (the Neocardioceras juddii Zone) is encompassed either in the Cenomanian (Wright, Kennedy, 1981), or in the Turonian (Hook, Cobban, 1981). From the latest papers concerning the Turonian biochronology based on ammonites, there results that the Mammites nodosoides Zone represents only the upper part of the Lower Turonian ; between this one and the Neocardioceras juddii Zone there is at least one more ammonite zone (Watinoceras coloradoense Zone) in the Western Europe.

In Romania, the only ammonite rich assemblage of the Lower Turonian is known in Maramureș (east of the Borșa Basin), which we attributed (Szász, a, in press) to the Paramammites polymorphum-Choffaticeras pavillieri Assemblage Zone. Taking into account the range of various species, we suppose that this assemblage represents the whole Lower Turonian.

The Middle Turonian in Romania is extremely poor in ammonites (*Romaniceras* cf. *kallesi* in the Babadag Basin) and does not offer a certain basis either for formations or for discussing boundaries between substages.

The Upper Turonian has supplied a richer ammonite assemblage in the Perşani Mts, belonging to the classical Subprionocyclus neptuni Zone. Elsewhere, the presence of the Upper Turonian is certified by *Subprionocyclus neptuni*, *Romaniceras* sp. (Haţeg Basin), or by *Lewesiceras mantelli* and *Tongoboryiceras* cf. *rhodanicum* (Babadag Basin).

The Turonian deposits of Romania are rich in inocerami which in some regions form the only macrofaunal criterium to establish their age. The inocerami assemblages are rather monotonous and there can be frequently identified only two assemblages : the former with *Mytiloides* ex gr. *labiatus* (with all „species" known in Western Europe), corresponding to the lower half of the stage, the latter with *Inoceramus* ex gr. *lamarcki* (with several closely related species) in the upper part of the level. We could not establish a finer zonation within the two assemblages after Kauffman's model used in different papers concerning America or Europe.

In Romania the last certain Turonian assemblage contains small sized inocerami with very widespread species which in the latest zonations of the German authors (Ernst et al., 1983) are placed in the Middle Turonian or in the lower part of the Upper Turonian. In Romania between this assemblage and the Coniacian base there are no other inocerami assemblages and the species used as zone indicators for the topmost Turonian in Germany (*Mytiloides striatoconcentricus*, *Inoceramus waltersdorfensis*) occur in Romania together with Coniacian ammonites or form an assemblage situated in the terminal Turonian-basal Coniacian, and therefore their value as zonal indicators is diminished.

As concerns *the Turonian-Coniacian boundary*, lately a very disputed subject, we have already expressed our point of view (Szász, 1981, 1982a, Szász, b, in press, Szász et al., 1978). We remid only that in Romania there are at least three areas (Babadag Basin, Getic Depression, Perşani Mts) where this boundary can be successfully discussed and where there are certain proofs that the whole Inoceramus schloenbachi Assemblage is of a Lower Coniacian age. At the same level there occur the various species of *Didymotis* which in the Babadag Basin are associated with Coniacian ammonites and in the Perşani Mts occur above the level with *Subprionocyclus*, never within it.

As regards the Coniacian zonation, according to ammonites and the separation of its substages, we mention that the Forresteria (Harleites) petrocoriensis Zone, which is now accepted as the first zone of the Coniacian of the classical zones in France (Willedieu), being equivalent with the whole Lower Coniacian (Kennedy, 1983, in litt.), is well represented in Romania, too ; in the Babadag Basin and in the

Getic Depression the index species occurs, while in the Perşani Mts, *Tissotioides haplophyllus* was found. In the Babadag Basin, even at the base of the Coniacian sequence, *Barroisiceras haberfellneri* occurs. Therefore, the Firresteria (Harleites) petrocoriensis Zone and the Barroisiceras haberfellneri Zone are synchronous or at least both index species appear even at the base of the Coniacian.

Both in the Babadag Basin and in France, over the F. (H.) petrocoriensis Assemblage there follows an assemblage with *Peroniceras* representing the Middle Coniacian (Kennedy, 1983, in litt.). As a zone indicator for this interval we use the *Peroniceras tridorsatum* species.

Among inocerami, the Inoceramus schloenbachi Assemblage is characteristic of the Lower Coniacian and the Inoceramus (Platyceramus) mantelli Assemblage is characteristic of the Middle Coniacian.

For the Upper Coniacian there are no characteristic ammonites, therefore the substage is attested by inocerami (*I. subquadratus, I.* ex gr. *undabundus*). In order to discuss the Coniacian/Santonian boundary we have not enough macropaleontological arguments.

Bio- and Chronostratigraphy Based on Planktonic Foraminifera

Cenomanian. At the Vraconian-Cenomanian boundary the planktonic foraminifera assemblage of the Carpathian regions in Romania, where this boundary could be studied, belongs to the Tethys province. It was so far recorded only in sequences where the Upper Vraconian has characteristic macrofauna. The beds with the fauna of the S. dispar Zone contain the assemblage with *Th. appenninica* and *Th. balernaensis* (the assemblage of the Th. appenninica Zone, cf. def. Dalbiez, 1955, redefined by Bolli, 1957). The following levels which are without a characteristic macrofauna contain *Th. brotzeni, Th. globotruncanoides, Th. greenhornensis, Th.* aff. *appenninica* (geronthic stage ?), *Th. marchigiana, R. montsalvensis.* Recent researches (Salaj, 1980 ; Robaszynsky, 1981 ; Bellier, 1983) have underlined that in the Tethys domain *Th. brotzeni* sometimes appears either above or below the Vraconian-Cenomanian boundary. Thus, for the Carpathian domain in Romania up to the finding of some sequences with characteristic macrofauna above and below this boundary, we can only arbitrarily admit that the appearance of *Th. brotzeni* and *Th. globotruncanoides* marks the beginning of the Cenomanian.

The lower part of the Lower Cenomanian corresponding to the M. mantelli Zone is characterized by the above mentioned assemblage with *Th. brotzeni, Th. globotruncanoides.* Upwards, probably in the interval corresponding to the lower part of the M. orbignyi Zone, *R. montsalvensis* is well represented, together with many species of Thalmanninella. This microfaunal assemblage characterizes the Th. brotzeni/Th. globotruncanoides Zone (cf. def. Lehmann, 1966 emend. Ion, 1978).

The terminal part of the Lower Cenomanian correlated to the final part of the M. orbignyi Zone, is characterized by the appearance of *Th. porthaulti* (= "*R. (Th.) reicheli* not typical form", J. Săndulescu, 1969), marking the Th. porhaulti Subzone (cf. def. Ion, 1983 = s. z. *R. reicheli* not typical, Ion, 1978). In our opinion, the Cn 2b Zone from the terminal part of the Lower Cenomanian of SE France (Porthault, 1974) is also the Th. porthaulti Zone ; its marker species represented by *R.* aff. *reicheli* being a junior synonymy of *Th. porthaulti*. Lately, in the boreal and Tethys domains (Robaszynsky, 1981), *R. reicheli* is admitted to appear in the topmost part of the M. orbignyi Zone. We do not know however if this distribution refers to individuals of typical *R. reicheli*. The same question arises with the individuals of *R. reicheli* and *deeckei* which were quoted as well in other places (Libia, Barr, 1973 ; Austria, Sturm, 1969) below the boundary of the Lower Cenomanian.

Starting from the Middle Cenomanian and during the most part of the Upper Cenomanian, the planktonic foraminiferal assemblages in the Carpathians and North Dobrogea have similar characteristics to those of Crimea, Caucasus and some Alpine basins from Central Europe. In these regions, rotalipores from the *cushmani-turonica* group do not appear during the Middle Cenomanian, but during the Upper Cenomanian, the Middle Cenomanian being characterized only by the appearance of *Th. Reicheli* and *deeckei*.

In Romania, the interval corresponding to the A. rhotomagense Zone is characterized by the Th. reicheli Subzone (cf. def. Ion, 1978) and that corresponding to the A. jukesbrownei Zone, by the Th. deeckei Subzone (cf. def. Ion, 1978).

The Th. deeckei Zone extends also in the base of the Upper Cenomanian, *R. cushmani* appearing only above some beds with *I.* cf. *pictus* and *I. ex. gr. crippsi* and with the reicheli-deeckei Assemblage. In the Carpathians and North Dobrogea *R. cushmani* occurs only in Upper Cenomanian strata (*E. pentagonum* beds), never lower than this substage. In our case, it seems therefore that for the Upper Cenomanian the change of the planktonic assemblage takes place somewhere above its lower boundary. It is clearly marked first by the speciation and development of Rotalipores (the *cushmani-turonica* groups) and by the appearance of *Dicarinella*, then of *Marginotruncana, Helvetoglobotruncana, Whiteinella, Pseudorotalipora, Pseudoticinella, Pseudothalmanninella* genera. This new assemblage forms the R. gr. cushmani turonica Zone (Malapris, Rat, 1961) and it is correlated to the lower part of the E. pentagonum Zone. Below its upper boundary established by the appearance of *W. paradubia*, taxa of *Thalmanninella* with a thorny outline appear.

The interval occupied by the upper part of the E. pentagonum Zone and partly by the M. geslinianum Zone is characterized by the W. paradubia Subzone (cf. def. Ion, 1978), which contains many species of *Whiteinella (paradubia, aprica, alpina)*, *H. praehelvetica* and, below its upper boundary, the first individuals of *P. oraviensis* and *P. oraviensis trigona*.

The final part of the Cenomanian, below the beds with *I.* ex gr. *labiatus*, is characterized by the lower part of the D. imbricata Subzone (as defined by Salaj, Samuel, 1966), probably equivalent to the W. archaeocretacea Zone which has been recently established for the whole boreal and Tethys domains (Robaszynsky, 1981). The D. imbricata Subzone is marked by the almost simultaneous appearance of *D. imbricata, D. indica (=hagni), D. canaliculata, D. elenae, A. cretacea ;* it still contains thalmanninellas, rotaliporas, pseudorotaliporas. Within the levels of the examined M. geslinianum Zone, we have not found a microfauna, but we have admitted that in our country, too, this assemblage with *imbricata* and *archaeocretacea* appears in the interval occupied by the M. geslinianum Zone, as in all regions of the boreal and Tethys domains.

Turonian. In Romania, the D. imbricata Subzone assemblage, still containing rotaliporas, thalmanninellas and pseudorotaliporas continues in the first levels of strata with *M.* ex. gr. *labiatus*, as well up to the appearance of *H. helvetica.* More than that, the thalmanninellas and especially the rotaliporas persist (in the East Carpathians and North Dobrogea) up to the levels with *M.* ex gr. *labiatus, H. helvetica* and (in the East Carpathians) with the first specimens of *M. schneegansi.* Therefore, the Cenomanian-Turonian boundary, marked by the appearance of strata with *M.* ex gr. *labiatus,* cannot be established on this basis of planktonic foraminifera. In our case, as for other regions of the Mediterranean domain, the top of the range of rotaliporas s.l. cannot be a criterium for this boundary, as it is very frequently used in the Anglo-Parisian Basin and North America.

Except for the first levels with *I.* ex gr. *labiatus* which contain the assemblage of the D. imbricata Subzone, the other strata with *labiatus,* as well as other sequences attributed to the Lower Turonian are characterized by the H. helvetica Zone (pro parte, as defined by Sigal, 1966), namely the H. helvetica Subzone — without *M. schneegansi* (= "only with *H. helvetica*". Sigal, 1955) and probably by the appearance of the M. schneegansi Subzone (as defined by Ion, 1983).

The first levels with *M.* ex gr. *labiatus* and the assemblage of the D. imbricata Zone from the basal Turonian in Romania can be correlated to those with *W. coloradoense* from the lower part of the Turonian in France and the neighbouring regions which contain *M.* ex gr. *labiatus* and the microfauna of the upper part of the W. archaeocretacea Zone. The other subzones are correlated with the rest of the *M.* ex gr. labiatus Zone and with the M. nodosides Zone respectively, with the mention that the M. schneegansi Subzone continues also in the Middle Turonian.

For the Lower Turonian-Middle Turonian boundary we have no criteria given by the planktonic foraminifera. The assemblage of the M. schneegansi Subzone occurs also in the Middle Turonian, with *M. scheegansi, M. sigali, Carpathoglobotruncana*[2] *marianosi, H. helvetica,* without rotaliporas and thalmanninellas and without species of gr. "*lapparenti*", but with a large number of species of *Dicarinella.* The *R. kallesi* beds in North Dobrogea contain the microfauna of this biozone.

The Middle Turonian-Upper Turonian boundary can be arbitrarily established on the basis of planktonic foraminifera. We have admitted as an index fossil for this boundary *M. coronata,* although in papers with a rigours parachronology (e.g. Porthault, 1974 ; Robaszynsky,

1983) it is known to appear in the final part of the Middle Turonian (R. ornatissimum Zone). The extinction top of the *H. helvetica* species cannot be a generally valuable criterium for this boundary, as it is known to occur up to the Lower Coniacian in various circum-Mediterranean regions and from the Pacific-Californian domain. We have found it as well in Romania, up to the strata with *I. schloenbachi, Barroisiceras haberfellneri, Forresteria petrocoriensis.*

What we know for sure for the Upper Turonian in Romania is the fact that the beds with *Subprionocyculs neptuni* or with microfauna of this zone contain the assemblage with *M. coronata, M. pseudolinneiana, M. sinuosa, M. "renzi"* (plane-convex ; not *paraconcavata), M. angusticarinata, M. marginata,* atrributed to the M. coronata Subzone (Ion, 1982). This assemblage is similar to that known in the stratotype Turonian region (Robaszynsky et al., 1983).

Coniacian. In the Romanian Carpathians and North Dobrogea the Turonian-Coniacian boundary is very well characterized by planktonic foraminifera, as the rich assemblages of ammonites and inocerami have allowed to obtain a rigorous orthochronology and parachronology. On this basis we could establish (Ion, 1979, fide 1983) that *M. tarfayensis* is the index species for the lower boundary of the Coniacian.

Unlike *M. „renzi", M. sinuosa, M. angusticarinata* and *M. coronata,* it was never found in strata with Upper Turonian macrofaunal assemblages. Porthault (1974) has proposed *M. tarfayensis,* but *M. sinuosa* as well, as indicators of the Turonian-Coniacian boundary in the SE France, but having an arbitrary value, as there was no fauna for parachronology in the respective levels.

In the upper part of beds with *Forresteria petrocoriensis, Barroisiceras haberfellneri* or/and *I. schloenbachi, Dicarinella concavata* appears and nearly at the same time *D. asymetrica (= carinata),* too.

Therefore, the Lower Coniacian characterized by the above mentioned zones of macrofauna is featured from the planktonic foraminifera point of view, by : the M. tarfayensis Subzone (cf. def. Ion, 1982) which contains together with the index species, all species characteristic for the subjacent M. coronata Subzone, with the mention that it still contains *H. helvetica ;* the appearance of the Dicarinella concavata Zone (cf. def. Sigal, 1955, emend. Ion, 1983) with the quoted species, and species of the M. coronata Subzone, but with *H. helvetica* only in its basis.

The Middle and Upper Coniacian are characterized as well by the assemblage of the D. concavata Zone, as seen from the direct correlation made with ammonite and inocerami zones or assemblages. The absence of a characteristic macrofauna for the lower boundary of the Santonian has impeded as well the possibility to find out some microfaunal changes at the Coniacian-Santonian boundary.

We have admitted that the D. concavata Zone extends up to the Lower Santonian, as in the Carpathian domain in Romania the microfaunal change caused by the appearance of *G. bulloides* takes place above the strata with *Texanites oliveti.* Vertical changes in the assemblage of the D. concavata Zone from the Romanian Carpathian domain and North Dobrogea were not interpreted yet as chronostratigraphic values because data are still insufficient or contradictory. For example, *M. paraconcavata* and *"G." fornicata* seem to appear in the basis of the D. concavata Zone, namely in the Lower Coniacian in the East Carpathians, while in North Dobrogea they were first found in strata with *Peroniceras.* What could be well argumented is that *D. asymetrica* appears in strata with *I. schloenbachi* and it exists as well in strata with *Peroniceras* and *I. mantelli*; therefore, the possibility of using this species as an index fossil for the Lower Santonian, as it is used in other regions in the world is certainly excluded.

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* *Carpathoglobotruncana* n. gen. — Ion, 1980 — in the summary of the Thesis of doctor's degree, lithographed.

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AMMONITE AND INOCERAMI ZONES AND ASSEMBLAGES AND PLANKTONIC FORAMINIFERA ZONES IN THE CENOMANIAN - CONIACIAN INTERVAL OF ROMANIA

Stages	Substages	Standard zonation used in this paper		Ammonite zones known in Romania (Szasz, 1982 a,b; 1983 a,b, in press)	Ammonite assemblages (Szasz)	Inocerami assemblages (Szasz)	Planktonic foraminifera zones and subzones Carpathians and N. Dobrogea (Ion, 1977, 1979, 1983)		Marker biohorizons
CONIACIAN	Upper	*Paratexanites (Parabevahites) serratomarginatus*		?		*Inoceramus subquadratus*	*Dicarinella concavata*		
	Middle	*Peroniceras tridorsatum*		*Peroniceras tridorsatum*	*Peroniceras tridorsatum, P. czoernegi, P. moureti, Nowakites karezi, Pachydiscus sp, Anapachydiscus arrialoorensis, Protexanites cf. strozzi, Gaudryceras glaneggensis Gauthiericeras roquei*	Assemblage zone with *Inoceramus mantelli*: *Inoceramus mantelli mantelli I. mantelli subrhenanus, I. mantelli beyenburgi*			
	Lower	*Forresteria (Harleites) petrocoriensis*		*Forresteria (Harleites) petrocoriensis* and *Barroisiceras haberfellneri*	*Forresteria (Harleites) petrocoriensis, Barroisiceras haberfellneri, Yabeiceras aff. orientalis, Nowachites spp. Pseudokossmaticera spp. Tissoides haplophyllum*	Assemblage zone with *Inoceramus schloenbachi*: *Inoceramus schloenbachi I. erectus, I. crassus, I. inconstans I. waltersdorfensis, I. rotundatus I. fiegei, I striatoconcentricus carpathicus, I. naumanni*	„grandes Rosalines plâtes"	*Marginotruncana tarfayensis*	*concavata* *tarfayensis*
TURONIAN	Upper	*Reesidites minimus*		No characteristic ammonites	*Subprionocyclus neptuni, S. branneri, S. cf. normalis, Lewesiceras mantelli Tongoboryceras cf rhodanicum, Scaphites ex. gr. geinitzi, Romaniceras sp.*	*Inoceramus apicalis, I. falcatus I. lusatiae, I. ex. gr. parvus-peshioensis I. ex. gr. costellatus-kleini, I. inaequivalvis*		*Marginotruncana coronata*	
		Subprionocyclus neptuni		*Subprionocyclus neptuni*					*coronata*
	Middle	*Collignoceras wollgari*	*Romaniceras (Yubariceras) ornatissimum*	?	*Romaniceras cf. kallesi*	*Inoceramus tenuistriatus*	*Helveto-globotr. helvetica*	*Marginotruncana schneegansi*	
			Romaniceras kallesi			*Inoceramus (Mytiloides) ex. gr. labiatus*			
			Kamerunoceras turoniense						
	Lower	*Mammites nodosoides*		*Paramammites polymorphum* and *Choffaticeras polymorphum*	*Paramammites polymorphum, Choffaticeras pavilleri, Spatites (Jeanrogericeras) reveliereanum, Kamerunoceras (Schindewolfites) inaequicostatus*			*Helveto-globotr. helvetica without M. schneegansi*	*schneegansi* *helvetica*
		Watinoceras coloradoense			*Fagesia peroni Mammites nodosoides zeibaensis*		„grandes Globigerines"	*Dicarinella imbricata*	
CENOMANIAN	Upper	*Neocardioceras juddii*		No characteristic ammonites		*Inoceramus pictus*			
		Metoicoceras geslinianum						*Whiteinella paradubia*	*imbricata*
		Eucalycoceras pentagonum		*Eucalycoceras pentagonum*	*Eucalycoceras pentagonum, E gothicum Calycoceras naviculare, Pseudocalycoceras thomeli, Forbesiceras bicarinatum*		*Rotalipora gr. cushmani-turonica*		*paradubia* *cushmani turonica*
	Middle	*Acanthoceras jukesbrownei*		*Acanthoceras jukesbrownei*	*Acanthoceras whitei, A. aff. jukesbrownei, Calycoceras newboldi, C. spinosum*		*Thalmanninella gr. reicheli*	*Th. deeckei*	*deeckei*
		Acanthoceras rhotomagense		*Acanthoceras rhotomagense*	*Acanthoceras rhotomgense sussexiense A. rhotomagense confusum, Calycoceras spinosum, Turrilites costatus, Sciponoceras baculoide, Mesogaudyceras leptonema*			*Th. reicheli* *Th. porthaulti*	*reicheli* *porthaulti*
	Lower	*Mantelliceras orbignyi*		No characteristic ammonites	*Turrilites costatus Mantelliceras spp.*		*Thalmanninella brotzeni/th globotruncanoides*		
		Mantelliceras mantelli		*Mantelliceras mantelli*	*Mantelliceras mantelli, M. cantianum M. saxbii, M. costatum, Stoliczkaia (Lamnayella) santaecatharinae, Hyphoplites spp., Hypoturrilites tuberculatus H. mantelli, H. tuberculatus, Neostlingoceras carcitaensis*	*Inoceramus crippsi*			*brotzeni globotr.*
ALB.	U.	*Stoliczkaia dispar*		*Stoliczkaia dispar*	*Stoliczkaia dispar, Ostlingoceras puzosianum Mortoniceras (Durnovarites) perinflatus*		*Thalmanninella appenninica*		

Imprim. Atel. Inst. Geol. Geof.

SYNTHETIC TABLE ON THE BIOSTRATIGRAPHY (PLANKTONIC FORAMINIFERA) OF THE MIDDLE CRETACEOUS IN ROMANIA (CARPATHIANS AND NORTH DOBROGEA)

Imprim. Atel. Inst. Geol. Geof.

Stratigraphie—Paléontologie

ASSOCIATIONS DE MOLLUSQUES ET BRACHIOPODES TRIASIQUES DES CARPATHES ORIENTALES ROUMAINES ET LEUR PLACE DANS LE CONTEXTE BIOSTRATIGRAPHIQUE GÉNÉRAL ALPINO-CARPATHIQUE

PAR

ILIE TURCULEȚ[1]

Les associations de Mollusques et de Brachiopodes triasiques, découvertes jusqu'à présent dans les Carpathes Orientales roumaines sont liées exclusivement aux formations développées dans la zone cristallino-mésozoïque. Cette zone est constituée de formations cristallophylliennes surmontées par des dépôts mésozoïques, en s'étendant des Monts du Maramureș, à travers les Monts de Rarău-Giumalău-Monts de Bistrița-Monts de Tulgheș-Monts de Hăghimaș-Monts Perșani, jusqu'aux Monts Bucegi-Piatra Craiului.

Dans le succession des formations mésozoïques de cet aréal les dépôts triasiques ont un développement relativement constant aussi bien dans la nappe bucovinienne que dans celles transylvaines.

Le Trias bucovinien est bien représenté dans la base de la séquence des dépôts mésozoïques, comportant des gréso-conglomérats, des calcaires et des dolomies à développement relativement constant. Le Trias des nappes transylvaines, généralement carbonaté, constitue des klippes, des olistolites et des blocs exotiques enrobés dans des formations plus jeunes, crétacées.

Le contenu biostratigraphique des formations triasiques des Carpathes Orientales est extrêmement riche. L'inventaire bionomique connu actuellement est le résultat des recherches entreprises le long de plus d'un siècle (Herbich, 1878 ; Mojsisovics, 1875 ; 1882 ; Paul, 1876 ; Uhlig, 1903, 1910 ; Merhart, 1910 ; Kittl, 1912 ; Atanasiu, 1927 ; Jekelius, 1936 ; Ilie, 1954 ; Turculeț, 1967, 1971, 1972, 1976 a, b ; 1979, 1980, 1981, 1982 ; Grasu, 1971 ; Săndulescu, 1975 ; Mutihac, 1968 ; Patrulius, 1969, 1976 etc.).

L'abondance des formes de Mollusques et Brachiopodes permet aujourd'hui de désigner des associations d'une réelle valeur biostratigraphique, à possibilités d'intégration dans les faunes triasiques alpino-carpathiques.

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Ces associations sont :

1. L'association à *Claraia clarai* (Emmr.), *C. tridentina* (Bittn.) et *Eumorphotis inaequicostata* (Ben.) mentionnée par Grasu (1971) dans les calcaires dolomitiques et les grès micacés des Monts Hăghimaș indique la présence du Griesbachien dans le Trias inférieur de la nappe bucovinienne.

2. L'association à *Eumorphotis venetiana* (Hauer) qui contient encore *E. inaequicostata* (Ben.), *Gervilleia incurvata* Leps., *Costatoria costata* (Zenk.), *Unionites fassaensis* Wissm., *Neoschizodus laevigatus* (Goldf.) est présente dans les calcaires gris, en plaques, en constituant les olistolites éotriasiques englobés dans le wildflysch éocrétacé du synclinal de Rarău (colline de Runcu) (Merhart, 1910 ; Turculeț, 1971 ; Patrulius, 1976). La présence de l'espèce *E. venetiana* indique l'existence du Dienérien et du Smithien dans le cadre des formations allochtones.

3. L'association à *Eumorphotis telleri* (Bittn.) et *Tirolites cassianus* (Qu.) contient encore : *Costatoria costata* (Zenk.), *Unionites fassaensis* Wissm., *Gervilleia bucovinensis* Turculeț, *Turbo rectecostatus* Hau., *T. lemkei* Wit., *Naticella costata* (Mnstr.) etc. Elle caractérise les marnes grises, jaunâtres, satinées, disséminées dans le wildflysch éocrétacé du synclinal de Rarău (Turculeț, 1971 ; Patrulius, 1976) aussi bien que dans les calcaires en plaques des Monts Perșani (défilé de l'Olt) (Ilie, 1954). La présence de l'espèce *E. telleri* ainsi que celle de *Tirolites cassianus* prouve l'existence du Spathien.

4. L'association à *Gervilleia modiola* Frech comporte également *G. mytiloides* Schl., *Hoernesia socialis* (Schl.), *Leptochondria alberti* (Goldf.), *"Myophoria" praeorbicularis* Bittn., *Entolium discites*, *Trigonodus sandbergeri* Alb., *Parallelodon beyrichi* Stromb. et d'autres. Elle caractérise autant le niveau des calcaires en plaquettes, logés immédiatement au-dessous des dolomies (mésotriasiques) de la nappe bucovinienne (Azodul Mare — Monts de Tulgheș — Atanasiu, 1927 ; Monts Ciofranca-Hăghimaș — Turculeț, 1967) qu'une série de calcaires et marnes satinées, „allochtones", disséminées dans le wildflysch éocrétacé des synclinaux de Rarău et de Hăghimaș (Turculeț, 1971) tout comme dans les Monts Perșani (Ilie, 1954) ou bien dans les Monts Bucegi (Gîlma Ialomiței) (Patrulius, 1969).

L'association à *G. modiola* a une position stratigraphique bien controversée. Certains géologues l'ont attribuée au Campilien supérieur, tandis que Ogilvie-Gordon (1927) souligne le fait qu'elle passe même dans la base de l'Anisien des Alpes Dolomitiques sud-tyroliens. Les dernières révisions faites par Patrulius (1976) soutiennent cette conclusion, étayée sur les données provenant des Monts Bakony et du bassin allmand. On pourrait conclure que cette association correspond au Spathien terminal et à une grande partie de l'Anisien inférieur (Aegéen).

En essayant d'intégrer les associations éotriasiques susmentionnées dans le contexte biostratigraphique alpino-carpathique, on constate que leur contenu ressemble à celui des associations des Alpes orientales du Sud, particulièrement aux associations des Alpes Dolomitiques, bien que lithologiquement les dépôts éotriasiques est-carpathiques rappellent les formations de Werfen des Alpes orientales du Nord.

5. L'association à *Daonella (Lömmelalla) lömmeli* (Wissm.), *Protrachyceras archelaus* (Laube) comprend encore : *Posidonia wengensis* Wissm., *Camptonectes (Annulinectes) concentrice-striatus* (Hörn.), *Leptochondria sarajebensis* Ki, *Daonella (Arzelella) arzelensis* Ki., *D. (A.) tyrolensis* Mojs., *D. (A.) indica* Bittn., *D. (A.) badiotica* Mojs., *Halobia bukowinensis* Ki, *Michelinoceras mojsisovicsi* (Salom.), *Protrachyceras furcatum* (Mnstr.), *Paratrachyceras basileus* (Mnstr.), *P. dichotomum* (Mnstr.), *Celtites evolutus* Salom., *Arpadites cinensis* Mojs., *Anolcites paronai* (Salom.), *Megaphyllites procerus* Arth., *Daonella (Pichlerella pichleri* Guemb., *D. (P.) pauli* Ki., et d'autres (Kittl, 1912 ; Turculeţ, 1967, 1979). Cette association se retrouve dans les calcaires rouge-jaunâtres de la klippe sédimentaire du ruisseau Pîrîul Cailor (Rarău). Les espèces-index montrent indubitablement la présence du Longobardien. Quoique les caractères lithologiques des calcaires ressemblent à ceux de Hallstatt des Alpes orientales du Nord, le contenu de l'association de Mollusques indique des affinités évidentes avec les couches de Wengen et les couches inférieures de St. Cassian des Alpes orientales du Sud. Ce fait a été observé dequis longtemps par Mojsisovics (1879), Arthaber (1906), French (1907), Renz (1910), Kittl (1912), Haug (1921).

Une association du même âge est celle de Halobiidae que renferment les calcaires noduleux, stratifiés, rouge-verdâtres, à silexites de la base de Piatra Zimbrului (Rarău). Outre les Halobiidae rencontrées sur le ruisseau Pîrîul Cailor, on a signalé encore : *Daonella (Grabella) grabensis* Ki., *Daonella (Arzellella bulogensis* Ki., *D. (A.) longobardica* (Mojs.) Ki., etc. (Turculeţ, 1972).

6. L'association à *Daonella lömmeli* Wissm., *Pachycardia plieningeri* Br., *Cassianella decussata* Mnstr. et *Coelostylina nodosa* Mnstr. contient une faune extrêmement riche (plus de 200 taxons d'où 100 espèces de Mollusques, 48 Brachiopodes et d'autres). On pourrait mentionner encore : *Amphijanira coronensis* (Jek.), *Antijanira alutensis* (Jek.), *Entolium nudiferum* (Bittn.), *Chlamys interstriatus* (Mnstr.), *Laubella delicata* (Laube), *Schizogonium serratum* Mnstr., de nombreuses espèces de *Pleurotomaria, Worthenia, Euchrysalis* etc., de *Pleuronautilus marmolatae* Mojs., *Hungarites elsae* Mojs., *Arcestes barrandei* Laube et d'autres (Jekelius, 1936). Parmi les Brachiopodes il y a bien des espèces de *Robinsonella. Caucasorhynchia, Trigonirhynchella, Pentactinella, Dioristella, Neoretzia,* etc.

Cette association provient des calcaires blanc-gris des environs de Braşov, en représentant le Longobardien-Cordévolien. Il s'y agit d'un faciès calcaire des couches de St. Cassian des Alpes méridionales.

7. L'association à *Trachyceras aon* (Mnstr.) renferme également *Anolcites armatum* (Mnstr.), *Eremites orientale* Mojs., *Clionites catharinae* bukowinensis Sim., *Protrachyceras furcatum* (Mnstr.), *Arcestes reyeri* Mojs., *A. gaytani* (Klipst.), *Cladiscites striatulus* (Mnstr.), *Megaphyllites jarbas* (Mnstr.), *Monophyllites agenor* (Mnstr.), *M. aonis* Mojs., *Coroceras hypsocarenus* Mojs., *Osthoceras politum.* Mojs. (Mojsisovics, 1879, 1882 ; Turculeţ, 1982 b).

Par son contenu l'association présente de réelles ressemblances avec les faunes des couches supérieures de St. Cassian des Alpes orientales du Sud, aussi bien que certaines influences du faciès de Hallstatt des Alpes septentrionales. Ce sont des blocs dans des calcaires rouges affleurant sur le ruisseau Pîriul Cailor (Rarău) qu'elle a rencontrés, en attestant de cette manière la présence du Cordévolien.

8. L'association à *Trachyceras aonoides* Mojs. et *T. austriacum* Mojs, renferme aussi : *Protrachyceras attila* Mojs., *Dittmarites ferdinandi* Mojs., *D. circumscissus* Mojs., *Protrachyceras roderici* Mojs., *Joannites johannis austriae* (Klip.), *J. diffissus* (Hau.), *J. compressus* Wang et He, *Paralobites tingriensis* Wang et He, *Coroceras delphinocephalus* Mojs., *C. laubei* Mojs., *Arcestes (Anisarcestes) periolcus* Mojs., *Megaphyllites jarbas oenipontanus* Mojs., *M. humilis* Mojs., *Sphingites stoppanii* Mojs., *Pompeckjites layeri* (Hau.) (Turculeț et al., 1979 ; Turculeț et al., 1979 ; Turculeț, 1982 a). Les taxons ont été déterminés des calcaires rouge-violacés du versant droit de la vallée du ruisseau Pîrîul Cailor (Rarău).

Grace au contenu bionomique et à la nature lithologique, le Trias de cette région a les particularités du faciès des calcaires de Hallstatt. La présence de certaines espèces décrites par Waug et He (1976) pourrait mettre en évidence quelques interférences avec les faunes himalayennes. L'âge indiqué est certainement julien.

9. L'association à *Halobia styriaca* Mojs. et *H. austriaca* Mojs. includ aussi : *H. očevjana* Ki., *H. superba* Mojs., *H. breuningiana* Ki., *Cardinia ovula* Ki. etc. Elle est connue dans les calcaires blancs et gris de la région de Rarău (Todirescu, Popchii Rarăului, Izvorul Malului et d'autres) et des Monts Perşani (Turculeț, 1971 ; Mutihac, 1968 ; Patrulius, 1969). L'âge est julien-tuvalien.

10. L'association contenant *Jovites dacus* Mojs. des calcaires rouges du ruisseau de Kovacs (Hăghimaş) et *Monotis digona* Ki. des calcaires gris de la colline de Hăghimaş (Rarău) est d'âge tuvalien (zone subbulatus) (Mojsisovics, 1875 ; Turculeț, 1979).

11. L'association à *Rhacophyllites neojurensis* (Qu.), *Pinacoceras postparma* Mojs. et *Distichites celticus* Mojs. comporte en outre *Distichites wulfeni* (Mojs.), *Placites subsimetricus* Mojs., *Cladiscites monticola* Mojs., *Halorites, Ectolcites, Paratisbites scaphitiformis* (Ha.) etc. (Mojsisovics, 1875). Elle apparaît dans les calcaires rouges affleurant dans les environs de Fagul Oltului (Hăghimaş) et indique la présence de l'Alaunien en faciès de Hallstatt (zone bicrenatus).

12. L'association à *Tragorhacoceras occultum* (Mojs.) comporte encore *Halorites alexandri* Mojs., *Megaphyllites insectus* Mojs., *Cladiscites quadratus* Mojs., *Placites polydactylus* Mojs., *P. myophorum* Mojs., *P. perauctum* Mojs., *Arcestes intuslabiatus* Mojs., *A. sisyphus* Mojs., *A. diphyus* Mojs., *Stenarcestes polysphinctus* Mojs., *Grypoceras mesodicum* (Ha.), *Paranautilus simonyi* (Ha.), *Juvavionautilus heterophyllum* Mojs., *Monotis haueri* Ki., *Halorelloidea rectifrons* (Bittn.), *Euxinella pamirensis* Dagys, *Mentzelia sinuata* Dagys, *Zugmayerella koessenensis* (Zugm.), *Oxycolpella oxycolpos* (Emmr.), *O. robinsoni* Dag., *O. kunensis* Dag., *Triadithyris gregariaformis* (Zugm.), *Athyris, Majkopella* etc. (Turculeț, 1976 a, 1976 b, 1980, 1983).

L'association est contenue dans les calcaires rouges, noduleux de la klippe des sources du ruisseau de Timon (Ciungi) — Rarău. La faune de Céphalopodes tout comme les caractères lithologiques permettent des comparaisons avec les calcaires de Hallstatt.

Certaines affinités fauniques peuvent être établies autant avec les calcaires de Bleskovy-Pramen (Carpathes occidentales) (Kollarova-Andrusovova et al., 1973) qu'avec les faunes à Brachiopodes du Sud de l'URSS décrites par Dagys (1963).

13. L'association à *Rhaetina gragaria* (Sss.) et *R. pyriformis* (Sss.) contient surtout des Brachiopodes : *Zeilleria elliptica* (Zugm.), *Z. norica* (Sss.), *Septaliphoria fissicostata* (Sss.), *Triadithyris* gregariaformis (Zugm.), *Zugmayerella koessenensis* (Zugm.), *Austrirhynchia cornigera* (Schaf.), *Euxinella pamirensis* Dagys et d'autres (Turculeț, 1971).

La faune est renfermée dans les calcaires gris, spathiques, lumachelliques qui affleurent comme blocs enrobés dans le wildflysch éocrétacé du synclinal de Rarău. L'association est caractéristique pour le Rhaetien et présente de grandes affinités avec celle du faciès de Kössen des Alpes ainsi que celles du Sud de l'URSS.

En synthétisant les données ci-dessus on peut conclure :

Dans la zone cristallino-mésozoïque des Carpathes Orientales roumaines, le Trias est présent dans toutes les unités tectoniques majeures qui renferment des dépôts sédimentaires.

Les dépôts triasiques des nappes transylvaines ont un contenu bionomique plus riche par rapport à ceux de la nappe bucovinienne.

Les dépôts éotriasiques tant de la nappe bucovinienne que des nappes transylvaines ont une nature lithologique voisine de celle des couches de Werfen des Alpes orientales du Nord, alors que les associations fauniques ont d'évidentes affinités sud-alpines.

Les formations mésotriasiques transylvaines comportent un faciès carbonaté nord-alpin (de Hallstatt), tandis que les associations de Mollusques sont nettement sud-alpines (couches de Wengen, couches inférieures de St. Cassian).

Les dépôts néotriasiques transylvaines présentent tous les caractères lithostratigraphiques et bionomiques du faciès de Hallstatt des Alpes orientales du Nord.

A certains niveaux stratigraphiques, les associations fauniques présentent des ressemblances avec les faunes des Carpathes occidentales, Caucase, Crimée et même avec celles himalayennes.

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Stratigraphie—Paléontologie

LES FACIÈS CARBONATÉS DU PALÉOZOÏQUE DE LA PLATE-FORME MOESIENNE (ROUMANIE)

PAR

CONSTANTIN VINOGRADOV, MIHAI POPESCU [1]

Introduction

La plate-forme moesienne est une unité géologique de l'avant-pays (Vorland) qui s'étend d'une part et d'autre du cours inférieur du Danube entre l'orogène carpathique au Nord et l'orogène balkanique au Sud. (Le secteur roumain est situé seulement au Nord du Danube).

Les formations paléozoïques constituant le premier cycle de sédimentation de la couverture de la plate-forme moesienne (fig. 1) se laissent grouper en trois séquences majeures : une mégaséquence terrigène inférieure, une séquence carbonatée et une séquence terrigène supérieure.

La mégaséquence terrigène inférieure d'âge ordovicien (cambrien ?)-eifélien est constituée d'ortoquartzites et quartzwackes (quartzites de Mangalia) ; argilites gris-noirâtres aux intercalations de grès, wackes quartzo-feldspathiques, siltites, biocalcarénites, métaarkoses et métatuffites (argilites de Ţăndărei) ; ortoquartzites, quartzwackes, grès et wackes quartzo-feldspathiques, argiles siltiques, sidérites (grès-quartzites de Smirna).

La séquence carbonatée d'âge givétien-dinantien comprend une large gamme de calcaires et dolomies, parfois associés aux anhydrites (formation de Călăraşi).

La séquence terrigène supérieure d'âge viséen supérieur-westphalien est constituée d'argiles siltiques charbonneuses, argiles sidéritiques ou limonitiques, marnes, siltites, wackes lithiques ou litho-feldspathiques, rarement des calcaires et dolomies ferrugineuses (formation de Vlaşin). Il est très probable que le complexe terrigène-volcanogène (Permien ?) qui surmonte la formation de Vlaşin appartienne également au cycle paléozoïque.

Le but de la présente note est l'étude sédimentologique de la formation de Călăraşi qui présente toutes les caractéristiques d'une plate-forme carbonatée.

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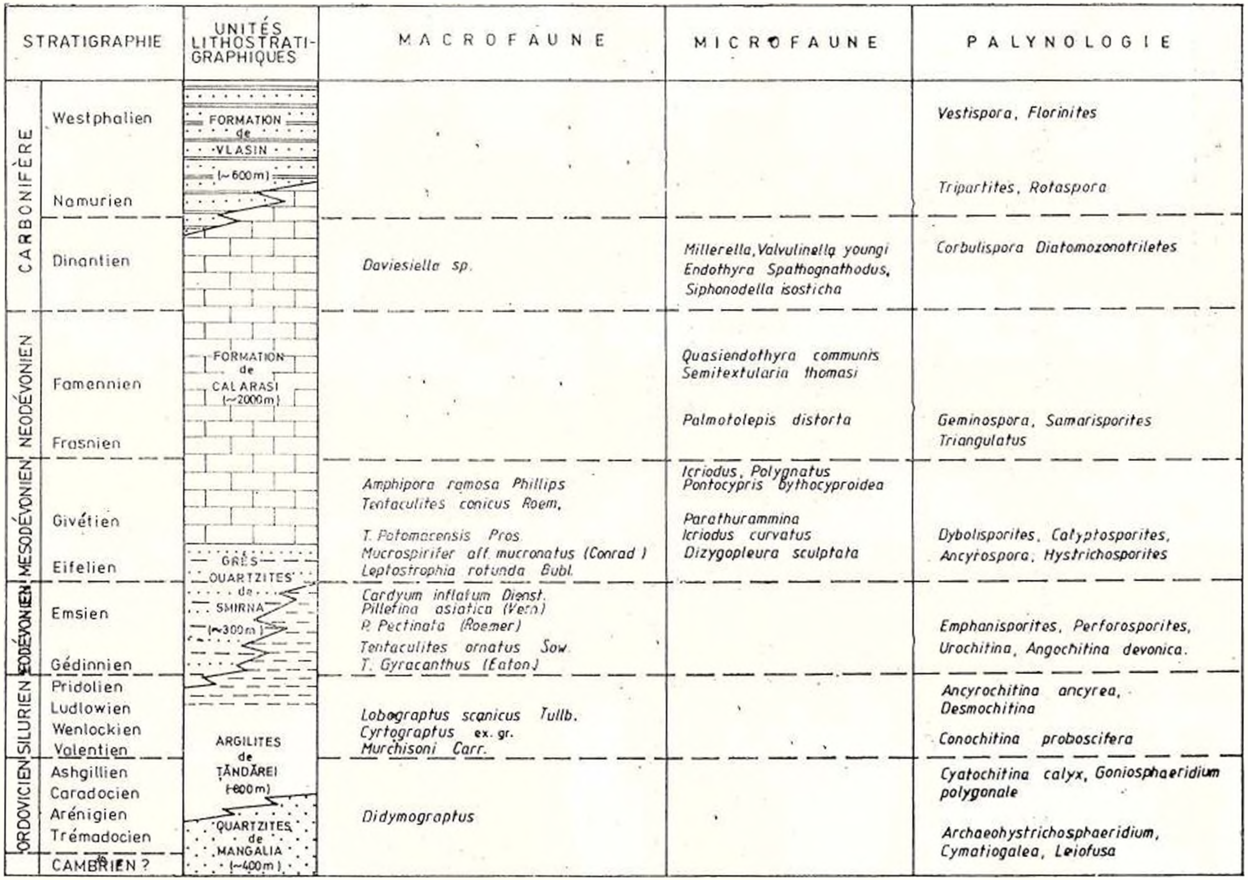
[1] Institut de Recherches pour Hydrocarbures, str. Toamnei 109, Bucureşti.



STRATIGRAPHIE		UNITÉS LITHOSTRATIGRAPHIQUES	MACROFAUNE	MICROFAUNE	PALYNOLOGIE
CARBONIFÈRE	Westphalien Namurien	FORMATION de VLASIN (~600m)			*Vestispora, Florinites* *Tripartites, Rotaspora*
	Dinantien	FORMATION de CALARASI (~2000m)	*Daviesiella sp.*	*Millerella, Valvulinella youngi* *Endothyra Spathognathodus,* *Siphonodella isosticha*	*Corbulispora Diatomozonotriletes*
NEODÉVONIEN	Famennien Frasnien			*Quasiendothyra communis* *Semitextularia thomasi* *Palmotolepis distorta*	*Geminospora, Samarisporites* *Triangulatus*
MESODÉVONIEN	Givétien Eifelien	GRÈS QUARTZITES de SMIRNA (~300m)	*Amphipora ramosa Phillips* *Tentaculites conicus Roem.* *T. Potomacensis Pros* *Mucrospirifer aff. mucronatus (Conrad)* *Leptostrophia rotunda Bubl.*	*Icriodus, Polygnatus* *Pontocypris bythocyproidea* *Parathurammina* *Icriodus curvatus* *Dizygopleura sculptata*	*Dybolisporites, Calyptosporites,* *Ancyrospora, Hystrichosporites*
EODÉVONIEN	Emsien Gédinnien		*Cardyum inflatum Dienst.* *Pillefina asiatica (Vern)* *P. Pectinata (Roemer)* *Tentaculites ornatus Sow.* *T. Gyracanthus (Eaton)*		*Emphanisporites, Perforosporites,* *Urochitina, Angochitina devonica.*
SILURIEN	Pridolien Ludlowien Wenlockien Valentien	ARGILITES de ȚÂNDĂREI (~800m)	*Lobograptus scanicus Tullb.* *Cyrtograptus* ex. gr. *Murchisoni Carr.*		*Ancyrochitina ancyrea,* *Desmochitina* *Conochitina proboscifera*
ORDOVICIEN	Ashgillien Caradocien Arénigien Trémadocien	QUARTZITES de MANGALIA (~400m)	*Didymograptus*		*Cyatochitina calyx, Goniosphaeridium* *polygonale* *Archaeohystrichosphaeridium,* *Cymatiogalea, Leiofusa*
	CAMBRIEN ?				

Fig. 1 — Formations paléozoïques de la plate-forme moesienne (secteur roumain).

Faciès et milieux dépositionnels de la formation de Călăraşi (Givétien-Dinantien)

A la partie basale de la plate-forme carbonatée (Givétien) trois secteurs distincts ont été mis en évidence : oriental, central et occidental (fig. 2).

Dans le secteur oriental c'est le milieu supratidal qui prédomine comportant également des épisodes intertidaux et sporadiquement des épisodes subtidaux. Le milieu supratidal a été déterminé selon les faciès dolomicrosparitique-anhydritique et mudstone-anhydritique, tous les deux ayant des intercalations fréquentes d'argiles gris-verdâtres à débris de plantes. D'habitude, ces faciès sont abiogènes, parfois à Ostracodes ou bien à nodules cyanophitiques. Les structures laminées, fenestrae et brèches de dessication ou de dissolution sont caractéristiques pour ce faciès. L'anhydrite est contenue dans des couches minces où elle apparaît comme des nodules en calcaires et dolomies. Toutes ces caractéristiques permettent de considérer que ce faciès a été généré dans un environnement particulier de type Sebkha. Les épisodes intertidaux sont mis en évidence par les faciès boundstone stromatoporoïdique et wackestone bioclastique à Gastéropodes, Lammelibranches et Calcisphères. Le passage des milieux supratidal/intertidal est réalisé par les faciès wackestone stromatolithique (tapis algaire) à Cyanophyceae et Ostracodes et wackestone pellétoïdal (pellets algaires) qui s'associent avec les faciès boundstone à *Stromatoporella*. Le passage des faciès intertidal/subtidal est réalisé par grainstone oncoïdal (onkosparite), grainstone-packstone à Crinoïdes et Brachiopodes, wackestone à Crinoides, Dasycladaceae *(Issinella, Nanopora)*, Gastéropodes, débris de Characeae, wackestone pellétoïdal.

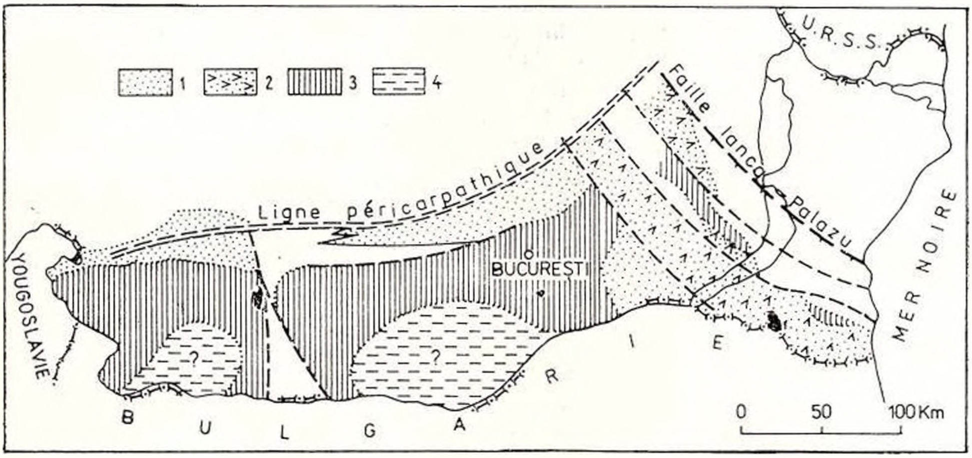


Fig. 2 — Schéma paléogéographique du Givétien.
1, plate-forme interne milieu supratidal ; 2, plate-forme interne milieu supratidal (Sabkha) ; 3, plate-forme interne milieu intertidal ; 4, plate-forme interne milieu subtidal.

Dans les secteurs central et occidental le milieu supratidal est restreint à leur partie septentrionale où se développent des faciès mudstone-dolomicrosparitique et dolomicrosparitique à fragments de schistes cristallins, niveaux bréchiques et croûtes limonitiques. Mais la plupart des secteurs central et occidental est occupée par des dépôts du milieu intertidal (faciès packstone à Crinoïdes, faciès boundstone stromatoporoïdique et faciès pelletoïdal à Ostracodes et Crinoïdes) ayant des passages vers les faciès supratidal (dolomicrosparitique laminés, packestone à Characeae et Ostracodes, grainstone pelletal-dolosparitique à rares Ostracodes) ou subtidal (boundstone à *Orthonella* et Annélides, wackestone stromatolithique à Ostracodes, Crinoïdes, Foraminifères et Calcisphères). Le milieu subtidal s. str. n'est que suposé dans la partie méridionale des secteurs central et occidental.

Pendant le Dévonien supérieur (fig. 3) le domaine supratidal (faciès mudstone à débris de plantes et Gastéropodes, faciès mudstone-dolomicrosparitique à Characées et nodules d'anhydrite, faciès dolomicrosparito-anhydritique) est bien restreint. Le passage des milieux supratidal/intertidal s'achève par le faciès wackestone stromatolithique (tapis algaire). Au contraire, les milieux intertidal et subtidal ont une grande extension. Les faciès packestone-grainstone à Crinoïdes, biscuits algaires, Tubiphytes, Moravamminidae, Gastéropodes, Lamellibranches et rares Characées *(Quasiumbella)*, boundstone à Stromatoporoïdes ou Tabulés et dolosparites caractérisent le milieu intertidal. Le passage vers le milieu subtidal est réalisé par les faciès packstone et grainstone pelletoïdal ou bioclastique à Crinoïdes, Brachiopodes, Foraminifères, Dasycladacées. La milieu subtidal est mis en évidence par les faciès wackestone et packstone à Foraminifères, Crinoïdes, Brachiopodes, Calcisphères, Ostracodes et Porostromates *(Orthonelle, Garwoodia)*.

Les faciès et les mileux y décrits (Givétien-Néodévonien) appartiennent à la plate-forme interne (marin restreint).

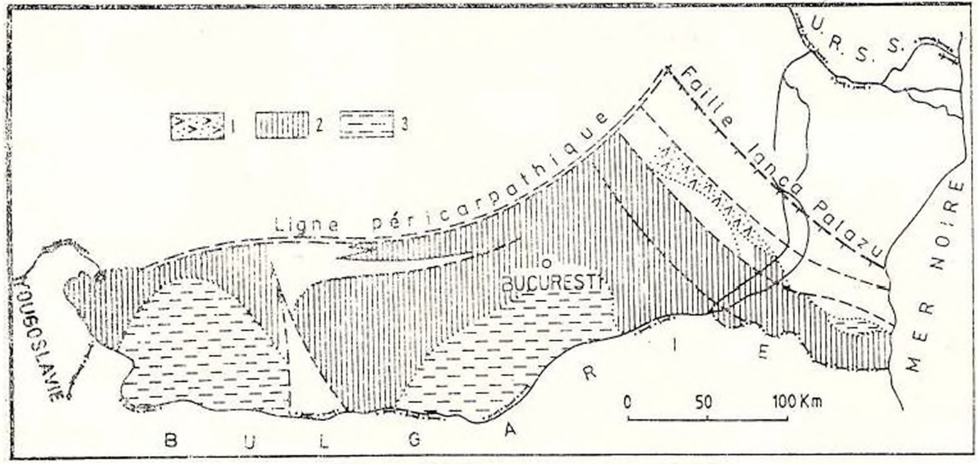


Fig. 3 — Schéma paléogéographique du Néodévonien.
1, plate-forme interne milieu supratidal ; 2, plate-forme interne milieu intertidal ; 3, plate-forme interne milieu subtidal.

A partir du Dinantien (fig. 4) la plate-forme interne est limitée aux bordures septentrionale et occidentale du domaine moesien, en échange la plate-forme externe s'étend sur de grandes aires. Dans le cadre de cette dernière plate-forme on peut distinguer des faciès de barrière (boundstone à Tabulés, Stromatopores, Rhodophycées accompagnés des faciès grainstone oncolithique, grainstone-packstone à Crinoïdes, Brachiopodes, Bryozoaires, Lamellibranches, Annélides, Tubiphytes, Moravamminidae, coprolithes de Gastéropodes) et des milieux marins ouverts relativement profonds (packstone à Crinoïdes, Brachiopodes, Bryozoaires ; wackestone à Crinoïdes, Brachiopodes, Bryozoaires, Ostracodes, spicules de Spongiaires ; wackestone spongolitique associé à silicolites spongolitiques et argiles).

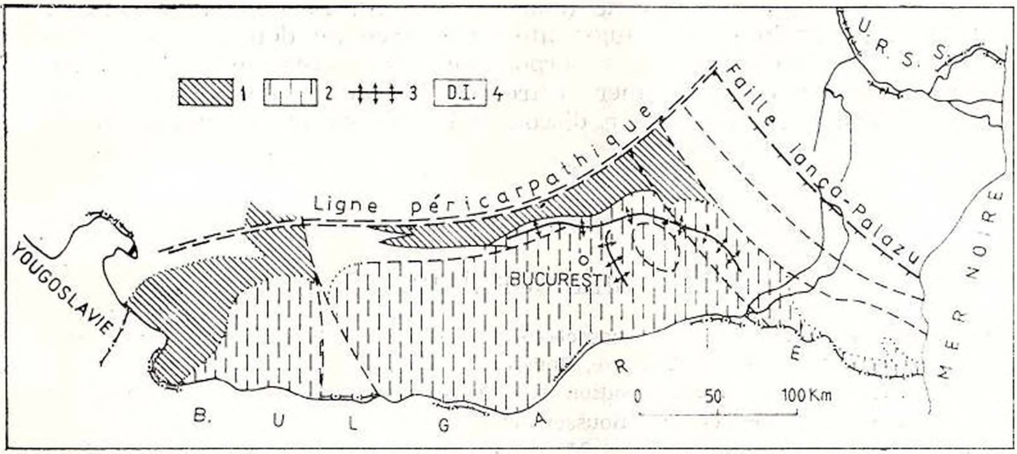


Fig. 4 — Schéma paléogéographique du Dinantien.
1, plate-forme interne ; 2, plate-forme externe ; 3, alignements récifaux ; 4, dépression d'Ileana.

Sur des aires restreintes du domaine moesien (dépression d'Ileana) la sédimentation carbonatée passe également dans le Namurien par des faciès subtidaux et intertidaux de la plate-forme interne qui font la transition vers les sédiments paraliques de la formation de Valșin.

Synthèse paléogéographique

L'étude sédimentologique du premier cycle de sédimentation (Paléozoïque), du domaine moesien a mis en évidence la plus ancienne plate-forme carbonatée (Givétien-Dinantien) de l'histoire géologique de cette unité. L'installation de la sédimentation de la plate-forme carbonatée a été précédée par le dépôt des sédiments terrigènes épicontinentaux ayant leur extension dominante pendant le Silurien moyen au moment où les associations palynologiques étaient dominées par le phytoplancton marin (acritarches, leiosphères) et chitinozoaires. Dans le Dévonien supérieur apparaissent des spores triletes de Ptéridophytes en coexistant avec les derniers représentants des chitinozoaires. C'est le début d'une phase régressive qui se continue jusqu'au Stéphanien. Les spores triletes de grande taille sont caractéristiques pour le Dévonien moyen, afin d'atteindre le maximum de développment durant le Viséen supérieur-Namurien.

La plate-forme carbonatée est le produit d'un domaine marin à régime de marée dont les oscillations sont ressenties sur de grandes surfaces. Les dépôts carbonatés et les évaporites dévoniennes appartiennent à la plate-forme interne (marin restreint). Pendant le Dinantien le domaine marin ouvert augmente en surface au détriment du domaine marin restreint. La superposition des dépôts de mer ouverte (Dinantien) sur celles de mer restreinte (Dévonien), en absence d'une phase transitive, indique une discontinuité sédimentaire au niveau du Dinantien basal (Tournaisien).

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PROBLEMS OF THE EUROPEAN CONTINENTAL MARGIN IN THE TRANSYLVANIAN-PANNONIAN AREA

BY

MARCEL LUPU[1]

Although the Penninicum and its eastward prolongation in the West Carpathians on the one side, and the Vardar Zone and its southeast prolongation on the other side, are unanimously accepted to represent the Tethys and, in this way to separate Eurasia from Africa, in the Transylvanian-Pannonian area there still exist some incertitudes concerning this boundary. The connection between the Penninicum and the Vardar Zone, in this area, is presumed to be represented by the Transylvanides (Săndulescu, 1980, 1983). Another link between the Penninicum and the Vardar Zone was also supposed to be the Intra-Pannonian ophiolitic Zone (Horvath et al., 1977) or the Mecsek Trough (Szepeshazi, 1979). Meanwhile according to Channel et al. (1979) the Balaton-Darnó area should represent, during the Jurassic, a gulf of the main Tethys, starting from the Vardar Zone.

In the present paper, the author will take into account the present-day geological framework, especially the ophiolitic rocks zones — possible sutures — and will thus discuss the main problems concerning the Mesozoic history of this territory together with an attempt to find an explanation for some of them.

Geological Setting

The following three major areas which are strongly involved in the geological evolution of the territory will be discussed : the Vardar Zone, the basement of the Pannonian Depression and the Transylvanian area (Fig.).

The Vardar starts its Mesozoic history with a Triassic carbonate platform sedimentation connected in the Anisian ?-Ladinian with early rifting products represented especially by keratophyres and porphyrites. The "diabase-chert" formation of partly volcano-sedimentary nature is characteristic of the Upper Triassic, while in Yugoslavia the melange character (first olistostrome and then tectonic) and a more comprehensive age were considered (Dimitrievič, Dimitrievič, 1976).

The pile of ophiolitic rocks which overlies the diabase-chert formation is of Lower-Middle Jurassic age and provides oceanic crust

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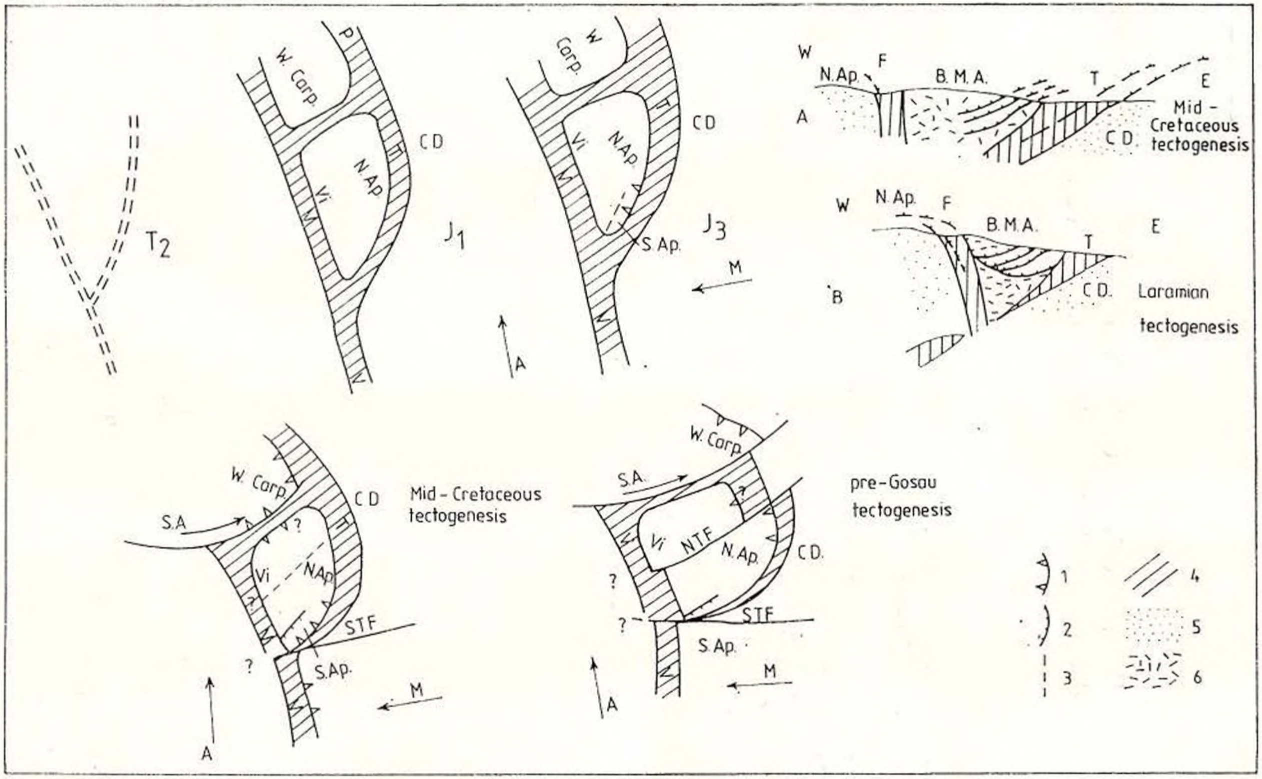

Fig. — Evolution of the East European continental margin in the Transylvanian-Pannonian area. A, B : Cross sections in the Transylvanian area.

W. Carp., West Carpathians ; Vi, Villány ; S. A., South Alpine ; N. Ap., North Apusenides ; S. Ap., Metaliferi Mountains (South Apusenides) ; C. D., Central Dacides ; V, Vardar Zone ; T, Transylvanian oceanic Zone ; P, Penninic Zone ; B. M. A., Bedeleu magmatic arc ; F, Feneş marginal Basin ; A, Adriatic promontory ; M, Moesian Platform ; N. T. F., North Transylvanian Fault ; S. T. F., South Transylvanian Fault.

1, active continental margin ; 2, minisuture zone ; 3, splitting area ; 4, oceanic floor area ; 5, continental crust area ; 6, magmatic arc area.

traits. The Upper Jurassic tectogenesis, the first important compressive stage, produced eastward vergent overthrusts and was accompanied by Paleoalpine metamorphism and also by granitisation. The Lower Cretaceous is marked also by some ophiolitic rocks and flysch formations. An important break representing a second tectogenetic phase appears at the end of Barremian. The sedimentation, usually marking a westward migration, starts again with the Aptian ?-Albian and in the northern part with the Cenomanian (Andjelković, Lupu, 1967). The Upper Cretaceous deposits feature a shelf or flysch-like facies.

Although in central and northern Greece several tectonic units have been distinguished, their northward correlation is still difficult, sometimes impossible.

The paleotectonic frame of the Vardar Zone continues to be a subject of various interpretations in the last years, most authors considering it as a complex area built up by marginal seas and island arcs (Dimitrievič, Dimitrievič, 1976, Mercier et al., 1975, Bebien et al., 1980, Karamata, 1930).

The prolongation of the Vardar Zone to the South Apuseni (Metaliferi) Mts in Romania was argued by Andjelković, Lupu (1967). Farther northwards, elements of the Vardar Zone are known as far as Zagreb where they are cut by an important fault — the Zagreb-Zemplin Line.

The Pannonian area consists of several tectonic units (Körössy, 1983) which can be divided — grosso modo — in three zones : the Transdanubian central range in the northwestern part featuring south Alpine facies ; the Köszeg hills situated west of the Rába Line are considered the easternmost Penninic Window ; in the Transdanubian range the Mesozoic is represented by full marine sequences of Triassic and Jurassic age (Transdanubian area).

In the east a zone with metamorphic basement, Paleozoic sequences, in which the Lower Triassic is detrital, becoming then limy-dolomitic. In the Mecsek Mts, the Middle Triassic — in which some tuffs are reported — features Muschelkalk similarities, the Upper Triassic is detrital, while in the Villány it misses. The Lower Jurassic, present only in the Mecsek Mts, is of Gresten type and is overlain by Fleckenmergel-like marls. The detrital-limy Middle Jurassic passes, in both areas, into a sometimes chert bearing limy facies. The lowermost Cretaceous of the Mecsek Mts is characterized by basaltic lavas and tuffs associated with detrital rocks. In the Villány some Neocomian bauxites (like in the Bihor Autochthon of the North Apuseni Mts) are overlain by trachydolerites. "Urgonian"- like Aptian limestones and Albian marls close the sedimentation in this area. Although stratigraphic and structural correlations between the Villány and the Bihor Autochthon have been made (Patrulius, Bleahu, 1967, Bleahu, 1976), as regards the Mecsek belt (Szepeshazi, 1979, Körössy, 1983) it has a NE prolongation situated westwards of the Apuseni ensemble.

The central part of the Pannonian area, which is bordered by the Balaton Line to the NW and the Zagreb-Zemplin Line (Horvath et. al., 1977) to the SW, is a complicated and disputed zone. Here, the "Igal-Bükk Eugeosyncline" (Wein, 1969) is either contested and interpreted only as a fracture zone (Kovacs, 1982) or admitted (Horvath et al., 1977, Körössy, 1983).

The Bükk Mts feature Carboniferous-Triassic marine sequences making thus accountable the attempts at paleogeographical and structural connections with the Dinarides (Wein, 1969, Channel et al., 1976, Horvath et al., 1977). In the Bükk Mts three magmatic complexes are known (Balla et al., 1981) : a first porphyritic one, of Middle Triassic age, a second one — diabase quartz porphyritic of Upper Triassic age, and a third one of Lower-Middle Jurassic age providing typical ophiolitic sequences associated with turbiditic ones. Anyway, the Hungarian geologists do agree that this zone separates two distinct realms in the west, a southern Alpine one and a more "European" one in the east, for which the name Tisia was also used.

The Transylvanian area consists of three main tectonic units, from W to E : Western Dacides, Transylvanides and Median Dacides (Săndulescu, 1980). The Supragetic nappes of the South Carpathians appear in the southern part.

The Western Dacides are represented by the North Apuseni Mts, where the Bihor Autochthon as well as the Codru Nappe and Biharia Nappe systems feature metamorphic covers of these units, as well as the stratigraphical or structural correlation with the West Carpathian Tatric nappes have been pointed out (Patrulius, Bleahu, 1967, Dumitrescu, Săndulescu, 1968, Săndulescu, 1972, Bleahu, 1976). This area has also been included in the Tisia (Kovacs, 1982). The tectogenetic phase which is responsible for the nappe structure is the pre-Gosau (pre-Coniacian) one.

The Transylvanides (Săndulescu, 1980) are considered to be built up of oceanic crust basement nappes. Northwards the Transylvanides link with the Penninicum (Săndulescu, 1980, 1983), while in the southwestern part the connection Metaliferi Mts (belonging to the Transylvanides) — Sumadija Zone is also known (Andjelković, Lupu, 1967).

The Transylvanides consist of two main nappe groups : in the east, the Transylvanian nappe system obducted eastwards over the Central East Carpathian nappe group during the Mid-Cretaceous (intra-Albian) tectogenesis (Săndulescu, 1980) ; the ophiolitic rocks of these nappes provide the Middle Triassic-Neocomian age (Doina Săndulescu et al., 1984). On the western side, in the South Apuseni (Metaliferi) Mts, the lower part of the ophiolitic complex is supposed to represent an oceanic crust (Herz, Savu, 1974), while Cioflica and Nicolae (1981) consider it as island arc magmatism, idea which is supported also by the author of this paper.

In this area the autochthon is represented by a metamorphic basement belonging to the upper Biharia nappes system units and supports two groups of nappes (Lupu, in Bleahu, 1981, 1983). The Criș nappes system consists of a pile of nappes with magmatic arc volcanics and marginal basin-type ophiolites basement supporting a sedimentary cover which starts with Callovian-Oxfordian jaspers and continues with Upper Jurassic limy deposits or, sometimes, calcareous flysch, Early Cretaceous flysch, locally spilites (in the Feneș Nappe). These nappes exhibit northward vergency, like in the Northern Apusenides, but the age of tectogenesis is Laramian. Evidence of reworked Mid-Cretaceous tectogenesis still exists in some of these nappes.

In the northeastern part of the Metaliferi Mts, the Bedeleu nappes group (Lupu, in Bleahu et al., 1981 ,1983) consists of several nappes featuring both metamorphic and Upper Jurassic keratophyric rocks basement. The stratigraphic sequence starts with *Aptychus*-type beds or massive Upper Jurassic limestones, followed by Early Cretaceous flysch-like deposits. These nappes featuring, as a whole, a magmatic arc environment were first thrust eastwards during the Albian (pre-Vraconian) tectogenesis. During the Laramian tectogenesis this group of nappes was overthrust westwards.

The Metaliferi Mts, as a whole, have been interpreted as the western active continental margin of the Transylvanian oceanic Basin (Lupu, 1983).

Suture Zones

The Transylvanian-Penninic (Săndulescu, 1980) and Metaliferi Mts-Sumadija-Vardar (Andjelković, Lupu, 1967) connections lead to the interpretation according to which the Transylvanides represent the main ophiolitic-Tethyan suture (Săndulescu, 1980, 1983). The Transylvanian segment of this suture zone provides some distinct traits with regard to the Penninic one. While in the eastern Alps the suture zone is overtaken by the overthrust of the Austroalpine nappes and appears only in some tectonic windows (the easternmost is the Rechnitz-Köszeg Window), in the Transylvanian area the oceanic basement nappes (Mid-Cretaceous) are overthrust eastwards. It is almost probable that this might be due to the steeper subduction plane in the Transylvanian area (starting south from the Zagreb-Zemplin Line or from the North Transylvanian Fault ?).

The eastern continental margin of the Transylvanian oceanic Basin is considered a passive one (Săndulescu, 1983), while the western one can be considered, at least from the Upper Jurassic, an active margin (Lupu, 1983). The magmatic arc character and the eastern Mid-Cretaceous vergencies of nappes within the Bedeleu Group, in the northern part of the Metaliferi Mts, as well as the magmatic arc and the related marginal basin behind it, of Middle ?-Upper Jurassic age confirm this supposition as well as the western dipping of the subduction plane (Rădulescu, Săndulescu, 1973). Southwestwards the subduction plane disappears along the Mureş Fault (South Transylvanian Fault — Săndulescu, Visarion, 1978). It is noteworthy that along this dextral fault disappear also the units belonging to the East Carpathian Transylvanides. The beginning of the movements along this fault is uncertain (Late Jurassic or earlier ?), but its ceasing is connected with the Mid-Cretaceous tectogenesis. This is proved by the unconformable overlying Upper Albian (Vraconian) deposits, in the Mureş Valley, on the ophiolitic rocks belonging to the Metaliferi Mts, as well as on the Supragetic metamorphic rocks of the South Carpathians. Southwestwards, the obliteration of the subduction area along the Mureş Fault is followed by the reappearance of the suture in the Sumadija-Vardar Zone.

Before pursuing the main suture zone in the Vardar area, it is to notice the presence in the Metaliferi Mts of two minisuture zones in which marginal basin-like floor was consumed during the Mid-Cretaceous events (actually the Criş and Feneş nappes). The NW vergencies are different from those of the Bedeleu Nappes and provide "Tatric" influence.

In the inner Vardar Zone, where the trough was closing during the Albian, the subduction plane features an eastward dipping, like all over the Dinarides. The ophiolitic nappes of this zone provide similarities with those of the East Transylvanides ones, exhibiting the same dipping of the subduction plane. Along the Vardar suture Zone dextral longitudinal movements are reported (Dimitrievič, Dimitrievič, 1976). Whether the Mureş Fault can be pursued to the Vardar Zone is still an unsolved problem ; anyhow these faults made difficult the correlations in this zone.

The changing of the subduction plane from the Transylvanian area to the Dinarides is due to many factors, among which the westward advancing Moesian plate on the one side, and the northward advancing Adriatic plate on the other side have to be considered.

In the basement of the Pannonian Depression the Igal-Bükk area constitutes doubtless a suture zone (Kovacs, 1982).

In this area comprised between the Balaton and Zagreb-Zemplin lines, there may exist two distinct elements : the Dinaric connected ones and the "autochthonous" Pannonian ones.

In the Bükk Mts area in which the Triassic volcanics and Jurassic Szarvasko Ophiolites as well as the south vergencies plead for a former Dinaric position (Kovacs, 1982) ; their present-day position is due to a SW-NE rotation along the Balaton Line, together with the Transdanubian area. Anyhow, the rotation is not earlier than the Lower Cretaceous because the south vergent structures in the Bükk area were, probably, created during the Upper Jurassic tectogenesis when they had their original position.

Balla (1982) argues for an westward subduction which started during the Late Jurassic in this Mid-Pannonian Trough, considering that the southeastern flank was passive. But the opening of the Mecsek Basin leads to the supposition of a possible active continental margin on the eastern flank of the mentioned trough.

Attempt at Geodynamic Interpretation

The Triassic Tethyan splitting and early rifting prefigured the areas in which later, during the Lower Jurassic, oceanic crust appeared. At that time, the North Apuseni Mts as well as the West Carpathians-Tatric-Choč ensemble and the Austroalpine realm must have belonged to a continental Intra-Tethyan Microplate, which may correspond — grosso modo — to the northern part of Kreios Plate (Tollman, 1978) with the difference that according to the author's opinion the North Apuseni Mts represent its southern boundary. This accounts for the similitudes of the Triassic-Lower Jurassic facies in the mentioned areas.

During the Lower Jurassic, the oceanic crust areas seem to have been well defined. Hence, starting from the south, the Vardar-Transylvanid-Penninic connection appears doubtless. A second, more northward situated bifurcation of the Vardar Zone which corresponds to the Igal-Bükk Zone detached the southernmost part (East Pannonian territory and North Apuseni Mts) of the above mentioned microcontinent.

Coming back to the expansion areas during the Jurassic the behaviour of the continental margins and that of compression are still uncertain. Thus, if we take into account that in the Greek Vardar Zone the Late Jurassic tectogenesis involves a complicated continental margin frame, featuring island arc and marginal sea (Mercier et al., 1975), it can be supposed that the active continental margin must have appeared earlier — even during the rifting ? Anyhow, on the western margin of the Transylvanian Basin (Metaliferi Mts) island arc tholeiites (Cioflica, Nicolae, 1981) are reported to the Lower-Middle Jurassic (Herz, Savu, 1974).

As in the Pannonian realm there are no data to provide any compressive movements, it is to suppose that during the Jurassic, expansion took place between two separate areas — the present-day East Pannonian area and a western one, probably the West Carpathian Austro-alpine units.

As mentioned before, during the Mid-Cretaceous tectogenesis the obliteration of the Transylvanian subduction plane and the forearc-trench system of the Metaliferi Mts (if the whole frame existed) was accomplished along the Mureș (South Transylvanian) transform Fault. The existing data support also the idea of a clockwise rotation of this territory.

In the Pannonian area it is possible that the northwestward subduction (Balla, 1982) was also related to a dextral movement along the Periadriatic-Balaton Line.

In the inner Vardar Zone compression took place, in the northern part, during the Aptian ?-Albian, followed by a westward migration of sedimentation.

During the pre-Gosau tectogenesis both the West Carpathians and the North Apuseni were overthrust northwards. In the East Pannonian area such structures are not known, but probably this territory, situated between the Zagreb-Zemplin Line and the North Transylvanian Fault (Săndulescu, Visarion, 1978) remained, more or less, in its original place (Bleahu, 1976). At that time, took probably place the most important movement along the Periadriatic-Balaton Fault which produced the replacement of the Austroalpine by the south Alpine zone. The presence of the Penninicum in the western part of Hungary suggests the possibility that the area involved in this northward movement of the south Alpine realm was bordered, in the west, by the Rába Line.

The Laramian tectogenesis which produced a strong compression in the Sumadija Zone is also responsible for the main nappe structure in the South Apuseni (Metaliferi) Mts. The west and northward vergencies of these nappes can be reported to the final consumption of the ocean-like crust of the two above mentioned marginal sea areas in which the subduction plane had a south and southeastern dipping. The small amplitude of these nappes expresses the small size compression and the lack of any influence in the North Apusenides.

By summing these data up the author considers that the realm under discussion belongs entirely to the Tethys which has to be interpreted as a complex realm featuring oceanic crust troughs, island arcs, marginal seas and including also microcontinents or small continental shelves (see also Tollmann, 1978).

Conclusions

The author considers that the Tatric-Austroalpine ensemble, the East Pannonian and the North Apusenides represent the southern part of a microcontinent bordered, on the one side, by the Penninic-Transylvanian, and on the other side, the Northern Vardar oceanic troughs. The East Pannonian and North Apusenides were detached from this microcontinent probably during the Lower Jurassic.

The western continental margin of the Transylvanides was an active one, at least starting from the Upper Jurassic, maybe also earlier. From this point of view the Transylvanides may include not only ocean crust nappes, but also tectonic units of magmatic arcs and marginal sea character.

Starting with the Transylvanides southwards the Vardarian character of the subduction plane does appear.

The suture zones of this area are frequently associated with fault systems, many of them longitudinal, the whole ensemble being affected by the movements of the Moesian and Adriatic plates.

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SOME CHARACTERISTIC FEATURES OF THE SOUTH CARPATHIANS

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The geology of the South Carpathians is well known. In the following pages we are going to present some new data and to discuss those elements of peculiar interest in the light of global tectonics theories.

Main Constituents of the South Carpathian Chain

The apparently unitary chain of the South Carpathians resulted from the joining of several microplates or blocks, with different structure and origin (Pavelescu, Nitu, 1977). The inlay aspect is blurred by several overthrusts of vast proportions ; however, it may be easily noticed by reconstructing the situation prior to Cretaceous diastrophism. Following a N-S line which includes also the adjoining structures, there are the following blocks, which, except one, coincide with the present-day structural units : Northern Apuseni Mountains ; Southern Apuseni Mountains ; Getic Domain ; Severin Domain ; a block supposed to occur between the Danubian Domain and the Moesian Platform ; Moesian Platform. (Fig.)

These blocks show alternatively continental and oceanic structures. The structure of the South Carpathians includes only the deformed remnants of the Getic, Severin and Danubian domains. The three northern continental blocks exhibit a structure typical of pieces torn off the Epi-Hercynian European Plattform. They show a basement built up of crystalline schists, which usually include metamorphic schists of Lower Paleozoic age to Lower Carboniferous inclusively. There is one instance, Drocea, where Hercynian granites are also present. These blocks exhibit a sedimentary cover which starts everywhere with Upper Carboniferous and Permian continental formations, considered to represent the Hercynian molasse. The cover continues

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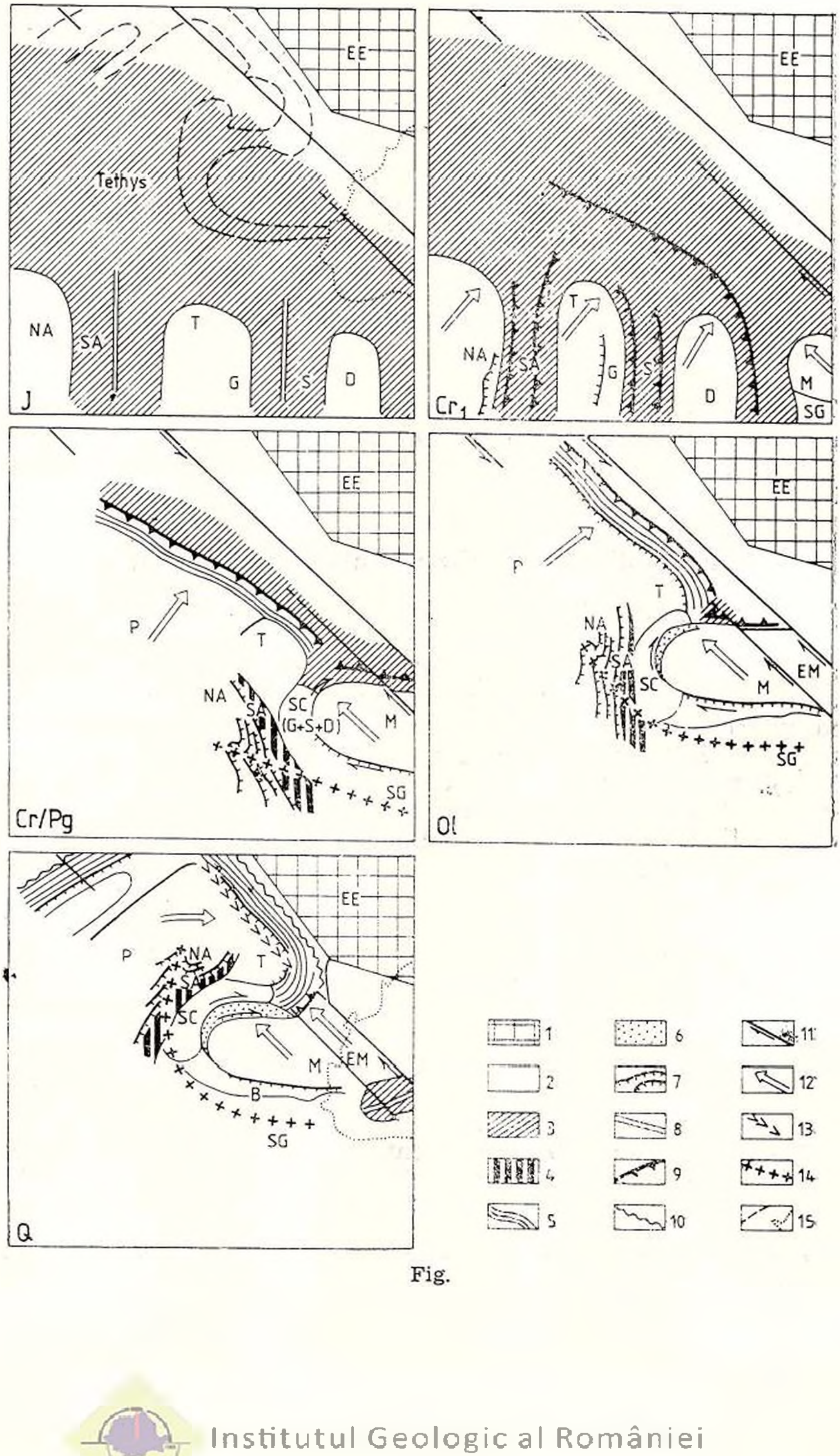
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Fig.

with mainly carbonate deposits, Mesozoic in age. It is worth mentioning that even after Alpine deformation, the cover of these blocks shows low dipping (the only important exception is represented by the Reşiţa-Moldova Nouă Zone in the Getic Domain).

It is difficult to state the source of these fragments. The basement and the Upper Paleozoic cover, although similar on the whole, show enough differences which account for stating that at the margin of the European Platform, where they come from, they were not adjacent. These blocks exhibit similarities with the structure of the Pannonian Massif, probably torn off the same platform.

The structure of the Moesian Platform is obviously different from the structure of other continental blocks. In its western area, the object of our study, the Precambrian metamorphic basement is overlain by a sedimentary cover 7—8 km thick, which includes an almost complete sequence of Paleozoic formations, starting from Cambrian, and Mesozoic ones. One should also note the epicontinental series of Upper Permian, Triassic or, according to some data, Lower Liassic age which includes detrital rocks, striped clays, dolomites, anhydrite, salt, some basic and acid lava flows.

There are obvious differences from the adjoining Danubian Domain. At the Lower Liassic level the rocks of the Moesian Platform and of the Danubian Domain point to areas with different climate. The variations of facies in these two units, Moesian Platform and Danubian Domain, are less intense than the differences between them. It is obvious that between the Moesian Platform and the Danubian Domain there was a space which disappeared due to its oceanic nature only (Niţu, 1977, Pavelescu, Nitu, 1977, Cioflica et al., 1980).

Practically it is impossible to determine the width of this area. The structure of the Moesian Platform points to the northern border of the African Platform, while according to other elements, it may not be located too far southwards. Thus we note : the occurrence of greenschists of Podolian type in the west of the platform (one may never say whether its two parts had ever been united), some facies

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Fig. — Supposed evolution of the Carpathian-Balkan area during Mesozoic and Alpine times (acc. to Nitu, 1977, modified).

SC, South Carpathians ; EE, East European Platform ; NA, North Apuseni Mts ; SA, South Apuseni Mts ; P, Pannonian Massif ; T, Transylvanian Massif ; G, Getic Domain ; S, Severin Domain ; D, Danubian Domain ; M, Moesian Platform ; EM, eastern area of the Moesian Platform ; B, Balkans ; SG, Srednegora.

1, thick continental lithosphere of the East European Platform ; 2, continental lithosphere ; 3, oceanic lithosphere ; 4, overthrust oceanic crust lambeau ; 5, folded flysch and molasse deposits ; 6, unfolded metamorphic rocks of the Marginal Depression in the South Carpathians ; 7, overthrusts ; 8, spreading ; 9, subduction ; 10, restriction of space due to lithosphere thickening ; 11, transform faults ; 12, relative direction of shifting of blocks ; 13, Neogene volcanics in the East Carpathians ; 14, chain of Laramian intrusions (banatites) ; 15, projection of present-day contour of the Carpathian-Balkan arc and of the Black Sea.

similarities with Paleozoic rocks in Moldavia, the occurrence of some European blocks to the south of the Moesian Platform.

Out of the three blocks which exhibit oceanic crust, the best represented one in the present-day geological structure is the Southern Apuseni Mts block. A remnant of this block, which occurs as a large ophiolite lambeau including Upper Jurassic to Senonian, inclusively, sedimentary rocks, in carbonate facies, in places flysch and wildflysch ones, overthrust northward the already burdened border of Northern Apuseni Mts.

The Severin Domain is represented by the nappe bearing the same name (Codarcea, 1940) which thrusts southwards over the Danubian Domain and is overlain by the Getic Nappe from the Getic Domain, recognized by Murgoci since 1905 (Murgoci, 1905). The Severin Nappe, restricted to a relatively small area, consists of some ophiolite bodies at the base and mainly of Lower Cretaceous flysch deposits.

No remnants are known from the block with oceanic crust which seems to have existed between the Danubian Domain and the Moesian Platform. They are probably covered by the Marginal Depression, filled by a pile of Tertiary sediments, 6—7 km thick.

Main Stages of Chain Building

The following three main stages are known : the tearing of continental blocks ; the joining of blocks and the subsequent development of the chain.

It is difficult to state the time of tearing of blocks. The oldest formations from the remnants of oceanic domains in the Southern Apuseni Mts are of Lower Jurassic age. Although the oldest oceanic formations of the block are not supposed to be preserved, this age should be viewed as upper limit.

The tearing of blocks is possible to have been preceded by the constitution of faults, grabens and rifts within an emerged continent, The Hercynian molasse, deposited by means of faulting on an uneven relief, may also account for a platform regime in course of reactivation. Upper Permian and Triassic formations do not occur, but for some places, in the Getic and Danubian blocks. It is possible that by that time the northern continental blocks, maybe the Moesian Platform as well (Pavelescu, Nitu, 1977), tore off.

The continental magmatic activity contemporaneous with the tearing of blocks is very reduced, however there are some alkaline intrusions in the east of Făgăraș, on the Lotru River and in the Mraconia Zone near the Danube.

The tearing of blocks probably coincides, on the whole, with the opening of the initial rift in the central area of the Atlantic Ocean, between North America and Africa and with the continuous stretching of the Tethyan Ocean (Pitman, Talwani, 1972 ; Zonenșain, Savostin, 1979). The blocks continued to be removed from their initial position during the Jurassic, concomitantly with the left direction shearing movements between Europe and Africa (Dewey et al., 1973).

We agree with other authors (Bleahu et al., 1973 ; Herz, Savu, 1974) that the reversed tendency from extension to restriction of space took place by the end of Jurassic. The Lower Cretaceous formations are frequently represented by turbidites characteristic of a very contrasting relief. They include, in places, olistoliths which point to the beginning of overthrusts and of compression. The features of some overthrusts, such as those of the ophiolites overthrusting the Neocomian flysch in the Drocea Mts, show that overthrusting began concomitantly with the constitution of overthrust sedimentary formations. Peive (1969) studies a wider area and assigns the beginning of intense tectonic movements, which altered the structure of the Tethys, to Lower Cretaceous. The restriction of the space continues as far as Upper Cretaceous.

It is to note that during the Cretaceous the coming together of the two continental blocks on the outer side of the South Carpathians, namely between the Northern Apuseni and the Moesian Platform, was very important.

The main restriction of space was obtained at the expense of the oceanic area which contained the continental blocks. The latter underwent also important narrowing. On the southern border of Northern Apuseni Mts the superposed overthrusts point to a narrowing of the block of at least 100 km. The Getic Domain was also narrowed, as shown by the occurrence of the Supragetic Nappe and of other overthrusts as well as by the foldings in the Reșița-Moldova Nouă Zone. The Getic Nappe overlies the Danubian Domain horizontally on at least 70 km. The Danubian Domain is the only one which seems not to lack in width.

One may consider that during the Cretaceous, the Moesian Platform came closer to the Northern Apuseni. as well as to the Pannonian and Transylvanian massifs, by more than 700 km or even more than 1000 km.

It is to note the tardy nature of the magmatic activity of the same date as compression. During the Cretaceous, prior to the Turonian, the continental blocks were practically attached to each other due to magmatic processes. It is during the Turonian-Senonian and the Paleocene that the effusive banatite series and the intrusive one respectively, of calc-alkaline nature, were formed. Rădulescu and Săndulescu (1973) are the first who considered them to be the result of subduction during Austrian and Laramian phases, respectively. The problem of a definite relationship with a certain subduction area is much debated upon. One should take into account the regional occurrence of banatites from Banat and Bulgaria to the Black Sea coast (Giușcă et al., 1966), their unconformable position to the ophiolite sutures, their retardation as compared to the movements during Lower Cretaceous, the fact that they were formed at the time when the region was a continent on the whole and finally the difficulty of agreeing on the subduction zone. Having these in mind, we do not relate the banatites to the disappearance of one of the three oceanic basins (Nitu, 1977), but to a wider background.

Anyhow, the lack of magmatic processes during the Cretaceous is still a problem to be solved. This may be due to several causes, as follows :

The northward advancement, by means of obduction, of the ophiolites in the Southern Apuseni Mts, was related to the southward subduction of the oceanic border of Northern Apuseni microplate (Nitu, 1977). Thus the magmatic products of subduction remained on the continental block. Otherwise, among ophiolites one notes a calc-alkaline sequence (Savu, 1980) which discloses subduction phenomena.

In the South Carpathians, the southward advancement of the Severin Nappe is also the result of obduction, probably encountered by the subduction in the opposite direction of the overthrust border.

The magmatic activity consecutive of subduction was delayed by important oriented pressures present at that time (Nitu, 1977). The occurrence of banatites in Srednegora is related, as already shown, to the generation of several deep faults during the relaxation of tension.

The second stage, of compression, coincides with the beginning of closing of the Tethys ocean in the early Cretaceous and with the main phase of closing of this ocean at the end of Cretaceous (Pitman, Talwani, 1972 ; Zonenșain, Savostin, 1979).

It is generally considered that the third stage of evolution of the South Carpathians lacks in important geological events. It has been shown (Pavelescu, Nitu, 1977) that this image of tectonic quiescence is delusive.

The effect of Tertiary movements on the inner structure of the South Carpathians is quite discreet. The presence of these movements is undoubtedly accounted for by the study of East Carpathian structure, which form an almost straight angle with the South Carpathians.

The characteristic feature of the East Carpathians is represented by a series of isoclinal folds and east trending overthrusts which wrap the flysch and molasse sediments of Cretaceous and Tertiary age, forming a complete stratigraphic sequence. The oceanic basement overlain by sediment piles disappeared by subduction (Bleahu et al., 1973 ; Rădulescu, Săndulescu, 1973 ; Herz, Savu, 1974). The sediments preserved are a rare marker of the disappeared area. The display of folds and overthrusts (Nitu, 1977 ; Pavelescu, Nitu, 1977) accounts for a restriction of space of at least 200—250 km, without taking into account the eroded areas from nappe fronts or their extensions at depth. Folding was accompanied by intense or quiet stages all the sedimentation stage through. Part of space restriction belongs to the Cretaceous and another part, of more than 150 km, to the Tertiary.

The Carpathian bend area represents, in a simplified manner, the place of triple junction among Transylvanian block, East European Platform and Moesian Platform. The boundary between the first two ones in a consumption boundary. The East European and Moesian platforms did not undergo movements of coming together during the Cretaceous and the Tertiary. The movement of the Transylvanian block and of Northern Apuseni Mts to the East European Platform must have taken place on the third boundary, representing a sort of transform fault on continental background.

During the Cretaceous, prior to the complete closing of oceanic basins in the South Carpathians, the faults in these basins could have favoured the right-side displacements of northern blocks. During the Tertiary, everything was completely altered. The horizontal shift on more than 150 km affected blocks with continental lithosphere in close contact, characterized by overthrusts passing from one block to another and covering the boundary in between under the conditions of continuous oriented pressure.

The Tertiary history of the Carpathians was influenced by the continuing closing of the Tethys Ocean, accompanied by gradual reduction of oceanic crust areas. Starting with the stage in which the reduction of space consumed on a contact, posterior to the collision of continental sides of moving blocks — in the South Carpathians. in our case — the narrowing continued, as late as the Tertiary, on other contacts with oceanic lithosphere remnants (East Carpathians), the first contact acting as transform fault.

Orientation of Structures in the South Carpathians

Just like in many other chains, the structures of the South Carpathians exhibit an orientation parallel to the main direction of the mountain arc. This is true of different structures (faults, cleavage schistosity, elongation of intrusive bodies, elongation of mountainous depressions) of most different ages. Even the Baikalian granite intrusions exhibit usually elongations following the chain direction and often crystalline schists enclaves follow the pluton elongation (Pavelescu, 1971). The pre-Alpine folds, such as the Tulişa Series syncline, exhibit the same trending. The Cerna Fault in the Danubian Domain and the faults in the Marginal Depression are curved and parallel to the chain. It is worth mentioning the coincidence between the trending of the chain and the schistosity trending of Danubian Crystalline. which is characterized by the oldest structures and was also sunk below the overthrust masses.

It is obvious (Pavelescu, Nitu, 1977) that the different structures of the chain exhibit Alpine orientation which discloses the results of rock displacements to the East Carpathians during the Tertiary. The eastward movement of the Transylvanian block and Apuseni Mts as compared to the Moesian Platform did not follow the direction of a unique fault, but resulted from the summing up of innumerable partial displacements. The basic rejuvenation of structures is also proved by K/Ar absolute age determination (Pavelescu, Nitu, 1977). It includes practically all metamorphic formations and is more important in the case of rocks with marked schistosity than in the case of those with massive texture.

The study of the movement of matter in the South Carpathians during the Tertiary makes us adopt Peive's (1967) concept of tectonic flowage. The flowage was more intense in some zones and very reduced in others, such as in the Tismana granitoid massif, characterized by isometric shape and massive texture.

The Carpathian oriented structures, more accentuated in the Danubian Domain, agree with the unconformable nature of surface and deep structures. The overthrusts do not undergo lateral compression, while the block situated at depth, the Danubian Domain, undergoes oriented pressure and then shearing.

The coincidence which is frequently present between the trending of old structures and the direction of young chains is probably due to the recent ranging of old structures under the influence of new tendencies.

Some Features of Vertical Movements

The elevation of the South Carpathians during the Tertiary represents a phenomenon consecutive to the thickening of the crust and lithosphere during preceding stages. At the same time, the features of rocks overlying the continental and oceanic blocks show that till the Cretaceous they had a position which did not break the isostatic equilibrium. During Upper Permian, Triassic and Lower Jurassic the continental blocks were partly emerged and the region formed an archipelago.

During the active stage of compression, there are instances in which deep sea sediments deposit on a metamorphic basement, within continental blocks. On the southern border of Northern Apuseni Mts flysch deposits, of Tithonian to Cenomanian age inclusively, belonging to the Bucium Unit and partly to the Barremian-Lower Aptian spillite olistostrome formation, the Aptian flysch and Albian wildflysch with ophiolite olistoliths, belonging to the Feneş Unit (Lupu, 1976) deposited on a basement of this type.

The Danubian Domain exhibits a lowered position during the Senonian overthrust of Severin Nappe and Getic Nappe, when the Mesozoic calcareous cover of this block was overlain by a Wildflysch Formation with big ophiolite blocks.

The Tertiary deposits of the Marginal Depression do not exhibit the features of a deep sea, but their thickness points to a significant lowering of the Moesian Platform border.

On the other hand, obduction was accompanied by or led to the elevation of the border of the oceanic lithosphere block. It is obvious that during compression stages, some blocks had, at least temporarily, an antiisostatic position. One way or another, this phenomenon discloses the primary part played in most cases by horizontal displacements and pressures in bringing about the vertical movements.

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Tectonique

COMPARED ALPINE GEOTECTONIC MODELS

BY

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Since the development of the plate-tectonic concept applied to the analysis of different alpine models, few authors insisted on the recalibration of the "classical" principles of geotectonics. Thus, some twenty years ago very common and circulated notions are practically out of interest now. In fact, working with the same folded chains implies a permanent comparison between the old and the recent geotectonic models. Such a compared analysis may be followed from different points of view ; we will try to enlarge upon some of these problems, as for instance the symmetry of the folded chains, the development of the crust in a mobile area, the structure and the morphology of continental margins of a mobile belt, etc.

The Symmetry of the Folded Chains — the Geosuture and the Vergency Problems

Well known are the classical models of Kober, which point out the bilateral building of the Alps, with two branches, versus Argand's model, stressing out the unilateral structure of the same chain. The dispute was mainly focussed on the geotectonic significance of the Ivrea and Gailthal fracture zones ; it still continues with different arguments and with a more general acceptance of Argand's model. Transfering the problem to some other areas, as for example to the Carpatho-Dinaric transversal, the bilateral symmetry is more easier to follow mainly concerning the structural vergency.

Transposing the problem to the plate-tectonic concept framework, one of the problems concerns the prior symmetry of the mobile zone before the compressive period which generated the actual main structural units of a folded chain. From this point of view the oceanic floor area and its continental margins are primarily important in the retrotectonic pictures of the mobile areas.

Following the above mentioned facts it is possible to distinguish : (1) a paleotectonic symmetry concerning the continental mar-

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gins reported to their adjoining oceanic areas and (2) a vergency symmetry. In fact, these two cases are two stages of the process of development of a mobile (geosynclinal) zone into its corresponding folded (orogenic) belt. In both situations the symmetry plane may be actually recognized in the ophiolitic sutures. In both stages the symmetry may be bilateral or unilateral.

The bilateral paleotectonic symmetry is well expressed in many retrotectonic pictures concerning practically all the segments of the Tethyan Alpine Chains (Fig. 1). Commonly the continental margins are, in these models, of Atlantic type. The convergent, subduction type, plate boundary is less frequent and usually locally developed during the spreading (distension) period. A peculiar bilateral paleotectonic symmetry is expressed by the intracratonic mobile belts, as for instance the North Dobrogea Orogen (Fig. 1). These are paleorift-valley areas set inside a continental plate, showing more (Red Sea type) or less (East African type) extension, crustal thinning and stretching.

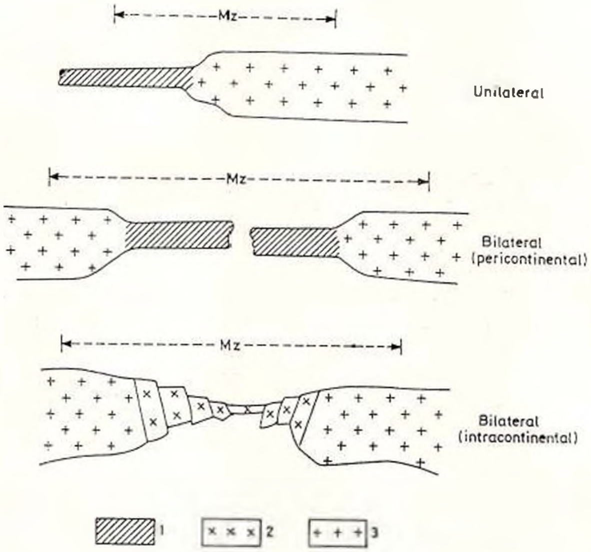

Fig. 1 — Types of paleotectonic symmetry.
1, oceanic crust ; 2, thinned crust ; 3, continental crust ;
Mz, mobile zone.

The paleotectonic bilateral symmetry may be expressed also in the detail structure of the continental margins. Following, for instance, the Carpatho-Dinaric transversal, inside the two continental margins developed intracontinental rift zones elongated generally parallel to the ocean/continent boundary.

The vergency symmetry (Fig. 2) of the orogenic belts may be also unilateral or bilateral. Both situations concern different paleotectonic pictures of the former mobile zones.

The unilateral vergency symmetry types are the Alps, with the suture zone sequeezed between two continental shearing groups of nappes (Fig. 2 b), or the Norhern Andes, with an ophiolitic nappe complex obducted on the continent (Fig. 2 a).

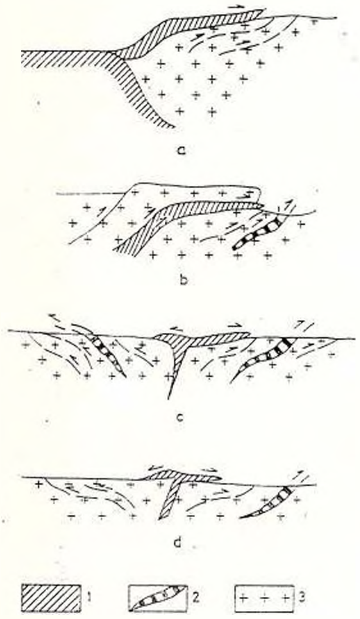

Fig. 2 — Types of vergency symmetry.
1, oceanic crust ; 2, thinned crust ; 3, continental crust : a, unilateral symmetry (North Andes model) ; b, unilateral symmetry (Alps model) ; c, bilateral symmetry (Carpatho-Dinaric model) ; d, bilateral symmetry (Balkano-Hellenic model).

The bilateral vergency symmetry is well expressed in the Carpatho-Dinaric or Balkano-Hellenic şectors, as well as in Minor Asia. There are different types of bilateral symmetry (Fig. 2 c, d) with respect to the more or less complex structure of the two continental margins involved in the orogenic chain deformations.

The above mentioned examples stress out that the symmetry problem is still important and that the solution which may be accepted takes into account the suture zones as reference planes.

Morphology and History of the Continental Margins

Following the retrotectonic pictures established in different sectors of the European Alpine Chains, it is possible to conclude that the continental margins of the oceanic Tethys show different morphologies and structures.

One of the most important facts is the presence inside the continental margin areas of the rift-valley structures (Fig. 3). They were

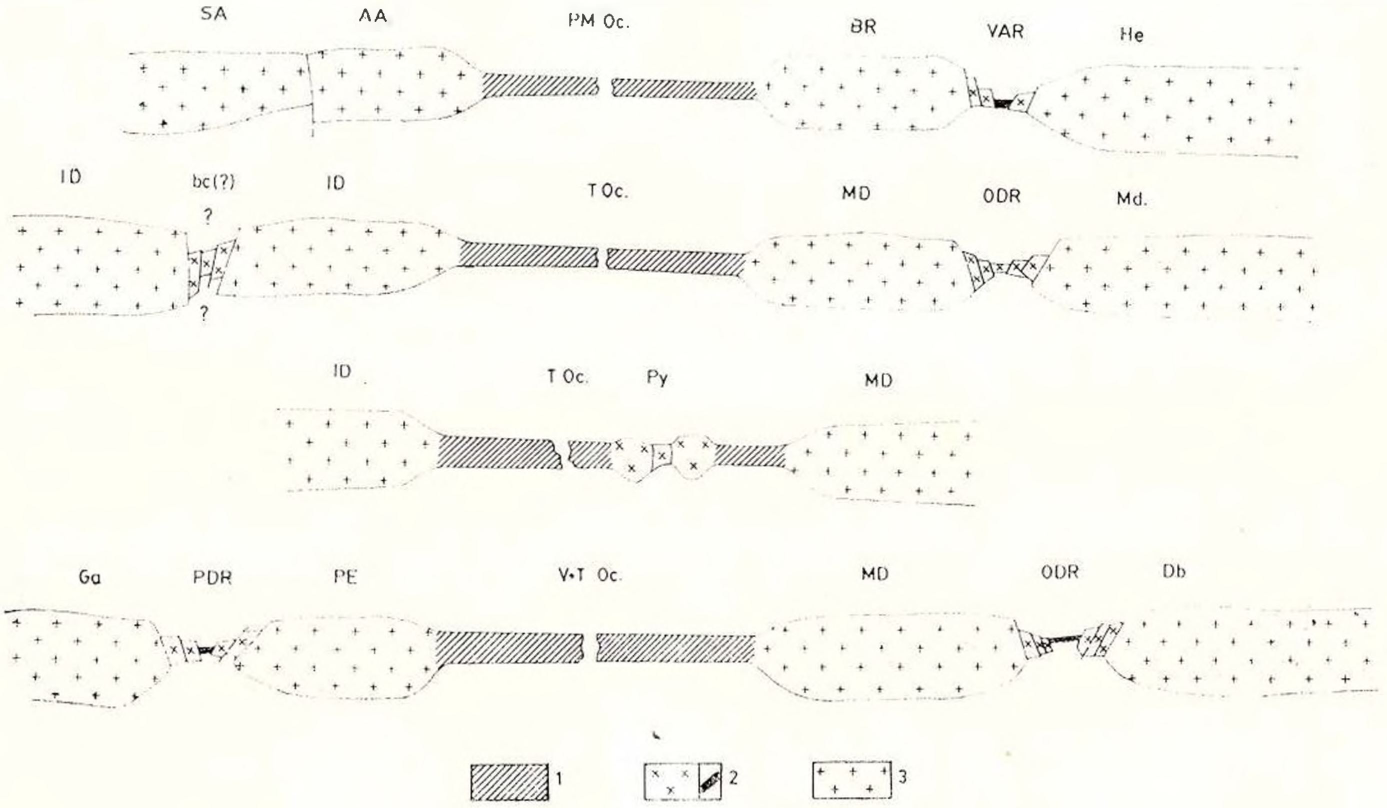

Fig. 3 — Morphology of the mobile zones.

1, oceanic crust ; 2, thinned crust (a, oceanic type) ; 3, continental crust ; SA, South Alpine ; AA, Austroalpine ; PM Oc, Piemontais Ocean ; BR, Briançonais ; VAR, Valais Rift ; He, Helvetikum ; ID, Inner Dacides ; bk, Bükk ; T Oc, Transylvanidian Ocean ; MD, Median Dacides ; ODR, Outer Dacidian Rift ; Md, Moldavides ; Ga, Gavrovo ; PDR, Pindus Rift ; PE, Pelagonian Zone ; V+T Oc, Vardar and Transylvanidian Ocean ; Db, Danubian ; Py, Pienyny (s.l.).

generated concomitantly with the opening and the spreading of the Tethyan Ocean (e.g. Pindus Rift in the Dinaro-Hellenic area) or are delayed with respect to the opening and contemporaneous with younger stages of spreading (e.g. Outer Dacidian Rift in the Carpatho-Balkan area).

Occurring in an Atlantic type continental margin, these (paleo) rifts may be compared with the actual Red Sea-Afars Rift or Nova Scotia Rift, which are also situated in the vicinity of the ocean/continent boundary.

These Tethyan (paleo) rifts hosted different types of sedimentation. Thus in the Outer Dacidian Rift only flysch sequences were sedimented, while in the Pindus Rift calcareous formations were also deposited together with flysch ones. Another model is the Valais Rift, defined in the Alps paleogeography where preflysch is followed by flysch formations.

The amount of distension in these Tethyan rifts was different along the same rift or from one rift to the other. Consequently, thinned crust or ocean type crust was generated. An interesting example is the Outer Dacidian Rift. In its East Carpathian sector (Black Flysch and Ceahlău nappes) it was characterized by thinned crust intruded and/or covered by basaltic igneous rocks (intraplate geochemical features). Along this sygmoidal-shaped rift, the picture changes in the South Carpathian segment where also oceanic type ophiolitic complexes are known (Severin Nappe) (Fig. 4). A similar situation seems to show the Pindus Rift on the opposite continental margin of the oceanic Tethys. Morphologically the two mentioned rifts are symmetrically situated (Fig. 3), their difference consisting mainly in the history of distension and sedimentary processes.

A peculiar rift-valley zone is that one corresponding to the North Dobrogea Orogen and mostly to its central unit, the Niculițel Nappe. Generally the mobile (geosynclinal) zone corresponding to this orogenic chain had an intracontinental (intracratonic) position (Fig. 1). In its central part is developed (Fig. 4) a rift with thinned crust and Ladinian-Carnian basaltic (intraplate features) effusions, followed by a flysch type sedimentation (Upper Triassic-Lowermost Jurassic). The differences between the rifts situated inside the continental margin area in the vicinity of the continent/ocean boundary and those situated farther inside the continental plate may be summarized as follows :

— the intracontinental margin rifts are involved, by tectogenetic compressions, in the same orogenic belt with the deformed neighbouring ocean crust ; they represent only a minor part of these complex folded chains ;

— the intracontinental (intracratonic) rifts represent the main part of the mobile zone from which were generated the intracratonic folded (orogenic) belts ;

— the suture corresponding to an intracontinental rift represents the symmetry element of the folded chain, while the suture corresponding to an intracontinental margin rift is parallel and collateral to the main ophiolitic suture of the folded chain ;

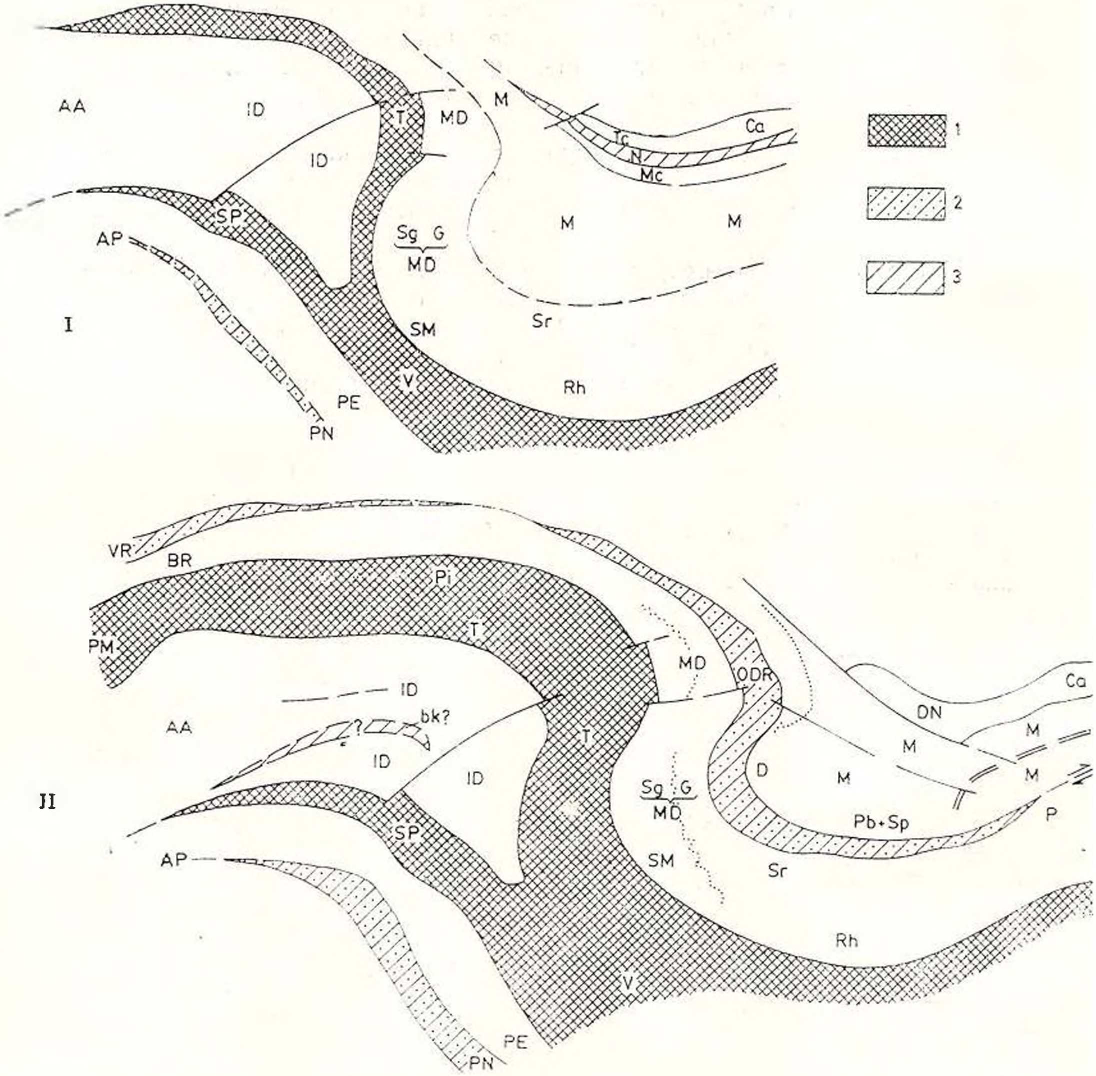

Fig. 4 — Palinspastic model of the central and southeast European Alpine Chains. I, Triassic ; II, Jurassic-Cretaceous boundary.

1, oceanic Tethys ; 2, intracontinental margin rifts ; 3, intracontinental (intracratonic) rifts.

AA, Austroalpine ; ID, Inner Dacides ; bk, Bükk ; AP, Apulia ; PN, Pindus Rift ; Pi, Pienyny (s.l.) ; T, Transylvanides ; V, Vardar ; SP, Pannonian Sphaenocasm ; MD, Median Dacides ; Sg, Supragetic ; G, Getic ; SM, Serbo-Macedonian ; Sr, Srednegorie ; Rh, Rhodope ; D, Danubian ; Sp, Stara Planina ; Pb, Prebalkan ; M, Moesian Platform ; ODR, Outer Dacidian Rift ; Mc, Măcin Unit ; N. Niculițel Unit ; Tc, Tulcea Unit ; Ca, Alpine Crimea ; DN, North Dobrogea ; P, Pontides.

— the tectogenetic history (compressional deformations) of an intracontinental margin rift is similar to or, at least, comparable with those of the neighbouring ocean, while the intracratonic rift tectogenetic development is independent.

Crusts of Different Types Involved in the Orogenic Belts

The analysis of the actual structure of the orogenic (folded) belts shows that it is possible to distinguish different types of compressive structures, namely the nappes: cover nappes, obduction nappes, shearing continental basement nappes. Excepting the first type (built only of sedimentary formations), the last two allow to establish the constitution of the crust and consequently of the lithosphere, before the compressive (tectogenetic) processes which generated these structures.

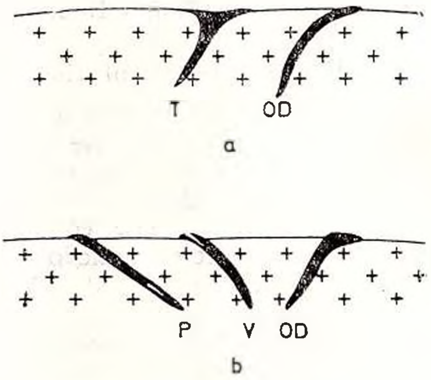

Fig. 5 — Mutual relations between consuming (subduction) paleoplanes: a, parallel; b, parallel and convergent.
T, Transylvanidian suture; V, Vardar suture; P, Pindus suture; OD, Outer Dacidian suture.

The obduction nappes may be divided into typical — which proceed from oceanic areas — and paratypical — which proceed from rift zones with thinned and/or oceanic type crust. Examples of the typical obduction nappes are known in the major Tethyan suture Zone of the Alps, the Carpathians, the Vardar Zone, etc. Paratypical obduction nappes occur in the Valais Zone (Alps), the Outer Dacidians (Carpathians) or the Pindus Zone (Dinarides).

Both types of obduction nappes are connected with important crustal shortenings along corresponding consuming paleoplanes (subduction-like processes). The fact that the most important crustal shortenings and the corresponding consumings took place in the former oceanic areas and rift-valley zones is a natural geodynamic phenomenon.

Taking into account the dipping of the consuming paleoplanes it is possible to distinguish (Fig. 5);

— parallel paleoplanes (Vardar and Pindus, Transylvanidian and Outer Dacidian) or

— convergent paleoplanes (Vardar versus Outer Dacidian).

The basement shearing nappes proceed from continental crust areas. They are well expressed in the Alpine folded chains as for instance the Austroalpine Nappes, the Median Dacidian Nappes (Central East Carpathians, Getic, Supragetic, etc.), the Pelagonian Massif, etc. This type of nappe is built up of pre-Alpine metamorphics and sedimentary, mainly Mesozoic, rocks.

The short comments above allow to stress out that three types of crust and consequently lithosphere may coexist in the mobile (geosynclinal) areas from which proceeded the folded belts: oceanic, thinned (rift-valley) and continental; the last two are developed on the continental margins of the former. This is the most complex situation. There are also cases when only oceanic and continental crusts or only thinned and continental crusts are involved in the folded belts (e.g. Northern Andes, North Dobrogea respectively).

Heterochronism of the Main Geotectonic Events

There are two major periods which may be recognized during the evolution of a mobile zone into the corresponding orogenic belt: the distension period and the compression one. The distension period is well expressed in the ocean opening and spreading as well as in the rift-valley genesis. The compression one is materialized in the tectogenetic events (tectogenetic phases, tectogenetic moments, etc.).

Comparing distant folded belts the heterochronism of these two periods is frequently evident. More interesting seems to be such an analysis along the same orogenic chain.

The heterochronism of the opening of the oceanic Tethys is enough documented along the Alpine mobile Zone of Europe (the future Alpine Chains of Europe). While the oldest Transylvanidian ophiolites are of Middle Triassic age, the Piemontais ones are of Lower Jurassic age. A similar heterochronism may be established between the beginning of the rifting processes in the Pindus Zone and the Outer Dacidian one (Fig. 4). This means that the two continental margins of the oceanic Tethys have reacted differently to the distensional strains.

More evident is the heterochronism of the tectogenetic events. For instance, the closing of the oceanic Tethys took place in the Carpathians during the Mesocretaceous, Mediterranean and Laramian tectogeneses, while in the Vardar Zone it began at the Jurassic-Cretaceous boundary (Neokimmerian) and ended during the Eocene (without important Cretaceous tectogenesis and crustal shortenings). Similar considerations may be emphasized for the two continental margins of the oceanic Tethys. The European one was deformed during the Mesocretaceous, Laramian, Styrian (old and young) and Moldavian tectogeneses. In turn, the Apulian continental margin was deformed mainly during the Tertiary tectogenesis (Pyreneean, Savian, Moldavian).

The areal heterochronism of the tectogenetic events does not contradict their temporal synchronism. In fact a certain tectogenetic event occurs wherever it is active during the same time span. The synchronism and the heterochronism of the tectogenetic events are not contradictory because these two meanings must be reported to different frameworks — temporal and areal respectively.

Conclusions

The compared analysis of different orogenic chains, of the same age, pointed out that :

— the geodynamic evolution is similar — a distension period is followed by a compression one ;

— the morphology of the mobile zone, generated during the distension period, may be different mainly with respect to the presence or the absence of intracontinental margin rifts ;

— the history of the main geotectonic events may be different, but follows similar ways.

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INTERPRETATION OF THE THRUST FAULTS GENESIS FROM AN ALPINE CHAIN SEGMENT — EAST CARPATHIANS

BY

MIHAI ŞTEFĂNESCU[1]

Introduction

The East Carpathians structure is characterized by the presence of the rootless nappes with high overthrusting amplitudes and with a remarkable directional constancy (Pl.). This complicated structure is the present-day stage of a long and agitated evolution.

The reconstruction of the mechanism which has led to such a structure is extremely difficult. Besides the proper features and data provided by the studied region, the results of researches made on the active continental margins with an intensely developing tectonics have supplied very useful information for such an attempt.

The aim of this paper is to try a reconstruction of the genesis mechanism of reversed faults with different amplitudes, starting from a comparison of the East Carpathians structure with that of the western margin of the North America. Within this last region, a successive and a progressive sediment accretion takes place at the continental margin, due to the underthrusting of their oceanic substratum. For the Carpathians, the underthrusting was admitted as the active folding element from the very beginning of this century (Murgoci, 1905 ; Mrazec, Popescu-Voiteşti, 1914). As concerns the substratum of the deposits in the compared zones, we have to mention : this one has an oceanic origin for sediments of trenches or those accumulated at the active continental borders ; the flysch zone substratum from the East Carpathians is not known anywhere today ; it was supposed to have an oceanic origin (Rădulescu, Săndulescu, 1973) for the innermost parts of the flysch zone — the Black Schists Nappe and the Ceahlău Nappe — as a result of an assymmetrical spreading (Ştefănescu, Ştefănescu, 1981) ; for the other nappes in the flysch zone, namely Bobu, Teleajen, Macla, Audia, Tarcău and of the Marginal Folds, it has

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always been admitted an initial substratum of a continental origin. Among the latter ones there are as well the units which constitute the examples used in this paper, namely the Teleajen Nappe, the Tarcău Nappe and the Marginal Folds Nappe.

Thrust Faults Genesis

The attempt to reconstitute the mechanism of the reversed fault genesis is based on the study of the internal structure of three of the most important structural units from the flysch zone in the East Carpathians.

As concerns the subject of this paper, the Tarcău Nappe is the most expressive, whose formation mechanism was minutely analysed for the first time by Dumitrescu (1962). It shows, together with more or less symmetrical folds, numerous reversed faults, but with different vergences both landward and seaward. While examining Figure 1, one can notice that these ones have an irregular distribution, either at the same parallel, or along the unit. Thus, south of the 46°N parallel, in the frontal part of the unit, the faults have inward vergences (landward), while south of the 48°N parallel, the faults in the frontal part of the unit have outward vergences (seaward). At the level of the 45°30′N parallel approximately, the vergences change several times, the fault groups with the same vergences being separated by convergent (upward) or divergent (downward) axes. These axes do not show a large continuity as they generally interrupt into transverse faults and rarely lose into folds.

Such a structural image characterized by reversed faults with different vergences is rather similar to those presented by Silver (1972), Seely (1977) and Barnard (1978) for the continental slope from the north-eastern Pacific. Both westwards the Washington State and southwards the Aleuthians, the interpretation of geophysical data (Seely, 1977) has shown that on the continental slope, starting from its lowermost part, there are some overthrusting faults with different vergences, as a result of the action of some compression forces which appeared due to the subduction of the oceanic crust.

The paleotectonic situation of the External Flysch Zone from the East Carpathians which preceded the beginning of the action of compression forces is simplified in section from Figure 1.

The action of the compression forces in the Moldavian Trench had as consequences : the décollement of the Cretaceous-Lower Miocene deposits from their substratum ; the formation of folds and faults with different vergences from the Tarcău Nappe ; the formation of folds and faults from the Marginal Folds Unit ; the thrusting of the Tarcău Nappe over the marginal folds and of these ones over the internal flank of the foreland (the Pre-Carpathian Molasse Nappe).

In order to explain the formation of the landward vergence (landward in our case), Seely (1977) uses the presence of some low shear strength zones which could be caused by overpressured clays due to the rapid accumulation of their overlying deposits. In these conditions, the clay shear strength (due to geostatic pressure) can approach zero and their cohesion is broken, clays having the physical conditions of a low viscous fluid (Seely, 1977). The part played by fluids and mainly by their pressure in rocks at the genesis of overthrusting faults was theoretically and also practically demonstrated and applied by Hubert and Rubey even since 1959.

Some rock packets containing important amounts of clays are found at different stratigraphical levels within the lithological columns of the flysch nappes from the East Carpathians, preceding or succeeding some thick deposit piles in a flysch facies. Deposits of the Tarcău Nappe are in the same situation.

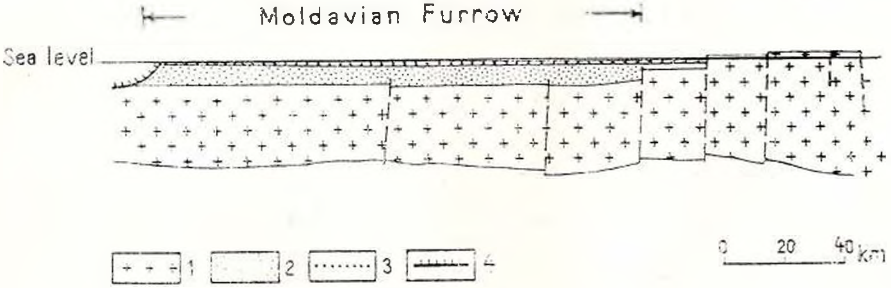

Fig. 1 — Paleotectonic situation at the Middle/Upper Miocene boundary. 1, continental type crust, including overlying sedimentary cover in non-flysch facies ; 2, mostly flysch deposits (Cretaceous-Lower Miocene) ; 3, molasse deposits (Lower-Middle Miocene) ; 4, older nappes.

As along the Tarcău Nappe there are two different structural aspects, we shall try to examine them in turn.

Between 45°10′N and 47°30′N the Tarcău Nappe has a structure with different vergence faults associated with folds. For this segment it may be admitted that at the same time with the beginning of the action of compression forces, the first to yield were the superpressured clay levels having the lowermost geometric position (Fig. 2 Aa). Thus, the whole sediment prism has taken off its substratum and subsequently, at the increase of the compression force intensity at the upper levels, it favoured the appearance of the basal planes of overthrust. At the same time with the horizontal shear phenomenon, the oblique shear planes of the future overthrusting faults started as well. The movement caused by the continuation of the compression (Fig. 2 Ab) has led on the one hand to the formation of overthrusting planes on initial shear planes, and on the other hand, to the appearance of folds in the outer part of the respective zone. The continuation of the movement has produced a backward curvature of the lower part of initial faults with a landward vergence, so that in the final stage they have the aspect of cylindrical faults.

North of the 47°30′N parallel, the Tarcău Nappe and its structural equivalent — the Skole Nappe — are characterized by an internal structure dominated by scales with overthrusting planes displaying only eastern vergences. In this case Hubert's (1951) and Seely's (1977) laboratory models show that such structures can appear when there is no horizontal shear. This is also the explanation adopted for the first step of evolution in this area. Subsequently there appear some shear

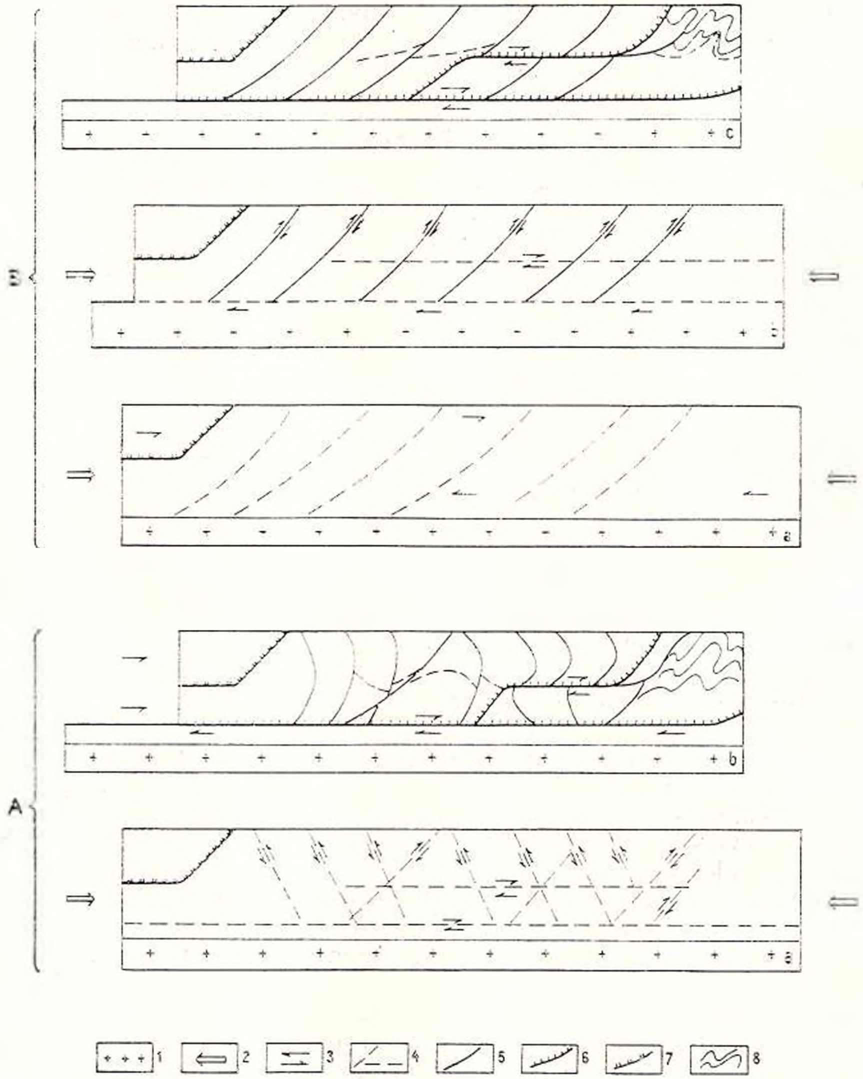

Fig. 2 — Geometrical simplified representation showing the possibilities of the thrust faults in the Moldavian Furrow.

A, case in which horizontal or quasihorizontal shear planes appear from the beginning; B, case in which horizontal or quasihorizontal shear planes subsequently appear; a, b and c represent evolution stages; Ab and Bc do not represent the final stages; 1, continental crust; 2, stress; 3, relative motion; 4, shearing planes; 5, thrusts; 6. overthrust; 7, older nappes; 8, folds.

planes parallel to stratification, due to overpressured clay levels, but probably having a lower water content because of the variation of their chemical content and thus with a pressure more similar to the geostatic one, which explains their shear delay. In both A and B cases, the subsequent evolution (2 Bc) leads to the detachment of the whole pile from its normal substratum, and therefore to the appearance of the overthrusting planes and to the formation of marginal folds. The formation of marginal folds can be explained by the existence of a morphological threshold at the limit between the depositional basin and the platform, a threshold which has played the part of a barrier which in the beginning was opposed to the seaward overthrusting tendency.

The appearance of the marginal folds even since the first moments of the structural evolution of the depositional basin (Moldavian Furrow) with morphological effects which created some subaerial erosion conditions for these deposits, explains the epiglyptical character of the Tarcău Nappe (Dumitrescu, 1962) as it overthrusts a domain which is already levelled by erosion.

The mechanism which was proposed for overthrusting genesis from the External Flysch Zone can be explained as well for the internal flysch. An example for the latter one is the Teleajen Nappe. It shows both overthrusting faults with contrary vergences (between 46°15′N and 46°35′N ; between 45°10′N and 45°20′N), and inclined folds, namely some structural aspects leading to the idea of obtaining overthrust from an overpressured clay level. The existence of some inclined folds in the frontal part of the Teleajen Nappe could be considered as a structural argument for the cordillera which is supposed (Ştefănescu, 1978) to exist outside its paleogeographical zone.

Besides the fact that they justify the overthrusting and the internal structure of nappes, the shears on lower levels (in pre-flysch facies) of overpressured clays explain as well the absence within the nappes of the complete normal substratum of flysch, including those from the neighbouring substratum, namely those of pre-flysch (except the Ceahlău Nappe where they are mainly marly). The appearance of the horizontal shear planes at higher stratigraphical levels explains the absence of pre-Senonian deposits at the frontal part of the Tarcău Nappe where these ones are left behind, at the "tail" of marginal folds. It is to be noticed that lability shown by deposits at about the same stratigraphical level with "diapir" contacts (Olteanu, in Mirăuţă, 1962), supports such an interpretation.

The existence of some low shear strength zones explains several structural aspects connected with flysch nappes, among which the existence of reversed faults with landward vergences. But in the East Carpathians there are as well some important retro-overthrusts (Săndulescu, 1964 ; Ştefănescu, 1976). They are located behind the active overthrust front, being more recent than the frontal overthrust planes of the affected units, but (probably) synchronous to some outside overthrusts which are more recent. Their genesis (besides other considerations — Ştefănescu, 1976) can be explained as well by using the existence of some landward vergence faults which are connected to more recent overthrust planes ; the older nappes overlying this one functioned as a single pile of deposits. It is very likeable that within this thickest and incompetent deposit pile which retro-overthrusted, the plane would correspond to an old fault with a landward vergence.

Instead of Conclusions

The existence of some overpressured clay levels, almost without resistence to shear efforts, can be used to explain both the genesis of the cover nappes (shearing and gravitational), and some of their internal structural features. But this explanation must be carefully used, taking into consideration some local peculiarities such as lithological contrasts, the sequence of the tangential effort stages and embrionary movements.

From the areal distribution of overthrusting faults from different units it is obvious that their occurrence is not uniform as in theoretical models which use uniform environments, but they mainly developed where at certain stratigraphical intervals there were some important changes (e.g. the Drajna Fault from the Tarcău Nappe, developed at the external limit of the Eocene sandy flysch facies).

While the basal plane of a unit is connected to a surface which is practically parallel to stratification, usually the frontal part obliquely crosses the stratigraphical succession of these deposits. It is obvious that the overthrust plane is oriented from the horizontal plane to a main reversed fault plane which in its turn originated in a strong lithological contrast. The passing of the movement from the horizontal plane to an oblique plane (of a reversed fault), is at the same time the genesis way for the digitations of the flysch nappes. These digitations show deposits which are different from their content point of view (Dumitrescu, Săndulescu, 1974).

Moreover, the lithological changings of pelitic levels can condition the water content variation and therefore the yielding speed or even the non-permission of yielding at tangential efforts, which implies the transmission of this movement on another plane (horizontal but located at another stratigraphical level, or oblique).

The embrionary movements originated within some depositional basins can generate large folds which allow water discharge from clay levels. Thus, they (totally or partly) transform a potential plane of a low shear strength in a plane insensible to similar efforts.

Polarity of tectonical events is well marked in the flysch zone of the East Carpathians. Moreover, it must be emphasized that the older internal planes were no more (generally) reworked during more recent movements which generated the external overthrusts. This situation can be explained as well by stabilization of equilibrium within clay levels through water loss during overthrusting, at least in the frontal part of the plane.

The genesis mechanism of overthrusting faults and of overthrusts shown in Figure 2 points out the possibility that from a certain vertical inward, some deposits of the initial stratigraphic column are absent towards the unit end. This possibility must be taken into account when one appreciates the possibilities that some deposits of economic interest can be found under the immediately upper overthrusting.

It is generally considered that cross faults are subsequent to the longitudinal ones, a situation which is perfectly suitable when they equally affect both overthrusted deposits and posttectonical cover ones. As already seen, there are cases when the axes separating longitudinal faults with different vergences stop in cross faults. Here, it is obvious that cross faults are syngenetical to longitudinal ones.

Although the two compared zones benefitted by different paleotectonic conditions, they show some similar structural aspects, due to overpressured clay levels which manifested themselves in the same way when a compression tangential force did appear.

If this comparison was useful to explain, at least partly, the genesis of reversed faults and of overthrust nappes, we think it can be used as well in the opposite direction, to explain seismic sections and theoretically to forecast the structural evolution tendency of continental margins.

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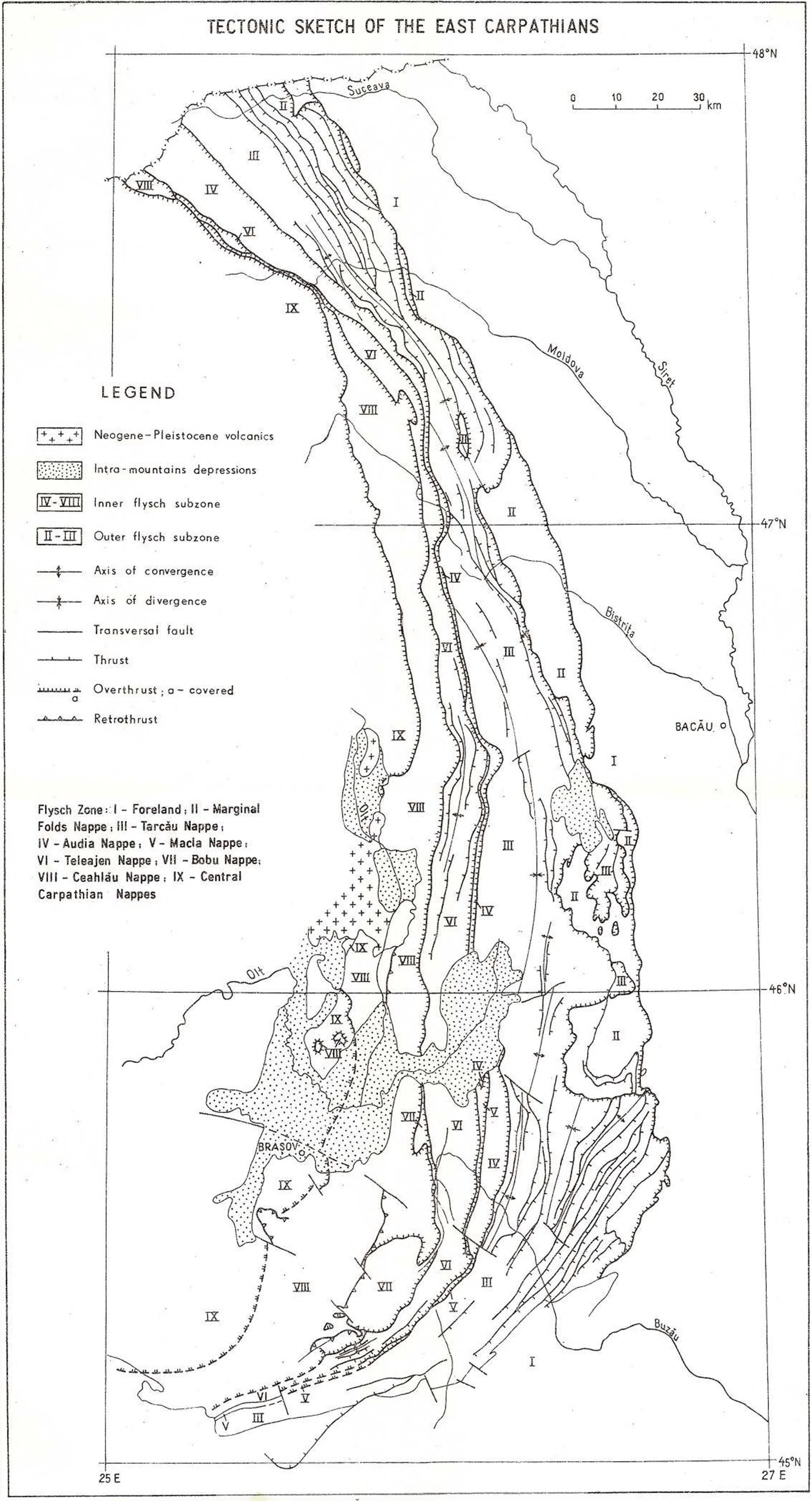

TECTONIC SKETCH OF THE EAST CARPATHIANS

THE REFLECTION OF THE BASALTIC LAYER IN THE REGIONAL ANOMALY OF THE GEOMAGNETIC FIELD ON THE ROMANIAN TERRITORY

BY

LUCIAN BEȘUȚIU [1]

Introduction

Emil Thellier would say about the geomagnetic field that "it reveals always too little even to its close friends" having "a strong reputation of being complicated and mysterious".

The image given by Romania's magnetic map reveals this very thing. Far from trying to decode the complete message offerred by this great amount of information, the present paper represents only an attempt of making clear some more general aspects which are to be found in this preliminary observation material.

Within the analysis carried out data provided by other geophysical methods (geothermic and seismic ones) have been used in order to give a more certain degree to the magnetic data interpretation.

Some Considerations on the Data Used

The Regional Anomaly of the Geomagnetic Field on the Romanian Territory

The regional vertical component magnetic contours have been drawn up by the author on the basis of the data provided by the vertical-component ground magnetic map of the Socialist Republic of Romania (Airinei et al., 1982).

The problem of separating the "regional" effects is still an open subject nowadays, first of all due to the relativeness of the "regional anomaly" concept. Its defining depends, first of all, on the interpreter of the data. The "personal factor" mentioned by Vajk is — in a modern

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language — the transfer function of the "filter" that the interpreter represents on the informational flow way. Willing or not he will define, by means of his own knowledge, the most probable source of producing the "regional" effects.

In our opinion, the regional anomalies separation is only a particular case of the general problem concerning the detachment of the "useful" signal from the "noise" accompanying it. Choosing one of the numerous filtering methods it becomes only a pure matter of option as long as we can master the meaning of the signal thus separated.

A handy economic method largely used was chosen — that of the "running average". The significance of the elements thus determined is quite simple, the mean of the field values in a certain domain, like the most probable value in that area, corresponding to the effect produced, as a rule, by large extended sources.

The sampling network spacing and the average device size have to be chosen according to the nature of the signal which must be separated. The wave length analysis of the broad scale magnetic anomalies on the Romanian territory has shown that at least a part of their sources are to be found at the depth of 13-15 km. In order to emphasize such effects, the variant of the regional anomaly got with the help of an average "window" having a 16 km side has been chosen for analysis. This does not mean that the signal thus "separated" does not also contain within it information from the sources situated nearer the surface. The simple average of the primary map cannot eliminate these effects (for this a similar procedure to the gravity "uncovering" would be necessary, which is almost impossible to achieve) but it assures a considerable improving of the signal/noise ratio.

It is worth mentioning that the arbitrary average of the signal produced by several small sources, not too deep, but with a wide spreading in the surface, can lead to the extended anomalies, similar to those produced by deep sources (as in case of the Neogene eruptive zone in the East Carpathian Chain). The confusion can be avoided by confronting the processed images with primary aspect of the magnetic map.

Geothermic Data

The main material having this nature used when interpreting the geomagnetic data is represented by the map of the regional distribution of the geothermal gradient on the Romanian territory (Paraschiv, Cristian, 1976). The results of the heat-flow measurements along the international seismic profile XI (Veliciu et al., 1977) have also been taken into account.

The geothermal gradient map has been drawn upon the basis of more than 3 000 measurements in wells from the main oil-bearing structures. The gradient values have been determined in stabilized conditions at the depth of 1 000 m under the sea level, that gives them a special accuracy.

The heat-flow determinations are more sporadic being affected by a raised noise level too, that gives them only an informative character.

Great difficulties have been recorded in establishing the temperature distribution at depth. The formulae used by different authors are approximative and tributary to a large number of parameters insufficiently determined for the Romanian territory.

Seismometric Researches

A series of seismometric researches at a great depth have been carried out on the Romanian territory in the last decade (Constantinescu et al., 1972 ; Rădulescu et al., 1976 ; Rădulescu et al., 1979 ; Rădulescu, 1981, etc.). Their results have been materialized in working out some sections through the Earth's crust along the international profiles (II, XI, XI_1, XII).

According to Rădulescu (1981) the quality of the information provided by this method is not always the same. Generally speaking, the Conrad discontinuity position has been well stated out (excepting perhaps profile XII), unlike that of the Mohorovičic surface from which the information has been sporadic.

Data Correlation and Their Interpretation

As already stated above the wave length analysis of the broad scale magnetic anomalies, has led to the conclusion that their main source is situated at the depth of 13-15 km. However, the seismometry shows that this is the main domain where the Conrad discontinuity is to be found, that is the basaltic layer top front credited with magnetic susceptibility included between 1 500-4 000 CGSu (Kurtihovskaya, Paschevici, 1976).

Admitting the basaltic layer as one of the sources of the drawn effects in the processed magnetic map and taking into account its continuity, an attempt was made to interpret this material starting from the regional magnetic minimums rendered evident. On this occasion two interesting things have been observed : correlation of geomagnetic minimums with maximum anomalies of the geothermal gradient, and superposition of some large geomagnetic heights over minimums of the geothermic data. The observation is in perfect agreement with the proposed model if it takes into consideration that the basaltic layer magnetic properties are done by the content in ferromagnetic minerals (mainly magnetite). Nevertheless the existence of the geothermic anomalies can "raise" or "lower" the 500°C isotherm position which represents the Curie point of the magnetite.

The resulting conclusion is that the broad scale magnetic anomalies rendered evident in the filtered map are conditioned by the ratio between the basaltic layer top front position and the depth of the magnetite Curie isotherm.

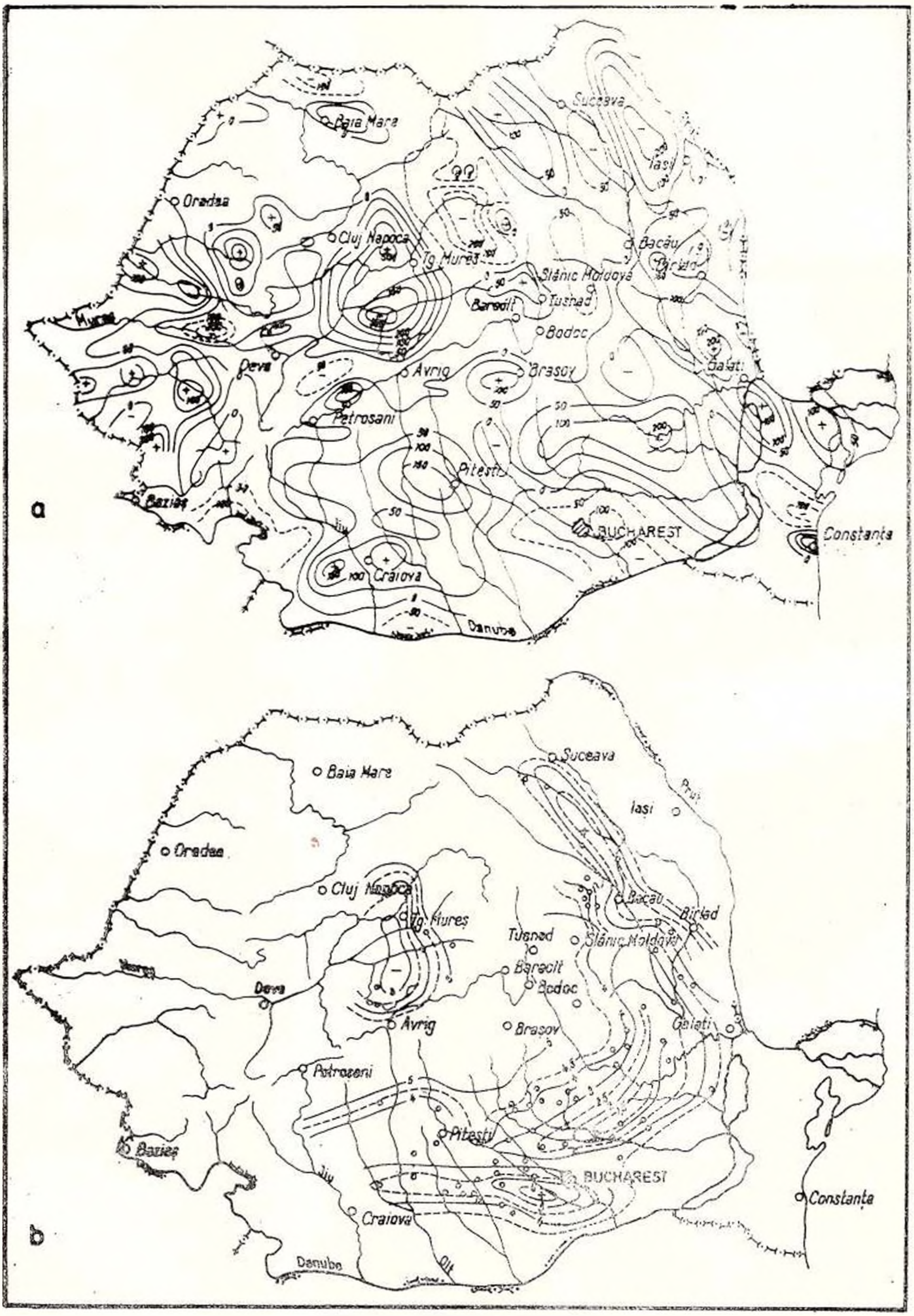

Fig. 1 — Field data used.
a, regional vertical component ground magnetic anomaly ; b, regional geothermal gradient data (after Paraschiv and Cristian, 1976)

Here are some practical examples of this interpretative criterion. The great magnetic anomaly in the centre of the Transylvanian Basin is undoubtedly the result of superposing the effects produced by multiple causes. A detailed analysis (Botezatu et al., 1976) has shown that the main sources producing it are situated at three levels : a. in the sedimentary cover ; b. at the level of the crystalline basement ; c. in a more profound area corresponding to a strong raising of the basaltic layer (partially confirmed by seismometry). This model cannot explain the more restricted area of the magnetic anomaly for a larger extention of the basaltic layer raising.

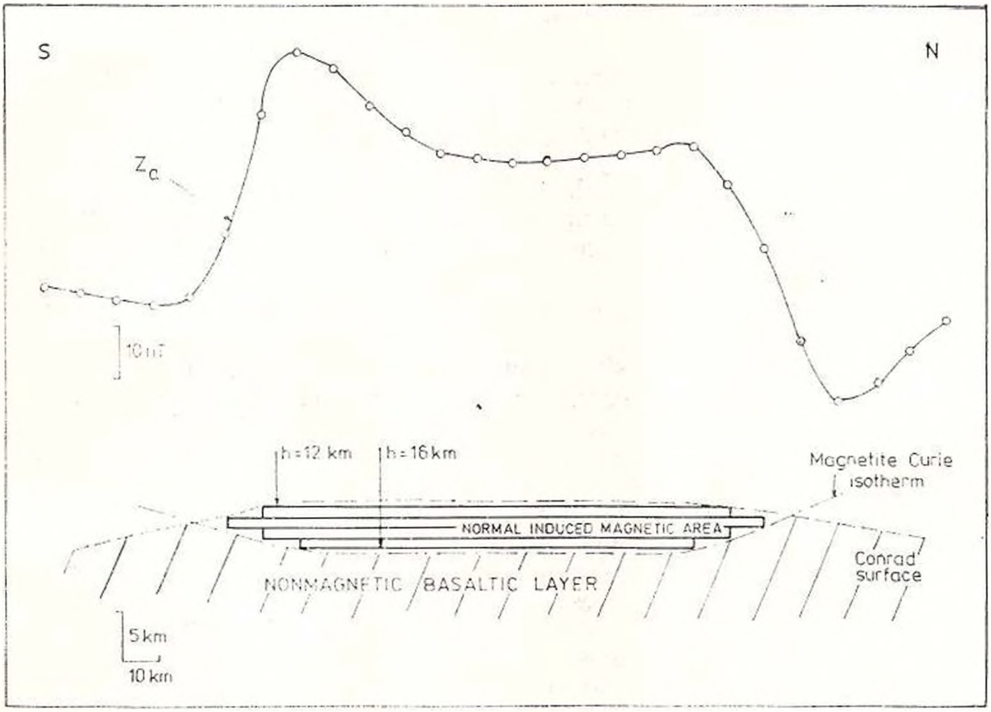

Fig. 2 — An example of the basaltic layer magnetic effect.

The interpretation can be improved if one takes into account the obvious correlation between the geothermic and magnetic data. Thus, the minimum of the geothermal gradient "pushes" the magnetite Curie isotherm, in the middle of the depression, at the depth of more than 17 km in an area where the seismometry has indicated about 12 km to the basaltic layer top front. As for the flanks of the basin, where the Conrad discontinuity lowering is not so important but a strong increasing of the temperature at the depth is pointed out by the geothermic data, the magnetite Curie point will be reached much nearer the surface. In this way a magnetic active area of the basaltic layer is detached between the basaltic layer top front and the magnetite Curie isotherm surface. Its lateral limitation, due to the faster temperature increasing on the flanks of the depression, explains the restricted area of the magnetic anomaly.

A similar case is present in the Moesian Platform (west of Piteşti), where the basaltic layer top front is situated at the depth of 14-15 km, above the magnetite Curie isotherm, a fact that allows the magnetic expression on the basaltic layer.

As for the correlation of the magnetic regional minimums of the same order geothermal gradient, the situation becomes interesting by their superposition on some known deep-crustal faults.

For example, the eastern flank of the Transylvanian Depression, an area in which the existence of a major tectonic accident is admitted, is characterized by a minimum geomagnetic correlated with an evident growing tendency of the geothermal gradient values ; Tîrnava Mică fault is recognized both in the relative geomagnetic minimum chute, which separates the main terms of the Transylvania anomaly, and in the tendency of maximum geothermic data, which tends to break in two the wide minimum in the centre of the depression.

Thus, an original criterion of detecting the deep crustal faults appears : the correlation of the regional geomagnetic minimum chutes with the anomalies of maximum geothermic of the same order.

Coming back to the geothermal gradient map, it should be mentioned that its authors have shown the existence of an interesting maximum anomaly in the Moesian Platform, much extended along the Turnu Severin-S Filiaşi-S Bucharest line with an orientation which could not be correlated with any of the structures of the sedimentary cover or the platform basement. The regional magnetic map indicates, in this region, a minimum magnetic chute with the same trending and whose intensity, in absolute value, increases in the area of Bucharest, where the geothermal gradient values raise too. According to the criterion proposed to be accredited there results that the source of these conjugated effects is a very old crustal shifting which does not appear in the superior structural level of the platform, but which hinders (due to the Curie point surpassing) the magnetic expression of the basaltic layer in this area.

The presence of another chute of minimum geomagnetic, this time trending NNW-SSE, on the Cîmpulung Muscel-W Bucharest line, superposed to an axis of maximum of the geothermal gradient shows that right near Bucharest there is a cross-section of crustal creases which would explain the "hot zone" with a maximum in the south-west of our capital. Besides, we incline to interpret the whole zone of a minimum geomagnetic surrounding Bucharest not through a petrographic differentiation or an inverted magnetization of the platform basement, as considered by other authors, but through the loss of the basaltic layer magnetic properties as a result of a considerable increase in the temperature at its level due to the existence of a crustal fault system which intersects this area. Unfortunately, the lack of the geothermal gradient data in the eastern part of the Moesian Platform does not allow their correlation with the geomagnetic ones.

Another interesting magnetic minimum chute starts from the Central and South Dobrogea (leaving the Palazu Mare anomalous zone aside) making for NNW along the western bank of the Siret. It is worth mentioning the fact that in its southern part this leading line seems to result from cumulating three minimum terms : two of them superposed to the known Peceneaga-Camena and Capidava-Ovidiu faults, with a NW-SE trending, and the third one, which seems to be the best developed one, placed between the above-mentioned ones, with a NNW-SSE trending like the Siret Fault. It seems difficult to maintain that there is a deep dislocation which extends this major tectonic accident, beyond the Danube up to Mangalia. However, it is sure that right along its axis appear the thermal springs from Hîrşova and Mangalia whose origin is not well stated so far ; recent geothermal researches (Polonic et al., 1981) have pointed out an important tendency of a maximum gradient west of the Danube, which superposes perfectly to the geomagnetic minimum.

Conclusions

The basaltic layer has been stated out like one of the main sources of the broad scale magnetic anomalies rendered evident on the regional magnetic map of the Romanian territory ; it is considered that its structure determines some characteristic features of the geomagnetic field.

This thing does not mean that it is the source of all the magnetic anomalies pointed out in the filtered map. There is a great variety of such causes and, moreover, not all the "regional" effects rendered evident have a deep origin. Their interpretation must be done carefully considering the limits of the filtering procedure used.

However, the correlation between the magnetic and geothermic data, which allows to establish an original criterion for deep crustal faults detection, is an important element.

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Géophysique

APPLICATION DE LA MÉTHODE DU PROFIL SLALOM DANS UNE ZONE COMPRISE ENTRE LA VALLÉE DE LA DÎMBOVIȚA ET LA VALLÉE DU BUZĂU (ROUMANIE)

BY

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La recherche sismique de la zone des plis diapirs et de la zone du flysch entre la vallée de la Dîmbovița et la vallée du Buzău a évolué allant du profilage simple (fig. 1 a) au recouvrement multiple de 2 400% (fig. 1 b) à groupements de géophones et sources de point de tir différentes (1, 3).

Dans la partie nord à relief accidenté on a appliqué la méthode d'observation discrète S.S.M. et S.S.M.M.

Tous ces travaux effectués jusqu'en 1980 n'ont pas réussi à résoudre intégralement les objectifs géologiques de cette zone constituée de dépôts miocènes inférieurs et paléogènes, du fait des insuffisances des méthodes appliquées, de la tectonique et de la morphologie compliquée (collines, forêts, localités).

Puisqu'on n'a pas pu exécuter des trajets droits dans cette zone de grand intérêt il a fallu utiliser le système d'observation type profil slalom appliqué pou la première fois en 1981 dans une zone au nord de Moreni sur la vallée du Cricovul Dulce (fig. 2, 3).

L'application de la méthode du profil slalom (5) a impliqué la réalisation d'un schéma de terrain homogène pour obtenir un diagramme de dispersion uniforme des points médians source-récepteur (scatérogramme) en bande de 500 à 600 m (fig. 4).

L'étude des travaux sismiques par la méthode du profil slalom a suivi les principes de la prospection sismique de réflexion, en impliquant l'utilisation des groupements de géophones et des puits pour atténuer le bruit organisé et aléatoire, l'augmentation de l'énergie au moment de l'explosion et finalement le calcul des paramètres du dispositif de réception (Δx, offset etc.) vu les conditions sismogéologiques de profondeur (fig. 3).

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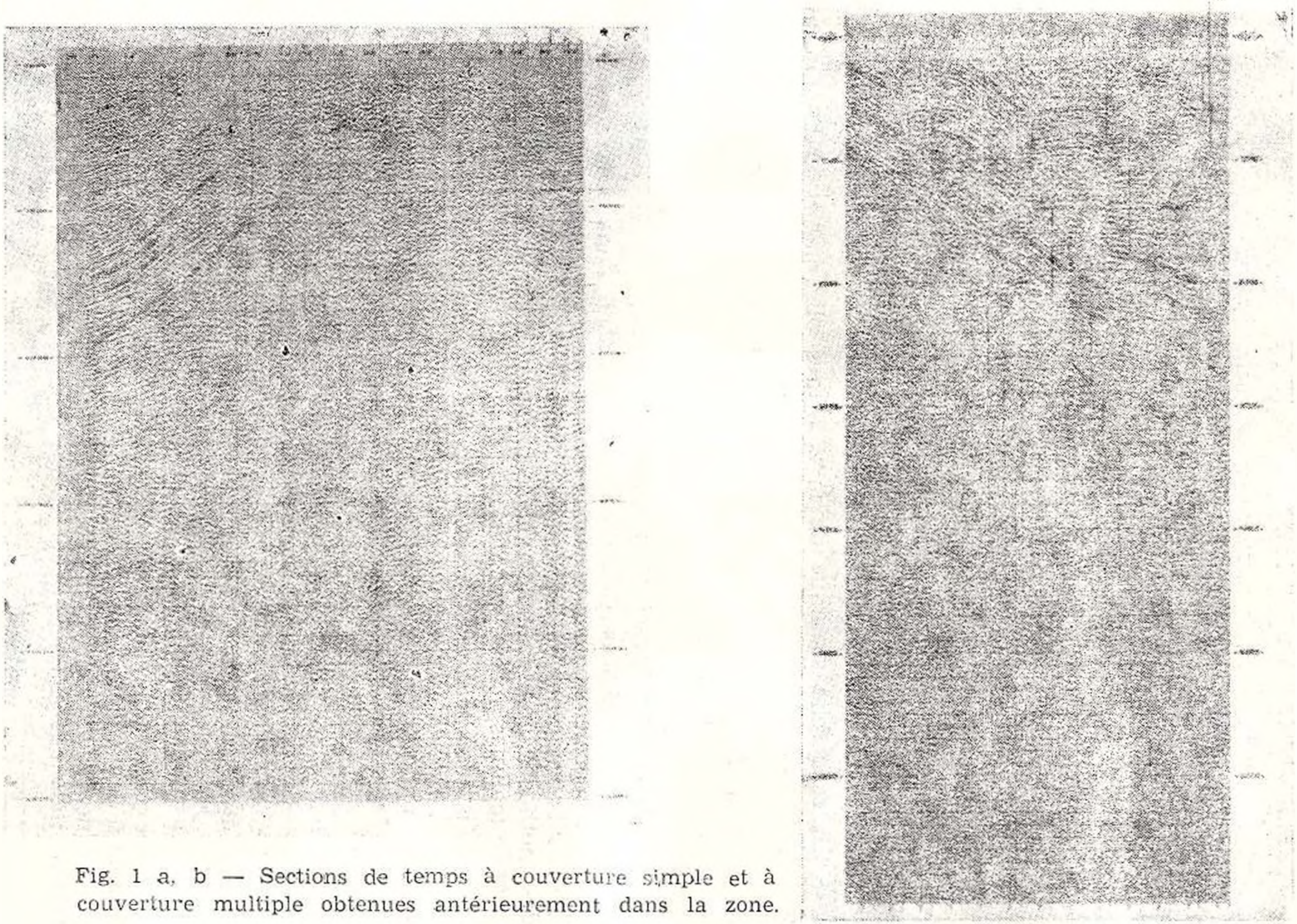

Fig. 1 a, b — Sections de temps à couverture simple et à couverture multiple obtenues antérieurement dans la zone.

Le programme expérimental, exécuté en zones où l'objectif géologique est situé à des différentes profondeurs, a eu comme but la connaissance détaillée des conditions sismogéologiques de la zone (analyse du bruit, études WZ, tests groupements optimums, filtres optimums etc.).

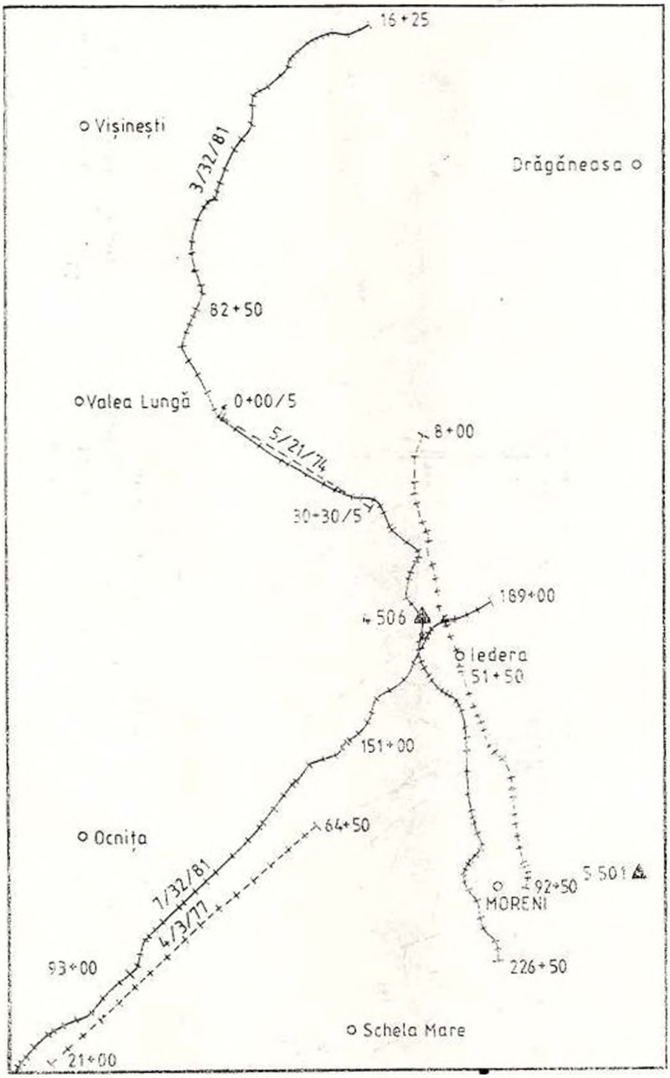

Fig. 2 — Plan de situation avec la positon des profils sismiques.

L'interprétation du grand volume de données expérimentales a été faite sur base des analyses du bruit aléatoire et des signes en amplitudes réelles, des spectres d'amplitude, des simulations de groupements, de la forme d'onde etc.

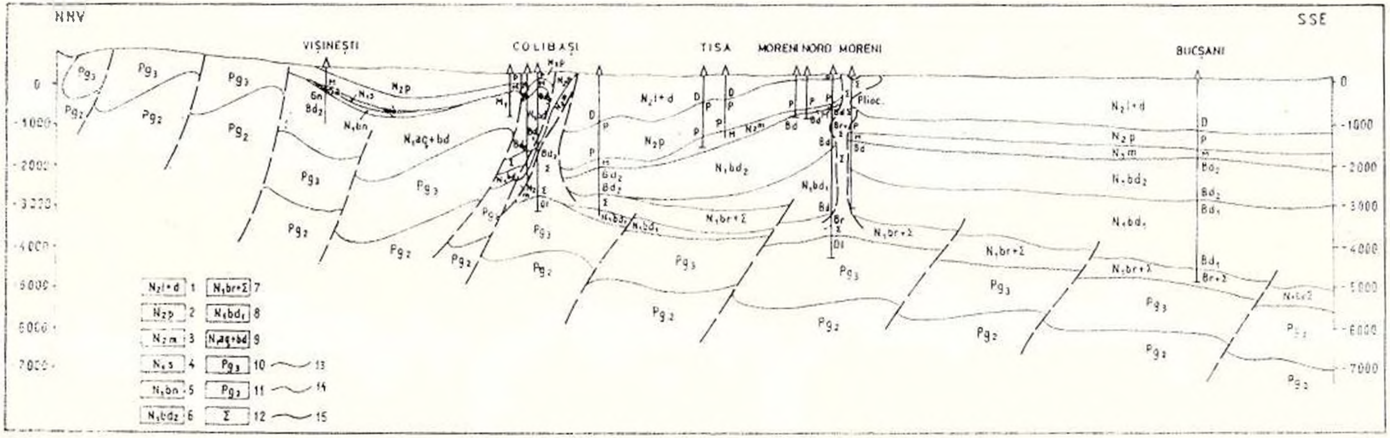

Fig. 3 — Coupe géologique sur Valea Cricovului Dulce.

1, Lévantin + Dacien ; 2, Pontien ; 3, Méotien ; 4, Sarmatien ; 5, Badénien ; 6, Burdigalien supérieur ; 7, brèche à sel ; 8, Burdigalien inférieur ; 9, Burdigalien + Aquitanien ; 10, Oligocène ; 11, Eocène ; 12, sel ; 13, limite normale ; 14, limite de transgression ; 15, faille.

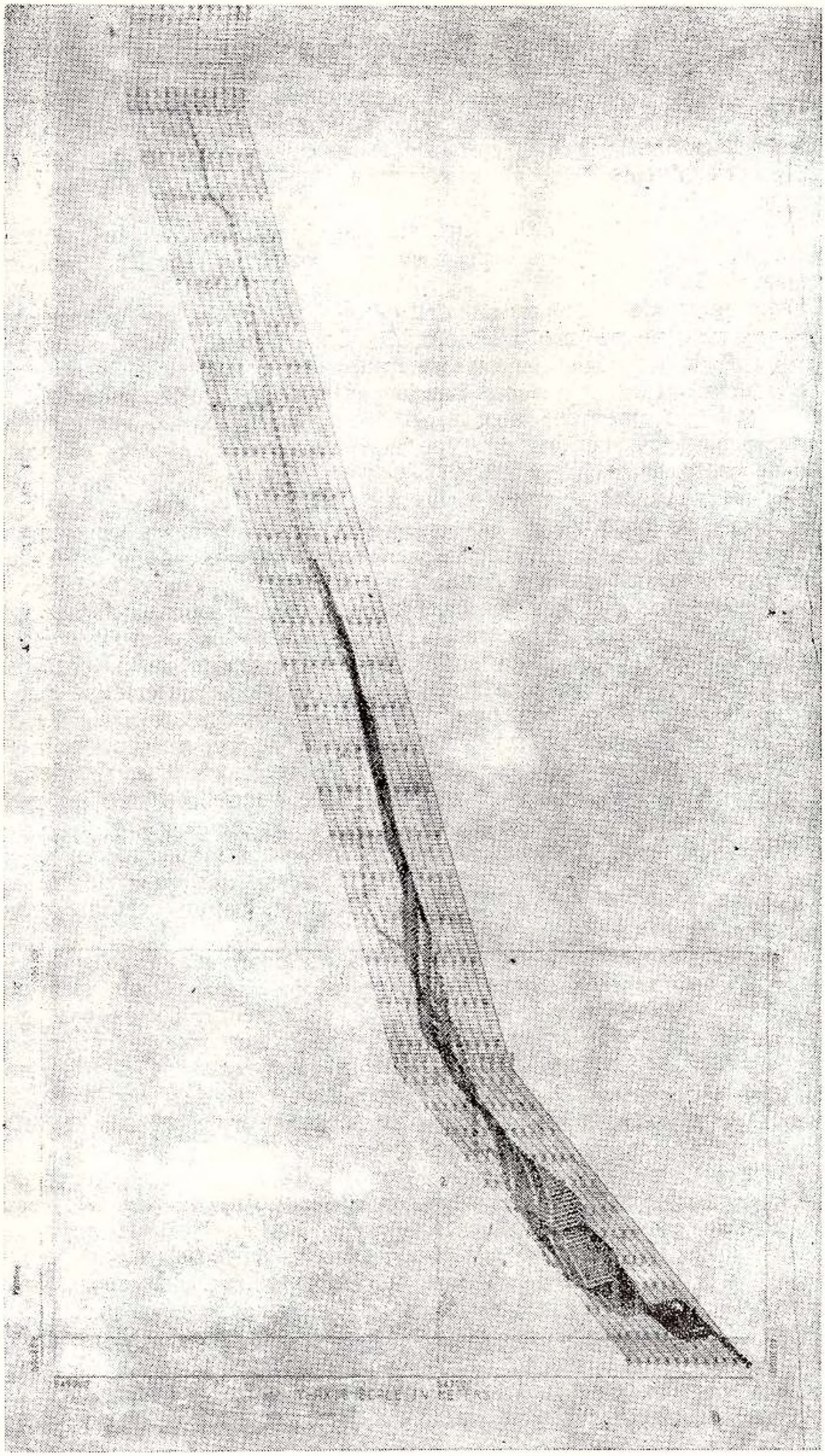

Fig. 4 — Diagramme de dispersion des points source-récepteur et des sous-bandes assimilées en tant que profils individuels.

Cette interprétation a mené aux suivantes conclusions méthodologiques : l'atténuation du bruit aléatoire est satisfaisante, dans les conditions de l'utilisation, à la réception, des filtres coupe-bas plus hauts que 12 Hz/24 db.

Afin d'atténuer les ondes superficielles identifiées sur les tableaux d'ondes (300 à 800 m/s) on a proposé un groupement de 22 géophones à distance de 50 m.

Cette méthode a été aussi vérifiée sur le terrain par l'enregistrement, dans la zone du profil slalom, d'un profil longitudinal superposé sur un profil antérieur de mauvaise qualité (fig. 5 a, b).

L'interprétation de la sous-bande centrale du profil slalom (fig. 4) a été faite conformément aux profils longitudinaux, se caractérisant finalement par des sections de type non migrées et migrées et par la section de profondeur (fig. 6 a, b, 7).

Les informations sismiques obtenues relèvent en grande partie les complications tectoniques de la zone déterminées par les phénomènes de diapirisme de Colibaşi et de Moreni, ainsi que par leur effet dans la structure tectonique des formations adjacentes, concrétisé par de nombreuses fractures et des pendages accentués des couches (fig. 6 b, 7).

Ces complications ont constituées des obstacles pour l'investigation sismique de la zone, en solicitant au maximum le potentiel de la technologie de terrain ainsi que l'interprétation, matérialisé par la grandeur de l'ordre de recouvrement (24), la diminuation de la distance entre les canaux (25 m), l'augmentation de l'énergie au point de tir, l'interprétation de la forme d'onde, les corrections d'amplitude etc.

La comparaison de la section de la sous-bande centrale du profil slalom avec les sections de temps des coupes antérieures (fig. 2) situées dans le proche voisinage dénote la qualité plus élevée prouvée par une meilleure distinction des horizons sismiques au niveau des formations sarmato-pliocènes et par la mise en évidence dans certaines zones des repères du Miocène inférieur et de l'Oligocène.

C'est ainsi que les sections de temps du profil slalom (fig. 6 a, b) relèvent d'une manière suggestive l'évolution des horizons sismiques du Nord au Sud, désignant le passage de la zone du flysch paléogène à la zone mio-pliocène des plis diapirs, là où apparaît le contour du diapir salifère. Sur le flanc sud du diapir apparaissent des horizons corrélables à des intervalles de 4 secondes et plus bas (fig. 6 b). Ces horizons manquent sur la coupe sismique du profil (fig. 1 b) situé parallèlement et très proche du profil slalom.

L'effort technologique dans les travaux de terrain est illustré également par la qualité de la coupe du profil longitudinal (fig. 5 b) enregistré au moyen de la méthode du profil slalom qui est nettement supérieure à celle du profil antérieur exécuté au voisinage (fig. 5 a).

En abordant le problème dans le contexte de la sismique stratigraphique, la coupe sismique du profil slalom comprend en grand deux séquences : une séquence supérieure des dépôts sarmato-pliocènes, caractérisée par un faciès sismique à réflexions parallèles, ondulations à fréquences moyennes et hautes et une séquence inférieure (Miocène et Paléogène) à réflexions généralement faibles, discontinues de fréquences moyennes et basses.

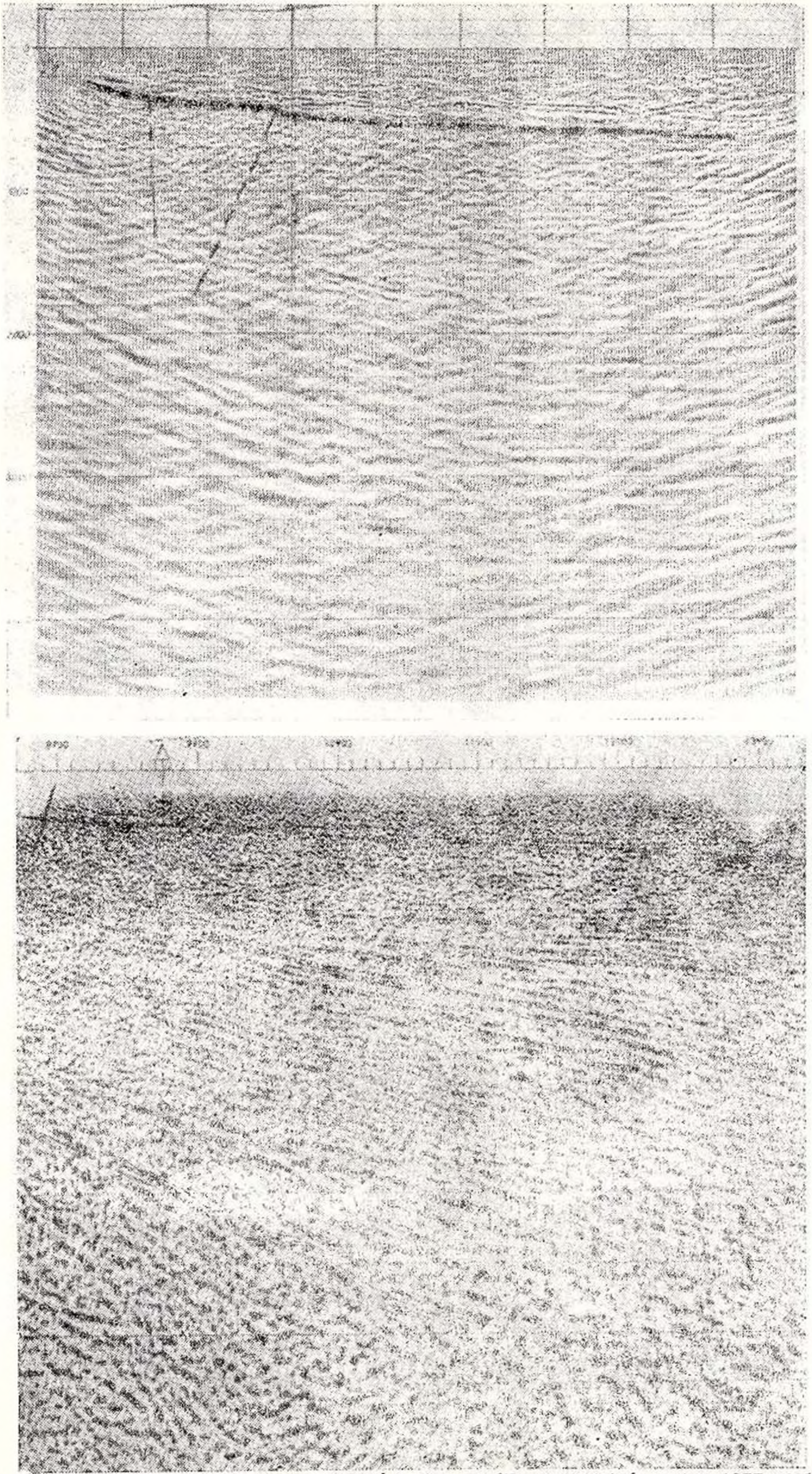

Fig. 5 a, b — Sections de temps migrées des profils enregistrés entre 1977 et 1981.

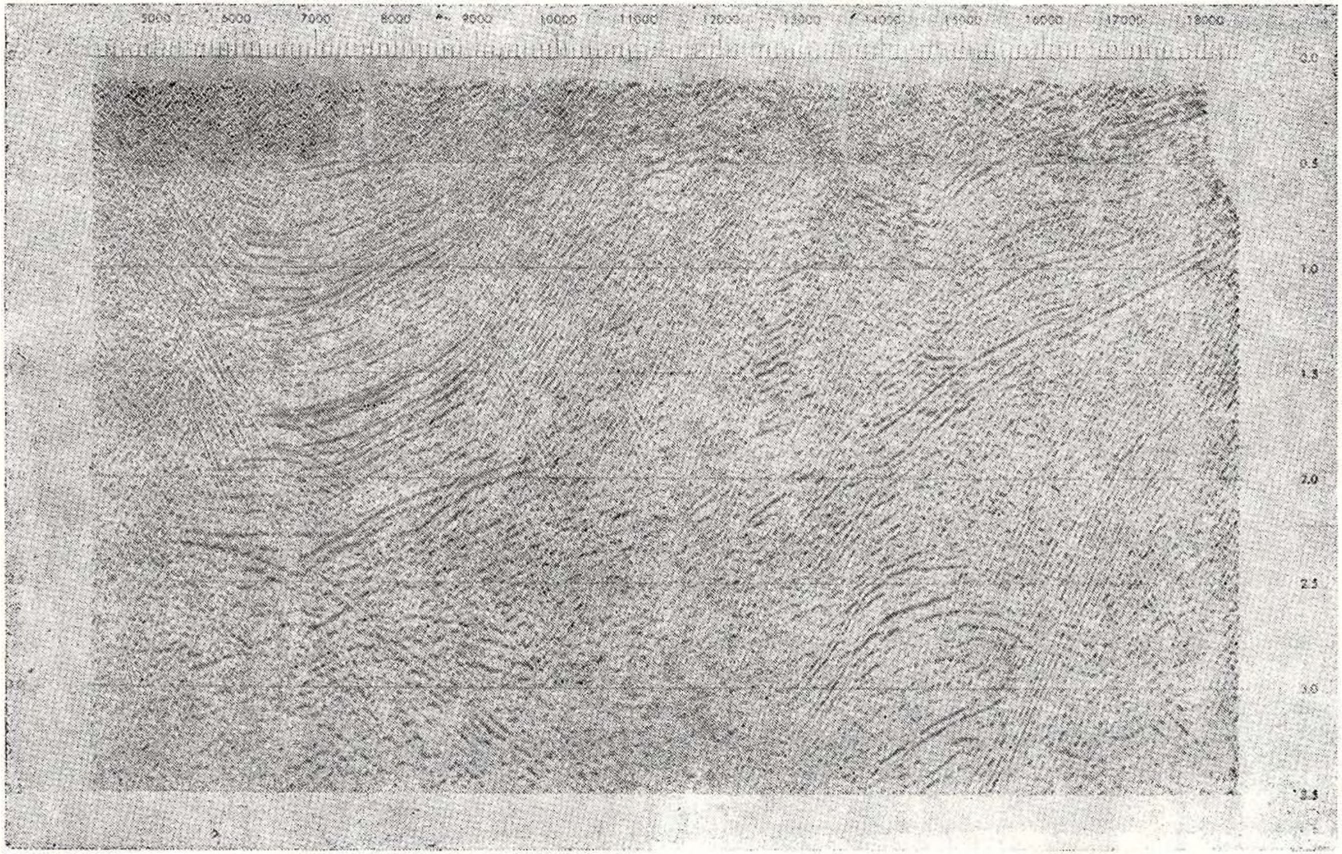

Fig. 6 a, b — Sections de temps non migrée et migrée du profil slalom.

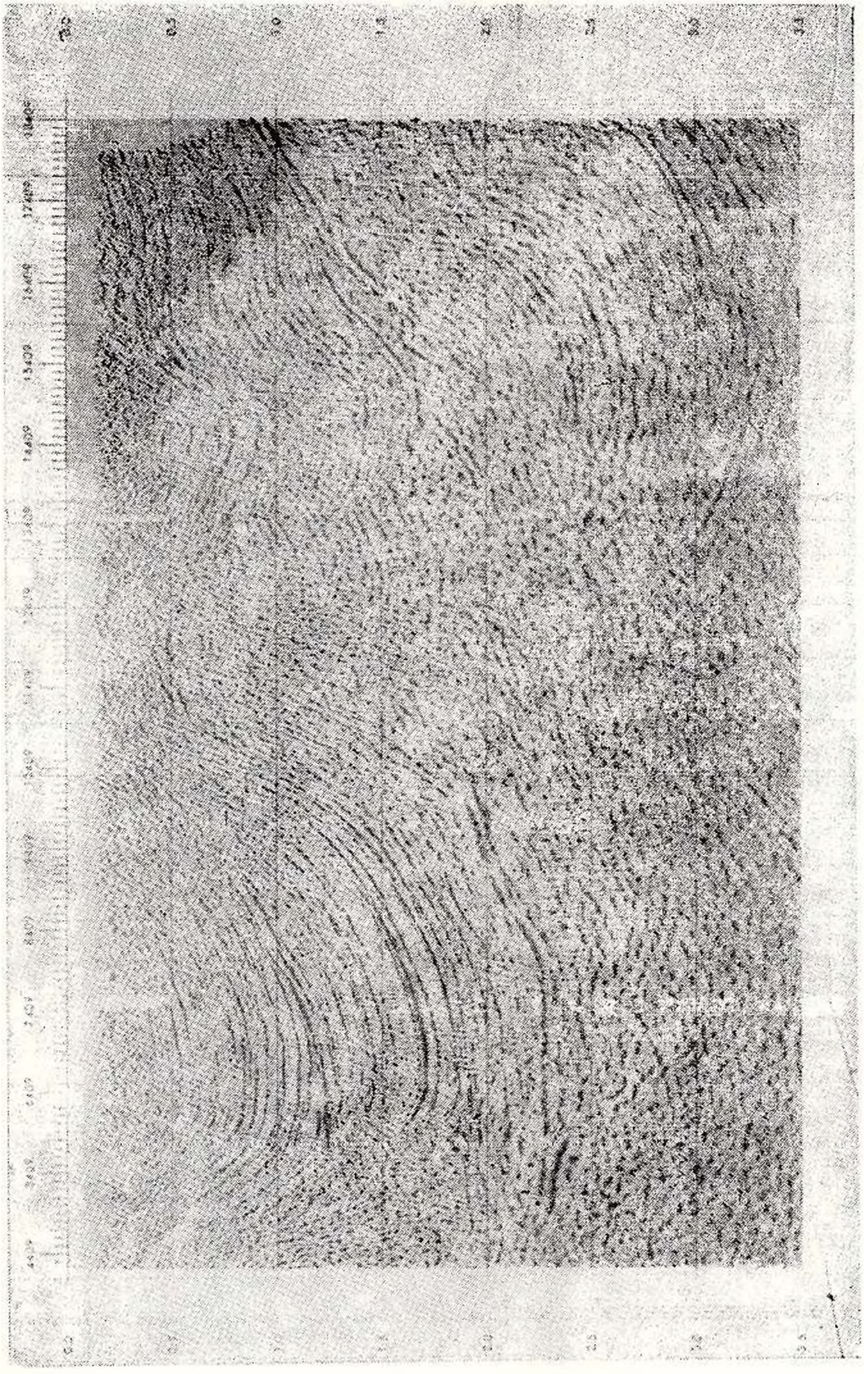

Fig. 6 b

Fig. 7 — Section de profondeur du profil slalom.

L'évolution du faciès sismique dans le cadre des séquences marque le passage du faciès structuré au-delà du diapir à un faciès de type chaotique à fragments de réflexions à inclinaisons ou absence de réflexions dans le proche voisinage ou bien à l'intérieur du corp diapir.

Une délimitation plus évidente des séquences sismiques est observable sur la coupe sismique du profil longitudinal (fig. 5 b) où apparaissent des éléments discordants (troncatures) à la limite supérieure de la séquence inférieure.

En guise de conclusion on peut apprécier que les résultats obtenus par l'expérimentation réalisée dans la zone de la vallée du Cricovul Dulce ont une double importance :

— au point de vue géologique, par les nouvelles informations obtenues, en des zones à structure tectonique très compliquée, comme la zone du flysch et celle des plis diapirs ;

— au point de vue technologique, par l'élaboration d'un nouveau système d'observation (profil type slalom), à grand ordre de recouvrement (24), pouvant donner des informations supplimentaires plus certaines dans les conditions d'une tectonique compliquée, méthode qui a ensuite (1982, 1983) ouvert le chemin vers la recherche des zones très compliquées de la zone du flysch (situé à l'Est de la vallée du Teleajen ou en Moldavie).

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GEOTHERMAL RESOURCES OF ROMANIA

BY

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Introduction

Romania's geographical location, overlapping the Carpathian folded chain, accounts for diverse and numerous occurrences of geothermal waters. Related also to a volcanic activity which ceased since the Lower Quaternary, the whole inner part of the East Carpathians is prospectively valuable for the existence of the "hot dry rocks".

Progress has been made during the past few years towards a better understanding from a geological and geophysical point of view, enough to improve significantly the basis for a reasonable assessment of the distribution, magnitude and recoverability of various categories of geothermal resources. Up to now only hydrothermal reservoirs have beend tapped in Romania for domestic heating and use in agriculture.

Geothermal resources exploration has been conducted along the same lines as those used for oil industry. Research methods were however modified by specific techniques which characterize the particular aspects of geothermal fields. Therefore, studies were carried out to obtain both the areal distribution and intensity of thermal anomalies through various techniques, i.e. regional heat flow density survey, geothermal prospecting by means of shallow boreholes, collection and interpretation of the temperature data from oil industry boreholes (Veliciu, Opran, 1983).

Complex geological, geothermal and hydrogeological investigations (Ghenea et al., 1980) indicate that the most favourable areas with respect to the geothermal resources are the Western Plain on the eastern limit of the Pannonian Basin and the Moesian Platform (hydrothermal convective systems) and the Neogene-Quaternary volcanic arc of the East Carpathians (mainly conductive dominated systems).

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As for the present economic use of geothermal waters, the wells drilled on the eastern limit of the Pannonian Basin exhibit proper temperatures and outflows. However, some problems have appeared during the exploitation due to the high degree of mineralization, composition and/or pollution.

Geological Setting, Regional Heat Flow and Geothermal Areas

From a geological point of view, the Romanian territory is dominated by the Carpathian folded chain (belonging to the Alpine orogenic belt) and its Foreland.

The Carpathians (East Carpathians, South Carpathians together with Apuseni Mts) have been divided into several major structural units (Fig. 1) which generally group together nappes of similar type and of synchronous age of tectogenesis (Dumitrescu, Săndulescu, 1968). From the last point of view the Carpathians show two main periods of compressions : Cretaceous and Miocene. The Cretaceous Carpathians group the Dacides, the Transylvanides and the Danubian. The Miocene Carpathians (Moldavides) cover the outer zones of the chain. A special belt showing both Cretaceous and Miocene deformations are the Pienides. Posttectonic covers of different ages are known mostly in the inner Carpathians. Two big Neogene molasse depressions (Transylvanian and Pannonian) and several smaller ones cover parts of the deformed inner Carpathians. A Neosarmato-Pliocene molasse foredeep borders the folded chain outwards.

Pre-Alpine crystalline formations crop out inside the Dacidian areas. Here they are generally covered by carbonate or detrital deposits.

The Transylvanides are the main suture zone of the Carpathians containing units with ophiolitic complexes and sedimentary formations. Outer Dacides are the second suture, showing flysch deposits and ophiolitic rocks. Large development of flysch is known inside the Moldavides. The Pienides consist of carbonate and/or flysch formations. Molasse or flysch deposits are mostly developed in the posttectonic covers.

Except the ophiolitic assemblage, the Alpine magmatic activity shows three igneous periods : an ensialic dominantly alkaline moment (Lower and partly Middle Jurassic) known in the Central East Carpathians units and in the South Carpathians (eastern part) and two subduction calc-alkaline moments, the first in the Upper Cretaceous and Paleocene time, the second during the Neogene time.

The Carpathian Foreland groups the platform areas of different ages. The oldest one is the Moldavian Platform (Precambrian) located in front of the East Carpathians partly covered by the Foredeep. The Moesian Platform develops south to the orogenic chain ; it is largely overlapped by the Foredeep and also underthrusts below the Moldavides.

Almost within each of these tectonic units (except the Transylvanian Basin and the Moldavian Platform) the concentration of geothermal resources was inferd either from surface manifestations as thermal springs or from deep boreholes exhibiting temperatures higher than "normal". In some areas it is too deep to reach economically a potentially useful energy source using current drilling technology. However, due to particular structural situations, a considerable amount

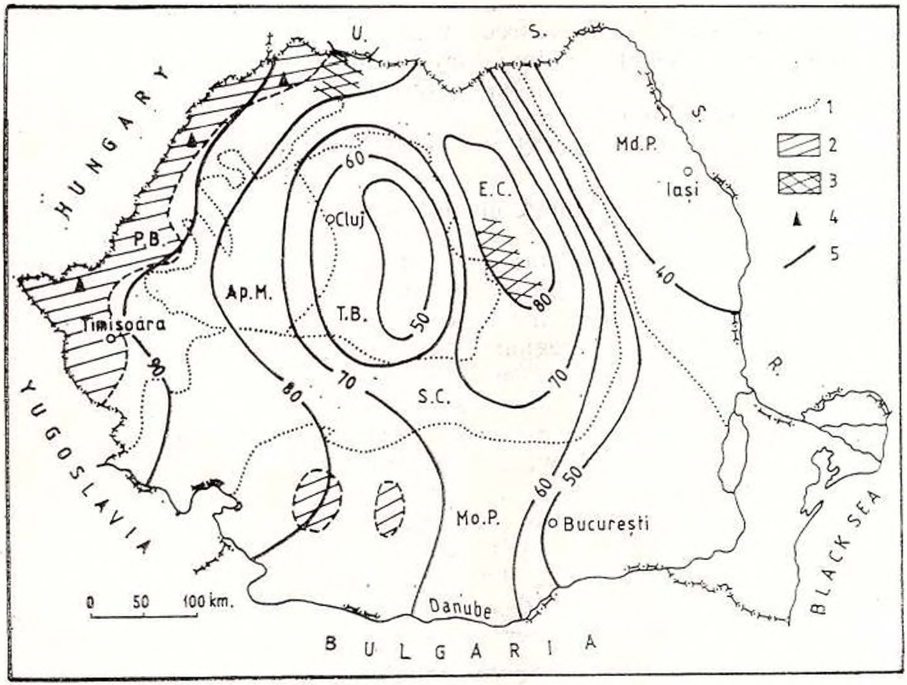

Fig. 1 — Areas as being prospectively valuable for geothermal energy in Romania. 1, limit between tectonic units ; 2, areas with hydrogeothermal systems ; 3, areas with hot dry rock ; 4, sites where geothermal energy is in use ; 5, terrestrial heat flow (mW m^{-2}) : Md.P., Moldavian Platform ; Ms.P., Moesian Platform ; E.C., East Carpathians ; S.C., South Carpathians ; Ap.M., Apuseni Mts ; T.B., Transylvanian Basin ; P.B., Pannonian Basin.

of heat exists near surface on the eastern limit of the Pannonian Basin, in the Moesian Platform, in the East Carpathians.

It is interesting to notice that the most important areas of concentration of geothermal resources are clearly outlined by the heat flow density map (Fig. 1). Accordingly, the eastern limit of the Pannonian Basin exhibits heat flow values exceeding 90 mW m^{-2} where numerous hydrothermal systems exist, and which are characterized by temperatures at the well head of 50-120°C and by yields of 5-30 l/s. The regional high heat flow anomaly was explained by the thinning of the lithosphere as a result of the extensional process produced since the Miocene time (Veliciu, Visarion, 1982).

As for the positive anomaly of 80-110 mW m^{-2} located at the inner part of the East Carpathians, the source of heat is related to the Miocene subduction from the Carpathian area ; the calc-alkaline volcanism was active in this region until the Upper Pliocene-Lower Quaternary.

The Moesian Platform is characterized by the presence of some Hercynian acid magmatic intrusions into its basement. These constitute local radiogenic heat sources which may sustain convective geothermal systems from the sedimentary cover.

The Moldavian Platform and the Transylvanian Basin are associated to the low heat flow values (average 40 mW m^{-2}) ; no geothermal manifestation has been noticed yet within these tectonic units.

Characteristics of the Convective Hydrothermal Systems

Eastern limit of the Pannonian Basin (Western Plain). According to the lithology and the structure of the water-bearing formations, in the Western Plain there have been distinguished four main hydrothermal structures with a regional extent exceeding 8 600 km^2 (Fig. 2). Two of these structures are situated in the northern and southern part respectively and they have been formed owing to the sedimentation of a thick sequence constituted by sands and sandstones, within a post-Senonian depression with a maximum depth of 2 000 m.

The central part of the Western Plain is underlain by the carbonatic deposits belonging to the Bihor Autochthon (Oradea zone) which constitute a distinct hydrogeologic unit. Southwards there developed another hydrogeologic structure of geopressured type, into conglomerates, sandstones and sands.

As for their thermal characteristics, the above mentioned hydrostructures have been separated as geothermal systems (Tab. 1) : with temperatures over 70°C (Biharia-Săcueni, Oradea-Borş, Ciumeghiu-Vărşand, western Banat) and temperatures under 70°C (Carei-Satu Mare, Mureş-Crişul Alb, eastern Banat). Except the geothermal system Carei-Satu Mare, all others manifest themselves as artesian with initial static pressures of 2 to 5 atm.

The yields of the wells from the geothermal systems connected with granular deposits of the Upper Pliocene are higher for the Biharia-Săcueni and western Banat where the permeabilities range between 400-600 μD and 800-1 000 μD respectively. Where the wells were spaced closer than 1 km, a decreasing of pressure, at a rate of 0.4-0.9 atm per mil. m^3 extracted water, has been observed during the simultaneous exploitation of five or seven wells. This phenomenon has been explained by the slow water circulation through granular sediments which behave themselves almost as a hydraulic structure. It is not the same case as for the Oradea-Borş geothermal system in fissured limestones and dolomites, where an active water recharge maintains the pressure in aquifer ; no pressure drop has been reported even for intensive exploitation of the geothermal wells.

In order to get an idea on the enormous heat contained in the geothermal systems from the Western Plain (eastern limit of the Pan-

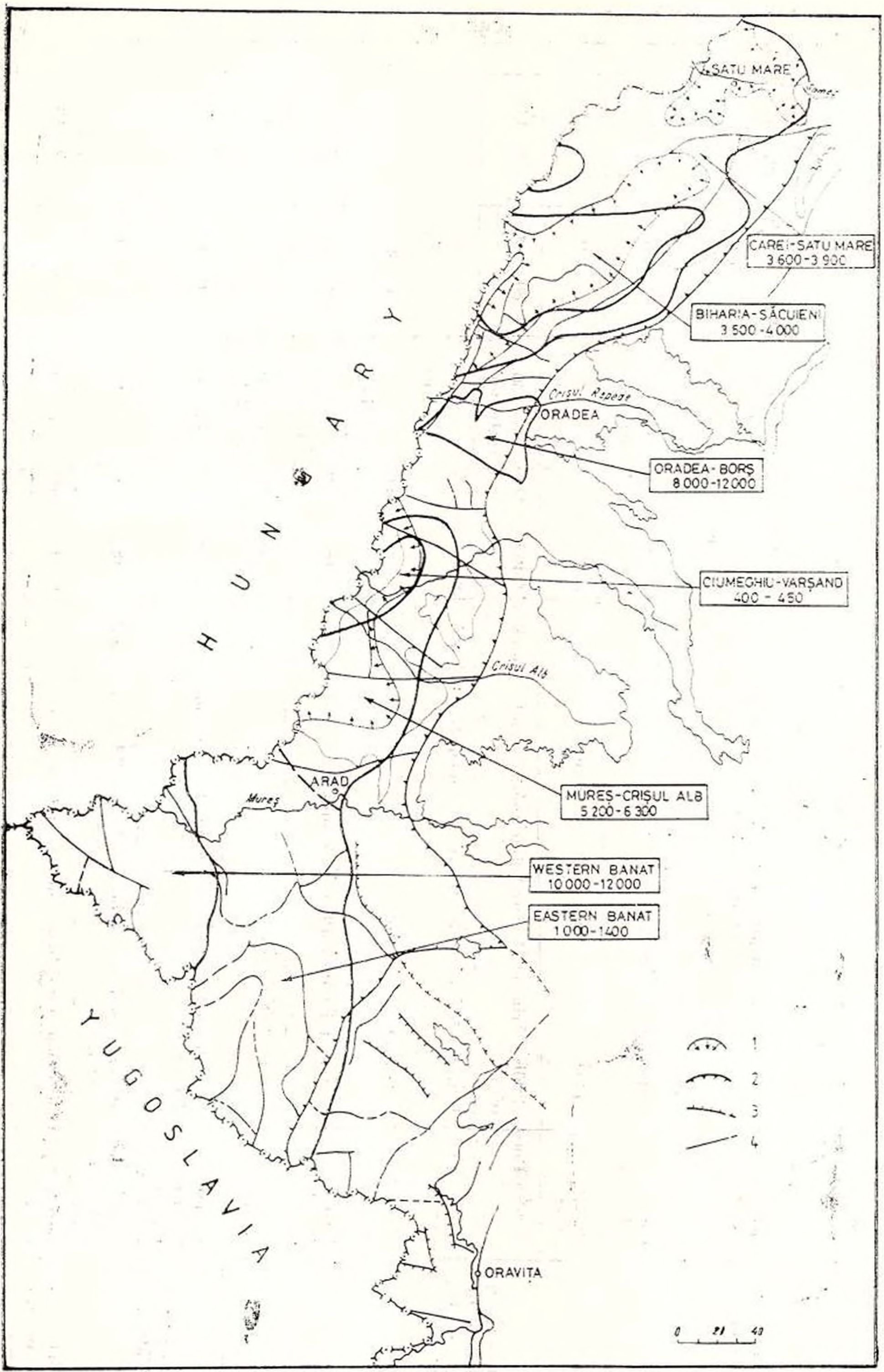

Fig. 2 — Heat (in kcal/sec) from hydrogeothermal systems in the Western Plain of Romania.

1, zone of maximum thickness of the Pannonian sediments ; 2, eastern limit of the Pannonian Basin outlined by geophysics ; 3, overthrusts in the basement ; 4, faults.

TABLE

Characteristics of the exploited convective geothermal systems in Romania

Tectonic units	Geothermal systems	Lithostratigraphic characteristics of the aquifers	Depth at the top of the aquifers (m)	Average yields per well (l/s)	Temperature at the well-head (°C)	Minerali-zation (g/l)
Eastern limit of the Pannonian Basin	Carei-Sătu Mare	Upper Pliocene (sands, sandstones)	1,000—1,400	5—15 (artesian)	55—60	3—4
	Biharia-Săcuieni	Upper Pliocene (sands and sandstones)	1,400—1,700	5—25 (artesian)	70—80	3.5—5
	Oradea-Borș	Middle Triassic (fissured limestones and dolomites)	2,400—3,000	15—30 (artesian)	80—120	1—14
	Ciumeghiu-Vărșand	Lower Pliocene (sandstones and microconglomerates)	2,000—2,600	8—15 (artesian)	80—90	5—7
	Mureș-Crișul Alb	Upper Pliocene (sands)	800—1,400	15—25 (artesian)	50—60	2—3.5
	Eastern Banat	Lower and Upper Pliocene (sands)	800—1,200	5—15 (artesian)	40—55	2—3
	Western Banat	Upper Pliocene (sands)	1,600—2,000	15—30 (artesian)	70—85	4—5
Moesian Platform	București-Otopeni Buftea	Jurassic and Cretaceous (fissured limestones and dolomites)	2,200—3,000	15—20	40—60	1.4—2.5
	Însurăței-Hîrșova	Jurassic and Cretaceous (fissured limestones and dolomites)	800—1,200	8—16	45—60	2—3.5
Getic Depression	Cozia-Căciulata	Senonian (sandstones, microconglomerates)	2,220—2,900	7—20	88	12—13

nonian Basin), it is mentioned that the estimate thermal energy (for 15°C reference temperature) may be of 7×10^9 Gcal or approximately $8\,600 \times 10^9$ MW(t).

Moesian Platform. The western part of the Moesian Platform is characterized by a higher terrestrial heat flow, but for its eastern part the thermalisation of the waters is refered to the deep circulation through numerous fractures, some of them of transcrustal importance. Exploratory drilling outlined till now two zones of interest for geothermal waters : București-Otopeni-Buftea and Însurăței-Hîrșova. Near București the drillholes opened in carbonatic formations of Upper Jurassic age (2 200-2 800 m deep) a thermal aquifer with remarkable dimensions and temperatures ranging between 42 to 60°C. The outflows obtained by pomping have been of 20-30 l/sec.

In the Însurăței-Hîrșova zone geothermal waters are contained in carbonatic deposits of Jurassic age situated in an elevated position of the basement (400-1 200 m deep). The aquifer exhibits an artesian character (yields of 8-16 l/sec.). From the point of view of thermalisation, this structure is associated to the deep water circulation on the Capidava-Ovidiu transcrustal fault. As a prospective zone, the whole Jurassic carbonatic formation, with large extent in the eastern part of the Moesian Platform, deserves further investigations.

Getic Depression. On the orogenic flank of the Getic Depression located between the Moesian Platform and the South Carpathians the geothermal waters have been infered from the boreholes drilled in the Căciulata-Cozia zone. Some artesian aquifers were previously encountered at depths of 200-1 200 m with temperatures up to 54°C. The later investigation of the deeper Cretaceous formations (Senonian), has identified a geothermal system with temperatures of 86-90°C and yields of 20 l/s per well. The isotopic study of the aquifer suggested the existence of an active hydrodynamic system.

South Carpathians. This tectonic unit has a limited value for the geothermal resources due to the structural and geothermal conditions. Emergences of thermal waters with moderate temperatures (30-60°C) from Herculane, Călan, Mehadica are in use for medical therapy.

Apuseni Mountains. Located between the hydrothermal regime of the Pannonian Basin and the low heat flow associated to the Transylvanian Basin, the Apuseni Mts exhibit some thermal springs with temperatures of 30-40°C. These springs emerge in fault zones which delimit the intra-mountain posttectonic depressions.

Regional Conductive Dominated Systems

Geothermal conductive dominated systems (hot dry rocks) are situated in areas that underlie most of the inner part of the East Carpathians, due to the regional high heat flow anomaly.

The observed surface heat flow density values constituted the basis for the maps showing the location of the hot upper crustal regions with temperatures higher than 150°C. Direct measurements of temperatures were possible only in boreholes or mines, but extrapolation of these data to greater depth had limited validity. More reliable temperature data for great depth have been obtained from heat flow values associated with heat generation and heat conductivity of rocks, which made it possible to calculate temperature-depth profiles (Haenel, 1979).

The calculated temperatures at a depth of 3 km for the East Carpathians are shown in Figure 3. The map suggested that the highest temperatures (more than 150°C) occur in a conductive dominated area of approximately 1 000 km^2 situated in the central and southern parts of the Harghita Mts. The total heat stored within the first 10 km of the crust may be here of 3.5×10^{20} Joules (Veliciu et al., 1984). Conductive dominated areas of economic interest area also disclosed around Oaș-Gutîi Mts, with a south-eastwards prolongation.

An investigation has been performed to study if any igneous-derived thermal anomaly may exist in connection with the Pannonian and/or Pliocene volcanic and/or subvolcanic structures. The age-volume data for each individual volcanic system showed the approximate present position of the system in relation to its probable cooling state. Due to the relatively old age of the last eruption (ranging between 7 and 1.5 m.y.) the calculations indicated that almost all igneous systems from the East Carpathians reached the ambient temperature or were very near of this. Consequently, from the point of view of the geothermal resources, the total heat preserved into the igneous systems (approximately 0.003×10^{20} Joules) is of very limited economic value.

Conclusions

The complex geological researches carried out in the last 10 years on the geothermally potential structures of Romania have led to the identification of important hydrogeothermal systems for which water reserves and heat content have been calculated and which are partially utilized in various purposes.

About 100 000 tec are expected to be spared by using geothermal waters for heating houses and greenhouses, drying seeds, cereals and pottery, and as warm household water.

While geothermal waters are increasingly largely used, researches are undertaken for solving important matters concerning the reinjection of already used waters for maintaining the reservoir energy as well as its impact on the environment.

The production of electricity out of geothermal waters with moderate temperature in the binary system, the gas recovery and the turning into account of useful mineral substances are experimented — pilot phase.

In Romania increasingly important attention is paid to the utilization of geothermal resources and the programs of future research will be focussed on the investigation of the possibilities to use thermal energy of hot dry rocks.

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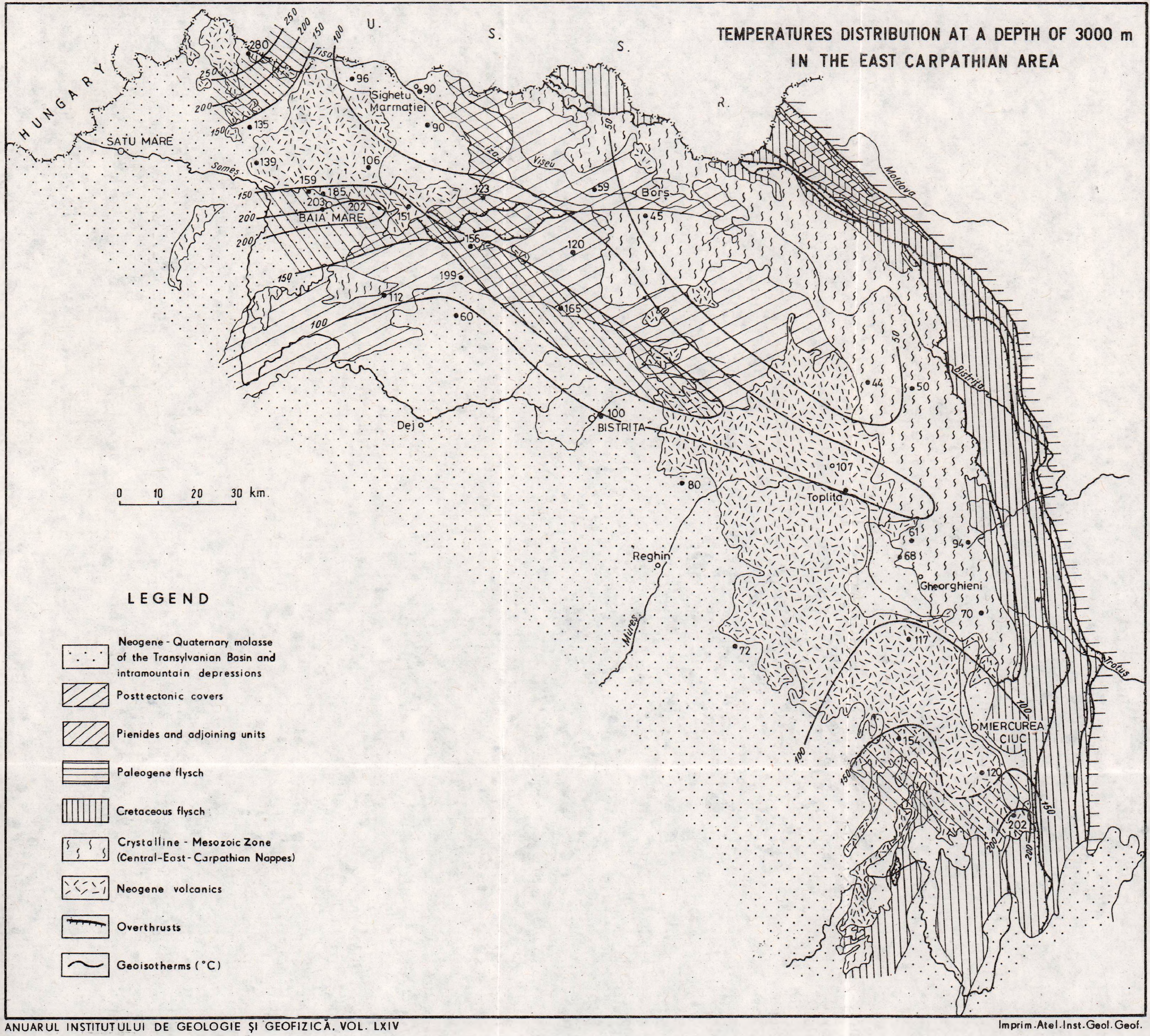

TEMPERATURES DISTRIBUTION AT A DEPTH OF 3000 m IN THE EAST CARPATHIAN AREA

Imprim.Atel.Inst.Geol.Geof.

Géophysique

PARADIGMES STRUCTURAUX-DÉPOSITIONNELS DES FORMATIONS PANNONIENNES DU SECTEUR ROUMAIN DE LA DÉPRESSION PANNONIENNE DÉDUITS PAR ANALYSE ET INTERPRÉTATION DES PROSPECTIONS SISMIQUES DE RÉFLECTION

PAR

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Le perfectionnement continu des technologies d'enregistrement et d'interprétation ont permis une meilleure valorification du contenu lithostratigraphique des coupes sismiques (Ionescu, 1967 ; Ionescu et al., 1979 ; Dicea et al., 1982 ; Lungu et al., 1983).

Les coupes sismiques effectuées dans les formations pannoniennes mettent en évidence des paquets de réflections dont l'amplitude, la fréquence, le cycle, la continuité et la configuration interne rendent clair les conditions de sédimentation des unités génétiques dépositionnelles, des paléoreliefs, des déformations structurales syn- ou postdépositionnelles et parfois de la lithologie et du contenu en fluides.

Rapports stratigraphiques entre les formations pannoniennes et prépannoniennes

Sur les coupes sismiques de temps représentées dans les figures 1 et 2 peuvent être délimitées nettement, par des surfaces de discordance, une séquence inférieure attribuée à partir des données de forage au Miocène moyen-supérieur (Badénien-Sarmatien moyen) et trois séquences dépositionnelles appartenant aux formations pannoniennes.

La séquence inférieure (Mi) est formée des réflections sismiques à faible continuité et configuration tantôt chaotique (fig. 1), tantôt parallèle-ondulée (fig. 2).

La carte à isobathes rapportés à la discordance prépannonienne (fig. 3) relève un relief fort, accidenté qui étant en même temps nivelé a été colmaté peu à peu par les formations pannoniennes. Le paléoréseau hydrographique représenté sur la carte a une orientation

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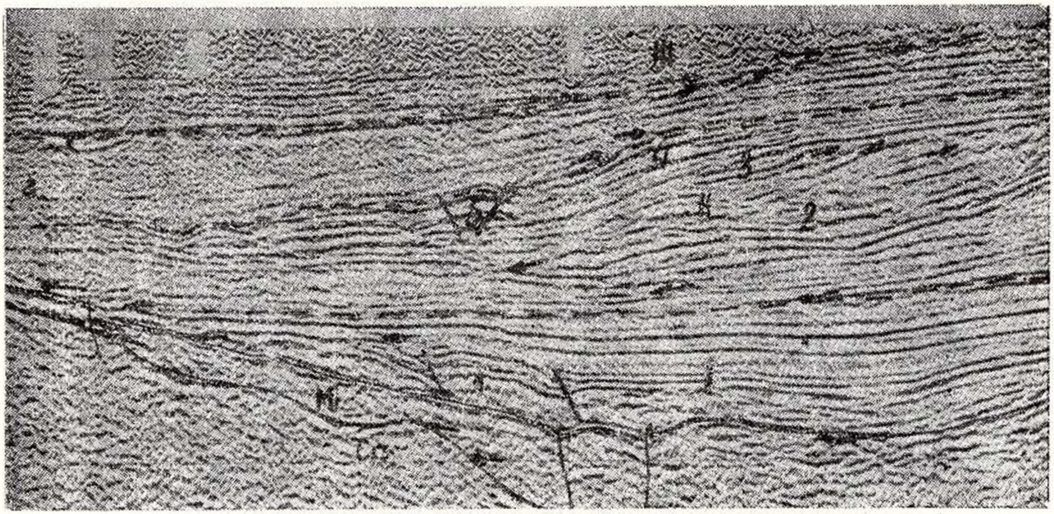

Fig. 1 — Coupe sismique de temps migrante de la zone d'Oradea. Mi, séquence Miocène moyen-supérieur ; Cr. soubassemennt cristallin ; I, II, III, séquences dépositionnelles en Pannonien.

1, glissements gravitationnels ; 2. configuration chaotique ; 3, configuration oblique ; 4. configuration sigmoïdale ; a, faciès de remplissage de la partie frontale et basse de la séquence de progradation, à configuration sigmoïdale.

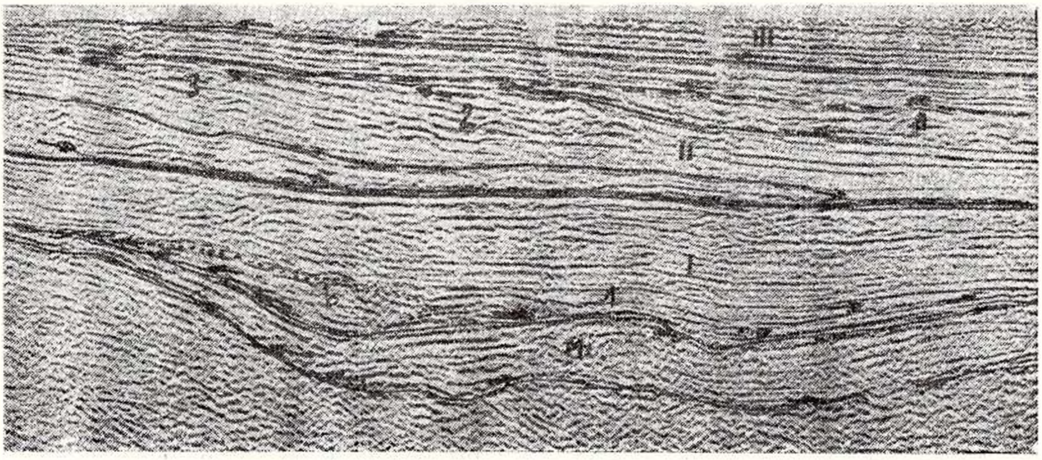

Fig. 2 — Coupe sismique de temps migrante de la zone d'Arad. Mi, séquence Miocène moyen-supérieur ; I, II, III, séquences dépositionnelles en Pannonien. 1, anticlinal de moulage ; 2. configuration chaotique ; 3, configuration oblique ; a, faciès de remplissage de la partie frontale de la séquence de progradation, à configuration oblique.

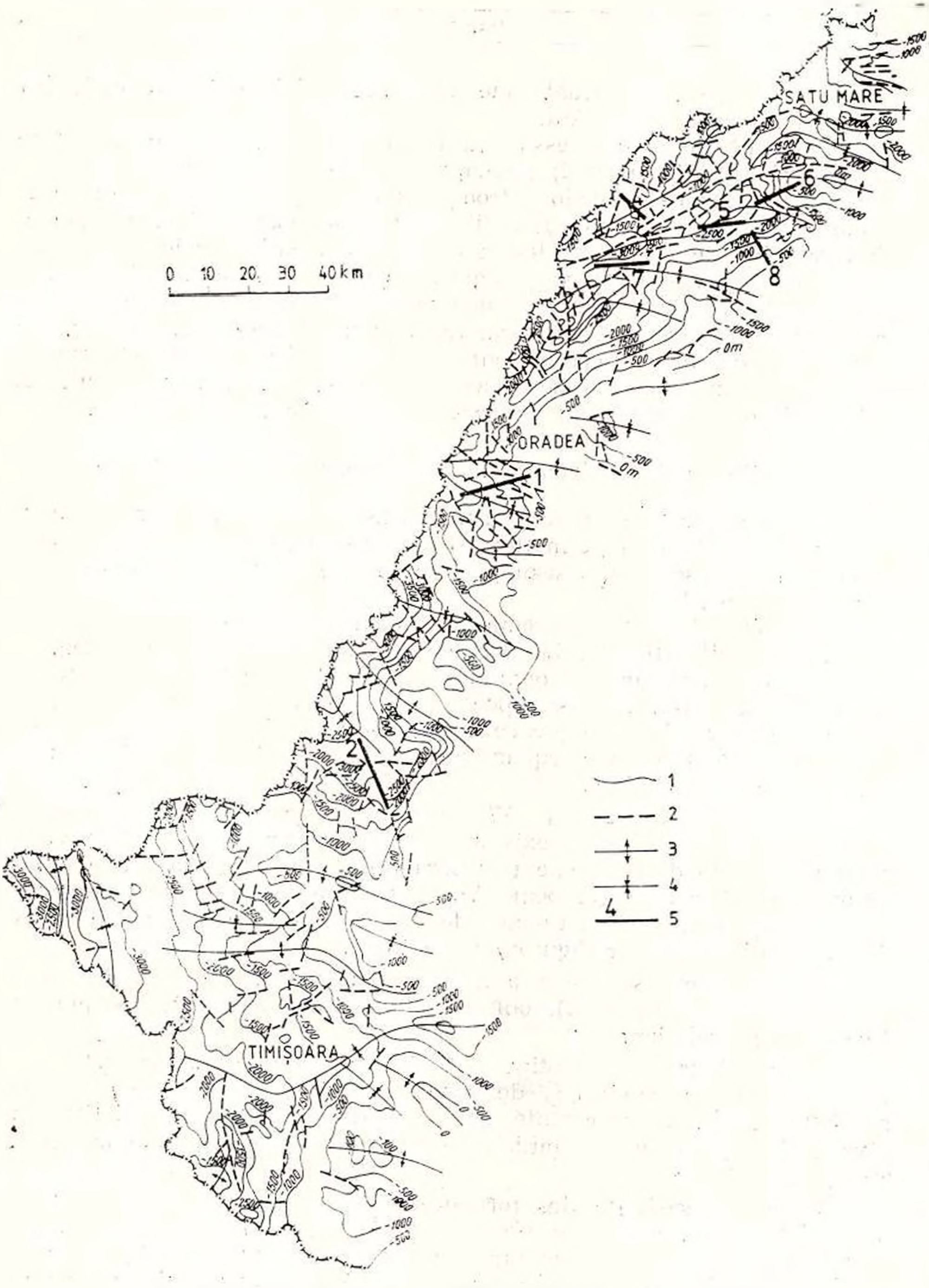

Fig. 3 — Carte de la surface du relief prépannonien.
1, isobathes du relief prépannonien ; 2, failles à effet dans le Pannonien inférieur ; 3, zone de soulèvement ; 4, zone d'abaissement ; 5, coupes sismiques présentées dans l'article.

presque est-ouest, montrant que la source d'alimentation était les chaînes montagneuses de l'est.

Les effilements progressifs du Pannonien inférieur sur la surface de discordance (fig. 1 et 2) („onlap" et „downlap" — Sangree et al., 1977) ainsi que les réflections tronquées de la partie supérieure de la séquence miocène démontrent qu'il s'agit d'un moment d'exondation et d'interruption de la sédimentation entre le Sarmatien et le Pannonien.

Comme on voit dans la figure 3, au niveau de la discordance prépannonienne apparaissent de nombreuses failles de soubassement ne traversant que rarement les formations pannoniennes (fig. 1 et 2). Au moment où ces formations en sont affectées, l'effet des failles s'atténue très rapidement dans la séquence inférieure. Un nombre réduit de failles du soubassement ont été actives durant tout le Pannonien.

Conditions de sédimentation des formations pannoniennes

La configuration et les terminaisons des réflections des coupes sismiques (fig. 1 et 2) permettent de séparer dans les formations pannoniennes trois séquences sismiques censées en tant que séquences dépositionnelles (I, II, III).

Les réflections de la première séquence dénotent quelquefois une bonne continuité (fig. 2). La partie inférieure de la séquence (fig. 1) présente une configuration onduleuse et chaotique (1), en relevant un glissement gravitationnel sur pente. Le long de ces deux coupes, la partie inférieure de la séquence a des contacts d'effilement progressif sur pente („onlap"), en marquant ainsi un hiatus de non-dépôt, généré par un colmatage graduel.

Dans la figure 2, les réflections de la partie inférieure de la séquence moule le relief préexistant, engendrant un anticlinal de revêtement (1). L'effet de revêtement disparaît peu à peu vers la partie supérieure. Sur cette coupe-ci peut être aussi observé l'effet de glissement et de remplissage gravitationnel des zones basses (b) représenté par l'aspect lenticulaire à configuration chaotique.

La deuxième séquence (fig. 1, 2) est formée des réflections à configuration chaotique (2), oblique (3) et sismoïdale (4), typique au faciès de progradation.

La configuration chaotique et oblique indique des moments de dépôt rapide en conditions de haute énergie. La configuration sigmoïdale (4, fig. 1) représente un moment de stagnation relative au niveau de la mer et un milieu de dépôt de faible énergie (Sangree et al., 1977).

Dans la succession des formations pannoniennes, la séquence de progradation marque un moment de soulèvement brusque de la terre ferme dû à la phase rhodanienne de l'orogenèse alpine. L'apport massif de matériel provenant des zones exondées se produisait au moyen de forts systèmes deltaïques, là où la sédimentation avait lieu au-dessus du niveau de base.

On constate dans les deux figures que sur la zone frontale de la pente de progradation (cliniforme) se développent des dépôts discordants qui s'effilent progressivement, en amont de la pente, pour revenir ensuite à une couche continue à configuration sigmoïdale. La discordance de la zone frontale met en évidence un nouveau hiatus de non-dépôt, déterminé par un moment de stagnation de l'apport de matériel et de la montée du niveau de l'eau. Les couches discordantes représentent des étapes d'avancement des eaux sur la pente de progradation.

La troisième séquence est représentée par des réflections à bonne continuité, dont la périodicité des cycles est relativement constante. Les réflections ne sont pas horizontales mais légèrement inclinées à configuration divergente.

Structure des formations pannoniennes

La structure des formations pannoniennes apparaît comme la conséquence des conditions de sédimentation du bassin pannonien, n'étant pas déterminée par une tectonique plicative. Les structures engendrées sont de type des anticlinaux de revêtement, de tassement ou bien de glissement gravitationnel („roll-over").

— Les anticlinaux de revêtement (fig. 2) se développent fréquemment dans la séquence inférieure (1) qui, en conditions de faible énergie, recouvrent le relief préexistant. L'effet du relief s'atténue au fur et à mesure que l'épaisseur s'accroît. Quand la différence entre la zone abaissée et la zone soulevée est plus grande intervient alors le phénomène de tassement différentiel (fig. 4).

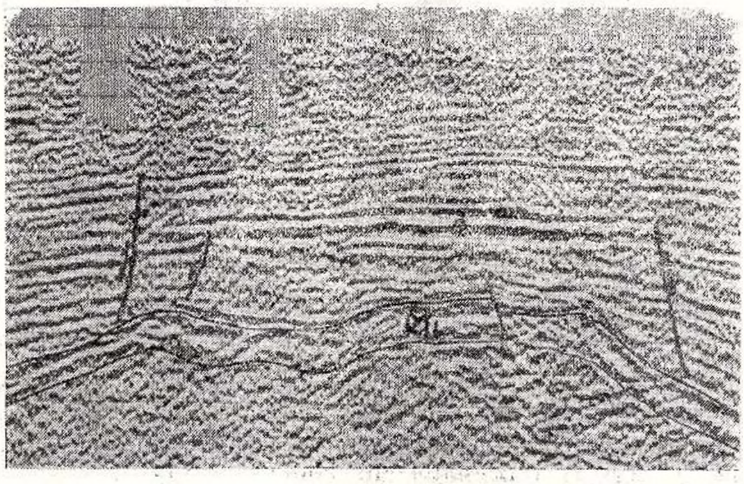

Fig. 4 — Coupe de temps migrante de la zone de Pişcolţ qui relève un anticlinal de tassement différentiel. Mi. séquence miocène ; a, couche de sable sommital à gaz („bright-spot").

— Les anticlinaux de tassement (fig. 4) sont favorisés par la présence du relief accidenté, surmontés par des sédiments déposés sous l'influence des courants de turbidité et de la gravitation. Par suite de ce fait, les zones basses du relief recoivent une quantité plus grande de sédiments, et dans la majorité des cas, plus pélitiques. Dans les zones soulevées, l'épaisseur des sédiments déposés est plus réduite, mais généralement ils sont plus grossiers (fig. 4). La puissance des sédiments des zones basses ainsi que la prédominance des pélites conduisent à une compaction plus grande de ceux-ci que dans les zones soulevées. Les couches se voûtent au-dessus des zones soulevées, processus qui diminue peu à peu, après le colmatage du relief.

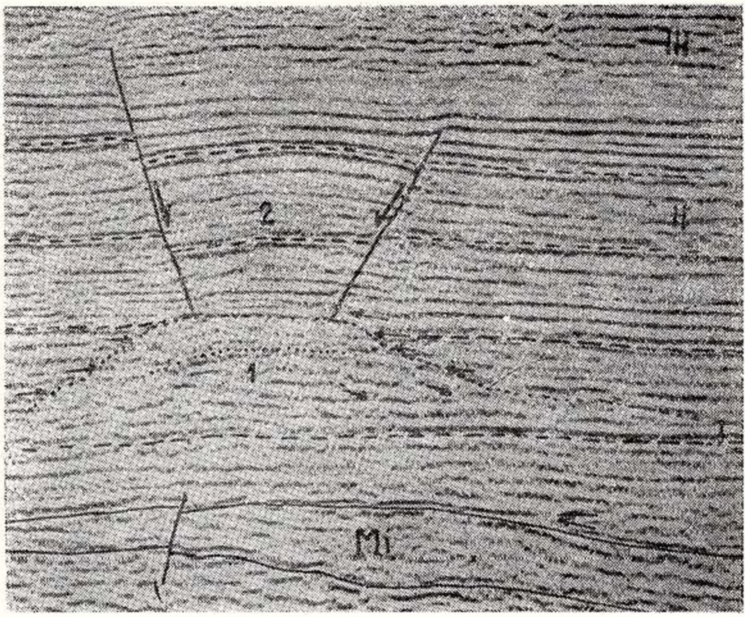

Fig. 5 — Coupe sismique de temps migrante de la zone de Carei qui relève un anticlinal de tassement différentiel. Mi, séquence miocène ; I, II, III, séquences dépositionnelles en Pannonien.
1, lentille d'argile formée par tassement différentiel ; 2, graben sommital de distension.

On remarque dans la figure 4, au-dessus du relief positif, une réflection de basse fréquence et d'amplitude grande (a). Cette anomalie, vis-à-vis des couches environnantes, est redevable à la présence d'une couche épaisse de sable, tandis que le contraste évident a été interprété par suite de la présence des gaz.

Un autre type d'anticlinal de tassement différentiel est représenté dans la figure 5. La pression différentielle exercée sur la couche pélitique (1) a déterminé une migration des argiles qui ont formé une lentille plane-convexe. L'effet de la lentille d'argile s'est transmis aux couches de dessus qui s'incurvent. Du fait de la distension de voûte a pris naissance le graben sommital (2).

— Anticlinaux gravitationnels. Dans la partie basale de la séquence inférieure (fig. 1) se développe un paquet de réflections à configuration chaotique et ondulée, affectée par des failles de glissement, dont la configuration est typique pour le glissement gravitationnel.

On présente dans la figure 6 l'image d'un anticlinal de coulement gravitationnel (roll-over). Les couches du compartiment supérieur de la faille subissent une rotation en aval sur le plan de la faille, due à la coulée de la couche d'argile sur pente.

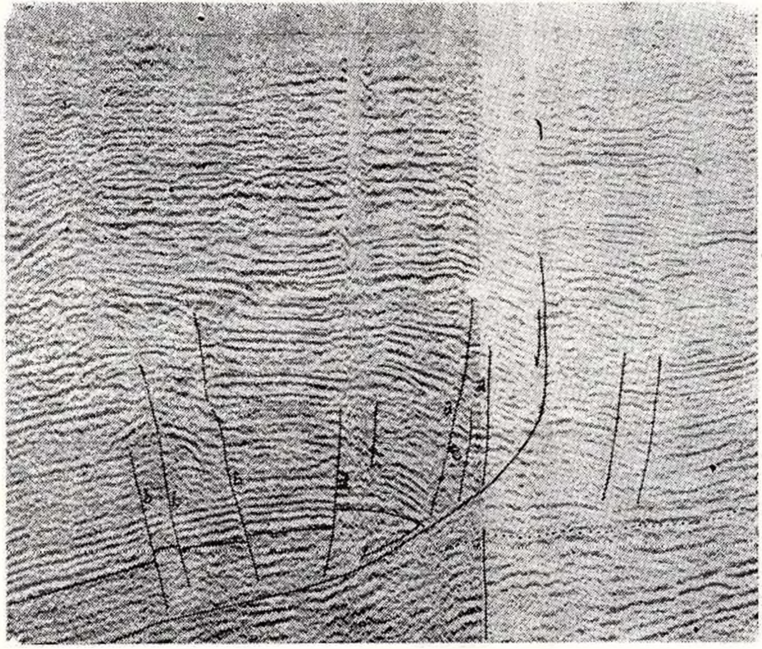

Fig. 6 — Coupe sismique de temps migrante de la zone de Tarcea qui relève un anticlinal de rotation („roll-over") dû à la coulée gravitationnelle des argiles sur pente dépositionnelle. a, failles synthétiques ; b, failles antithétiques.

Comme effet compansatoire, la voûte est affectée par des failles de distension. Les couches du compartiment inférieur de la faille subissent elles-aussi un liage en raison du poids des couches du compartiment supérieur qui glissent. L'effet de suction du compartiment supérieur est neutralisé, en temps, par le dépôt des couches sableuses.

Un autre exemple d'anticlinal redevable à la coulée gravitationnelle des argiles sur une pente est illustré dans la figure 7. Sur cette coupe on observe le corp d'argile épaissi, au-dessus duquel les couches recouvrantes se voûtent et engendrent des failles, en formant des grabens de voûte. Dans ce cas, il semble que le mouvement vertical différentiel des blocs du soubassement a joué un rôle important.

Dans la figure 8, la faille principale affecte tant le soubassement que le Pannonien. Dans le compartiment abaissé prend naissance un

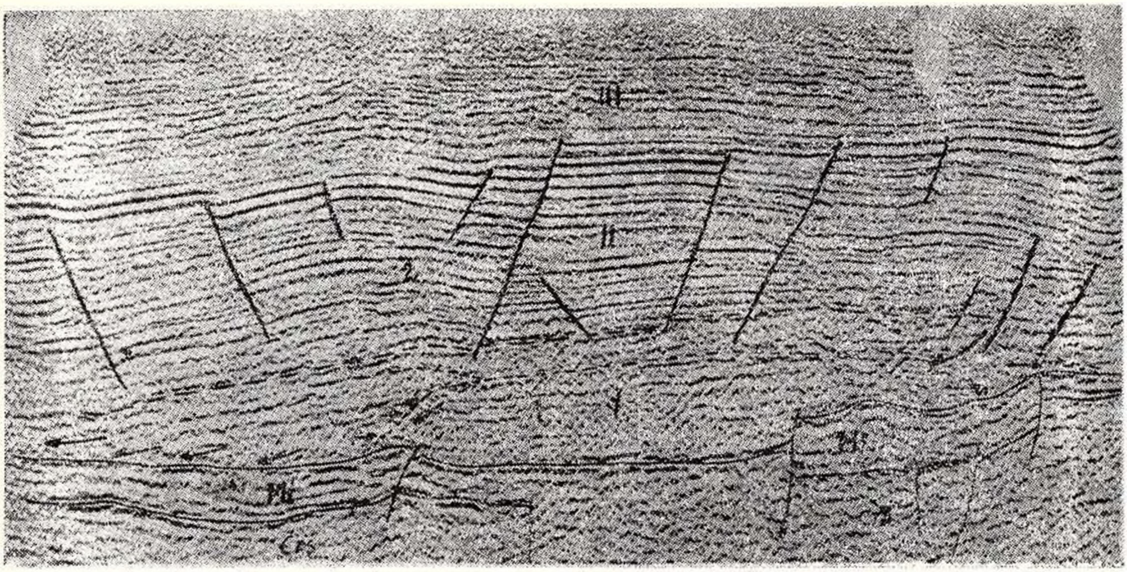

Fig. 7 — Coupe sismique de temps migrante de la zone de Tarcea qui relève un anticlinal de rotation („roll-over“) dû à la coulée des argiles sur pente dépositionnelle en Pannonien.
1, lentille d'argile formée par coulée gravitationnelle ; 2, graben sommital de distension.

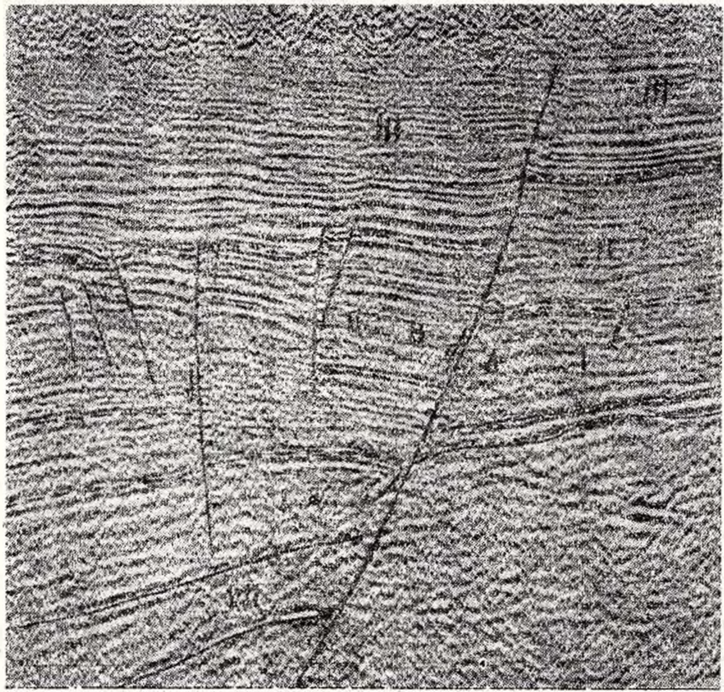

Fig. 8 — Coupe sismique de temps migrante de la zone de Mecenţiu qui relève un anticlinal de rotation („roll-over“) dû aux failles syndépositionnelles, à la coulée gravitationnelle et au tassement différentiel des argiles. Mi, séquence miocène. A, compartiment soulevé de la faille ; B, compartiment abaissé de la faille ; a, séquence argileuse déformée par glissement gravitationnel et tassement différentiel.

anticlinal de rotation („roll-over“) dû à la coulée de l'argile de la Ière séquence. La faille a été active même au cours du dépôt de la IIIe séquence supérieure.

Conclusions

A la suite de l'analyse des coupes sismiques de la dépression pannonienne on conclut que :

Les formations pannoniennes se sont déposées sur un relief fortement accidenté, à la surface duquel affleuraient des formations miocènes, crétacées, jurassiques, triasiques, paléozoïques et du soubassement.

Entre les formations du Badénien-Sarmatien et celles du Pannonien a existé, au moins pour le secteur roumain de la dépression, un moment d'exondation et d'interruption de la sédimentation.

Dans les formations pannoniennes se différencient, à partir de la configuration et des terminaisons des réflections sismiques, trois séquences dépositionnelles, en relevant les conditions spécifiques de sédimentation.

Le développement de ces trois séquences varient beaucoup suivant le faciès et la puissance, elles étant séparées par des moments de hiatus dépositionnel, du fait du changement continu des rapports apport du matériel/subsidence et de la variation du niveau marin.

On ne constate pas dans la structure des formations pannoniennes l'influence des forces diastrophiques tangentielles.

Les failles qui affectent les formations pannoniennes sont généralement des failles gravitationnelles, de croissance ou bien des failles dépositionnelles. Les failles dépositionnelles sont la continuation des failles du soubassement.

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EVALUATION OF THE SLIDING RISK OF SLOPES IN CONSOLIDATED CLAYS BY STOCHASTIC SIMULATION

BY

PETRE BOMBOE [1], FLORICA STROIA [1]

Introduction

It is a well-known fact that in a mass of clay rocks the material properties, particularly the cohesion and the angle of friction, should be regarded as stochastic variables because their values fluctuate within the rock body. In a deterministic analysis the stability of the slope in a homogeneous clay is calculated by means of given values of the cohesion and angle of friction, which could be the mean values of the experimental data in accordance with the maximum shear strength or residual shear strength assumption. But any mean value has a standard deviation, which represents the variations of the cohesion or angle of friction within the rock mass. In consequence the factor of safety for a given slope should be regarded as a probabilistic value, depending on the fluctuations of the shear strength properties of the rock along the critical sliding circle.

Geologic Environment and Shear Strength Experimental Data

The stratigraphic sequence outcropping in the investigated zone includes Barremian limestones, overlain transgressively by Aptian sandstones and a thick complex of Aptian clay rocks, covered by a blanket of Quaternary loess (Fig. 1).

The designed slope will cut entirely the Aptian clay complex, which includes few transitions to sandy-clay and marls. The attitude of the layers within the clay-complex is mostly horizontal, without a definite stratification, but with frequent microjoints randomly orientated. Extensive geological investigations have been carried out by exploration drillings and a complete field-test programme concerning the rock shear strength have been performed.

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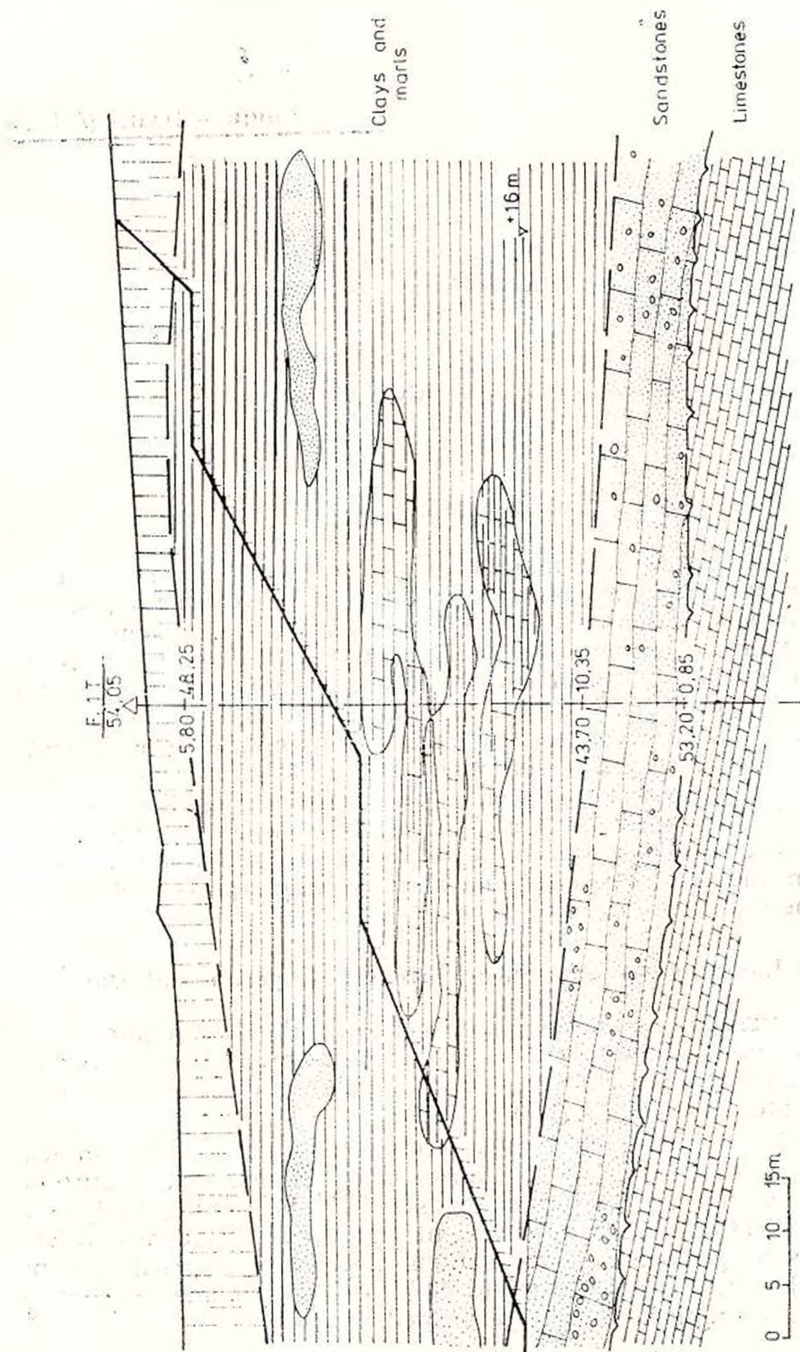


Fig. 1 — Geological sections through the designed slope.

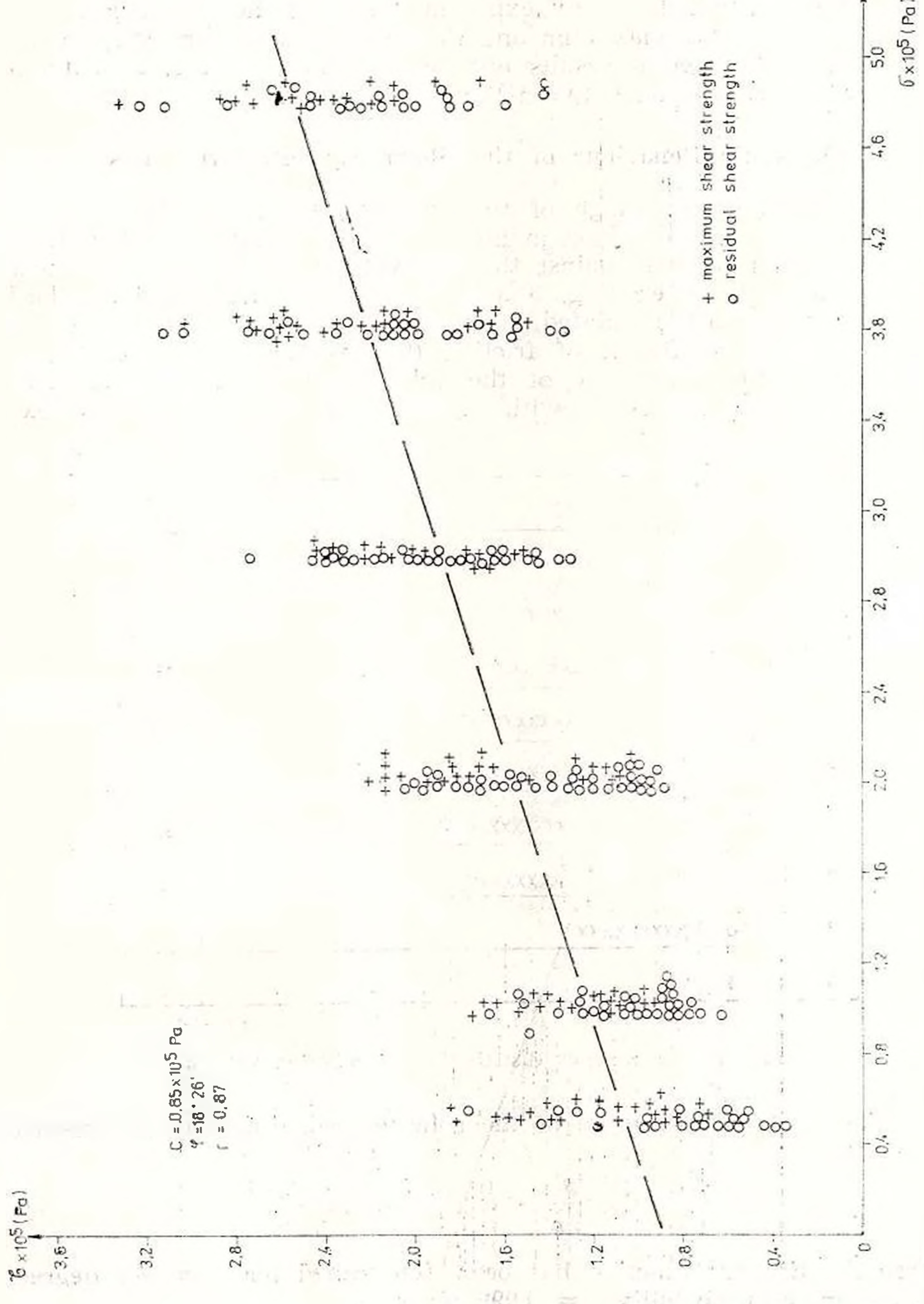


Fig. 2 — Distribution of the shear strength experimental data.

In accordance with the experimental data, the shear strength of the investigated Aptian clays variates between $\tau = 3.5$ and $\tau = 0.4 \times 10^5$ Pa. Although at any experimental point there is a good distinction between the maximum and the residual shear strength, on the whole the experimental results are very fluctuating (Fig. 2) without a distinct discrimination between the two levels of shear strength.

Stochastic Behaviour of the Shear Strength Properties

By dividing the range of the experimental values for cohesion and for coefficient of friction in intervals and finding out the number of observations plotted against the interval, the distribution of shear parameters is described (Fig. 3 and Fig. 4). The mean and standard deviation have been calculated for cohesion ($c = 0.86 \times 10^5$ Pa ; $s_c = 0.369$) and for coefficient of friction ($\mathrm{tg}\varphi = 0.29$; $s\varphi = 0.134$). It follows that the variability of the cohesion and friction coefficient exceeds 42%, in accordance with the stochastic behaviour of the experimental data.

CLASS	INTERVAL c (× 10^5 Pa)	10 20 30	Frq (n_i)
1	0 - 02	XXXX	4
2	0,2 - 0,4	XXXXXXXXXXXXX	13
3	0,4 - 0,6	XXXXXXXXXXXXXXXXXX	18
4	0,6 - 0,8	XXXXXXXXXXXXXXXXXXXXXXX	23
5	0,8 - 1,0	XXXXXXXXXXXXXXXXXXXXXXXXXXX	27
6	1,0 - 1,2	XXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXX	32
7	1,2 - 1,4	XXXXXXXXXXXXXXXXX	17
8	1,4 - 1,6	XXXXXXXXX	9
9	1,6 - 1,8	XX	2

$\Sigma n_i = 145$

Fig. 3 — Frequency distribution of cohesion values.

The confidence interval of the cohesion calculated by the known inequality

$$\bar{c} - t_{(\nu,\ p)} \frac{s_c}{\sqrt{n}} < c_{\mathrm{real}} < \bar{c} + t_{(\nu,\ p)} \frac{s_c}{\sqrt{n}} \tag{1}$$

where the Student function has been determined for $\nu = n\text{-}1$ degree of freedom and probability $P = 0.999$, gives :

$$0.752 < c_{\mathrm{real}} < 0.962$$

By a similar procedure it could be proved that the real mean value of the friction coefficient of the investigated clays fluctuates between

$$0.253 < \mathrm{tg}\,\varphi_{\mathrm{real}} < 0.327$$

The deterministic analysis of the slope stability using experimental mean values $\bar{c}$ and $\mathrm{tg}\varphi$ will give a solution with low confidence level since shearing parameters can take any value between the limits of the confidence intervals.

CLASS	INTERVAL tg φ	10 20 30 40	Frq. (n_i)
1	0 - 0.1	xxxxxx	6
2	0,1 -0,2	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	33
3	0,2 - 0,3	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	47
4	0,3 - 0,4	xxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxxx	34
5	0,4 - 0,5	xxxxxxxxxxx	11
6	0,5 - 0,6	xxxxxxxxxxxx	12
7	0,6 - 0,7	x	1
8	0,7 - 0,8	x	1

$\Sigma n_i = 145$

Fig. 4 — Frequency distribution of friction values.

The experimental shear strength equation could be identified by regression and correlation analysis.

The scattering diagram of the experimental data in σ, τ coordinates (Fig. 2) advances a suitable linear correlation, established by the regression coefficients, given by

$$a = \frac{\sum_{i=1}^{n} \sigma_i^2 \sum_{i=1}^{n} \tau_i - \sum_{i=1}^{n} \sigma_i \tau_i \sum_{i=1}^{n} \sigma_i}{n \sum_{i=1}^{n} \sigma_i^2 - \left(\sum_{i=1}^{n} \sigma_i\right)^2} \tag{2}$$

$$b = \frac{n \sum_{i=1}^{n} \sigma_i \tau_i - \sum_{i=1}^{n} \sigma_i \sum_{i=1}^{n} \tau_i}{n \sum_{i=1}^{n} \sigma_i^2 - \left(\sum_{i=1}^{n} \sigma_i\right)^2} \tag{3}$$

where n is the number of the shear tests and σ_i, τ_i are experimental values at any *i* experimental point.

In accordance with the available experimental data, from equation (2) and (3), results $a = 0.85$ and $b = 0.34$, so that the equation of the shearing strength given by the equation of the regression line will be

$$\tau = 0.85 + 0.34\,\sigma$$

The significance of the shearing strength equation is tested by the correlation coefficent of the experimental data, given by

$$r = \frac{n \sum_{i=1}^{n} \sigma_i \tau_i - \sum_{i=1}^{n} \sigma_i \sum_{i=1}^{n} \tau_i}{\left\{\left[n \sum_{i=1}^{n} \sigma_i^2 - \left(\sum_{i=1}^{n} \sigma_i\right)^2\right]\left[n \sum_{i=1}^{n} \tau_i^2 - \left(\sum_{i=1}^{n} \tau_i\right)^2\right]\right\}^{\frac{1}{2}}}$$

The result $r = 0.87$ points out a good linead correlation, which motivates a better confidence in the determined shear strength equation than in the mean values.

The concordance of the observed frequency distributions (Figs. 3, 4) with the normal distribution has been tested by chi- square (χ^2) method [1] ; which comes to the inequality

$$\chi^2 = 12.02 < \chi^2_{(\nu = k-3, P = 0.99)} = 16.8$$

for cohesion, certifying the normal distribution of the c_i experimental values. By a similar procedure has been tested also the normal distribution of the experimental values for the coefficient of friction within the investigated clay formation.

TABEL

$\bar{c} = 0.86$; $s_c = 0.369$

k	c_i	n_i	$t_i = \frac{c_i - \bar{c}}{s_c}$	$F(c_i)$	$p_i = F(c_i) - F(c_{i-1})$	$n_i^* = n \cdot p_i$	$\frac{(n_i - n_i^*)^2}{n_i^*}$
1	0.1	4	−2.058	0.0197	0.0197	3	0.33
2	0.3	13	−1.517	0.0643	0.0445	6	8.16
3	0.5	18	−0.975	0.1660	0.1017	15	0.60
4	0.7	23	−0.433	0.3336	0.1676	24	0.04
5	0.9	27	0.108	0.5438	0.2102	30	0.30
6	1.1	32	0.650	0.7422	0.1984	29	0.31
7	1.3	17	1.192	0.8686	0.1264	18	0.05
8	1.5	9	1.734	0.9582	0.0896	13	1.23
9	1.7	2	2.276	0.9884	0.0302	4	1.00

$$\chi^2 = \sum \frac{(n_i - n_i^*)^2}{n_i^*} = 12.02$$

Knowing the regression equation, the intervals of confidence and the probability density function of the shearing parameters the stochastic simulation of the sliding risk can be initiated.

Determination of the Safety Factor by Stochastic Approach

Accepting the shear strength parameters determined by the regression equation, the factor of safety for any circular arc of sliding is calculated by one of the known deterministic methods (Fig. 5).

Since the intervals of confidence show that cohesion and friction coefficient can take lower values, the stability along the critical arc has been checked also by deterministic method using the lowest limit of the confidence intervals for c and tgφ. The outcome $F = 1.142$ is below the stability mark, but says nothing about the level of the sliding risk. If the sliding risk is extremely low the practical reliability on the designed slope should be not eliminated.

Therefore it is necessary to evaluate the safety margin or the sliding risk by stochastic simulation.

If cohesion and friction coefficient through the homogeneous slope are random variables with normal distribution, the derived factor of safety *(F)* must be also random variable with normal distribution given as :

$$F = S(c,\ \mathrm{tg}\ \varphi) \tag{6}$$

where S is the set of the known deterministic equations by which the safety factor is calculated. If the probability distribution of the input variables $P\ (c,\ \mathrm{tg}\varphi)$ is known, the probability distribution $P\ (F)$ of the safety factor can be determined [3].

For each circular failure arc accepted the random variables c and tgφ are sampled from their experimental distribution and transferred to a random sampling of the safety factor, by one of the deterministic functions of the sliding analysis.

If the sampling number of the input variables is large enough the outcomes will be also large enough to generate a histogram of the calculated factors of safety (Fig. 6).

For a statistical description of the histogram of F values the mean standard deviation, skewness, and kurtosis of determined factors of safety are calculated. Standard deviation evaluates the dispersion of the individual values from the mean, whereas skewness is a measure of the symmetry about the mean, and kurtosis is a measure of peakedness. Skewness is zero or near zero and kurtosis tends toward 3.0 if the calculated F values follow a symmetric normal distribution.

The probability of sliding is estimated form the F calculated distribution as the probability of the factor of safety falling below the accepted limiting equilibrium condition Fc.

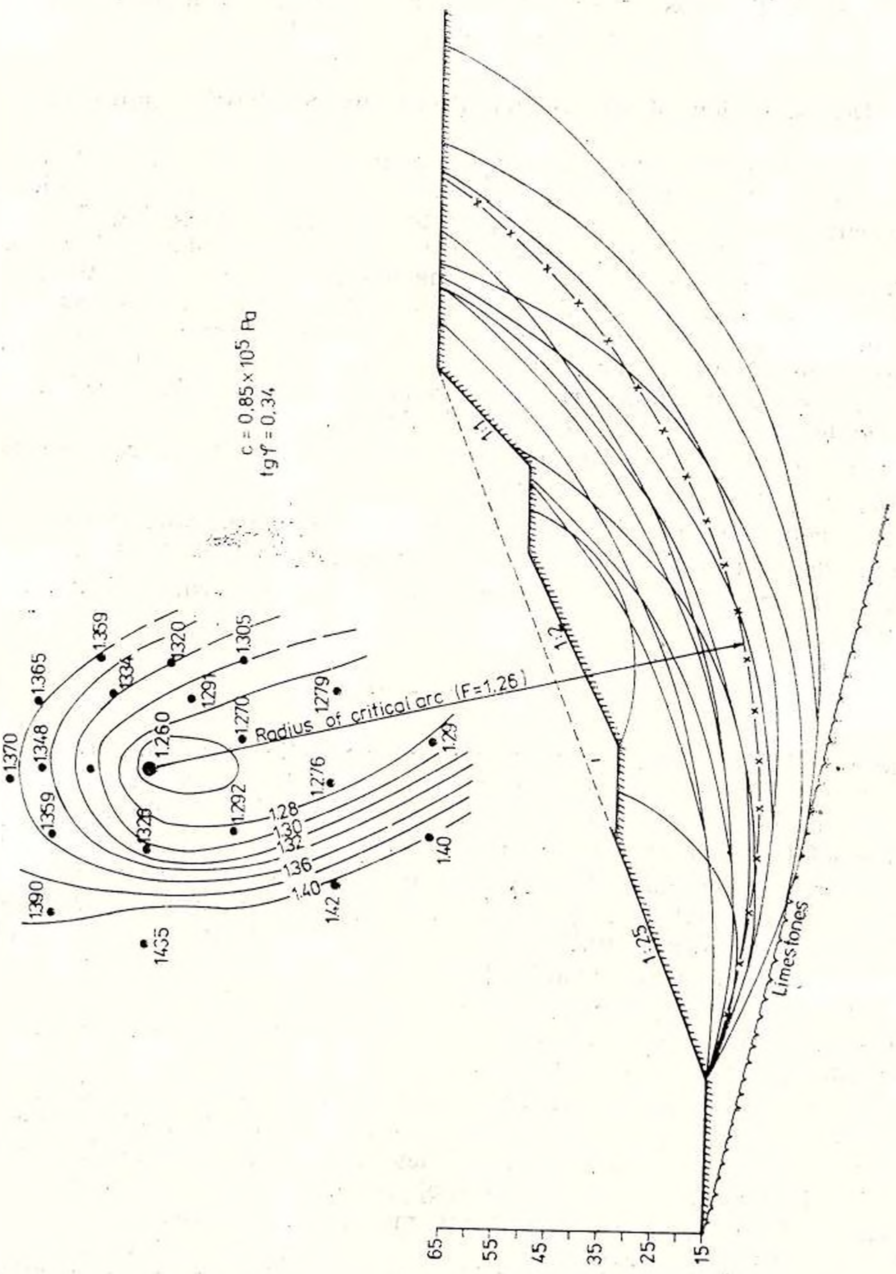
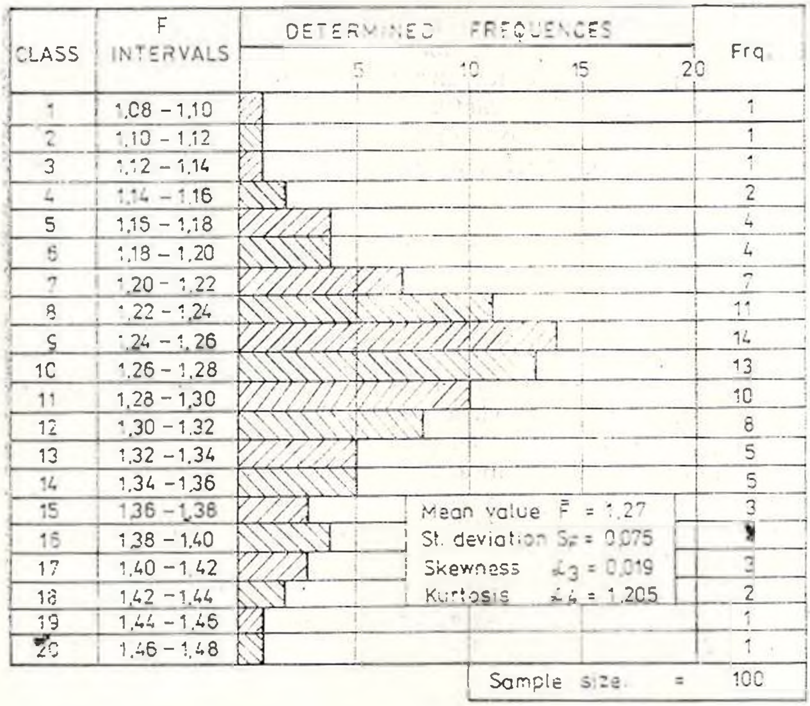


Fig. 5 — Tested circular arcs and stability solution by deterministic methods.

If the normal distribution of the F values is established by chi-square method, the probability of sliding can be determined by normal distribution function

$$P(F)=\frac{1}{\sqrt{2\pi}\,s_F}\int_{-\infty}^{F_c} e^{-\frac{(F_c-\bar{F})}{2s_F^2}}\cdot dF \tag{7}$$



CLASS	F INTERVALS	Frq
1	1,08 – 1,10	1
2	1,10 – 1,12	1
3	1,12 – 1,14	1
4	1,14 – 1,16	2
5	1,16 – 1,18	4
6	1,18 – 1,20	4
7	1,20 – 1,22	7
8	1,22 – 1,24	11
9	1,24 – 1,26	14
10	1,26 – 1,28	13
11	1,28 – 1,30	10
12	1,30 – 1,32	8
13	1,32 – 1,34	5
14	1,34 – 1,36	5
15	1,36 – 1,38	3
16	1,38 – 1,40	[illegible]
17	1,40 – 1,42	3
18	1,42 – 1,44	2
19	1,44 – 1,46	1
20	1,46 – 1,48	1

Mean value $\bar{F}$ = 1,27
St. deviation S_F = 0,075
Skewness α_3 = 0,019
Kurtosis α_4 = 1,205

Sample size = 100

Fig. 6 — Ideogram of the calculated factors of safety for a given failure arc.

which evens out the irregularities of the distribution generated by the random sampling techniques. The integral can be evaluated by polynominal approximation, getting, for the analysed slope, a probability of failure $P = 0.086$, or a sliding risk of 8.6% which is rather high.

A more confortable method determines the probability of failure by the ratio between the number of the factor of safety values below the limiting equilibrium condition and the total sample size.

In accordance with the results presented in Figure 6 the probability of failure is $P = 0.13$ pointing out a higher sliding risk.

The difference in these two estimates of the probability of failure indicates either a slight deviation of the F distribution from the normal distribution or the inacceptancy of some input values.

Conclusion

The evaluation of the slope stability by deterministic methods, using mean values or regression values for experimental shearing parameters, is statistically unjustified in a lithologic complex with variable cohesion and friction coefficient. Since the factor of safety can be ascertained only with some degrees of uncertainty it cannot be definitely stated that the analysed slope will be either stable or unstable.

The stochastic approach by a probabilistic simulation of the distribution of the safety factors on each potential failure circle allows a justified evaluation of the probability of failure, fitted to the uncertainties of the experimental input variables. The large number of the calculations of the safety factors are confortably solved by available computer programmes for any slope geometry and structural conditions testing as many sliding circles as necessary to find out the most critical slip surface.

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Géotechnique—Hydrogéologie

ASPECTS SUR L'ANALYSE SISMIQUE DES BARRAGES (ROUMANIE)

PAR

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Introduction

La Roumanie située au SE de l'Europe est traversée par les Carpathes qui sont connues comme une unité géologique sismique active. La zone de Vrancea, la plus active des zones sismiques de Roumanie, est située à la courbure des Carpathes Orientales. Les tremblements de terre de profondeur intermédiaire de cette zone ont eu des magnitudes montant jusqu'aux valeurs de 7,5 (fig. 1).

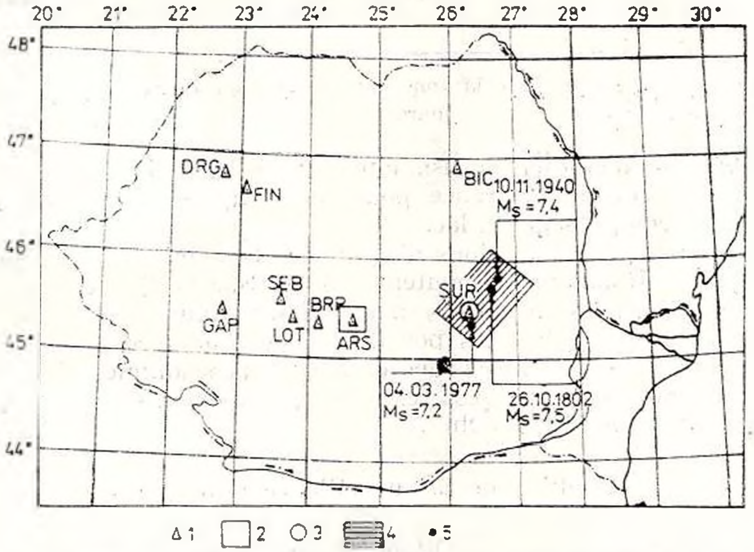

Fig. 1 — Equipements sismiques installés aux barrages de la Roumanie. 1, équipements à enregistrement direct ; 2, réseau télémétrique ; 3, équipement digital ; 4, zone sismique de Vrancea h = i ; 5, séismes à M < 7.

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La surveillance sismique des barrages ou bien des emplacements de prochains barrages a comme but le mesurage de la sismicité locale naturelle et induite aussi bien que l'analyse et l'interprétation des données. Elle apporte également de nouvelles contributions concernant l'étude de la sismicité régionale. Cette surveillance avant le remplissage des lacs en vue d'étudier la sismicité induite offre la possibilité d'obtenir des données de référence pour les comparer à d'autre données ultérieures au remplissage du lac.

TABLEAU 1

Equipements sismiques installés aux barrages et caractéristiques des accumulations

Barrage	Haut. du barrage	Volume de l'accumulation (10^6 mc)	Année de la construction	Roche du fondement	I_B	Date de l'installation de l'équipement
Izvorul Muntelui	127	1230	1961	grès quartzitique	6	04.11.1974
Vidra	121	340	1973	grès	6	07.11.1974
Vidraru	167	465	1965	grès granitique	7	04.04.1975
Fintinele	92	225	1978	schistes cristallins	6	20.05.1978
Oașa	90	132	1982	grès	6	22.05.1978
Surduc	192	155	1985	grès	7–8	18.09.1978
Gura Apelor	168	225	1985	schistes cristallins	6	01.01.1979
Brădișor	62	39	1983	scristes cristallins	6–7	03.03.1983

Note : I_B = intensité dans la zone du barrage en degré MSK-64 d'après la zonation microsismique du pays, standard.

Les premières installations d'équipements sismiques dans les zones de barrages de Roumanie remontent à 1974. Sur la figure 1 sont illustrées les équipements sismiques pour la surveillance des barrages et dans le tableau 1 les données portant sur les barrages surveillés. Pour cette surveillance sont utilisés des équipements sismiques fixes et portables, équipements télémétrés, réseaux d'équipements télémétrés, systèmes digitaux et accélérographes.

Résultats de la surveillance sismique

Le barrage d'Izvorul Muntelui est logé dans le flysch des Carpathes Orientales dans une zone où antérieurement au remplissage du lac on n'a pas signalé d'activité sismique. 15 ans après le remplissage du lac durant la surveillance sismique, trois séismes de magnitudes entre 3,9 et 4,5 se sont produits. Ils ont eu des distances épicentrales par rapport à l'accumulation allant de 27 km à 41 km, sur un alignement orienté NS, correspondant à la direction des structures tectoniques majeures. Il est possible de corréler cette activité sismique avec la présence de l'accumulation dans cette zone.

Le barrage de Vidra est situé dans une zone sismique stable suivant les données sismo-tectoniques connues. Au cours de la période de surveillance sismique on a mesuré une activité sismique locale relativement réduite ; la corrélation entre le niveau de l'eau du lac et le nombre de chocs locaux a été insignifiante pareillement au barrage d'Izvorul Muntelui (tabl. 2).

TABLEAU 2

Sismicité locale ($D < 40$ kP) dans les zones de quelques barrages en activité

Barrage	Durée de l'observ. (ans)	$\overline{N}$/an	M_L max	$r_{(N,H)}$	b
Izvorul Muntelui	8,5	3,4	3,6	+0,10±0,10	0,71
Vidra	8,5	36	2,5	−0,14±0,10	0,88
Vidraru	8	143	3,3	−0,36±0,09	0,97
Fintinele	2	3	1,7	−0,27±0,17	—
Oașa	1	—	—	—	—

— D — distance hypocentrale ;

— N/an — nombre moyen de chocs locaux par an ;

— M_L — magnitude locale ;

— $r_{(N,H)}$ — coefficient de corrélation linéaire entre le niveau de l'eau du lac (H) et le nombre de chocs locaux (N) ;

— b — pente de la droite correspondante à la relation fréquence cumulative-magnitude.

L'accumulation de Vidraru est localisée dans les Carpathes Méridonales dont la tectogenèse alpine est bien connue. Le barrage a été construit sur des roches cristallines. En amont, dans la zone du lac il y a un contact de faille abrupte entre les formations cristallines et celles sédimentaires. La faille marquant le contact cristallin/sédimentaire (formée à la suite du soulèvement continu du cristallin) est censée comme active.

Des données sur la sismicité observée il en résulte la production, dans la zone, du tremblement de terre du 26 janvier 1976 à $M_s = 6{,}4$ et $I_0 = 8$.

Pendant la période de surveillance sismique on a mesuré l'activité locale représentée par un nombre assez grand de séismes, mais d'un niveau modéré des magnitudes maxima produites. On a constaté une corrélation inverse relativement intense et significative entre le nombre des chocs locaux et le niveau de l'eau du lac (tabl. 2), (fig. 2).

L'analyse de la distribution épicentrale dénote une augmentation de la densité épicentrale dans le voisinage du lac. Une diminuation de la densité épicentrale (fig. 3, 4, tabl. 3) a lieu durant le remplissage

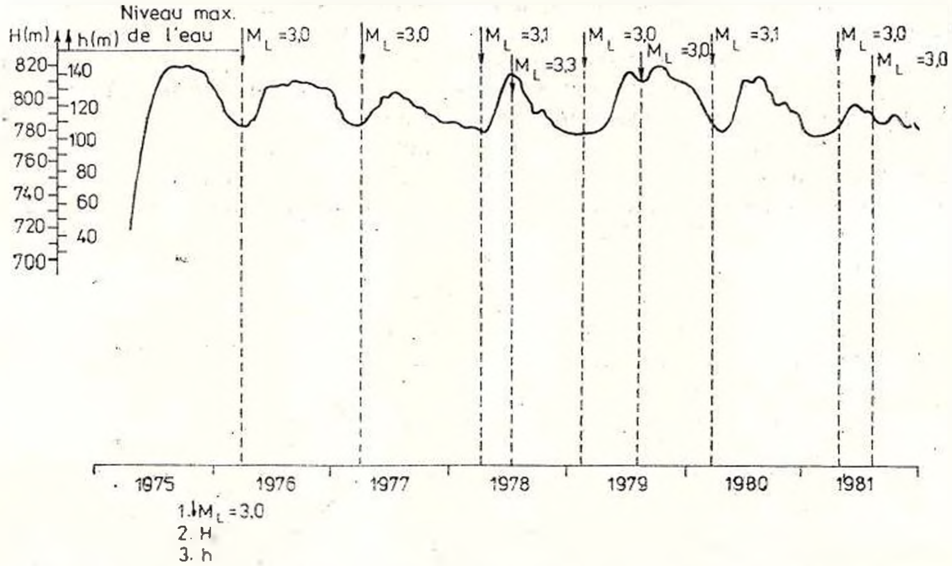

Fig. 2 — Barrage de Vidraru. Niveau de l'eau et tremblements de terre locaux. 1, M_L = 3,0 — magnitude locale pour séismes à des distances hypocentrales inférieures à 40 km ; 2, H — niveau de l'eau ; 3, h — hauteur de la colonne d'eau.

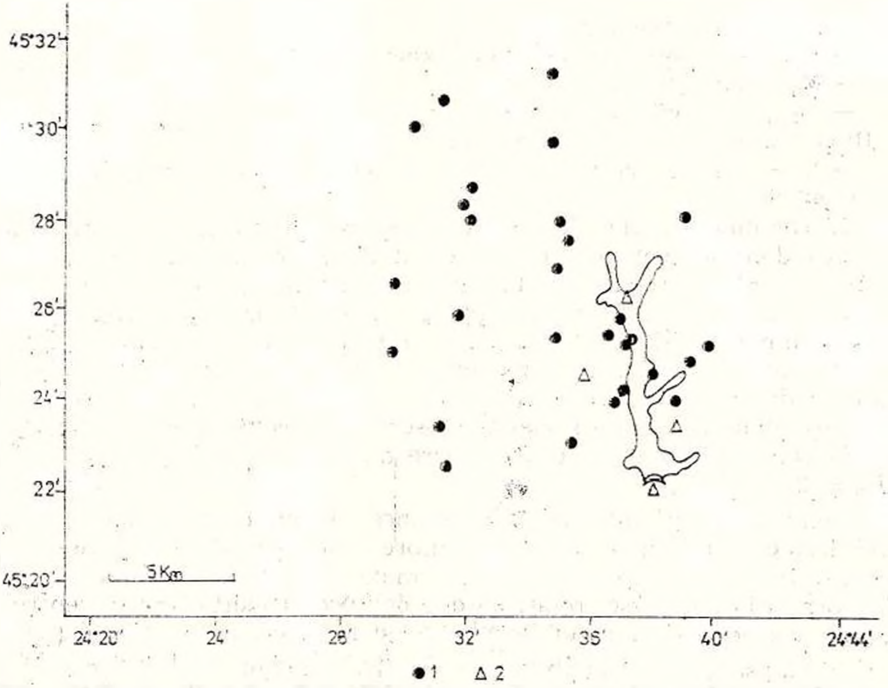

Fig. 3 — Barrage de Vidraru. Période du premier avril 1978 au 31 juillet 1978. Niveau du lac 779 à 809 m.
1, épicentre ; 2, équipement sismique.

et le maintien du lac à des niveaux maxima aussi bien que dans le voisinage du lac.

La distribution des profondeurs focales (fig. 5) montre que le remplissage du lac a comme effet la supression des chocs locaux à de petites profondeurs entre 0 et 8 km.

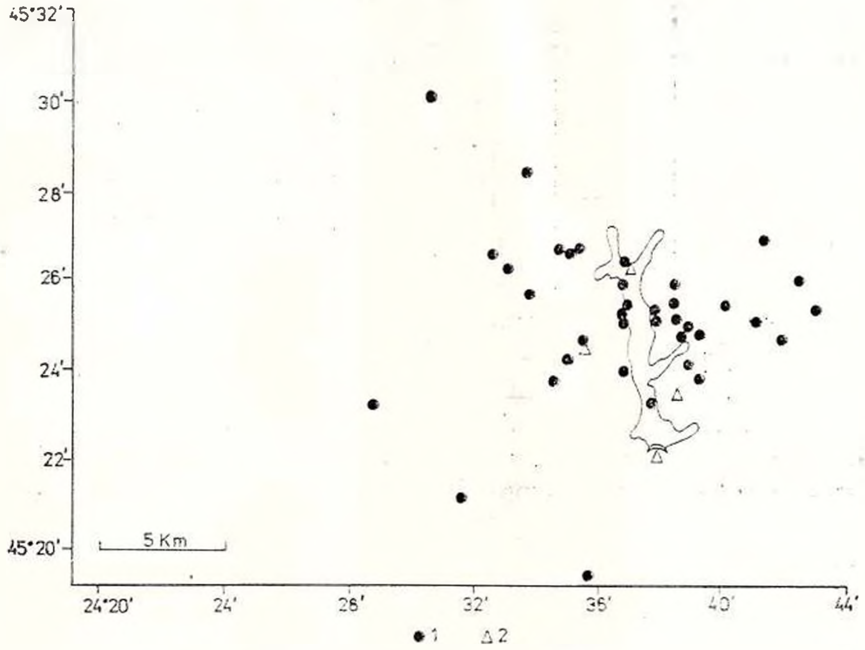

Fig. 4 — Barrage de Vidraru. Période du premier août 1978 au 30 novembre 1978. Niveau du lac 808 à 780 m.
1, épicentre ; 2, équipement sismique.

TABLEAU 3

Accumulation de Vidraru; densité épicentrale (chocs an/km²) entre 01.04.1978 et 30.11.1978

Surface Cycle	Surface du lac + 2 km à partir du bord du lac	Surface entre 2 et 8 km à partir du bord du lac
Remplissage	0,53 1,06	0,18 0,17

Les solutions des mécanismes en foyer (fig. 6) relèvent que la plupart des séismes se sont produits sur des failles de décrochements, le reste sur des failles normales, les deux types de rupture étant en accord avec le stress principal minimum orienté généralement NS et le stress principal maximum orienté généralement EO ou vertical.

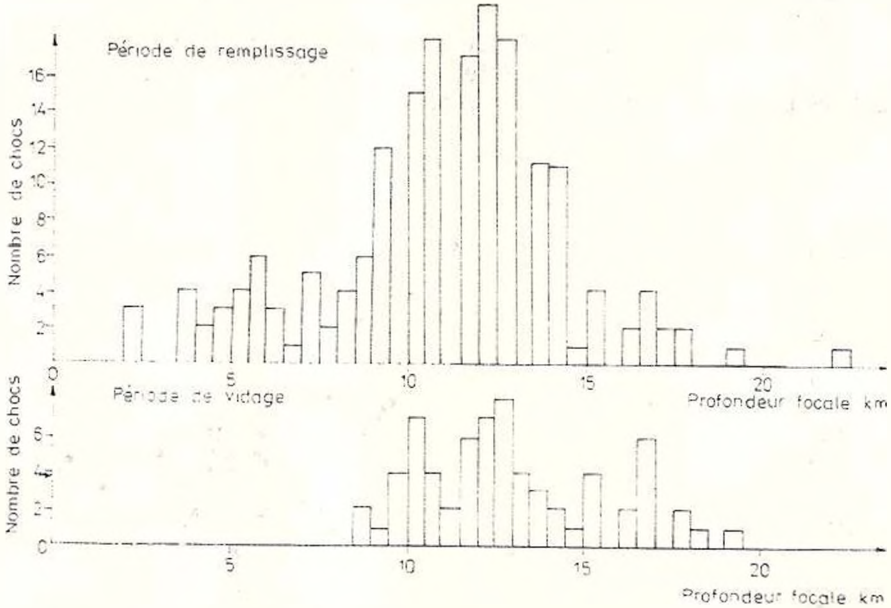

Fig. 5 — Barrage de Vidraru. Distribution des profondeurs des foyers. Période 1977—1981.

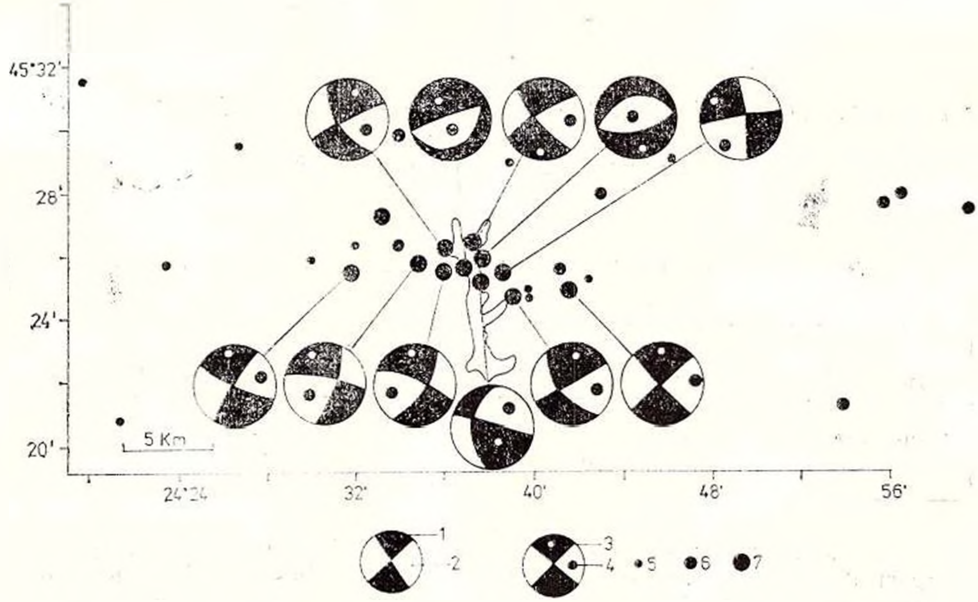

Fig. 6 — Barrage de Vidraru. Période du 5 avril 1982 au 22 mai 1982. Solutions de plan de faille.

L'orientation verticale du stress principal maximum correspond au soulèvement du cristallin et à la genèse de la faille au contact avec le sédimentaire.

On constate la production assez systématique des chocs locaux de magnitudes au-dessus de 3, pendant les périodes à niveau minimum de l'eau du lac, la plupart des magnitudes maxima apparaissant à de courts intervalles lorsque l'eau baisse au-dessous du niveau seuil de 744 m (fig. 2).

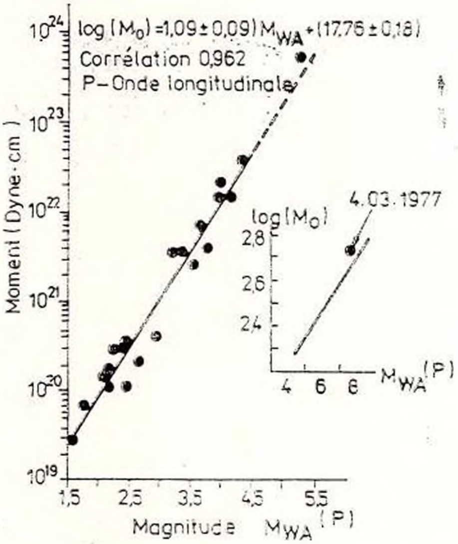

Fig. 7 — Tremblements de terre intermédiaires de la région de Vrancea.

L'augmentation significative de la densité des épicentres dans le voisinage du lac suggère qu'une partie de la séismicité locale est induite. La corrélation inverse entre le niveau de l'eau et l'activité sismique a été expliquée dans bien des études portant sur la production des séismes sur des failles inverses (par exemple, Snow, 1972). Mais autant les solutions de mécanismes en foyer que les données géologo-tectoniques infirment la présence des failles inverses dans la zone de l'accumulation de Vidraru.

Les barrages de Fîntînele et d'Oaşa sont situés dans les zones à activité sismique très réduite, conformément aux données sismo-tectoniques connues.

Après une surveillance sismique de presque 2 ans au barrage de Fîntînele on a mis en évidence un nombre bien réduit de chocs locaux de magnitudes très petites (tabl. 2).

Dans le barrage d'Oaşa, en cours d'une année de surveillance sismique on n'a pas enregistré d'activité sismique locale.

Le début de la surveillance sismique du barrage de Brădişor a coïncidé avec le remplissage du lac ; la période de surveillance est trop courte pour obtenir des données significatives.

La surveillance sismique des barrages de Surduc, Gura Apelor et Drăgan a commencé avant le remplissage des lacs en vue d'obtenir des détails concernant la séismicité locale de la zone.

Les emplacements de Surduc et Siriu sont situés tout près de la zone des séismes à profondeur intermédiaire de Vrancea (fig. 1).

Les données fournies par le système digital de la zone de Surduc ont permis de déterminer la relation moment sismique magnitude pour magnitudes locales Richter M_{WA} = 1,5-5,5 (fig. 7). Les données du tremblement de terre du 4 mars 1977 confirment la valabilité de l'extrapolation de la relation à des grandes magnitudes de plus de 7.

Activité sismique aux barrages après le tremblement de terre du 4 mars 1977 de Vrancea

Dans la période aussitôt suivante au tremblement de terre du 4 mars 1977 on a observé une augmentation générale de l'activité sismique pour toute la zone des Carpathes roumaines que dans les zones limitrophes. Les données obtenues ont confirmé des augmentations de l'activité sismique dans les zones des barrages d'Izvorul Muntelui, Vidra, Vidraru. De ce fait, entre le 4 mars 1977 et le 4 juin 1977 on a enregistré à ces barrages les suivantes augmentations en comparaison avec la période antérieure du 4 décembre 1976 au 4 mars 1977 :

— rapport de croissance du nombre de chocs de 2 à 17 ;
— croissance de magnitudes moyennes de 0,1 à 0,6.

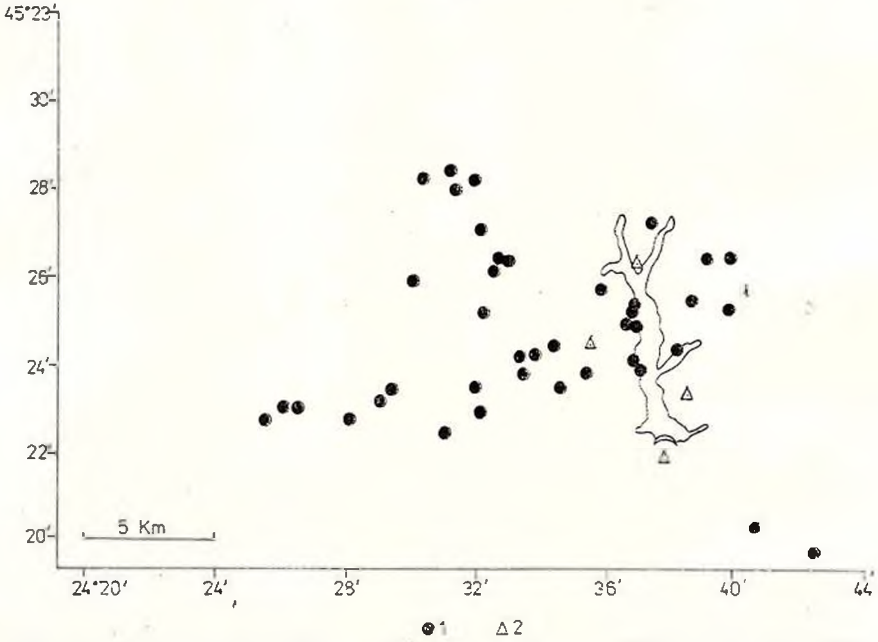

Fig. 8 — Barrage de Vidraru. Période du 4 mars 1977 au 3 juin 1977.
1, épicentre ; 2, équipement sismique.

Entre le 4 mars 1977 et le 4 juin 1977, période d'activation sismique, se sont produites des modifications de la pente b entre 8 et 32$^{0}/_{0}$, par rapport à la période antérieure. Ces modifications significatives de la pente b dénotent des modifications des processus d'élibération des tensions élastiques dans les zones focales ou bien l'apparition des zones focales nouvelles à d'autres caractéristiques. Il en est de même pour la période d'activation du 4 mars 1977 au 4 juin 1977, lorsqu'on a observé des modifications significatives des distributions des distances hypocentrales ainsi que des distributions épicentrales. Ainsi, dans la zone d'Argeş la distribution des épicentres entre le 4 mars 1977 et le 4 juin 1977 présente une tendance de disposition le long d'un alignement EO, tendance plus aiguë qu'en d'autres périodes (fig. 8 en comparaison avec les figures 3 et 4). La solution composée du plan de faille pour les séismes de la période d'activation (fig. 9) correspond à la faille qui sépare le cristallin du sédimentaire, en représentant la structure tectonique principale de la zone.

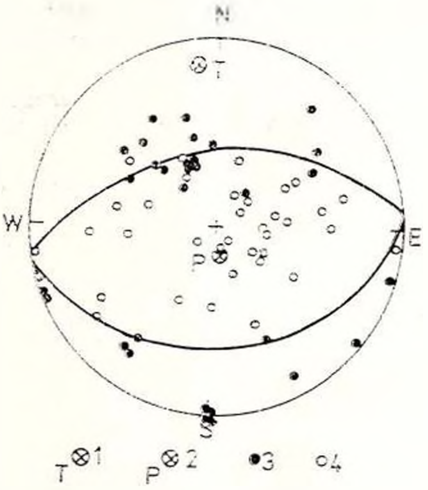

Fig. 9 — Barrage de Vidraru. Solution composée du plan de faille pour 22 tremblements de terre entre le 1er janvier 1977 et le 4 juin 1977.

1, axe de la compression ; 2, axe de la tension ; 3, compression ; 4. dilatation.

Poursuite du comportement des constructions hydroénergétiques pendant les séismes

Aux équipements sismiques installés aux barrages on enregistre quotidiennement et on calcule des paramètres des séismes (magnitude, distance hypocentrale, atténuation) afin de signaler à temps la production des séismes importants.

On observe également le déclenchement des accélérographes. On signale surtout les séismes locaux de magnitudes au-dessus de 4 ou bien l'apparition, dans la zone des barrages, des intensités sismiques de plus de 5 générées par des séismes locaux ou régionaux. Dans ce cas, on effectue un renforcement et un surcroît des mesures et des observations sur l'ensemble terrain/construction.

A la suite de l'analyse des données concernant le comportement des constructions hydrotechniques au cours du séisme du 4 mars 1977 il s'ensuit que :

— le contrôle par observations directes n'a pas mis en évidence des situations anormales ou des phénomènes montrant un comportement anormal des constructions ;

— les enregistrements et les mesurages effectués avec des appareils de mesure et contrôle dans le corp et la fondation des barrages après le tremblement de terre n'ont pas présenté des variations ou bien des modifications essentielles envers les mesurages d'avant tremblement de terre ;

— on a signalé au barrage d'Izvorul Muntelui des infiltrations qui ont augmenté après le séisme, mais ne dépassant pas les limites des valeurs mesurées antérieurement et qui ont revenu à normal (en 10 jours). Les débits mesurés sont les suivants :

Date	Débit des infiltrations en litres/minute
02.03.1977	50—55
04.03.1977:	tremblement de terre produit en Vrancea, dans la zone du barrage I 6 degrés MSK 64
05.03.1977	103
07.03.1977	75
14.03.1977	53

— des 1 750 km de digues situés en des zones intensément affectées par le séisme comme Moldavie, Munténie, Olténie 25 km de digues ont subi des fissures, dislocations de talus, tassements.

Conclusions

La surveillance sismique des barrages permet une meilleure connaissance de la séismicité locale et des relations séismicité/aspects tectoniques géologiques de la zone aussi bien que de l'activité sismique antérieure.

Par suite du grand nombre des données peuvent être réalisées des réconsidérations du risque sismique dans les zones des barrages.

La détermination opérative des paramètres des tremblements de terre capables de produire des effets significatifs pour l'étude du comportement au cours des séismes offre la possibilité de prendre des décisions immédiates en vue d'intensifier les mesures et les observations de l'ensemble construction/terrain.

La poursuite de la relation accumulation/séismicité de la zone complète les données nécessaires pour la détermination du risque sismique dans les zones des barrages.

MODELS FOR INTERPRETING SOME HYDROGEOLOGICAL AND THERMIC CHARACTERISTICS OF GEOTHERMAL STRUCTURES SITUATED IN THE EAST OF THE PANNONIAN DEPRESSION

BY

ALEXANDRU GHEORGHE [1], PETRE CRĂCIUN [2]

Introduction

As part of the investigation programme of regions including geothermal structures from Romania, the eastern area of the Pannonian Depression is a major object of study by drilling in view of outlining and revaluating the thermal water accumulations.

By means of several boreholes for hydrocarbons or thermal waters, geophysical investigations have been carried out in order to assess the fluid content of different geological formations ; thermic measurements have also been performed, most of them recording maximal temperatures at the bottom of the borehole.

The investigation by means of drilling pointed out the presence of two high hydrothermal potential sectors, namely the Săcuieni-Marghita-Tăşnad zone situated north of the Crişul Repede River and the Sînnicolau Mare-Tomnatec-Şemlac zone situated south of the Mureş River (Fig. 1).

The relatively insufficient primary data on the hydrodynamic characteristics of aquifers in drilling areas make difficult the analysis of geothermal structures. One of the problems raised by hydrogeological interpretation regards the assessment of potentiometric surface within the thermal aquifer zone. The present paper is an attempt at estimating the reservoir pressure based on well head records, for each borehole. The convective circulation regime on the whole of investigated geothermal structures is also in our concern.

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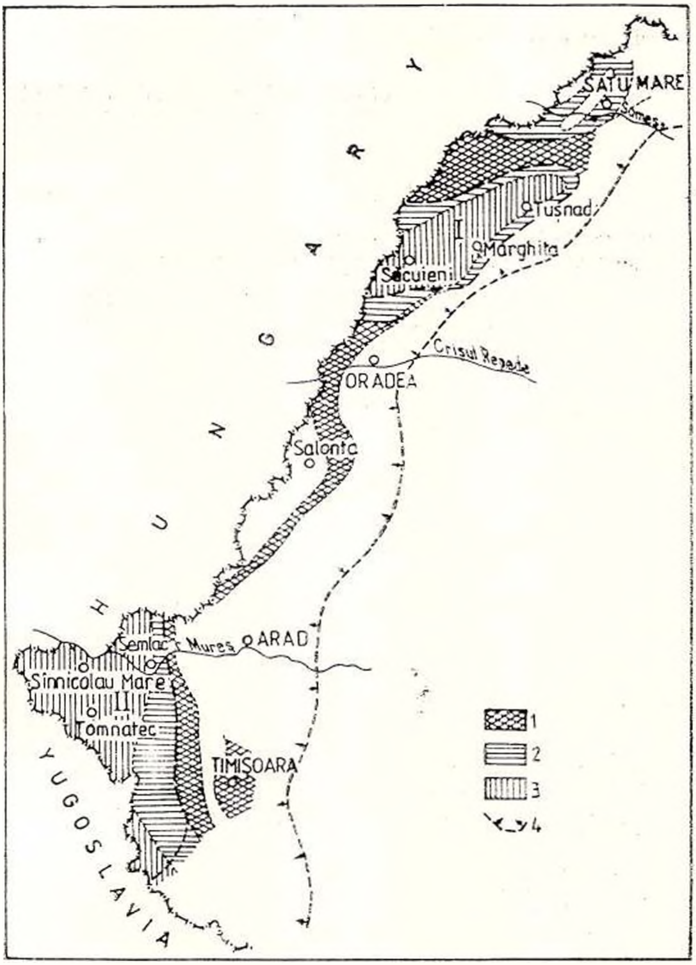

Fig. 1 — Thermal zonation of the aquifer system in Upper Pannonian formations, East Pannonian Depression. 1, temperature of 50-60°; 2, temperature of 60-80°; 3, temperature of 80-100° ; 4, eastern border of the Pannonian Depression ; I, Secuieni-Marghita-Tăşnad zone ; II, Sînnicolau Mare-Tomnatec-Şemlac zone.

Geological and Structural Setting

The area of the Pannonian Depression situated on Romanian territory consists of a sequence of post-Cretaceous deposits which overlie a Mesozoic sedimentary basement in the region of Oradea and Sînnicolau Mare and a metamorphic basement north-east of Oradea and east of the Sînnicolau Mare structure (Ali-Mehmed et al., 1978 ; Visarion et al., 1979). The Pannonian deposits which include the main aquifer of this region are the most widespread, mainly in the western sunk areas.

The Pannonian formation includes two lithological series : a Lower Pannonian marly one and an Upper Pannonian (= Pontian) sandy complex. The sandy series is the main reservoir of thermal waters.

According to the known structural feature (Ali-Mehmed et al., 1978) the blocks with sunk basement situated north of the Crişul Repede

River include two main depressions. Thus, to the north of the Săcuieni-Marghita structure (Fig. 2) lies the Galoșpetreu-Mecențiu Depression, bordered northward by the main Dragoș Vodă Fault on an area of 800 sq km. East of the Săcuieni area lies the Sînnicolau de Munte Graben.

South of the Mureș River (Fig. 1) the Neogene cover becomes thicker from east to west and the maximum subsidence zone is developed along the Sînnicolau Mare and Beba Veche hemisynclines.

Geothermal and Hydrogeological Characteristics of the Pannonian Geothermal Reservoir

Information on the intensity of the geothermal field in the investigated area was almost entirely provided by temperature records in boreholes for hydrocarbon or thermal water. The temperature was measured as punctiform maximal records at the bottom of the borehole and only in a few instances as differential curves. The areal distribution of temperature points measured at depth is relatively ununiform. The thermal measurements were performed at different depths in different boreholes.

The geothermal gradients show higher values as compared to the normal ones, in the Tășnad, Săcuieni, Sînnicolau de Munte sectors (5-6.5°C/100 m) and lower values in the other sectors (Ghenea et al., 1980).

In the southern sector, the vertical gradients are smaller and the frequency of values is of 4-5°C/100 m. The maps of the thermal field at the base of the thermal aquifer complex (Figs. 2, 3) show that the geothermal structures are characterized by some peculiar features. Thus, it is to note the presence of maximum thermal fields of 80-100°C in the Sînnicolau de Munte, Cherechiu, Galoșpetreu and Mecențiu depressions. On the eastward elevated flanks of the depression basement, the temperature decreases gradually, at the same level, from 80° to 50°C.

A similar feature is noted in the region situated south of the Mureș River (Fig. 3). Here, the maximum thermal field, at the base of the Pontian aquifer, reaches the values of 90-95°C in the Sînnicolau Mare-Tomnatec depressions. Eastward, the temperature decreases gradually to less than 50°C.

The aquifers which form the Pontian hydrothermal system are represented by sandy horizons, each 2-30 m thick or even thicker in places ; they alternate with sandy clays. The thermal complex consists of ten aquifers in the south-east of the Săcuieni area, fifteen in the west and about eighteen in the north. South of the Mureș River occur 7-12 aquifers, most of them below the depth of 1500 m.

As regards the petrophysical characteristics of reservoir formations, there are relatively few available data as compared to the dimensions of the hydrothermal system. According to the measurements performed in some geothermal boreholes, the transmissivity and permeability of thermal aquifers have been estimated. The processing of tests assumed an isothermal regime. The Table is an illustration of some of these results.

The Pontian sands exhibit moderate permeabilities and at the same time there are significant variations from one zone to another.

The processing of some geophysical diagraphs from the boreholes at Săcuieni has shown that the porosity of Pontian sands is generally between 14-25%, characteristic of fine-medium grained sands.

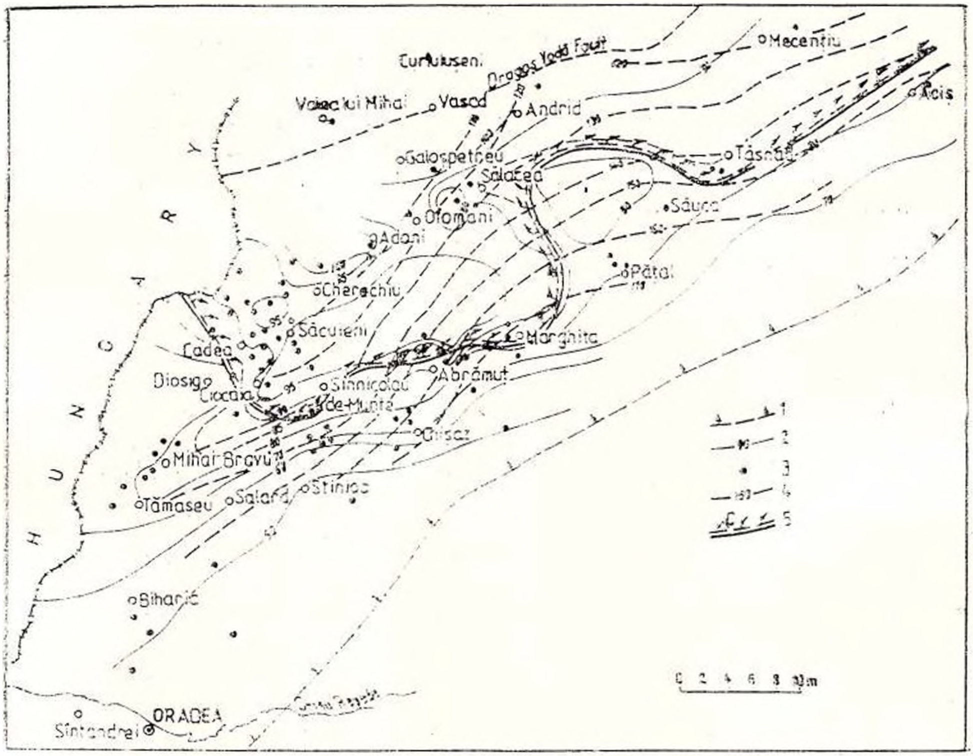

Fig. 2 — Piezometric surface and temperature distribution map of the Pannonian hydrothermal system, Săcuieni-Marghita-Tășnad Basin.

1, eastern border of the Pannonian Depression ; 2, isogeotherm of 90°C at the bottom of the thermal aquifer ; 3, well with recorded temperature ; 4, contour showing equivalent piezometric level above sea level ; 5, contour of the region where the Rayleigh criterion indicates the likelihood of mixed convection (on side "C").

In order to determine the regional hydrodynamic features of the geothermal system under natural conditions, it is necessary to study the distribution of reservoir pressure. As far as both the reservoir temperature and the thermal profile of the water column in the borehole differ from one place to another (Fig. 2, 3), the local, physical hydraulic heads should be converted to equivalent heads.

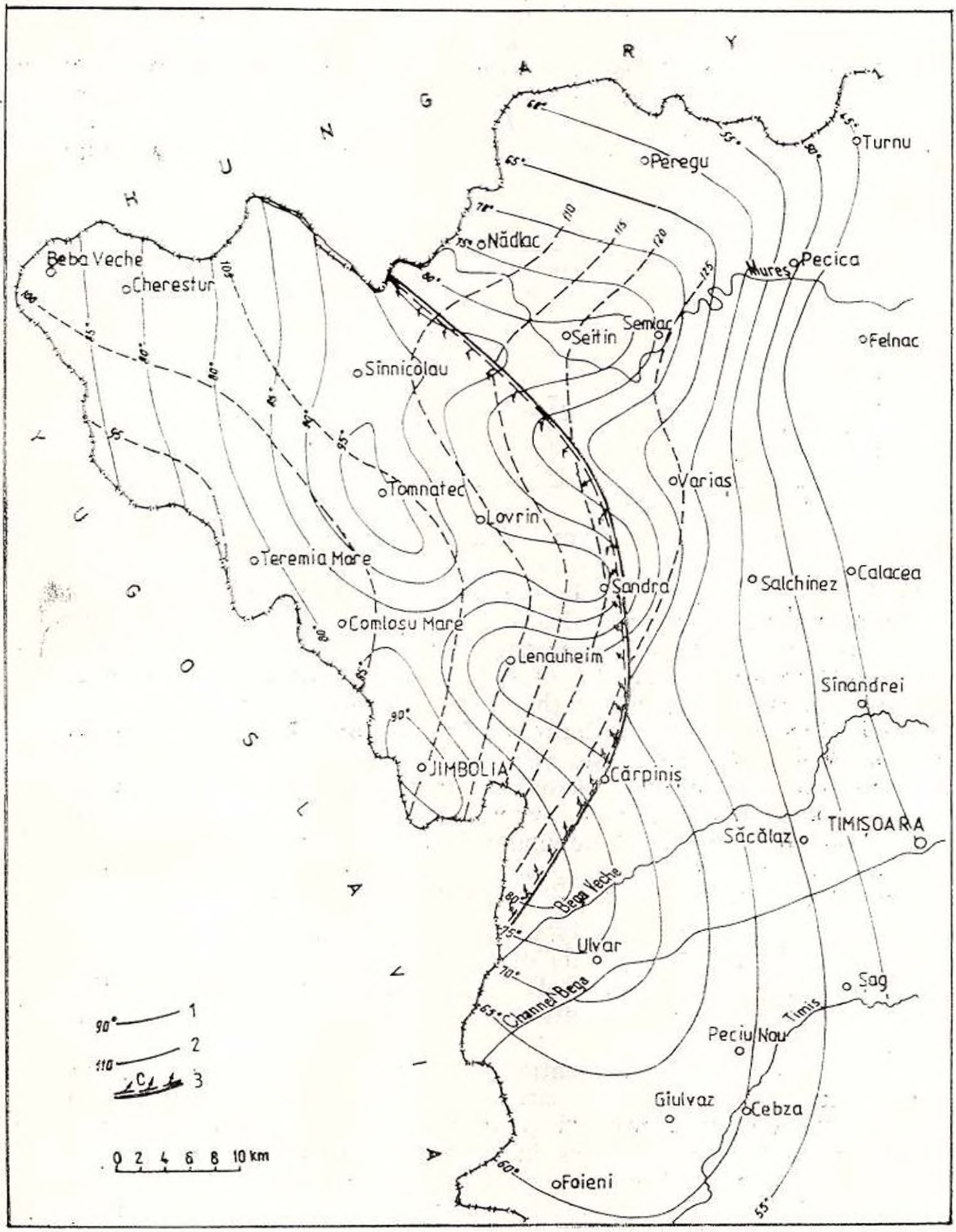

Fig. 3 — Piezometric surface and temperature distribution map of the Pannonian hydrothermal system, Sînnicolau Mare-Tomnatec-Șemlac Basin.

1, isogeotherm of 90° at the bottom of the thermal aquifer ; 2, contour showing equivalent piezometric level above sea level ; 3, contour of the region where the Rayleigh criterion indicates the likelihood of mixed convection (on side "C").

The equivalent or "cold water" heads represent a water column at a constant conventional temperature (e.g. 20°C), the same for the whole region. It exerts at the base a pressure equal to the reservoir pressure at the place under discussion. According to this assumption, it is to note the relation $Z + \frac{Pr}{\gamma 20} = Hp$ in which Z stands for the elevation of the base of thermal aquifer, Pr — reservoir pressure under natural hydrodynamic conditions, γ_{20} — specific weight of water at 20°C and Hp — equivalent head. In order to state the equivalent head of hydrogeological characterization of geothermal systems. Thus, the borehole flowing stopped. For the most boreholes these were not recorded. However, we dispose of some measurements of static pressures at the well head. By using the maps of distribution of temperatures at the base of thermal aquifers (Figs. 2, 3) the reservoir pressures in boreholes with well head records were stated.

TABLE

Locality	Depth of the aquifer	Permeability mD	Transmissivity mDxm
Săcuieni	1514 – 1691	300 – 400	16000 – 21000
Mihai Bravu	803 – 924	62 – 65	2500 – 2600
Adoni	1008 – 1547	100 – 125	9100 – 11250
Marghita	985 – 1376	200 – 208	17400 – 18100
Tăşnad	936 – 1354	140 – 150	12600 – 13500

The automatic calculus of the formation temperature was based on the equation of state of water which relates the fluid density to temperature. Thus, the following relation was used :

$$\rho = \rho_0 \left[1 - \beta_1(T - T_0) - \beta_2(T - T_0)^2\right]$$

in which ϱ_0 — water density at reference temperature (T_0), β_1, β_2 — thermal expansion coefficients of first and second order, T — current temperature. By using the data of the automatic calculus of Hp, for each borehole, the equipotentials were drawn for the two geothermal areas (Figs. 2, 3).

Although the above mentioned image is not considered to be the most accurate one, it points satisfactorily to some useful data in view of hydrogeological characterization of geothermal systems. Thus, the arrangement of contour lines in Figure 2 discloses the occurrence of a source area on the eastern border of the depression and a regional flow trending WNW. The frequency of these lines does not reveal any discontinuity within the determined geothermal areas. As regards the structure situated south of the Mureş River (Fig. 3) the equipotential lines point to a natural flow from east westwards and a source area to the eastern border of the depression.

In view of determining the possibility of generating free convection currents in thermal aquifers, the dimensionless Rayleigh parameter was estimated for several points belonging to the two thermal water sectors.

Rayleigh parameter was calculated according to the relation (Sorey, 1975) :

$$Ra = \frac{\beta g K L \Delta T}{Km/\rho_0 \, C\mu/\rho_0}$$

in which β — thermal expansion coefficient of water, g — gravity acceleration ; K — permeability of the aquifer, ΔT — difference of temperature between the aquifer bed and aquifer top ; Km — mean thermal conductivity of porous rock ; L — aquifer thickness ; ρ_0, C — density and specific caloric power at mean temperature (between those ones of the top and bed) ; μ — dynamic viscosity of water.

As regards the horizontal aquifers with large lateral extension, by taking into account that the physical properties of water depend only on temperature, cellular convection may exhibit Rayleigh *(Ra)* value greater than 70 (Sorey, 1975).

According to *Ra* values estimated, each of the two geothermal areas includes one sector in which free convection is possible. These sectors occur in the western area of the region, within sunk zones of the aquifers (Figs. 2, 3).

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HYDROGEOLOGICAL AND PROBABILISTIC METHODS USED IN THE STUDY OF DEWATERING SYSTEMS FOR COAL DEPOSITS

BY

ALEXANDRU GHEORGHE [1], DANIEL SCRĂDEANU [1]

The optimum exploitation of coal deposits characterized by difficult hydrogeological conditions implies the elaboration of some general methods in view of estimating the efficiency and evolution of dewatering process.

Dewatering is mainly determined by the aquifer and the drainage network. A complete image of dewatering is offered by global evolution of the unitary system made up of the sub-systems : aquifer and drainage network. The complex character of the system is due to the great number of variables during dewatering : natural hydrodynamic conditions (limit conditions, recharge, natural drainage, etc.) and the components of the drainage network (number and location of drainage wells at work, pumped discharge).

By taking these into account it is obvious that the study of dewatering should be based on global parameters, while elementary parameters (hydraulic conductivity, radius of influence of the well, hydraulic diffusivity coefficient, etc.) need schematization, sometimes forced, of natural conditions. This schematization may lead to important deviation from real circumstances brought about by dewatering.

A frequently used global parameter is represented by the speed of reducing the piezometric level, reported from piezometer wells in the mining field. The results obtained allow the qualitative estimation of the efficiency of dewatering.

In the Roşia de Jiu open-pit, from the Rovinari Basin, where lignite is exploited, the aquifer between coal layers V and VII is being dewatered and the artesian aquifer from coal layers IV—V undergoes tension release. The mean drained flows between October 1980 — December 1981 are of 213.2 l/sec for complex V—VII and of 19.4 l/sec for complex IV—V.

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The study of piezometric level variation in piezometer wells (300, 310, 318 for complex V—VII and 105, 106, 108, 111 for complex IV—V) has pointed to the following reduction speeds (Fig. 1, 2) :

— for complex V—VII : $\upsilon\ \mathrm{E}[-4.50;\ +1.80]\dfrac{\mathrm{cm}}{\mathrm{day}}$;

— for complex IV—V : $\upsilon\ \mathrm{E}[-4.50;\ +3.00]\dfrac{\mathrm{cm}}{\mathrm{day}}$.

The same value of the piezometric level at the beginning and the end of the studied interval shows that the drainage network allows the elimination from aquifers of the water supplied by dynamic reserve (surface and recharge).

These qualitative conclusions are the result of processing of routine measuring performed in the drainage network of a mining field undergoing dewatering.

The method of processing the same data (measurements of piezometric level) is proposed and supplementary information on the system aquifer-drainage network is obtained. According to this method one states the quantitative criteria useful for :

— estimating the distribution of different stages of dewatering in the mining field ;

— estimating the stabilization of dewatering ;

— determining the most favourable stage for inferring, on statistic grounds, of a law of the evolution of dewatering.

The method proposed, typical of information theory, implies the calculation of some dynamic aleatory variables which describe accurately the evolution of the system aquifer-drainage network.

These dynamic aleatory variables associated to each piezometer well are of the type :

$$x : \begin{pmatrix} t_1 & t_2 & \ldots\ldots\ldots & t_i, & \ldots\ldots & t_n \\ P_1 & P_2 & \ldots\ldots\ldots & P_i, & \ldots\ldots & P_k \end{pmatrix} \qquad (1)$$

in which :

$t_1,\ t_2, \ldots,\ t_i \ldots .\ t_n$ — stages of system evolution ;

$P_1,\ P_2, \ldots,\ P_i \ldots . P_k$ — probabilities of achieving discreet stages (1, 2, . . . k) of the system.

In order to exemplify this, the dewatering system of pilot-mine Prunişor East in the SE of Mehedinţi district was used.

The pilot-mine cuts the first coal layer (Upper Dacian lignite) the bed of which contains a confined aquifer while the overlier includes an unconfined aquifer. The piezometric level is studied by means of six piezometer wells located in the aquifer from the bed of the coal layer.

To each piezometer associated a dynamic aleatory variable in which :

— time interval ($\triangle\ t$) between moments $t_1, \ldots t_n$ is $t = 10$ days ;

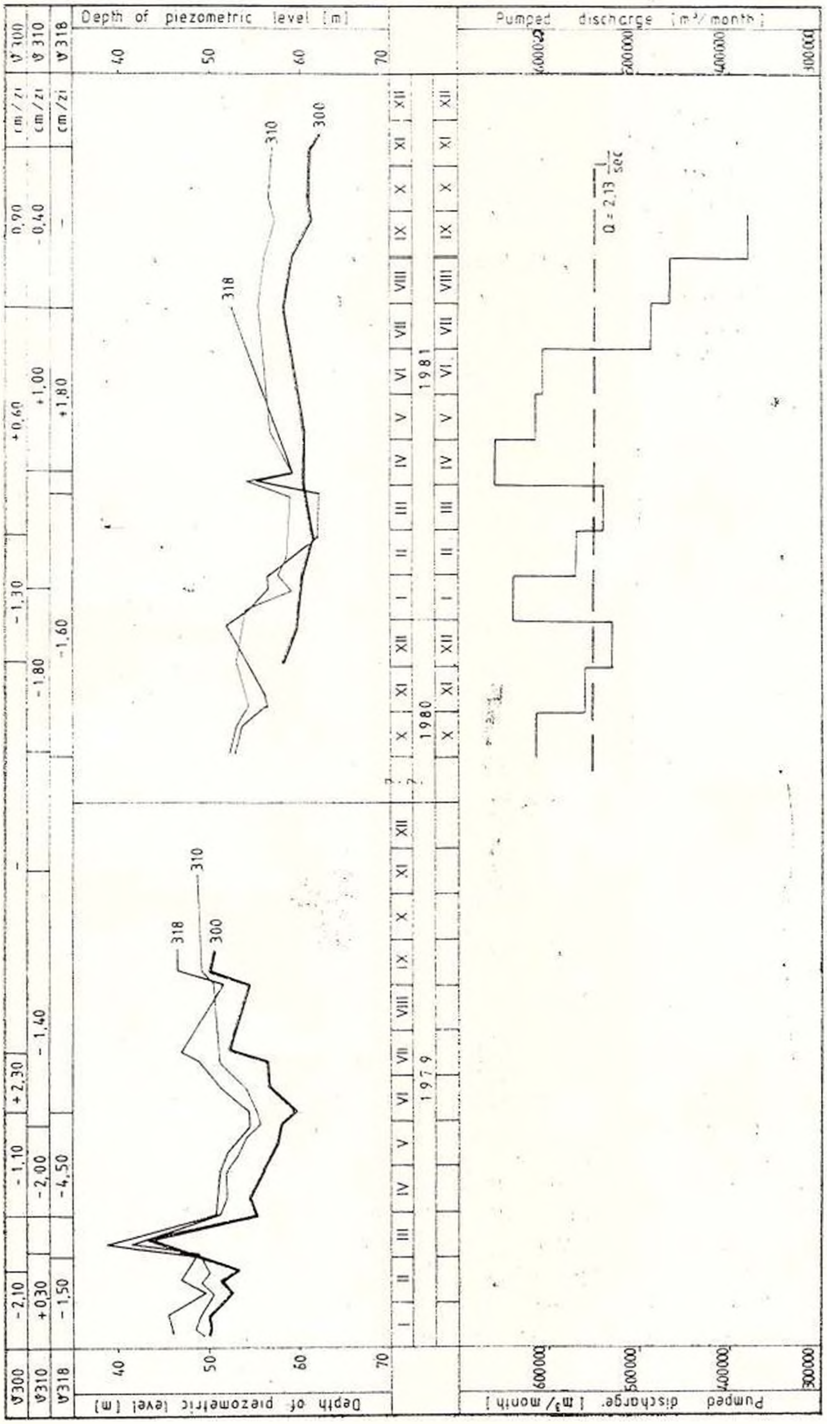

Fig. 1 — Evolution of hydrodynamic elements of dewatering for the aquifer V—VII coal, Roşia de Jiu open-pit.

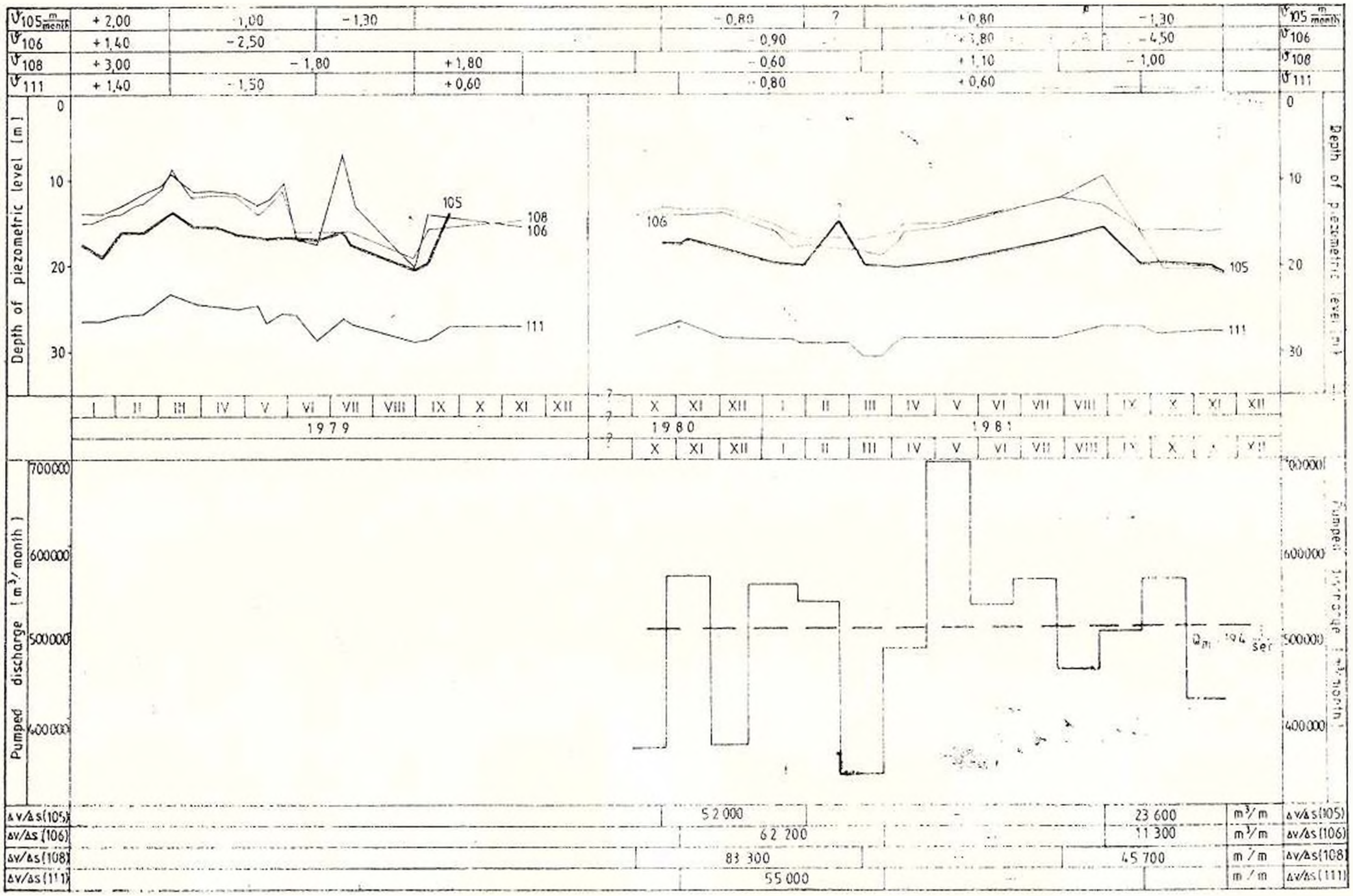

Fig. 2 — Evolution of hydrodynamic elements of drainage for the aquifer situated at the bottom of Roşia de Jiu open-pit (IV—V coal).

— possible states of the system for which $p_1, \ldots p_k$ are calculated are the following :

state 1 : increase of piezometric level ;
state 2 : constancy of piezometric level ;
state 3 : decrease of piezometric level.

All these aleatory variables describe the evolution of the system the state of which at time "t_i" depends on the state at time "t_{i-1}"

This process of stochastic type is called markovian process and acts at discrete time intervals "n" in discrete states. The model of "one step" markovian process adopted is described by means of a transition matrix, of the type :

$$M : \left\| \begin{matrix} P_{11} & P_{1j} & \ldots\ldots\ldots & P_{1n} \\ P_{i1} & P_{ij} & \ldots\ldots\ldots & P_{in} \\ P_{n1} & P_{nj} & \ldots\ldots\ldots & P_{nn} \end{matrix} \right\| \qquad (2)$$

in which P_{ij} — probability of transition from state "i" to state "j".

In the case of piezometer well 2HE, by using piezometric measuring during the interval December 1979 — November 1980 (Fig. 3), the following transition matrix is obtained for the quantification error [—0.10 ; 0.10]

$$M_{2HE} \left\| \begin{matrix} 0.38 & 0 & 0.62 \\ 1.00 & 0 & 0 \\ 0.28 & 0.11 & 0.61 \end{matrix} \right\|$$

Thus the system aquifer-drainage network was turned into a complex of aleatory variables and the markovian model was adopted for the evolution of dewatering. It should be mentioned that the adoption of the markovian model implies the approximation due to the variation in time of causes determining dewatering.

By reduction of the deviation from the markovian model, the capacity of well $\left(q = \frac{Q}{S_0}\right)$ was used and was noted a slight deviation as compared to the results obtained by taking into account the unevenness only (S_0). The unimportant difference is accounted for by global processing of measuring for about one year lapse.

According to these, the hydrodynamic complexity of dewatering is calculated by using the entropy inferred from the relation :

$$H_i = - \sum_{i=1}^{n} P_i \log_2 P_i ; \text{ [bit]} \qquad (3)$$

in which : P_i — probability of achieving state "i" of the system ; H_i — entropy of the process in which a certain state is achieved independently of the other possible ones of the system.

The entropy of each piezometer is calculated according to quantification errors corresponding to the accuracy of measuring (Tab.).

The maximum value of entropy "H_i" is calculated by using the relation :

$$H_{i_{max}} = - \log_2 n \text{ [bit]} \tag{4}$$

and corresponds to the case in which all "n" states of the system are equally possible. Thus, there is maximum incertitude of achieving one of the "n" states.

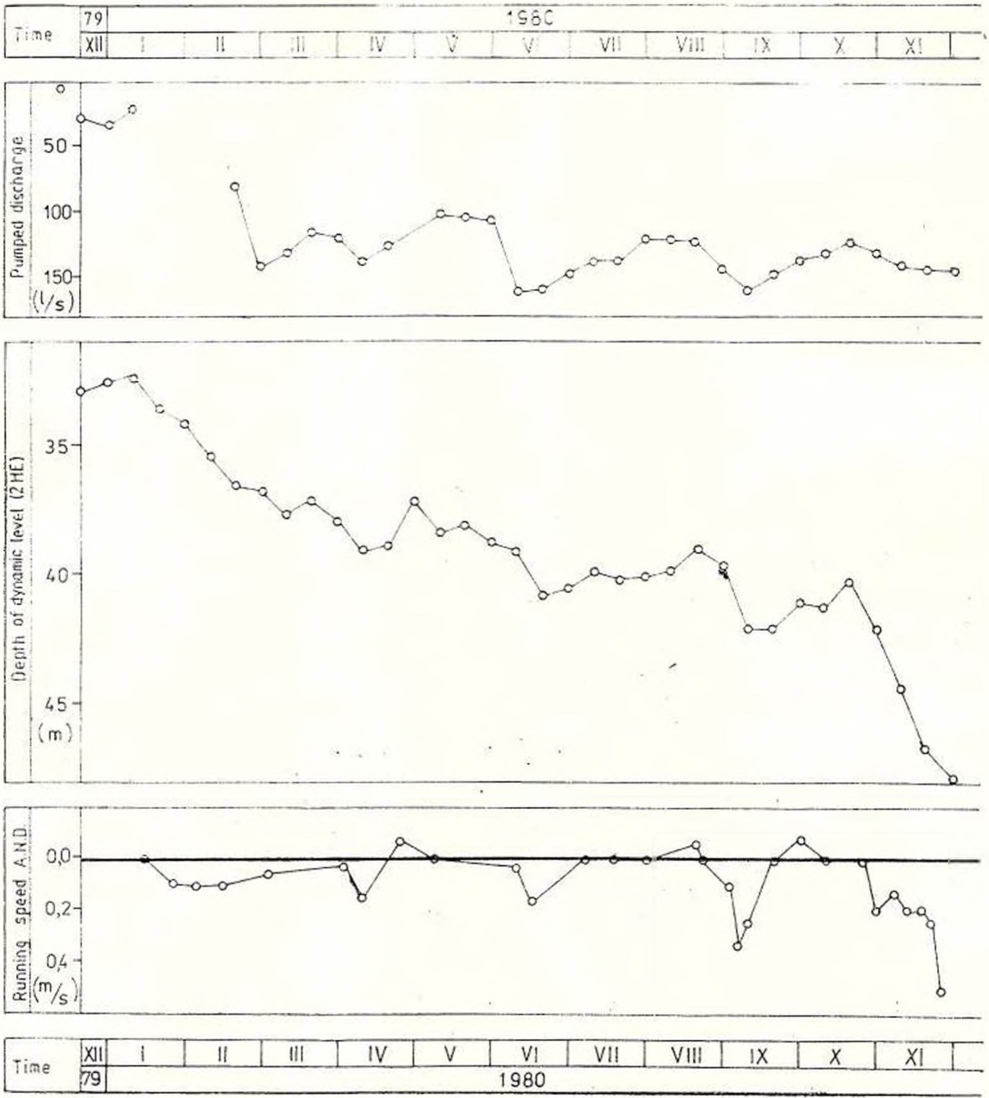

Fig. 3 — Evolution of hydrodynamic elements of dewatering for the aquifer in Prunişor pilot-mine (B 2HE).

The system used by us, with three possible states, exhibits the following maximum entropy :

$$H_{i_{\max}} = -\log_2 3 = 1.58$$

The values of entropy H_i from Table are smaller than the value of maximum entropy, pointing to a certain order quantitatively defined as a ratio$(H_i/H_{i_{\max}})$ or a difference $(H_{i_{\max}} - H_i)$.

TABLE

Quantif. value	Quantification error	Entropy	Piezometer wells					
			2HE	3HE	4HE	5HE	7HE	8HE
h	−0.05↔0.05	H_i	1.15	1.44	1.15	1.30	1.14	1.39
		H_d	1.08	1.05	0.80	0.97	0.94	1.28
		ΔH_{i-d}	0.07	0.39	0.35	0.33	0.23	0.13
	−0.10↔0.10	H_i	1.24	1.52	1.26	1.30	1.35	1.39
		H_d	1.29	1.08	0.91	0.97	1.21	1.28
		ΔH_{i-d}	0.15	0.44	0.35	0.33	0.14	0.13
	−0.20↔0.20	H_i	1.44	1.55	1.41	1.50	1.35	1.50
		H_d	1.55	1.37	0.87	1.37	1.20	1.39
		ΔH_{i-d}	0.09	0.18	0.54	0.13	0.14	0.11
q	−0.05↔0.05	H_i	1.36	1.48	1.49	1.56	1.47	1.58
		H_d	1.30	1.38	1.41	1.52	1.26	1.54
		ΔH_{i-d}	0.06	0.10	0.08	0.04	0.21	0.04
	−0.10↔0.10	H_i	1.35	1.55	1.55	1.58	1.58	1.37
		H_d	1.14	1.44	1.48	1.55	1.54	1.48
		ΔH_{i-d}	0.41	0.11	0.07	0.03	0.04	0.09
	−0.20↔0.20	H_i	1.54	1.57	1.55	1.53	1.44	1.55
		H_d	1.35	1.44	1.50	1.44	1.43	1.47
		ΔH_{i-d}	0.19	0.13	0.05	0.09	0.01	0.08

The entropy of the markovian process adopted by us, is calculated according to the relation :

$$H_d = -\sum_{i=1}^{n} P_i \sum_{i=1}^{n} P_{ij} \log_2 P_{ij} \tag{5}$$

The values of H_d(Tab.) are less than H_i values pointing to a more "organized" system by adopting the markovian model. This additional organization proves the amount of information used for stating the interdependence of the values of aleatory variables.

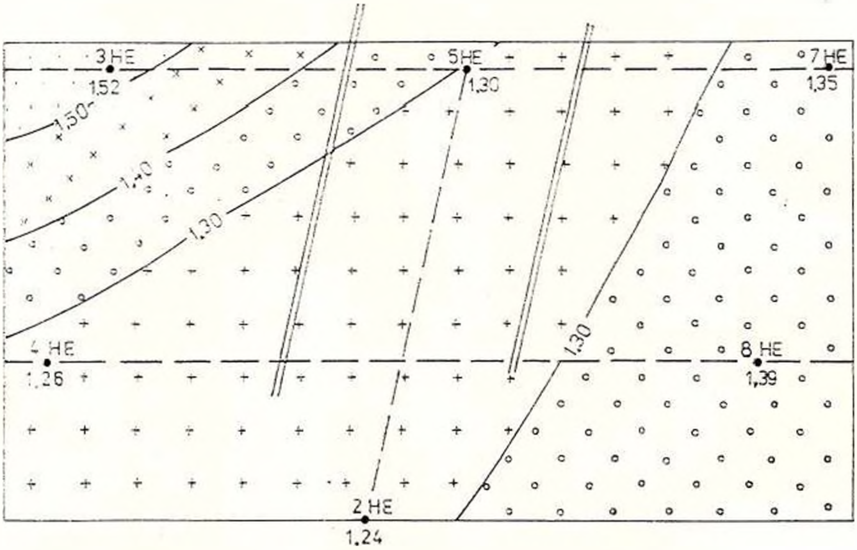

Fig. 4 — Distribution of unconditioned entropy (H_i) in the area of Prunişor pilot-mine.

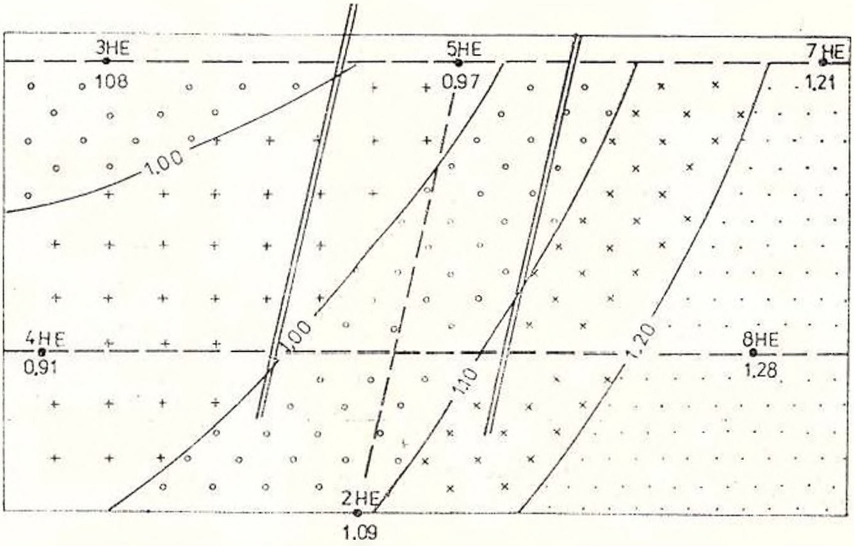

Fig. 5 — Distribution of transition matrix entropy (H_d) in the area of Prunişor pilot-mine.

By using the entropy values (H_i, H_d, $H_i - H_d$) calculated by liniar interpolation between piezometers, equal entropy lines for pilot-mine Prunişor East were stated (Figs. 4, 5, 6).

The direction of equal entropy lines points to the preponderent role of the drainage network as regards the distribution of entropy in the mining field. The drainage network is located on two alignments which form an angle of 75°, one of them trending eastwestwards. The equal entropy lines form an angle of 40°, trending EW. so that they are parallel to the vector that shows the trending of maximum share of drainage in the mining field.

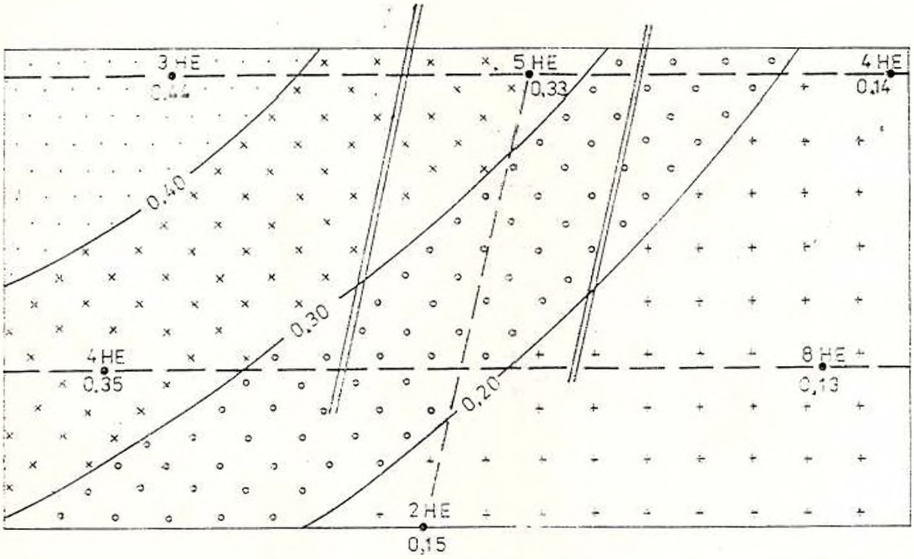

Fig. 6 — Distribution of residual entropy ($H_i - H_d$) in the area of Prunişor pilot-mine.

The division of the mining field in zones of different entropies points to the different evolution of dewatering. It is difficult to delimit quantitatively the role of the aquifer from that one of the drainage network on the complex nature of dewatering ; it implies the detailed study of one of the two sub-systems.

The stabilization of dewatering is inferred from the evolution of system entropy. The evolution of dewatering in the pilot-mine Prunişor East marks an increase of entropy between July-November 1981 as compared to January-June 1981.

The distribution of entropy dynamics in the mining field is similar to that one of entropies H_i, H_d, H_{i-d}. According to this analogy, the main source of disturbances within this system is due to the drainage network.

Dewatering is marked by stabilization in a narrow area between boreholes 4HE and 5HE (Fig. 4). This axis forms an angle of about 40°, trending E-W. Consequently, dewatering did not reach stabilization all over the dewatered mining field. Thus it is impossible to state the laws of the evolution of the piezometric level necessary to foreseeing the evolution of dewatering under known drainage conditions.

By means of entropy, one may determine the areal distribution and evolution in time of dewatering. A great amount of data may be processed by using a very simple calculation program.

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STRUCTURAL AND TEXTURAL STUDY CONCERNING THE CRYSTALLINE AND MAGMATIC ROCKS IN THE SOUTH CARPATHIANS.
ENGINEERING-GEOLOGICAL SIGNIFICANCE

BY

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The mobile zone of the South Carpathians supposes a long geological evolution which was better outlined during the Upper Precambrian (about 1600 m.y.), and accomplished during the last movements of the Alpine cycle. The various orogenic phases have affected this mountainous edifice, by leaving their marks on the rock structure, texture and mineralogical content.

The most peculiar geological element of the South Carpathians is given by the coincidence between various structural elements (faults, folds, cleavages, foliations, etc.) and the general direction of this mountainous chain, no matter the very old or very new foundations. One can notice, as well, that most of the granitoid intrusions (Assyntic, Caledonian) show a prolongation, following this chain direction. Therefore, we must admit that the old structures have suffered a radical reorientation and an essential rejuvenation during the alpine movements, which were proved by radiometric measurements.

Even since 1905, Murgoci has separated two large tectonical units in the South Carpathians : a Proterozoic crystalline basement with its Paleozoic-Mesozoic sedimentary cover, which he called the South Carpathians Autochthonous and a overlapping this one, called the Getic Nappe. This unit is formed of Proterozoic crystalline basement and a Paleozoic-Mesozoic cover.

From the lithostratigraphic, metamorphic conditions and tectonical points of view, three crystalline series were separated : the Drăgșan Series with two complexes (amphibolites and sericite-chloritic), the Lainici-Păiuș Series and the Tulișa Series. The first two series are considered to be formed during the Assyntic-Caledonian orogenesis (850 ±

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50 — 375 m.y.) in the metamorphic conditions of the facies of amphibolites with almandine, albite-epidote-amphibole respectively.

Within the Drăgșan Series there are massive or banded amphibolites with garnets and epidote, amphibolic gneisses, crystalline limestones, sericite-chloritic schists, quartzites, chloritic gneisses, chloritic schists with epidote, albite and actinote and several concordant intercalations of serpentinites, metadiorites and metagabbros. The granitoid rocks which cross the crystalline schists of this series as veins, generally granitize them. These rocks are marked by retromorphism.

The texture patterns and mineralogical parageneses of the Lainici-Păiuș Series crystalline schists underline the fact that they were formed on the basis of some epiclastic deposits in the amphibolic facies conditions.

The Tulișa Series (Lower Paleozoic) basically starts with metaconglomerates, then quartzites, crystalline limestones, tuffogene green rocks with intercalated cipolinic limestones, serpentinites, and ends with various phyllite types.

The first two crystalline series contain several massifs of sinkinematic and postkinematic granitoid rocks of 650—104 m.y.

The crystalline of the Getic Nappe is formed of two crystalline series : the Sebeș-Lotru Series (1600—850 m.y.) and the epimetamorphic series (320 m.y.). The crystalline schists of the Sebeș-Lotru Series are represented by various types of gneisses and micaschists with intercalated quartzites, amphibolites, pegmatites, aplites, migmatites, serpentinites, eclogites and granitoid rocks.

The epimetamorphic series is formed of sericite-chloritic schists, tuffogene rocks with intercalated crystalline limestones, quartzitic, amphibolitic schists, etc. The granitoid massifs inserted among crystalline schists of the Sebeș-Lotru Series, by methods U/Pb and Pb/Pb have given and age of 620 m.y., 328 m.y. respectively.

The microgranoblastic, lepidoblastic and porphyroblastic structures dominate the crystalline schists of the Lainici-Păiuș Series. Their texture is schistous, lenticular or banded. The crystalline schists of the Drăgșan Series are dominated by the same structural and textural types (microgranoblastic, nematoblastic, lepidoblastic and porphyroblastic, namely by schistous, lenticular or banded textures). The following structures are found in granitoid rocks : the hypidiomorphic grainy cataclastic or granoblastic structures ; textures are massive up to oriented (gneissic) ones.

The following structures are found in the Tulișa Series : the granoblastic, lepidoblastic, heteroblastic and microgranoblastic. The textures are dominated by schistous-phyllitic and banded ones.

The crystalline schists of the Sebeș Series are characterized by granoblastic, granolepidoblastic, porphyroblastic, poikiloblastic, nematoblastic and nematogranoblastic structures, while textures are represented by the massive, schistous, banded, ocular, lenticular textures. The structures and textures of the crystalline schists of the epimetamorphic series are similar to those of the Drăgșan Series.

The metamorphic rocks are formed of aggregates of minerals and each mineral ratio together with the structure, texture and degree of transformation form the basis of their geotechnical classification.

The mineral types composing the crystalline schists of the Autochthonous or of the Getic Nappe mainly belong to silica : neosilica (garnet, zircon, olivine, garnets, zircon), neosubsilica (kyanite, chloritoid, staurolite, titanite), sorosilica (epidote, zoisite, allanite, lotrite), cyclosilica (cordierite, tourmaline), inosilica (diopside, omphacite, tremolite, actinote, hornblende, glaucophane), phyllosilica (prehnite, talcum, muscovite, biotite, vermiculite, chlorite, antigorite, cristotile), tectosilica (orthose, microcline, plagioclases, zeolites), among oxides (quartz, rutile, ilmenite, magnetite, hematite), among carbonates (calcite, dolomite, ankerite), among phosphates only apatite, among sulphides, mainly pyrite.

Rocks containing quartz, calcite and oligoclase in their mineralogical composition are the most competent because of the competence degree of these minerals. These minerals impress massive-granoblastic structures and textures to the above mentioned rocks.

Generally speaking, these rock complexes (from the Autochthonous or the Getic Nappe) contain nearly the same main minerals ; they vary only as frequency from one rock type to another. For example, mixt gneisses contain (in lowering frequency) : microcline, plagioclases, quartz, biotite and muscovite, in a total of about 95%, of which microcline, plagioclase and quartz are about 75% from the rock mass. This prevalent mineralogical content gives a more massive, hard character to the rocks being very resistant at compression and friction. These rocks belong to the category of competent, quasielastic massive rocks E_d = 7000 MPa, E_s = 10 000 MPa, E_{din} = 35 000 MPa, K_o = 1000 daN/cm^3 and tg φ = 0.7—1.00.

Almost the same minerals, but in other ratios as frequency, compose micaschists : plagioclases about 15%, quartz about 30%, biotite about 25%, muscovite about 20%. Thus, stable minerals totalize about 31.5%, while unstable minerals reach about 56.5%. To these ones there add incompetent secondary minerals such as : chlorite, sericite, epidote, zoisite ; all these ones give the rocks a character with many discontinuity surfaces, with altering possibilities. These rocks belong to the category with middle to weak competence, elastic to nonelastic ones : E_d = 1 500—2 000 MPa, E_s = 3 500 MPa, E_{din} = 9 000 MPa and K_o = 350 daN/cm^3.

As the unstable minerals ratio increases, the rock competence decreases.

According to this fact, there were determined the technical factors of the Autochthonous rocks (crystalline schists of the Drăgșan Series, granitoid rocks and crystalline schists of the Tulișa Series), as well as from the Getic Nappe crystalline (the Sebeș-Lotru Series).

According to engineering-geological and geotechnical characteristics, four zones were separated on rock categories : the A zone which contains : granitoids, gneisses, quartzites, pegmatites, limestones, rocks with rare ruptural tectonic lines and with upper geotechnical characteristics. Within this zone, the presence of lower category rocks is mini-

TABLE 1

Technical coefficients

Rock type at gallery level	Elasticity modulus E, MPa	Coefficient of elastic resistance (deformation way) K_0, daN/cm³	Coefficient of rock hardness f(Protodyaconov)
Crystalline of the Getic Nappe (Sebeș-Lotru Series)			
Biotitic-muscovitic gneisses with pegmatite injections	1000 – 9000	100 – 900	1 – 9
Muscovitic-biotitic schistous gneisses with pegmatitic injections	1000 – 5000	100 – 700	1 – 7
Biotitic-muscovitic microcrystalline gneisses	1000 – 8000	400 – 800	3 – 10
Ocular gneisses with intercalated micaceous gneisses	3000 – 10000	300 – 1000	2 – 10
Amphibolic gneisses	6000 – 7000	700 – 800	6 – 8
Biotitic-muscovitic macrocrystalline gneisses with garnet micaschist intercalations and pegmatitic injections (the whole complex is strongly tectonized)	1000 – 6000	100 – 600	1 – 6
Biotitic-muscovitic gneisses with intercalations of schistous muscovitic gneisses and pegmatite injections	3000 – 10000	100 – 1000	1 – 12
Autochthonous *Crystalline schists of the Tulișa Series*			
Green schists with serpentinite lenses	1700 – 7000	150 – 800	1.5 – 9
Green schists with intercalated limestones	1000 – 5000	100 – 600	1 – 6
Blastodetrital schists with intercalated graphitic phyllites, metaconglomerates	1000 – 4000	100 – 600	1 – 6
Quartzitic schists with sericite	2000 – 3000	200 – 500	2 – 5
Granitoid rocks			
Laminated granitoids with intercalated green schists	2000 – 7000	200 – 800	2 – 8
Massive granitoids with mylonitization zones	3000 – 8000	300 – 900	3 – 9
Massive granitoids	1200 – 10000	800 – 1200	8 – 12
Granitoids with inclusions of amphibolites	5000 – 10000	500 – 1000	5 – 10
Granitoids in gneissic and massive facies	7000 – 9000	800 – 900	10 – 12
Schistous granitoids	2000 – 5000	200 – 400	2 – 5
Cataclastic granitoids	3000 – 5000	300 – 400	3 – 4
Very fissured granitoids	4000 – 7000	500 – 600	5 – 6
Porphyroid granitoid	6000 – 9000	700 – 900	7 – 8
Granitoid with quartz and aplitic veinlets	6000 – 8000	700 – 800	7 – 8
Partly altered porphyroid granitoid	3000 – 6000	400 – 500	4 – 5
Altered, fissured and faulted granitoid	2000 – 4000	300 – 400	2 – 4
Crystalline schists of the Drăgșan and Lainici-Păiuș Series			
Sericitic-chloritic schists with intercalations of graphitic schists and limestones	5000 – 6000	500 – 600	5 – 6
Amphibolites	8000 – 9000	800 – 900	8 – 9
Micaceous schists	2000 – 6000	200 – 400	3 – 5

TABLE 2

Engineering-geological and geotechnical characteristics on rock categories

Rock category	Mineralogical content	Structural and textural characteristics	Tectonical and microtectonical characteristics	Alteration characteristics	Geotechnical characteristics				
					$F_{prot.}$	E_d MPa	$\frac{E_s}{E_d}$	σa MPa	tg φ
I	quartz, feldspar plagioclase, microcline, orthose	massive, granolepidoblastic	weakly tectonized	unaltered, weakly alterable	6	5000	$\frac{1}{1.2}$	1.0	0.75
II	feldspars, micas, quartz, chlorite, epidote	schistous granolepidoblastic	tectonized, fissured	unaltered, alterable	5–6	3000- 5000	$\frac{1}{1.9}$	2.5 3.0	0.65
III	hornblende, micas, feldspars, quartz chlorite, sericitee pidote, argillous minerals	schistous, lepidoblastic	tectonized, fissured, exfoliated	altered, alterable	3–5	1500 3000	$\frac{1}{2.3}$	1.2 1.8	0.55
I	micas, chlorite, epidote, sericite, feldspars, quartz hornblende, argillous substance*	schistous, lepidoblastic	very tectonized and fissured	altered and alterable	1–3	1000 1500	$\frac{1}{2.9}$	0.8 1.2	0.40 0.45
V	graphite, chlorite, epidote, sericite, quartz, argillous substance**	lepidoblastic, schistous cataclastic, mylenitic	very tectonized	very altered	1	1000	$\frac{1}{3.2}$	0.4 0.8	0.3 0.4

* argillous minerals 25–30%
** argillous minerals 60–80%.

mum (on the fault zones) or they are absent ; the B zone is characterized by unaltered but alterable schistous rocks which are fissured and have middle geotechnical characteristics, such as : paragneisses, amphibolic gneisses chloritic-sericitic schists, quartzous-feldspathic schists ; the C zone is characterized by fissured, weakly altered and alterable schistous rocks with lower geotechnical characteristics, such as : amphibolic schists, chloritic-sericitic schists, serpentinites (very fissured), micaschists, chloritic schists, serpentinites (very fissured), micaschists, chloritic schists, and the D zone, with very schistous rocks which are very tectonized and altered, such as crystalline schists of the Tulișa Series.

From the above mentioned data one can conclude that within the zonal classification, the lithological, structural and textural characteristics are very important ; the tectonical characteristics are accidental events (gneisses, no matter if they belong to the Autochthon or to Getic Nappe, if they have the same mineralogical content, the same structure and texture, the same geotechnical characteristics, they are encompassed within the same zone). From the above tables it results that within one zone several rock categories are separated (Tab. 2), which can coexist in different ratios. The large variation is for transition rocks (categories II—III).

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SOURCE-AREAS OF THE ASSYNTIC FLYSCH DEPOSITS IN THE CENTRAL DOBROGEA MASSIF (ROMANIA)

BY

NICOLAE ANASTASIU [1], DAN JIPA [2]

Although apparently monotonous, the Assyntic flysch deposits in Central Dobrogea ("greenschist series") group a wide range of terrigenous deposits, the main constituents of which — quartz granoclasts, feldspars, chlorite and lithic fragments — offer valuable information on the source-areas ; their granulometric and morphometric parameters prove to be useful for defining the transport and sedimentation conditions in the former Assyntic basin.

Petrographic Data

From petrographic point of view the greenschist formation is of terrigenous, siliciclastic nature and consists of rocks characterized by a large granulometric and mineralogical variety : rudites, arenites, siltites and lutites of polymictic and rarely oligomictic constitution.

Depending on the amount of their detrital matrix, rudites become orthorocks and pararocks respectively. They consist of magmatic (55%), metamorphic (30%) and sedimentary (15%) lithoclasts and quartz, feldspar, mica (muscovite, biotite, chlorite), heavy mineral (zircon, sphene, staurolite, garnet, zoisite, epidote, etc.) granoclasts. The binding material is represented by neoformation chlorite-rich orthomatrix. The rocks are characterized by a low degree of sorting ($\sigma = 1—2$), while the grain size parameters occur within variable limits ($R_0 = 0.1—0.3$ to $0.8—0.9$; $S = 0.4—0.7$). Diagenetic transformations alter, especially, the groundmass constituents.

Arenites, from coarse- to fine-grained, include all petrographic types : greywackes, arkoses, lithic sandstones and quartz sandstones. Greywackes abound in the middle series of lithologic columns of Assyntic flysch deposits ; their composition varies from lithic (chlorite, biotite) to

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feldspathic (plagioclase An$_{5-20\%}$) ($\pm$ sphene, staurolite, zircon, clinozoisite) and the binding material is represented by a chlorite-rich siliciclastic matrix ; Q : F : L ratio varies continuously (Fig. 1). They are characterized by a low sorting degree (σ = 0.7—2). Diagenetic transformations are also varied (anchimetamorphic recrystallization, silicification, epitaxy, authigeneses — pyrite, chlorite, chalcedony). As concerns the

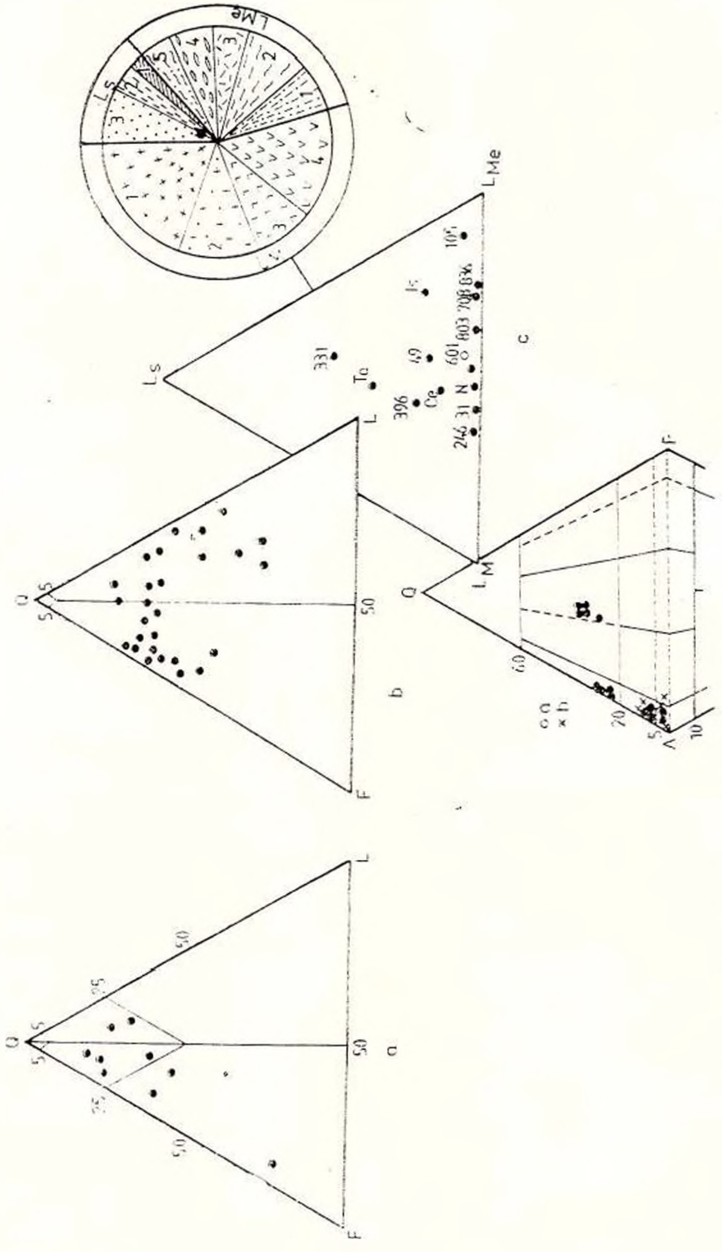

Fig. 1 — Percentage composition in granoclasts (Q — quartz ; F — feldspars) and lithoclasts (L) of arenites (a, b) and rudites (c) of Assyntic flysch deposits (Central Dobrogea Massif — "greenschist" zone) ; diagram Q-A-P shows also the composition of igneous lithoclasts.

arenitic series, the variation of quality in sandstones is reduced ; the individualization of above mentioned petrotypes is the result of quantitative variation of Q, F, L (Fig. 1) ; the same constituents are to be found in siltites or even lutites.

Quartz occurs as granoclasts with straight and undulatory extinction and oriented inclusions. Usually feldspars are represented by low alteration degree and in places by orthoclase or fresh microcline. The lithic fragments are of the same nature as those of rudites.

The roundness of granoclasts is often of reduced value (R_0 = 0.5—0.4) and the sphericity is not a significant parameter (S = 0.2—0.8). The sorting is variable (σ = 0.5—0.75 in the case of quartz sandstones and 0.7—1.5 in the case of arkoses and lithic sandstones).

The binding material of arenites is represented by siliceous (chalcedony) or chloritic (recrystallization chlorite) cement.

Mineralogical Data

Nature of granoclasts (quartz, feldspars, chlorites)

Quartz is abundant (60—90%) in all epiclastic types and occurs, almost usually, as granoclasts with undulatory extinction which points to a source-area containing rocks that underwent mechanic deformation or others with straight extinction of magmatic origin (with zircon inclusions).

Feldspars, frequently occurring in conglomerates, greywackes and arkoses constitute several associations :

— plagioclase (An_{5-18})-microcline, in terrigenous series at Istria, Taşaul, Crucea, Saraiu, Războieni (Fig. 2) ;

— plagioclase (An_{5-10})-orthoclase (2V = 63—68°) in arenites at Beidaud, Ciamurlia, Runcu, and exclusively as

— plagioclase (An_{5-15}) in sandstones and conglomerates at Sibioara, V. Alecsandri, Războieni, Rahmanu. Secondary transformations (kaolinization, sericitization) affected partly the plagioclases, the orthoclase in places and never the microcline.

Plagioclase is twinned according to Albite and Albite-Carlsbad laws and does not exhibit zoning. Its optical properties assign it to low temperature elements, characteristic of plutonics and "low grade" metamorphic rocks.

Orthoclase, partly Carlsbad twinned, shows optical features characteristic of subsequent granitoids.

Microcline is always easy to identify due to its quadrille structure and is characteristic of gneisses and crystalline schists associated gneisses. No feldspar granoclast allows the identification of features characteristic of pyroclastic rocks.

Chlorite, as granoclasts with obvious deformation traces (curved cleavages, anomalous extinction) occurs in all the rocks accompanied by biotite or muscovite ; its features of standard mineral of low grade metamorphic facies are obvious ; it also occurs as secondary product on biotite.

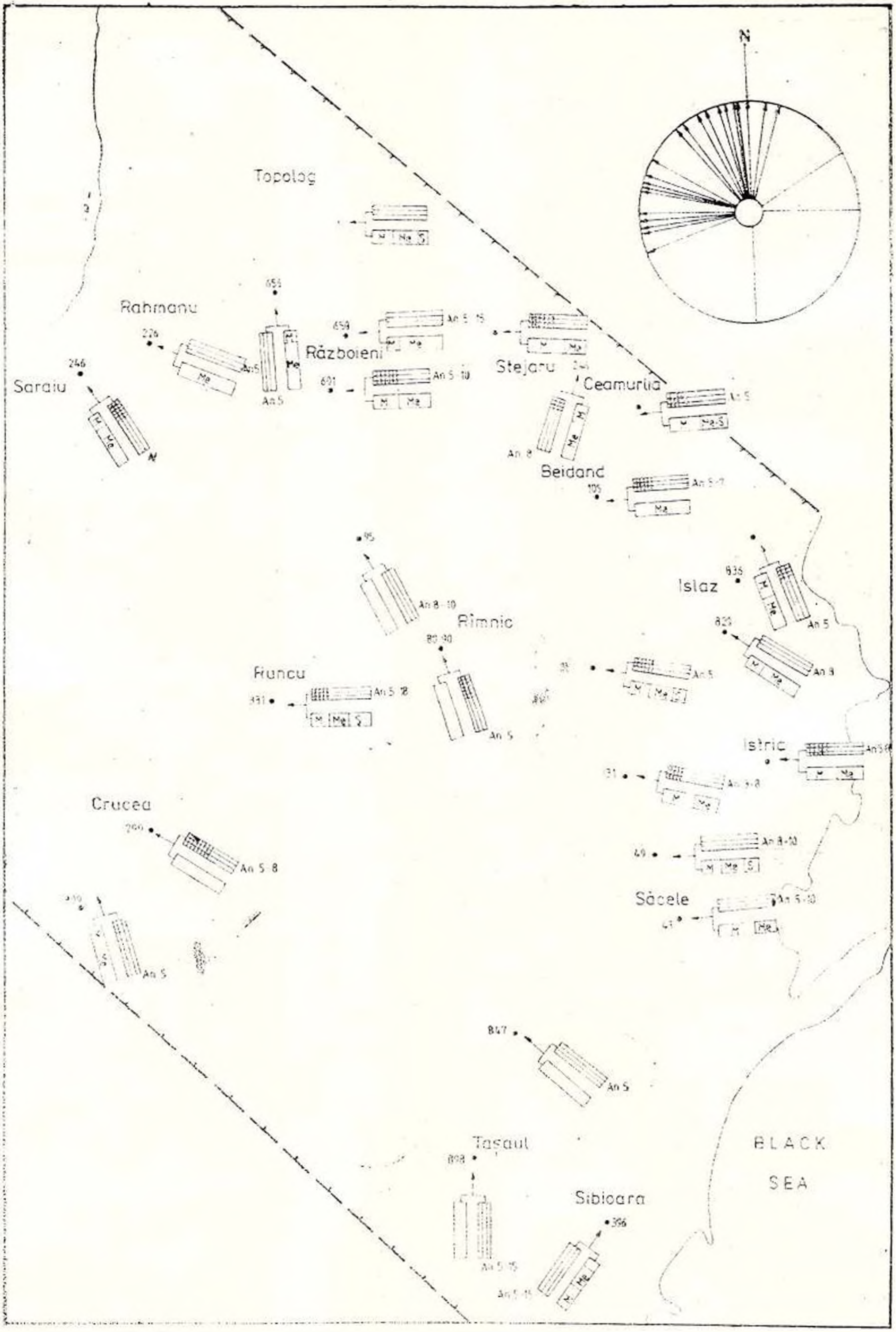

Fig. 2 — Main transport directions of feldspars (plagioclase represented by rectangles with horizontal lines, microcline represented by grating) and lithoclasts (M — igneous ; Me — metamorphic ; S — sedimentary) in Central Dobrogea Massif ("greenschist" zone).

Nature of lithoclasts

The lithoclasts of coarse-grained arenites and rudites may be determined and belong to rocks of different origins (plutonic, volcanic, metamorphic and sedimentary).

The lithoclasts which represent plutonite fragments are the most frequent and exhibit the following mineralogical associations and structural features :

— plagioclase (An_{5-10}) + quartz, in plagigranites with allotriomorphic granular structure and massive texture ;

— orthoclase + albite + quartz, in pegmatites with graphic structure ;

— microcline, perthite microcline + plagioclase (An_{10-15}) + quartz, in calc-alkaline granites with medium allotriomorphic and microgranular structure (in aplite facies) ;

— plagioclase (An_{15}) + orthoclase + quartz + biotite, in granodiorites with hypidiomorphic medium granular structure ;

— plagioclase + quartz in myrmekite intergrowths.

The volcanics fragments are scarce and belong to tachylytic rocks, of albitophyre type (plagioclase — An_{5-8} + orthoclase in radial intergrowths) with partly porphyry structure and flow-bostonitic texture in the Taşaul and Izlaz conglomerates, dacites (plagioclase phenocrysts and corroded quartz in a microcrystalline groundmass) with porphyry structure in the Sibioara Conglomerates.

The lithoclasts which represent metamorphite fragments are less frequent and include mineralogical associations typical of :

— gneisses (microcline + oligoclase + quartz + biotite ± staurolite) in the Sibioara Rudites ;

— micaschists (biotite + muscovite + quartz ± oligoclase) ;

— quartzites in rudites and lithic sandstones at Sibioara and Ciamurlia ;

— phyllites (chlorite + sericite + quartz) in rudites at Izlaz and Palazu Mic ;

— retromorphics (biotite + chlorite + oligoclase + albite + quartz) in the Palazu Mic Epiclasts.

The lithoclasts resulting from sedimentary rocks are quite relevant and belong to highly diagenized quartzose sandstones (the granoclasts of which exhibit suture joints) at Taşaul, Sibioara and Palazu Mic, to subarkoses, at Taşaul and Izlaz and to siltic rocks, rich in quartz and mica, with obvious profound anadiagenetic transformations (it is difficult to assign them to silicolites or to metasiltites ; these polymetamorphic particles may stand for siliceous remobilizations of diagenetic nature abounding in chalcedony ; they often exhibit concave-convex joints with surrounding granoclasts and lithoclasts). Some rudites contain shale pebbles which represent resedimented lutites of intrabasinal origin.

The binding material of epiclastic rocks is of matrix and cement nature. The matrix (ortho- or epimatrix) includes chlorite, sericite, iron hydroxides and a few carbonates. The cement is siliceous (opal and/or chalcedony) and impure due to iron hydroxides. Structurally it is a

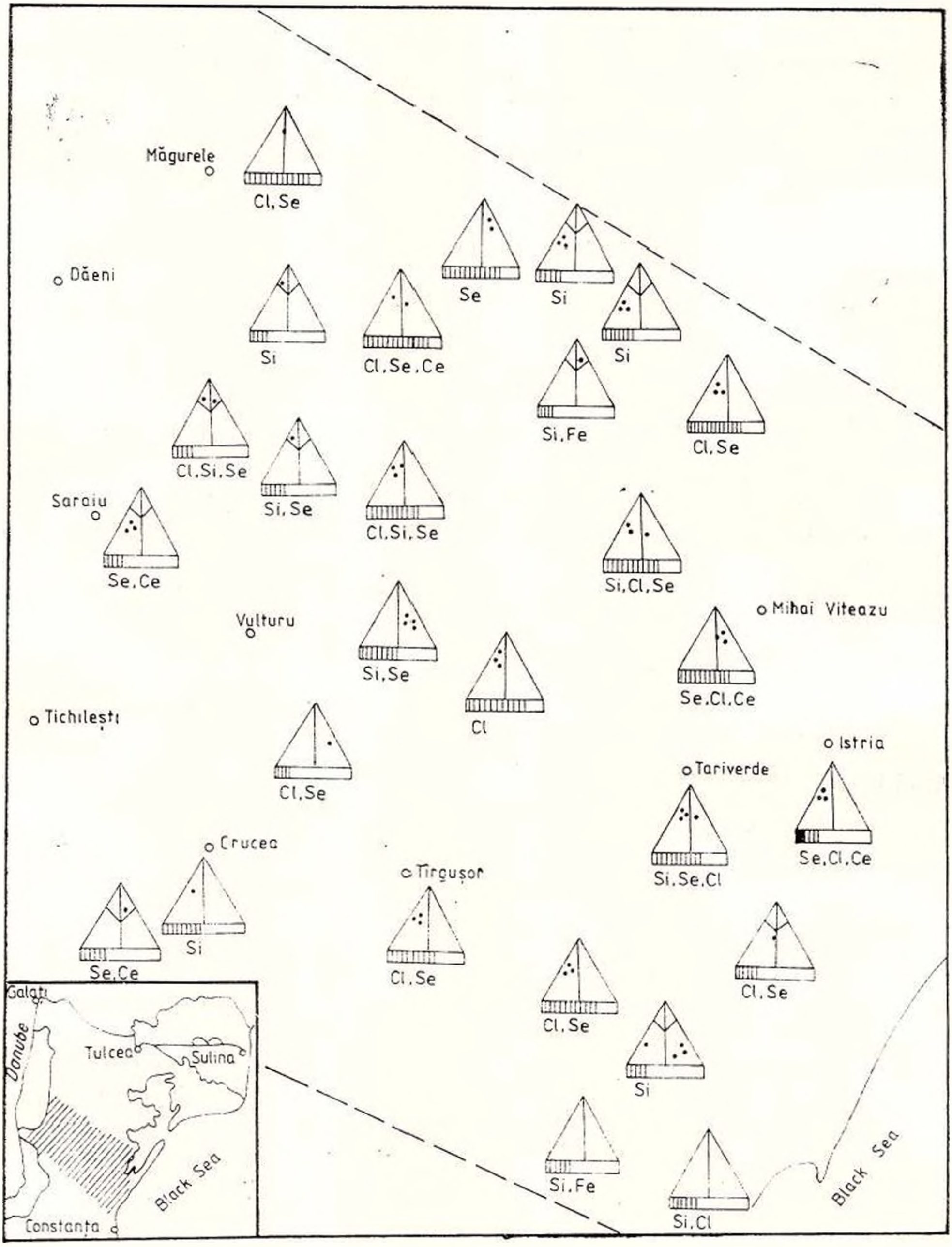

Fig. 3 — Spatial distribution of arenite types in Central Dobrogea Massif. At the base of each Q-F-L triangle, the binding material (black)-grains (white) ratio is shown ; on the left side of the triangle — main constituent minerals of the binding material (Se — sericite ; Cl — chlorite ; Si — opal or chalcedony ; Fe — iron hydroxides ; Cc — calcite). Other signs : ∅, Md or mean diameter of grains ; So, grain size sorting as standard deviation ; Ro, grain roundness index.

basal binding material, partly film-like (when micas occur round granoclasts and lithoclasts ; the increased frequency of these particles leads to the decrease of interstitial areas and thus to the constitution of pore cement). Within a complete epiclastic sequence one notes this tendency starting from base to its top. The transition from basal to pore binding material is graded within complete sequences starting with rudites, continuing with greywackes and arkoses and ending with quartzose sandstones ; the areal distribution of rocks (Fig. 3) exhibits no principle of frequency of binding material or sedimentary particles.

Sedimentary Structures

The terrigenous sequences in the "greenschist" area exhibit frequently a rhythmic character and an arhythmic one at Sibioara. The rhythms are incomplete, normal and accidental and exhibit centimetric and metric thicknesses ; complete rhythms and macrorhythms are rare (Jipa, 1970).

Within rhythmic sequences the depositional structures consist of simple, normal graded-beddings — either continuous or discontinuous — cross laminations and rarely convolute laminations (at Năvodari) and frequent current ripples ; imbrications and sliding structures were identified at Sibioara. Erosional structures are represented by directional markings due to paleocurrent erosion (flute marks) or to the objects transported (drag marks, chevron marks).

Direction measuring of cross laminations, current ripples and sole markings (especially flute marks and drag marks) account for three paleocurrent systems : east to west, south to north, west to east and for the role of current transport (suspension and traction) and mass transport in the dynamics of sedimentary material (Jipa, 1970).

Reconstruction of Source-Areas

The petrographic study of epiclastic formations correlated with the sedimentary structures allows a discussion on possible source-areas and on their position as compared to the Assyntic "greenschist" basin.

By reconstructing current directions one distinguishes three possible sources of terrigenous material (Fig. 4) :

— a western source, the farthest one, which supplied the material to the northwestern area of the basin ;

— a southern source, probably the nearest one, which generated the coarse-grained material and

— a northern source which supplied the fine-grained material.

Their petrographic nature was heterogeneous from the very beginning and their erosion was rapid. Extrabasinal sources are characterized by highly metamorphosed crystalline schists (staurolite gneisses, garnet micaschists), low metamorphosed ones (phyllites, quartzites) and associated plutonics of quartz-feldspar composition (granitoids, pegmatites). The nature and optical features of feldspars (absence of plagioclases that exceed 20% An) account for a source-area in which basic rocks were absent and volcanics occurred in places. The occurrence of the latter in the source-area is obvious (trachyte and dacite lithoclasts) and points to probably pre-Assyntic volcanic activity ; the structural facies of flow and porphyry volcanics accounts for the effusive nature of rocks from the source-area. The alkaline trend of lithoclasts-sienites/trachytes points to consanguinity relations between plutonic and volcanic products.

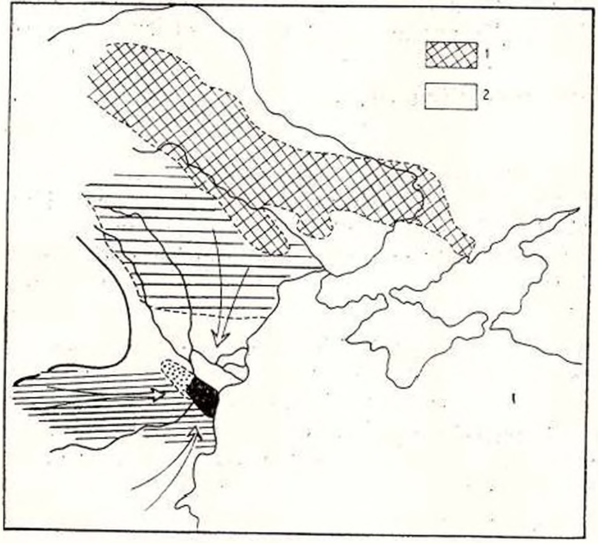

Fig. 4 — Location of source areas in respect of the Assyntic basin.
1. Scythian Platform; 2, Moesian Platform.

The sedimentary rock lithoclasts (quartzose sandstones, arkoses, siltites) in "greenschist" rudites and arenites prove the conservation in extrabasinal source-areas of a sedimentary cover emerged during the Assyntic and accumulated prior to it ; this is one of the few proofs of the nature and occurrence of some Precambrian unmetamorphosed sedimentary rocks, in adjacent areas. These lithoclasts seem to be the oldest sedimentary rocks reported from a geological formation in our country.

This petrographic configuration corresponds partly to the Nipru-Bug Series (granitoids) and the Krivoirog Series (phyllites and retromorphics) and/or to the Ovruci Series (sandstones) in the Scythian Platform (Ucrainian craton) and to their Middle or Lower Proterozoic equivalent from the basement of the Moesian Platform.

The variation of grain size (Md, σ) and morphometric parameters (Ro, S) of terrigenous deposits in Central Dobrogea Massif may be used, together with the other structural features, as marker of the location of the source-area related to the basin and of the dynamics of sediments to or within the basin. Thus, proximal, coarse facies prevail in the neighbourhood of source-areas, in the south and south-east of Central Dobrogea Massif, while distal, fine-grained ones occur in the north and northwest ; from petrographic point of view, the terrigenous deposits exhibit a lithic and a feldspathic facies. According to the degree of roundness, some quartz, feldspar, heavy mineral granoclasts point to the reworking of former sedimentary material, while the lithoclasts with no petrographic equivalent in supposed sources, show small values of the roundness degree which point to the transport from an intrabasinal area.

The correlation of data — great thickness of Assyntic flysch deposits (5000 m), swift transition from proximal to distal facies, occurrence of shale pebbles, volcanic lithoclasts and matrix as main binding material of terrigenous rocks — shows that sediments accumualted in a basin larger than present-day Central Dobrogea and of variable depth. The coexistence of various current systems in the same area accounts for the presence of a flat accumulation area in central zones. The evolution of Central Dobrogea Massif — inserted between two important crustal sectors : Scythian Platform and Moesian Platform — may be regarded as starting from a probably intracontinental basin with inactive margins, below which incipient subcrustal erosion processes led to the thinning out of lithosphere and the initiation of active subsidence (which favours the accumulation of thick sediment piles).

The morphological configuration of this basin should also include intrabasinal elevations which acted intermittently as sources of material and influenced the dynamics of currents and mass transported material. The alkaline volcanic lithoclasts should be considered as reworked products from neighbouring areas, already stable, and not as an expression of intrabasinal manifestations.

Sédimentologie

LARGE SCALE PROGRADATION STRUCTURES IN THE ROMANIAN CARPATHIANS : FACTS AND HYPOTHESIS

BY

DAN JIPA[1]

Investigation of the sedimentary structures advanced rapidly. During evolution all efforts have been directed to the description and interpretation of the small scale features. The very large sedimentary structures belong to the field of megasedimentology. As "microsedimentology" still have much knowledge to accumulate, megasedimentology is totally neglected.

The present author had the opportunity to identify several sediment accumulations produced by large scale progradation. In this way he felt stimulated to approach the difficult field of megasedimentology.

Progradation Structures Recorded in Romania

Bucegi Conglomerates (Albian)

Located at the bend of the Romanian Carpathians (Fig. 1 A), the Bucegi Mountain is geologically made up mostly of an Albian conglomeratic sequence (Patrulius, 1969). Known as the Bucegi Conglomerates this sequence consists of a monoclinal alternation of conglomeratic, sandy and silty lithologic horizons.

The Bucegi conglomeratic Complex is supported by an Aptian detrital formation mainly consisting of an alternance of thin bedded sandstones and marls. At the upper part of the Aptian formation Patrulius (1969) separated a distinctive intercalation of calcareous rudites, named the Raciu Breccia. This intercalation discontinuously occurs at the same stratigraphic level all over the Bucegi Mts area. Due to detailed lithologic mapping (Patrulius, 1969 ; Jipa, 1982), it has been pointed out that the various major lithologic horizons representing different levels of the Albian sequence are coming in direct contact with the underlying Aptian deposits (Fig. 1 B). Moreover, most of these lithologic horizons rest on the

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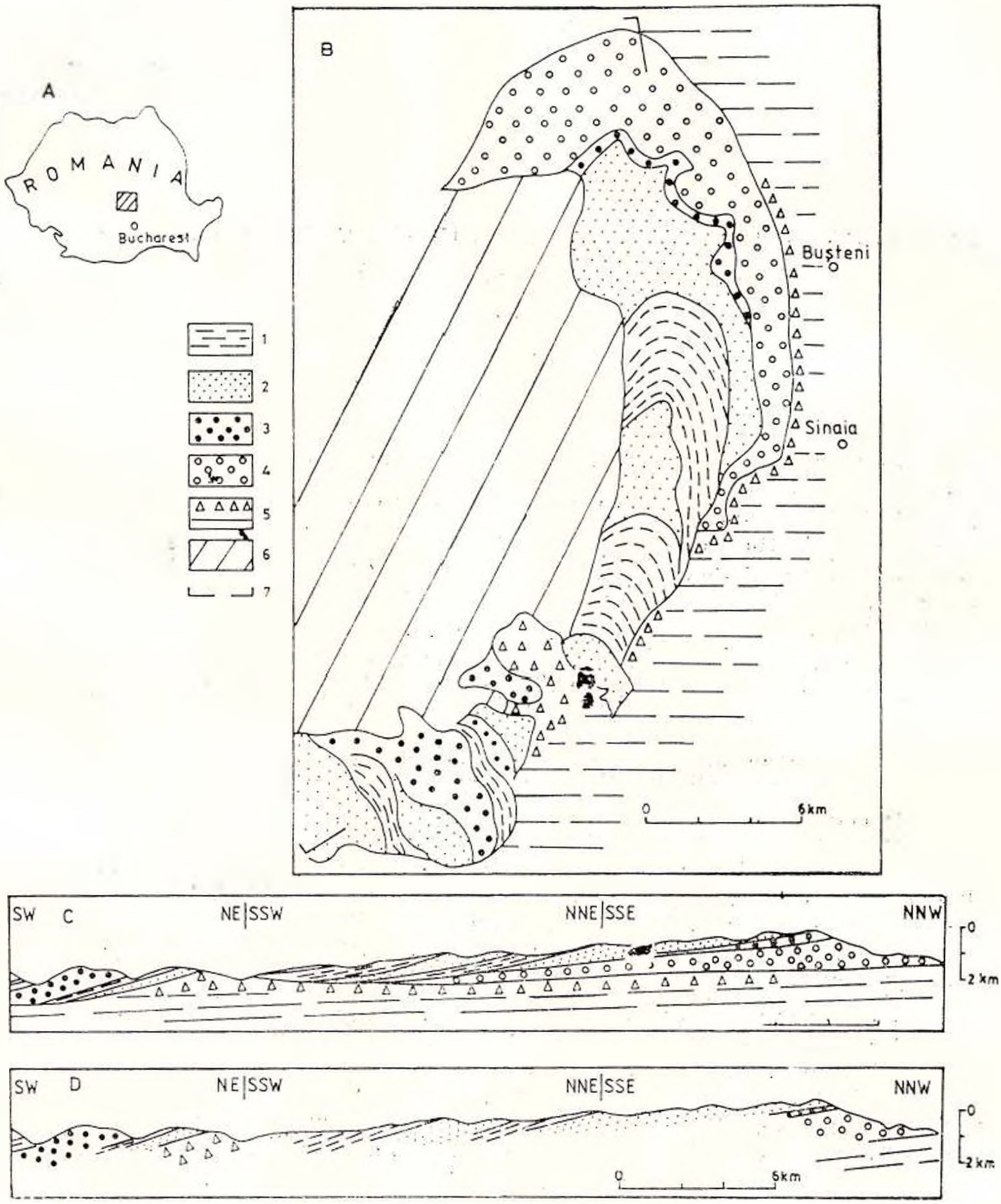


Fig. 1 — Broad primary structural features of the Bucegi Conglomerates. A, index map ; B, geological sketch of the Bucegi area (largely simplified, after Patrulius, 1969 ; Jipa, 1982).

Albian deposits : 1, siltites ; 2, sandstones ; 3, upper level conglomerates ; 4, lower level conglomerates ; 5, Aptian deposits (triangles = Raciu Breccia) ; 6, crystalline rocks ; 7, direction of geological section : C, longitudinal geological section : D, same section in the concept before progradation was documented.

same Upper Aptian level, materialized by the Raciu Breccia. Relying on these data, Jipa (1982) concluded that the Bucegi Conglomerates represent a very large, oblique bedded formation (about 40 km long and more than 1 km thick). Every lithologic Albian horizon consists of prograding deposits lying down on the inclined surface of the previously accumulated sediments, reaching beyond these sediments the basal accumulation surface (represented by the Aptian/Albian limit) (Fig. 1 C).

The documentation of the progradation structure of the Bucegi conglomeratic sequence resulted in the modification of the previous ideas concerning the thickness of the sequence. Before realizing the progradation action, the Albian deposits appeared as a simple monoclinal sequence (Fig. 1 D) ; consequently, its stratigraphic thickness amounted more than 8000 m. The real thickness of the prograded Albian formation is to be measured perpendicularly on the basal surface (that is the Aptian/Albian limit). In this way the thickness of the Bucegi conglomeratic complex is only 1200—1500 m.

Lower Red Clay Complex

The Paleogene deposits from the north-western Transylvania make up two rhythmical megasequences. Each sequence originates with red continental deposits, continues with evaporitic facies and ends with marine deposits (Răileanu, Saulea, 1956). The lower Red Clay Complex (Paleocene-Ypresian) consists of red lutites, arenites and rudites. The lower red deposits nearby the Agîrbiciu village (about 25 km west of Cluj ; Fig. 2 A) are dominantly coarse grained, made of rudaceous beds (up to 5 m thick) with red silt intercalations. The coarse grained red deposits at Agîrbiciu are overlain by a dolomitic oolite intercalation (Popescu, 1976) known as the Agîrbiciu Limestone.

The detailed survey of the Red Clay Complex at Agîrbiciu revealed that the rudite beds are oblique as compared with the almost horizontal Agîrbiciu Limestone (Fig. 2 B). Consequently, within the investigated area the lower Red Clay Complex represents a body (of about 2 km visible length) with large scale oblique stratification.

The existence of this oblique megastructure indicates that at least a part of the Paleocene-Ypresian continental deposits have been accumulated as alluvial fans, at the base of the marginal slope of the Transylvanian Basin (Fig. 2 D).

The detection of the progradation effects is also important from the structural viewpoint. It was assumed so far that the Paleogene stratigraphic units extended throughout the whole (or most of) the Transylvanian Basin (Fig. 2 E). The progradation structure suggests that the red Paleogene facies is restricted to the marginal area of the basin (Fig. 2 E).

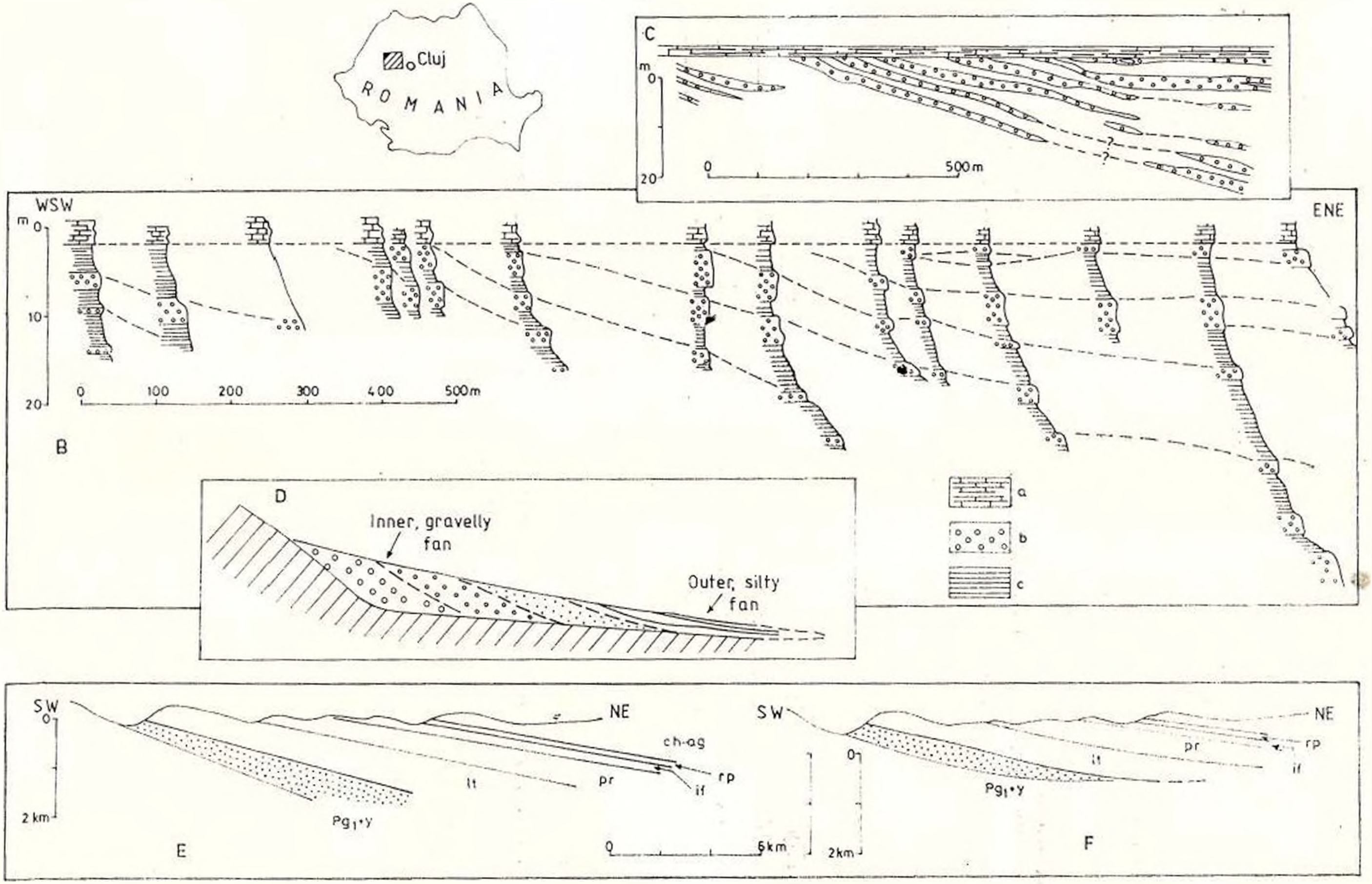


Fig. 2 — Primary structural features of the lower Red Clay Complex of north-western Transylvania. A, index map ; B, rudaceous beds of the lower Red Clay Complex in the Agîrbiciu area (a, Agîrbiciu Limestone ; b, red rudites ; c, red siltites) ; C, detailed cross-section at Agîrbiciu : D, interpreted accumulation environment of the red clay deposits ; E, simplified, classic geological cross-section through the Paleogene deposits of north-western Transylvania ; F, same section in the progradation concept of the lower Red Clay Complex (Pg_1-y).

Getic Paleogene Deposits

The Getic Depression is the term applied to the area with mostly Tertiary deposits located to the south of the eastern part of the South Carpathians (Fig. 3 A). The basal, conglomeratic facies of the Getic Eocene shows important thickness variations. Two large rudaceous bodies occur at the eastern and western extremities of the Getic Paleogene area. The thick body (700—800 m) in the eastern end is rapidly thinning out toward the west, to about 50 m of rudites. Farther westwards, another smaller (300—400 m thick), lens-like body of rudites is shaping out (Fig. 3 B). Between the two extreme, important rudite accumulations there is a zone where arenaceous sediments completely replace the rudites. In some well exposed and large enough outcrops, cross stratification at the scale of the whole rudite body was observed. Transport directions of the rudaceous material are dominantly towards south-east (Fig. 3 C). All these data indicate that the rudaceous Eocene material accumulated at the shelf margin as marine shallow water fans.

Within the concept of Paleogene deposition as sedimentary cones the basal conglomerate facies represents the inner fan zone, the overlying sandstone facies appears in the middle fan zone, the marly facies occurring in the outer fan and the fan plain zone (Fig. 3 D).

The occurrence of coarser gained deposits — for example the Oligocene Corbi Sandstone — within the marly Paleogene facies, with local distribution (Murgeanu, 1941), is also explained by the fan sedimentation concept. Such coarse grained deposits represent the clastic material accumulated on the distal slope of the fan (Fig. 3 E). The Corbi arenaceous body was laid down through only one channel distribution net, belonging to a single, active fan. This explains the local occurrence of the Corbi Sandstone and its wedge-like shape.

Hypothesis on Very Large Scale Progradation in the Romanian Carpathians

During their postgeosynclinal rising, in front of the Carpathians there accumulated large quantities of clastic sediments. According to the thickness determined by the usual stratigraphic standards, huge sedimentary piles — many kilometers thick — have been accumulated in very shallow sedimentary basins (several metres or tens of metres deep). Consequently, a conflict between the quantity of sediments versus the housing capacity of the sedimentary basins is clearly shaping out. To explain this conflict the subsidence phenomenon is invoked. An alternative,

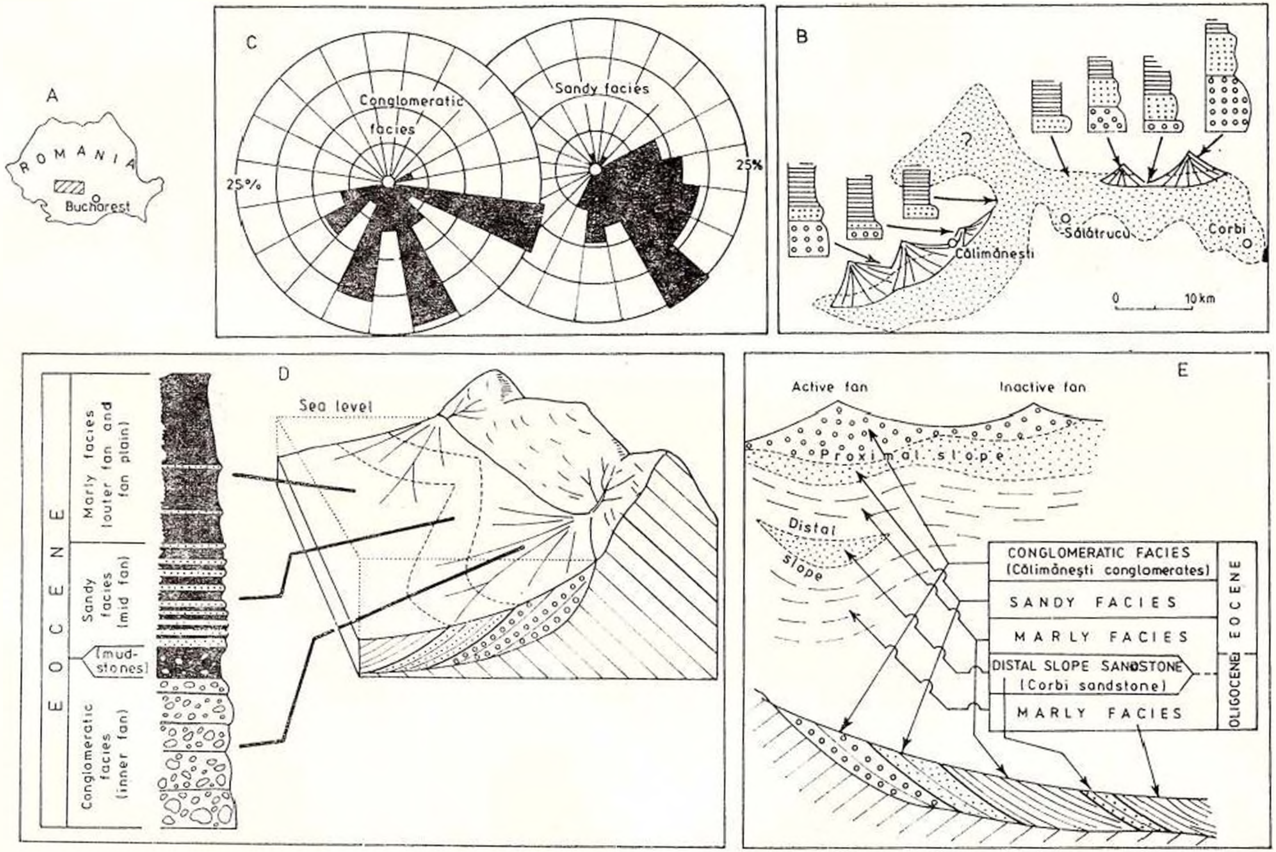


Fig. 3 — Primary structural features of Paleogene deposits (Getic Depression). A, index map ; B, thickness variation of the Eocene conglomerates and the presumed locations of the main rudaceous fans ; C, paleocurrent directions ; D, genetic interpretation of Getic Paleogene facies ; E, genetic interpretation of genesis and distribution of the Corbi Sandstone (Oligocene).

rational explanation is now accessible, taking into account the possibility of progradation action developed at the scale of the mountain chain. When progradation is demonstrated it leads to an entirely new concept on the geological structure of the investigated area.

The following two hypothetical large scale applications of progradation to explain the accumulation of some Carpathian sedimentary sequence are based on two different progradation mechanisms.

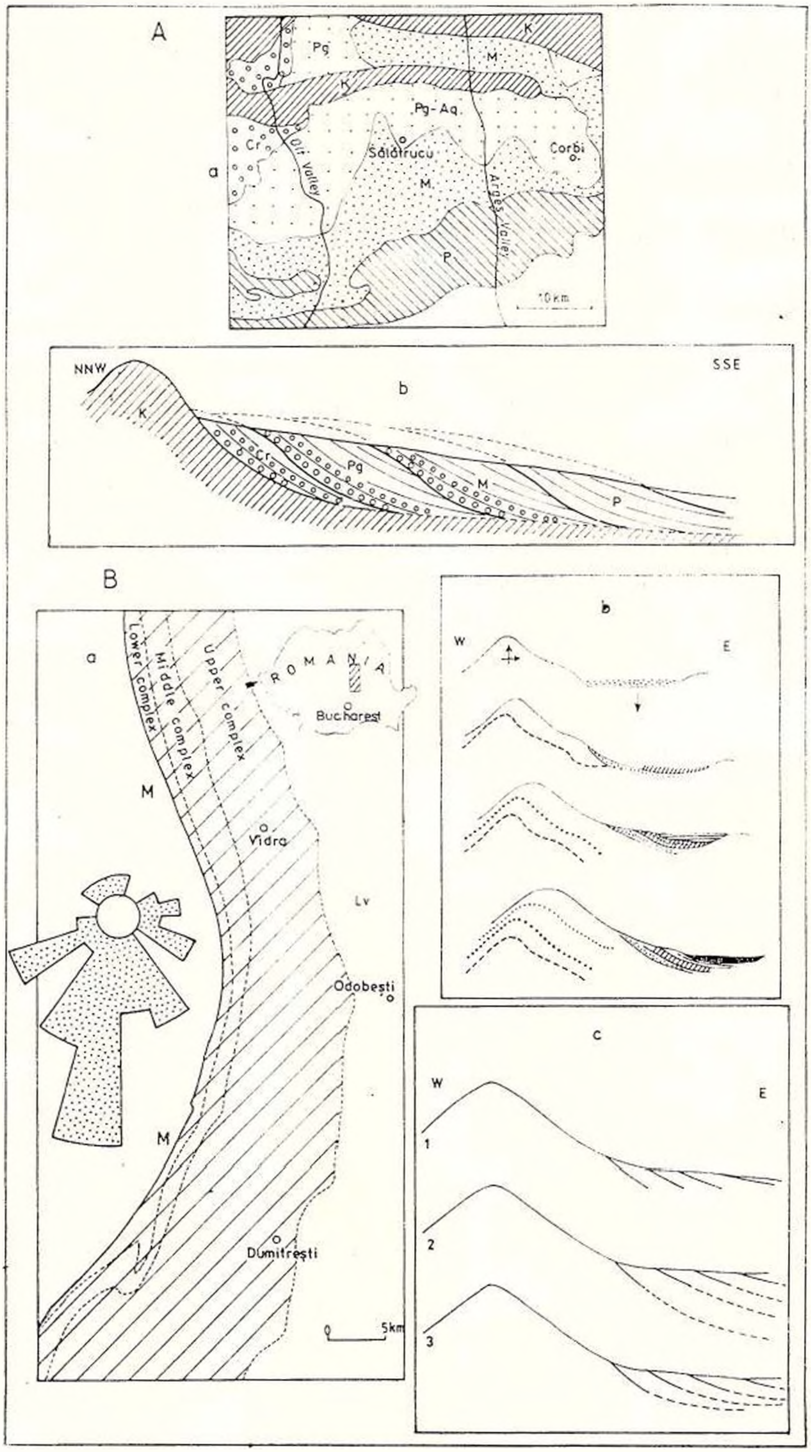
Sediments of the Getic Depression

According to the above presented arguments, the accumulation of the Paleogene deposits in the Getic Depression took place by progradation. The Getic Upper Cretaceous to Pliocene deposits make up — together with the Paleogene deposits — a homoclinal series dipping away from the crystalline core of the South Carpathians (Fig. 4 A). This feature suggests that progradation might have been active for all Getic, Cretaceous to Pliocene deposits. In such circumstances, the sediment accumulation front advanced laterally, away from the source area. Consequently, a constructional shelf area appeared (Fig. 4 A), producing the southward migration of the axis of the sediment acumulation zone. The current directions recorded in the Getic Cretaceous to Pliocene deposits are in agreement with the progradation direction.

Milcov Formation

Sarmatian-Pleistocene deposits occurring in front of the East Carpathians (Fig. 4 B), between the Trotuș and Buzău valleys, have been named the Milcov Formation (or Beds). This formation, constantly dipping eastwards, is involving the same conflict between its stratigraphic thickness (about 10 000 m) and depositional depth (several metres). It appears, then, that the accumulation of this formation could be analysed in connection with the progradation process.

The paleocurrent directions of the Milcov Formation are mostly longitudinal towards the south. This feature does not agree with the supposed eastern progradation of the formation. But the Sarmatian-Pleistocene sedimentation is synchronous with the very active uplift of the Carpathians. A horizontal, eastern component of the rising movement was for a long time considered (Lăzărescu, Dinu, 1983). The rising and lateral shifting of the Carpathian source area resulted in an eastward migration of the sedimentary basin, possibly leading to a progradation structure (Fig. 4 B). This is a special type of progradation, governed by the direct intervention of tectonic forces.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a
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Pg-Aq.
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Olt Valley
Sălătrucu
Corbi
M
Argeș Valley
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10 Km
NNW
SSE
b
K
Cr
Pg
M
P
B
a
Lower complex
Middle complex
Upper complex
ROMANIA
Bucharest
M
Vidra
Lv
Odobești
M
Dumitrești
0
5km
b
W
E
c
W
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1
2
3

Fig. 4

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Fig. 4 — Hypothetic progradation structure of some Carpathian sedimentary units. A, Getic Depression : a, geological sketch of the Getic Depression (K, crystalline basement ; Cr, Cretaceous ; Pg-Aq, Paleogene-Aquitanian ; M, Miocene ; P, Pliocene) ; b, idealized cross-section of the Getic Depression assuming progradation action (not to scale) ; B, Milcov Formation : a, geological sketch, after Macarovici et al., 1967 (M, Miocene ; hatced = Milcov Formation — Sarmatian to Lower Levantine ; Lv, Upper Levantine ; paleocurrent diagram of the Milcov Formation.) ; b, presumed mechanism of progradation determined by the rising and shifting of the East Carpathians (not to scale) ; c, interpretation of a symbolic cross-section of the Milcov Formation (1) in the classic way (2) and in the progradation concept (3).

A FLUVIAL SEDIMENTATION MODEL — THE DANUBE DELTA

BY

NICOLAE MIHĂILESCU[1], CONSTANTIN ROGOJINĂ[1]

The Danube Delta is bordered northwards by the Bugeac Platform, south-westwards by the North Dobrogea hills and south-eastwards and eastwards by the Black Sea. It has a surface of about 5460 km^2, of which 60—70% is covered by water.

The multiannual flow discharge value, at the delta top (Ceatal Izmail), is of 6 300 m^3/sec., while the maximum levels in May reach 10 000 m^3/sec. and the minimum levels in October lower to 3 250 m^3/sec. (Diaconu et al., 1963 ; Bondar, 1972). The flow discharge is distributed as follows : 62.5% along the Chilia branch and 37.5% along the Tulcea branch ; at Ceatal Sf. Gheorghe, it amounts to 22% for the Sf. Gheorghe branch and to 15.5% for the Sulina branch.

The Danube Delta is geographically and geologically a terminal plain, a result of the variation of the Black Sea level during the Postglacial age. The river, the sea, the wind and the vegetation exercised a combined action related to the variation of the Black Sea level, and the detrital deposits were set on a loess plain (Mihăilescu, Banu, 1957).

The Lower Quaternary deposits (Liteanu et al., 1961) of the delta basement are overplaced on a relief (Fig. 1 a) formed of an old hydrographic network (the Danube, Katlabug, Ialpug and Kitaj valleys). They are formed of gravels and boulders (Fig. 1 b), 0.70—37.70 m thick (Liteanu, Pricăjan, 1963).

The drillings have shown (Liteanu et al., 1961, 1963) the presence of loessoid deposits northwards (Stipoc) and southwards (Razelm Lake) the delta, over the gravels. In the central zone of the delta the loessoid deposits were subsequently eroded (Fig. 1 c). The loess of the northern zone of the delta (Chilia Veche) is a continuation of that from the Bugeac Platform and southwards the delta (Mihăilescu et al., 1984) the continuation of loessoid deposits from North Dobrogea. These deposits, of a continental origin, are those where Antipa (1912) noticed some fragments of *Elephas primigenius* Blumb. and *Rhinoceros antiquitatis* Blumb. (Sulina canal — Mile 12) in their basement.

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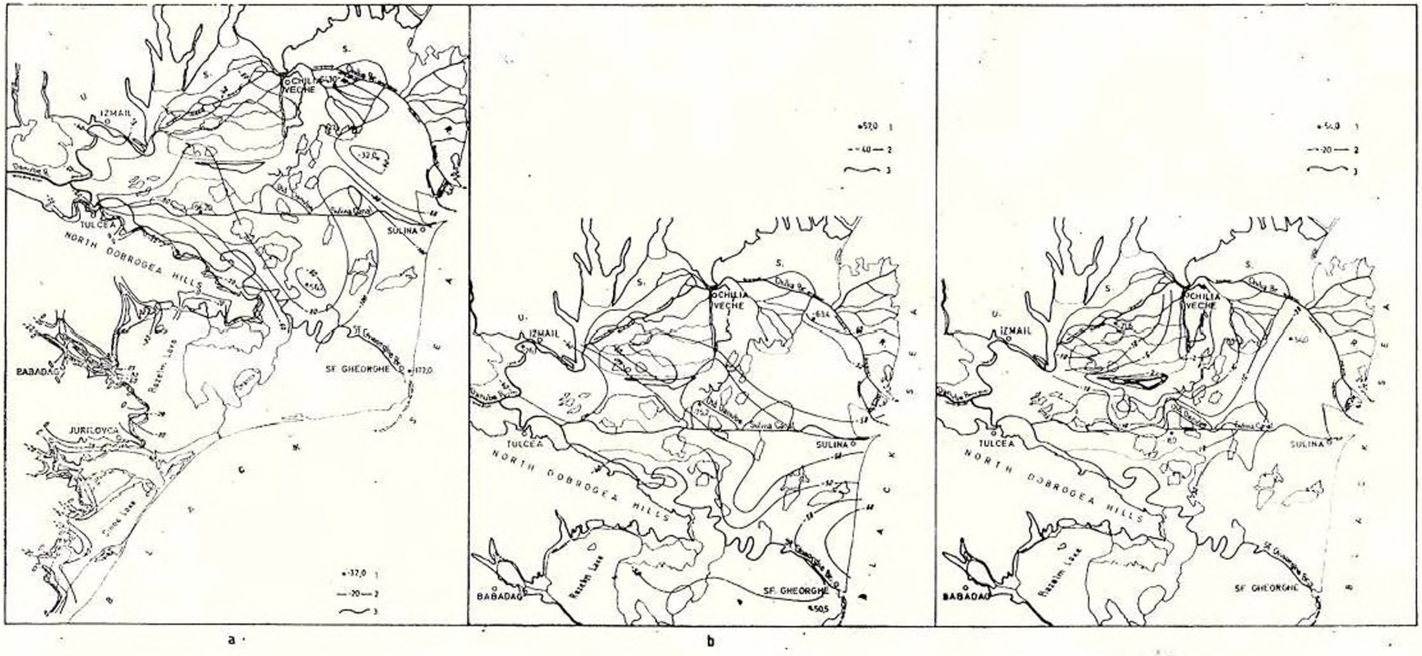

Fig. 1 — Structural maps of the Danube Delta (after Liteanu, Pricăjan, Baltac, 1961 ; Liteanu, Pricăjan, 1963 ; archives of the Ministry of Geology and IGCL, Tulcea).

a, pre-Quaternary basement ; b, Lower Pleistocene deposit top ; c, loessoid deposit top ; 1, limit of the Danube Delta territory ; 2, isobaths ; 3, deposit altitude.

The core samples prelevated from the lakes northwards the Sulina canal have shown the presence of loessoid deposits (Fig. 1 c) under the present-day delta structure (Antipa, 1912 ; Mihăilescu, 1984). These deposits contain rare terrestrial gastropods ; they are completely lacking in foraminifera and contain rare fresh water algae and spore granules similar to those from the loessoid deposits of the Moldavian Platform [2].

The deltaic deposits are represented by levels and lenses of sands and silts, in a succession with numerous lateral varieties of facies (Liteanu et al., 1961 ; Liteanu, Pricăjan, 1963). They were previously considered of the Upper Pleistocene (Mutihac, Bandrabur, 1967) or Holocene (Panin, 1974) age.

At the beginning of the Holocene age the "initial spit" was formed (Vâlsan, 1934). This spit separates two sedimentary domains : a fluvial-lacustrine domain westwards and a marine domain eastwards. Our available data (drillings, core samples) show that both marine and fluvial-lacustrine sediments overlie :

— coarse-grained deposits formed of gravels and boulders (mainly on the borders of North Dobrogea hills) ;

— loessoid deposits (northwards the delta and on the borders of the Razelm-Sinoe lakes).

Therefore, we think that the "initial spit" was not accidentally formed here, as it is based on loessoid deposits (Antipa, 1912 ; Murgoci, 1912 ; Mihăilescu, 1984). In places where the base was not "sufficient", this spit was pierced by the Sf. Gheorghe branch.

At the end of the Neoeuxin stage (Upper Pleistocene-Lower Holocene) the "cuspate" Sf. Gheorghe I Delta was formed (Panin, 1974). The Caraorman Formation represents the northern part of this delta (Panin, 1974) ; this formation overlies loessoid deposits (Mihăilescu et al., 1974). The southern part of this delta, in the Sf. Gheorghe branch-Razelm zone, is also superposed on loessoid deposits (Mihăilescu et al., 1984).

The Phanagorian regression, during which the Black Sea level went down to 3—4 m, had as a result the formation of the Sulina Delta structure (Panin, 1974).

At the beginning of the Histrian transgression (Bleahu, 1963), the flow and solid discharges of the Sf. Gheorghe branch grew and the Sf. Gheorghe II Delta (Panin, 1974) was formed. At the same time the Chilia branch started the formation of its own delta.

The present-day morpho-hydrographic features will be discussed in relation to the planar geometry of the meandering channels. Measurements of morphometric parameters were made on maps 1 : 25 000 and 1 : 50 000.

1. The behaviour of the meander curvature of the Sf. Gheorghe branch is compared to the harmony imposed by a "sinus-generated" curve (Leopold, Langbein, 1966). It is illustrated by the projection of the angular deviation of the small successive segments of the channel from the average direction of the meandering "valley" [3], in relation to the distance measured along the river bed (Fig. 2). The study of this diagram has pointed out the existence of four segments having characteristic configurations : a) between km 15—22 a limited meandering segment

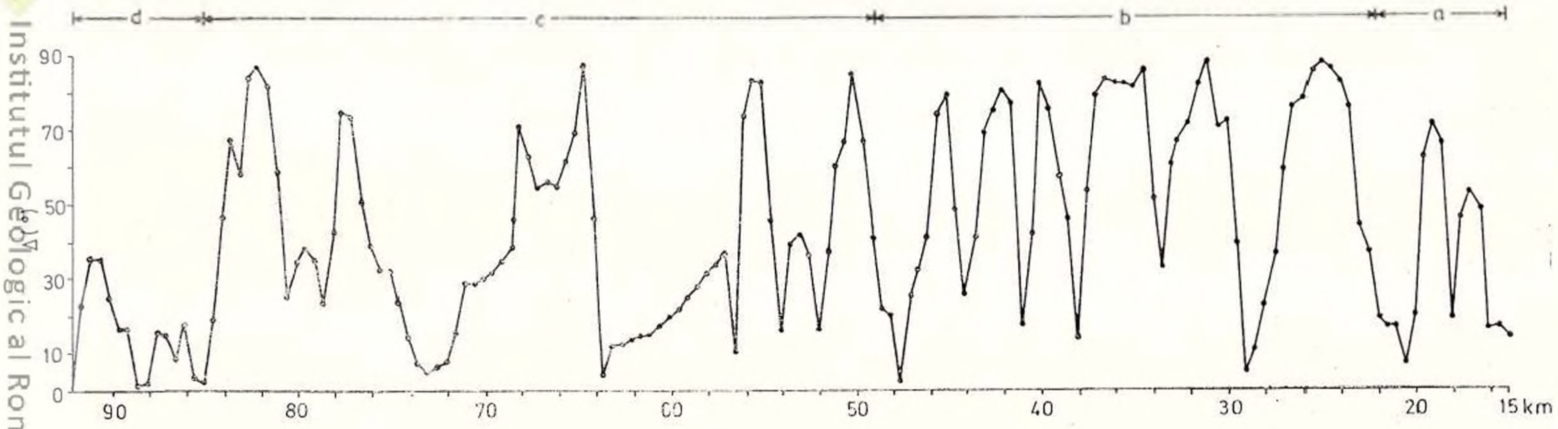

Fig. 2 — Angular deviation of the channel direction as compared to the average direction of the meandering valley (Δ) along the Sf. Gheorghe branch (km 16—92). Each point represents the angular deviation of a 500 m long segment along the channel axis.

(Panin, 1976) ; b) between km 22—49, a free meandering segment, in accordance with the present-day direction of the meandering "valley" ; c) between km 49—85, a free meandering segment, not following the present-day direction of the meandering "valley" ; d) between km 85—92, a limited meandering segment.

TABLE 1

Average morphometrical parameters and relationships among these parameters for different reaches of the Danube Delta

Reach	N	r_c (m)	w (m)	r_c/w	N	(m)	A (m)	λ/w	A/w	λ/r_c
Chilia (km 30—70)	12	1679	363	5.1	7	4282	803	11.6	2.2	2.7
Chilia (km 74—116)	12	1250	400	3.0	6	5204	1858	13.0	5.0	4.2
Paleo-Sulina	8	731	142	5.3	8	4897	1554	40.3	21.0	4.9
Tulcea	5	1260	420	2.8	3	3775	766	8.9	1.8	3.3
Sf. Gheorghe (km 25—44.5)	13	985	242	3.9	4	3005	2926	13.6	13.8	4.3
Sf. Gheorghe (km 44.8—84)	16	875	289	2.8	5	3315	2320	11.7	8.6	4.9
Sf. Gheorghe (km 23—84)	29	930	266	3.3	9	3160	2630	12.5	11.0	4.6

Symbols: N = number of measurements; r_c = radius of curvature; w = width of channel; λ = wavelength; A = wave amplitude.

The Sf. Gheorghe segment between km 15—92 is generally a free meandering channel, slightly limited at extremities. The following morphometric analysis (Tab. 1) will deal with this segment only [4].

Important significances have been attributed to the r_c/w relationship [5]. The values of this relationship are between 0.6—8.8 with a theoretically determined average of 2.7 and a statistically determined average of 3.3. These data are very similar to those presented by Leopold and Wolman (1960) for rivers meandered in non-deltaic environments.

The constancy of the λ/r relationship along the whole course of a meandering river determines the regular aspect of meandering, this relationship being quoted with an average value of 4.7 (Leopold, Wolman, 1960 ; Leopold, Langbein, 1966). For the meandering reaches of the Sf. Gheorghe branch, values from 3.1 to 7.3 were obtained with an average of 4.6 (Tab. 1).

The λ/w ratio has rather constant values for a large meander spectrum, being mentioned with average values of 6.6 (Inglis, 1949, in Leopold, Wolman, 1960 and 10.9 (Leopold, Wolman, 1960). For the Sf. Gheorghe branch meanders an average value of 12.5 (9.2—16.4) was obtained, with a lower significance as compared to the other relationships mentioned above.

The above analysis shows that the Sf. Gheorghe branch, with a freely meandered bed in a "valley" unrestricted by major geomorphological features can be characterized by a series of morphometrical parameters with well established relationships between them.

2. The Chilia branch has apparently the configuration of a large meandered channel. The sedimentological and morphological analyses show that only a few upstream reaches freely meandered. Relationships among morphometrical parameters for this meandered reach of the Chilia branch (km 74—116) are defined by the average values $r_c/w = 3.0$, $\lambda/w = 13.0$ and $\lambda/r_c = 4.2$. These values are very similar to those obtained for the Sf. Gheorghe branch (Tab. 1).

The Chilia branch reach between km 30—70 is characterized by the ratio $\lambda/w = 11.6$ (4.2—18.4). This value resembles the values of the meandered reaches of the Chilia branch (km 116—74) and the Sf. Gheorghe branch. The other parameter ratios have different values as compared to the meandered reaches values (Tab. 1). While comparing the data, one can notice that the r_c/w and λ/r_c ratios can characterize the freely meandered channels and can differentiate the freely meandering channels from the sinuous, apparently meandered channels.

Paleogeographically speaking, these data raise the question if this freely meandered segment of the Chilia branch (km 116—74) is not the ancient bed of a former river.

3. The Tulcea branch crosses a meander "valley" with a well developed divagation zone on the right side of the valley (Panin, 1976). The meandering evolution on the right side of the valley is limited by the lithological and morphological features of the geological unit of North Dobrogea.

The r_c/w ratio with an average of 2.8 is very similar to our values for the Sf. Gheorghe branch (Tab. 1). But meandering limitation influences both meander amplitude and wavelength as we can see while comparing the values λ/w, A/w and λ/r_c d of the Tulcea branch and of the Sf. Gheorghe branch (Tab. 1).

4. Meanders of the paleo-Sulina branch are located in a well limited divagation zone (Panin, 1976). The process of meander abandonment took place artificially during this century, following the rectification works of the Sulina branch, between 1868—1902 (Petrescu, 1957). By comparing the existing maps, one can notice that a filling up took place there, from upstream to downstream of the abandoned river channels. Now there are wholly or partly filled up river channels. The partly filled up beds have greatly diminished discharges, water flow being connected to the natural and artificial secondary channel network.

The morphometrical parameters mentioned below concern only partly abandoned meanders, which are still occupied by water. The r_c/w, λ/w and A/w ratios have much larger values than those of the same ratios of the Chilia (km 116—74), Sf. Gheorghe and even Tulcea branches (Tab. 1). These anomalous values are mainly imposed by the decrease of present-day channel width, by discharge decrease and by filling up. The $\lambda/r_c = 4.9$ ratio is very near to the values of meandered reaches in the Danube Delta (Tab. 1, 2), and proves to be less affected by recent modifications of the hydraulic geometry.

The morphometric characters of the abandoned meanders were presented by Panin (1976). For the Erenciuc Lake, our measurements have shown that the r_c /w ratio varies between 2.1—5.3, with an average of 2.8. Despite recent modifications of hydraulic geometry, the r_c /w value is well correlated with the values obtained for the Sf. Gheorghe branch (Tab. 1).

5. The field reaches and the interpretation of maps have pointed out the existence of some secondary channels which seem to represent meandering beds (Fig. 3 a).

reach

TABLE 2

Average morphometrical parameters and the λ/r_c ratio for the main secondary channels of the Danube Delta

Channel	N	r_c (m)	N	(m)	A (m)	λ/r_c interval	λ/r_c average
** Pardina	37	713	17	839	184	1.5—7.6	3.2
** Tătaru	12	765	7	2021	312	2.0—5.2	3.1
** Gotca	14	432	8	1019	186	1.8—5.7	3.0
*** Păpădia-Mitchina	22	246	14	754	204	1.9—5.5	3.5
* Sontea	18	234	9	891	231	2.6—8.4	4.4
* Lopatna	60	170	29	514	162	1.8—5.8	3.8
*** Litcov—Perivolovca	25	442	11	1380	315	2.3—8.1	3.4
* Dunavăţ	34	301	20	1423	882	2.8—7.8	5.1
* Dranov	17	281	7	1121	444	2.4—6.2	4.2
*** Zencova	9	238	5	1135	402	3.4—8.0	6.0

* typical meandering channels; ** sinuous channels; *** partly meandering channels
The other symbols are the same as in Table 1.

The correlation diagram λ/r_c for these channels (Fig. 3 b) shows an increase of point scattering, according to the increase of wavelength (for λ/2 200 m). Between the λ and r_c values of a large number of rivers, some approximately linear relationships were noticed (Leopold, Wolman, 1960). The best-fit lines of λ/r_c values, together with field observations (Fig. 3 a) have led to the separation of three types of secondary channels :

a — Typical meandering channels : Dunavăţ, Dranov, Şontea, Lopatna (Fig. 3 b). These river beds have a good apparent correlation of values λ/r_c = 3.8—5.1. Small zones of meander divagation are preserved within adjacent areas of river beds ;

b — sinuous channels : Pardina, Tătaru, Gotca. The λ/r_c values have a large dispersion (Tab. 2), and the average values of these ratios are between 3.0 and 3.2, similar to the value of λ/r_c = 2.7 of the sinuous reach of the Chilia branch (km 30—70) (Pl.). Developed levees and crevasse-splay deposits are present in the adjacent areas where divagation zones are absent. These channels belong to the distributary system of the Chilia branch, and are grouped within the depression between the Stipoc Formation and the Chilia branch ;

c — partly meandered channels : Păpădia-Mitchina, Litcov-Perivolovca, Zencova (Pl.). Dispersion of λ/r_c values (Fig. 3 a) and the average values of this ratio (Tab. 2) have an intermediary character as compared to the first two groups. On certain reaches of these channels restricted divagation zones have been preserved.

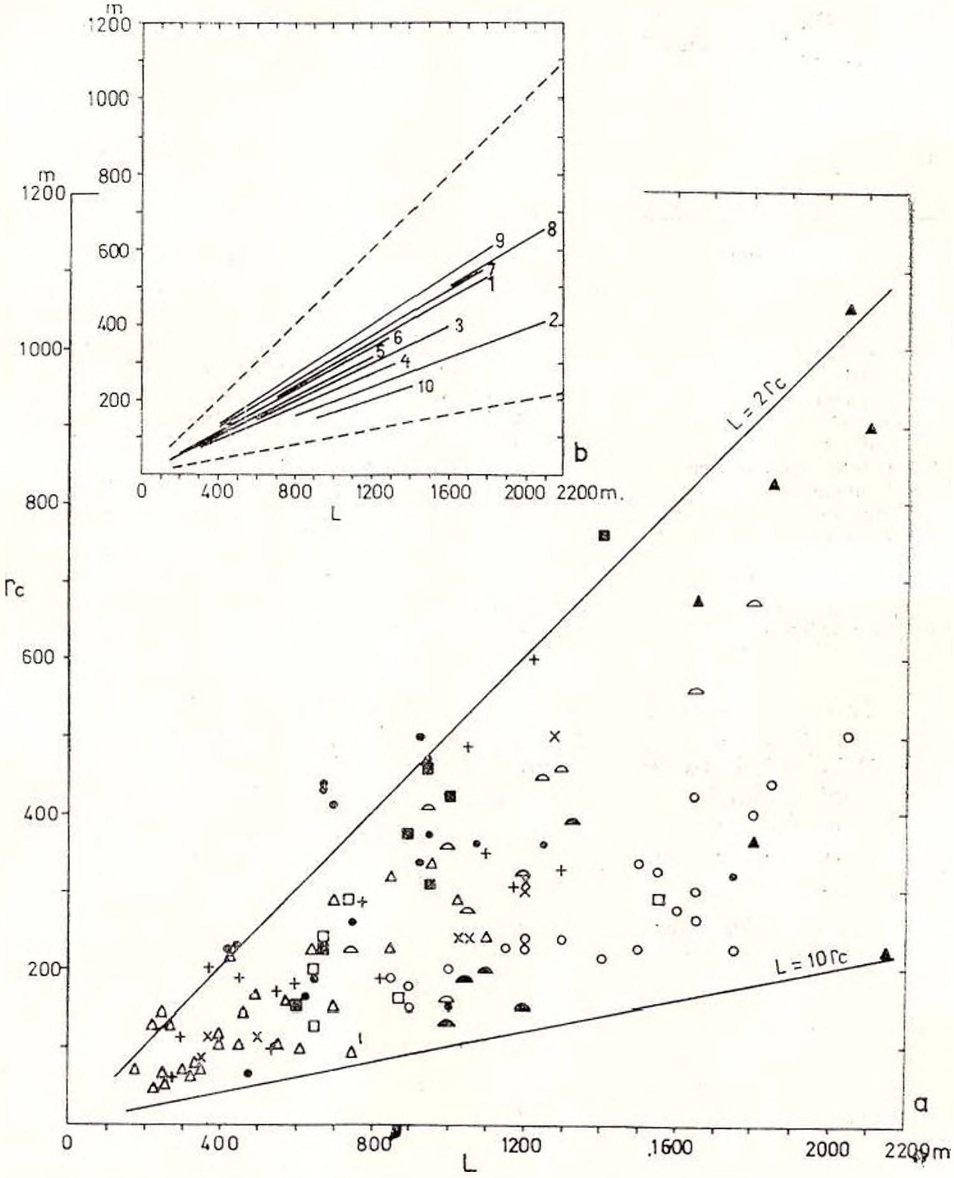

Fig. 3 — a, λ/r_c diagram for secondary channels of the Danube Delta ; b, best-fit lines of the λ/r_c values for secondary channels of the Danube Delta ; dashed lines delimit the field of point scattering.

6. The depressionary zones closed by the "initial spit" changed into flood basins. Subsequently they were divided by the levees of the forming hydrographic network. The lacustrine sedimentation of the Danube Delta is located within zones which "become" depressionary due to enclosures made by levees or by fossil littoral belts.

The lacustrine sedimentation is controlled on the one hand by the solid discharge favoured by the main branches of the Danube and on the other hand by the composition of the lake basement. The lacustrine sediments of the delta are most of all the result of some processes of removing and redeposition of detritus from their basement and of some diagenesis processes, connected to the chemical composition of the older marine, brackish or continental deposits.

Conclusions

At the beginning of the Pleistocene, both the Danube and the Ialpug, Katlabug and Kitaj were forming a hydrographic network which concentrated towards a central depressionary zone (Fig. 1 a).

During the Pleistocene, the lowering of the base level of those rivers flowing into the NW of the Black Sea allowed the activation of this hydrographic network, having as a result the display of a gravel nappe (Fig. 1 b). A loessoid deposit cover was formed in parallel and covered the whole NW area of the Black Sea border.

During the Holocene, several partly superposed deltaic structures were formed on the present-day area of the Danube Delta, due to the repeated variation of the Black Sea level (Degens, Ross, ed., 1974).

During the Holocene time, a WNW—ESE oriented hydrographic network was formed on the relief formed of loessoid deposits. This is located between the North Dobrogea hills and the Stipoc-Chilia Veche zone (Fig. 1 c). In the meantime, on the eastern border of the loessoid deposits the "initial spit" started to form, having behind a fluvio-lacustrine sedimentary environment.

The hydrographic network of the fluvio-lacustrine delta was drained by a river bed which, at the same time with the formation of the Sf. Gheorghe I Delta, was going to get the orientation of the present-day Sf. Gheorghe "valley" (Pl. I). In this period a rotation took place in the direction of the Sf. Gheorghe branch "valley", accompanied by a partial, gradual avulsion of the former river beds. The Dunavăț, Dranov and Zencova channels can stand for the possible stages of this migration (Pl.).

During the Phanagorian regression, when the sea level lowered, the Sulina Delta formed. The piercing of the "initial spit" and the modification registered in water flow distribution at the delta top (as a result of rotation in direction of the Sf. Gheorghe branch "valley"), have favoured the transformation of a secondary channel (paleo-Sulina) into a main branch of the Danube.

As compared to the formation time of the Chilia Delta, the Chilia branch proves to have functioned as a main distributive branch of the Danube, after the formation of the Sf. Gheorghe and Sulina branches. The river bed configuration, the morphometric data and the morphostructural characteristics of the adjacent zones suggest that this branch is made of reaches which were formed during stages and at different times.

The Danube gulf "became" a gulf only at the end of the Pleistocene by the inundation by the sea of an older hydrographic network. This gulf was dammed by a first marine formation behind which a vast fluvio-lacustrine zone was formed.

This fluvio-lacustrine zone developed in parallel to the evolution of the marine zone of the delta front. The main delta channels have divided the primary lake, have compartimented it, while the fluvial deposits (levees, accretion zones, divagation zones) have formed some emerged regions and the secondary channels have strengthened this process.

The configuration of the present-day depressionary zones of the Danube Delta (Pl.) is outlined by several morphological elements of Upper Pleistocene (loessoid deposits) and Holocene ages :

— relief of loessoid deposits (Chilia Veche, Stipoc) ;

— relief of divagation zones of the Sf. Gheorghe, Sulina and Chilia branches ;

— relief of fossil littoral bars which belong to various formations in time and

— relief of present-day littoral bars.

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[2] Samples determined by St. Roman (I.G.G.).

[3] The divagation zone of the Sf. Gheorghe branch (Panin, 1976) is considered in this case as a meander "valley".

[4] Brice (1974) found out that the planar evolution of meanders is generally reduced to a descendent migration, an amplitude growth and a "cut-off". These modifications are well characterized by the following morphometrical parameters : radius of curvature (r_c), bed width (w), wavelength (λ) and meander amplitude (A) (as defined by Leopold, Langbein, 1966, with specifications made by Hickin, 1974, and Brice, 1974).

[5] The relative constancy of the r_c /w ratio for all meandered rivers is considered to make them seem very similar on maps. On the other hand, Hickin (1974) and Hickin and Nanson (1975) have shown that the evolution and the migration ratio of these meanders are controlled by certain values of the r_c /w ratio.

[6] Relationships among morphometrical parameters have been defined by equations of $P = kR^{\alpha}$ type, where P is λ or A, R is w or r_c and k is a constant. As the α exponent varies between 0.98—1.10 in all equations (Leopold, Wolman, 1960), it could be approximated as unitary and thus relationships become linear (Fig. 3 b).

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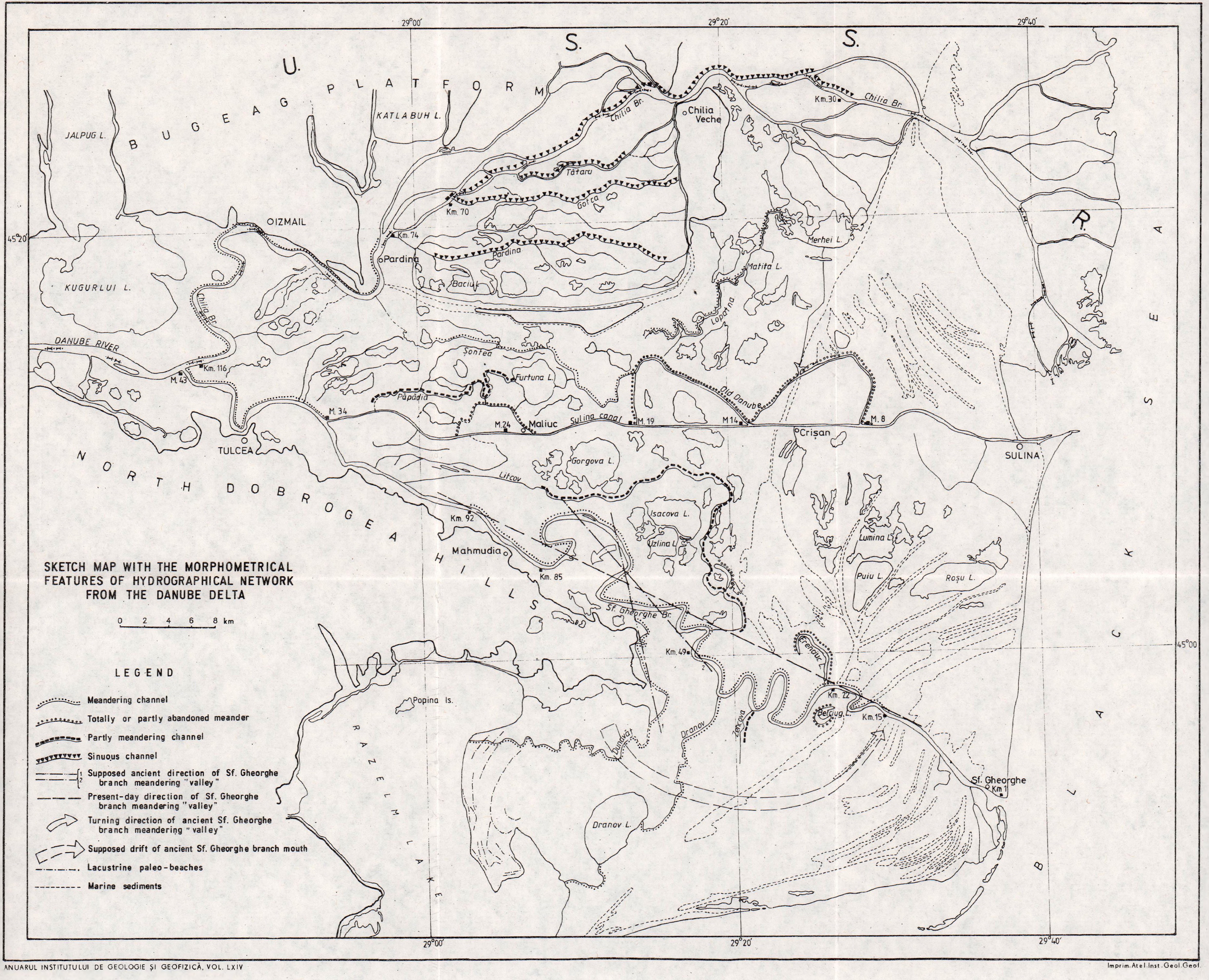

SKETCH MAP WITH THE MORPHOMETRICAL FEATURES OF HYDROGRAPHICAL NETWORK FROM THE DANUBE DELTA

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