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Cover illustration: Veined rock from the stockwork mineralization at Roșia Poieni. Picture taken by Anne-Sylvie André-Mayer.



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Fissural permeability in the Roşia Poieni copper deposit: Influence on ore repartition at the open pit scale

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Abstract Fracture permeability is a key parameter in fluid flow through the crust. In a hydrothermal system, fluid flow is dominated by the vein network. A quantification of paleopermeability has been performed at the scale of the hydrothermal system of Roşia Poieni, a Romanian porphyry copper. A suitable method for the quantification of permeability in this type of hydrothermal environment is used based on a discontinuous environment, which consists of the rock as a homogeneous medium (matrix) + N different elements (fractures) that can be described by structural analysis.

Paleopermeability heterogeneities have been observed: lower levels of the intrusion are characterized by higher paleopermeabilities generated by thick and clustered veins, whereas higher levels are characterized by a more pervasive paleopermeability with thin veins that have a homogeneous spatial repartition. At each level of the open pit, specific zones can be defined in relation to the density and spacing of veins, leading to variations in paleopermeability that are correlated with two types of mineral filling: (1) quartz-pyrite and (2) copper minerals.

Keywords: Porphyry copper, fissural permeability, fluid flow, mineralization

1. Introduction

Fracture permeability is a key parameter in the understanding of fluid flow through the crust. The fractures, which may present a great extent and width, act as preferential channels in rocks of low permeability and thus control both the intensity and the direction of the fluid flow. The high permeabilities described in hydrothermal systems are directly produced by a fracture permeability that is much greater than the low matrix permeability of the host rocks, which ranges from 10^{-4} to 10^{-17} darcy in crystalline rocks (Brace et al., 1968) and from 10^{-2} to 10^{-11} darcy in sedimentary rocks (Norton and Knight, 1977; Curewitz and Karson, 1997). Therefore, in a hydrothermal environment, permeability must be estimated by modeling an equivalent porous matrix crosscut by anisotropic 3D fractures that concentrate fluid flows.

Following Scheidegger (1974), the modeling of rock permeability can be performed using two main types of models:

Geometrical models use the concept of equivalent porous media and require the definition of a size parameter that is defined by an equivalent hydraulic radius (Walsh, 1981; Brace et al., 1968; Gangi, 1978; Brown, 1987). This approach requires the geometrical characterization of the main drains (extents and widths) and pore sizes. Each object, pore, and fracture is represented in the model and acts as an individual source of permeability.

Statistical Models are based on statistical laws that describe the geometrical characteristics of the pore and fracture networks. No direct or deterministic representations of pores and fractures are used (Guéguen and Dienes, 1989). When precise observations and measures are sufficient to describe the fracture network geometry, geometrical models are better suited

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for characterizing the permeability developed by a fracture network in a massive rock. Within these geometrical models, two types of porous media can be distinguished:

The first model offers a continuous approach, which maintains that the rock porosity is defined by a single homogeneous medium constituted by a matrix that is crosscut by dense fractures. A homogenous porous medium is defined by a mean porosity and permeability that integrate both the pore matrix and fracture hydraulic contributions (Snow, 1969; Oda, 1986; Ababou, 1991; Vuillod, 1995).

The second approach defines a discontinuous media and maintains that fractures are individual channels characterized by a specific porosity and permeability that differ from those of the fracture matrix. This approach requires a precise structural analysis and a deterministic characterization of the fracture network. The identification of discrete fractures in the rocks allows the assessment of the flow anisotropy and the definition of a permeability tensor. Permeability is related to the geometrical parameters of fractures, such as the direction, density, spacing, lengths and widths. Therefore, this approach is more appropriate for characterizing the paleopermeability of a vein system when its geometrical characteristics can be described by a field study or drillhole logging.

The fracture permeability can therefore be quantified using these geometrical parameters, and several different reports in the literature use a classical “cubic law” of width approach (Snow, 1969; Oda, 1986; Ababou, 1991; Vuillod, 1995). For example, equation 1 determines the conductivity C_m of the fracture set m and relates this conductivity to the mean spacing S_m and mean cubic width w_m of the fracture set, which is represented by k fractures. C_m (units of m/s) is the maximum conductivity because the fracture set is assumed to be infinite in this model.

$$C_m = \frac{g}{12 \cdot \nu} \cdot \frac{w_m^3}{S_m} \quad (1)$$

$$\text{with } S_m = \frac{1}{N_m} * \sum_{K=1}^{N_m} S_{m,k} \quad (2)$$

$$\text{and } w_m = \frac{1}{N_m} * \sum_{k=1}^{N_m} w_{m,k}^3 \quad (3)$$

where g : acceleration due to gravity [$m.s^{-2}$]

ν : kinematic viscosity of the fluid [$m^2.s^{-1}$].

w_m : mean width of fracture set [m]

$w_{m,k}$: thickness of the k^{th} fracture [m].

S_m : mean spacing of the fracture set m [m]

$S_{m,k}$: spacing between fractures $k-1$ and $k+1$ [m]

N_m : number of fractures in the fracture set m

The fracture set conductivity C_m is related to the permeability K_m (m^2) by the following relation:

$$K_m = C_m \cdot \frac{\mu}{\rho \cdot g} \quad (4)$$

where μ is the fluid dynamic viscosity, ρ is the fluid density [kg/m^3] and g is the acceleration due to gravity [m/s^2]

The permeability quantification method is based on equation 1, which was proposed by Vuillod (1995). This method uses a cubic law of fracture width and the spacing parameter between fractures within a fracture set. Indeed, even if large and well-developed drains exist within the rock matrix, the final equivalent permeability of the rocks is not always important. This model uses the spacing parameter as a weight of the permeability values and then defines a hydraulic influence of each fracture of width w (Fig. 1).

This work will discuss the use of a geometrical model to characterize the paleopermeability in a hydrothermal system. The model has been applied on a porphyry copper deposit (Roşia Poieni, Romania), and this paper will discuss the correlation between the estimated permeability (and its variation at the open pit scale) and the spatial distribution of the copper mineralization.

2. Geological setting

The southern part of the Apuseni Mountains (West Romania) exhibits a variety of mineral occurrences (Cioflica et al., 1973; Ianovici et al., 1977; Borcos et al., 1983; Udubasa et al., 1992; Heinrich and Neubauer, 2002) that are related to Neogene volcanic activity (Lemne et al., 1983; Pécsay et al., 1995; Rosu et al., 1997; Alderton et al., 1998). The Roşia Poieni porphyry copper deposit, which is the largest such deposit in the Apuseni Mountains, belongs to the metallogenic district of Bucium-Baia de Arieş (Fig. 2). The basement consists of pre-Mesozoic metamorphic rocks intruded by magmatic bodies and covered by pyroclastics of variable composition (Fig. 2).

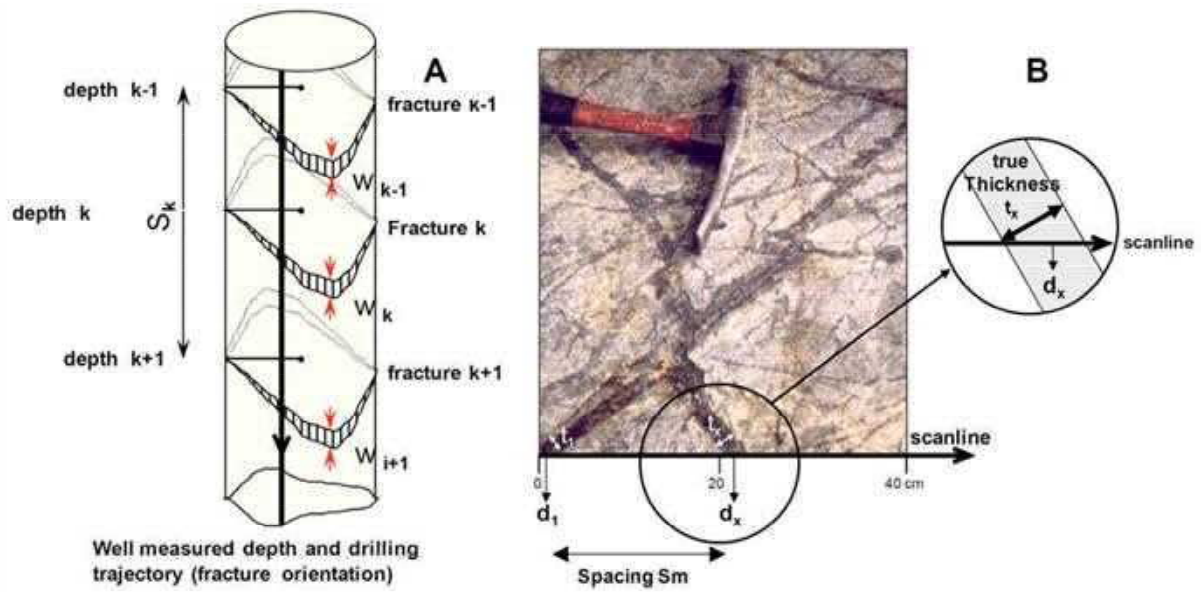


Fig. 1. A. Geometrical parameters describing a fracture set m that is determined by structural studies and that can be used for the quantification of the equivalent fracture conductivity (from Vuillod, 1995). S_k represents the mean spacing between fractures ($k-1$ and $k+1$) within the fracture set m . B. Stockwork organization and methodology of sampling along a scanline in an open pit: distance from the beginning of the scanline ($D_1, \dots, D_x, \dots$), true thickness ($t_1, \dots, t_x, \dots$), strike, dip and mineral assemblage of each vein are plotted.

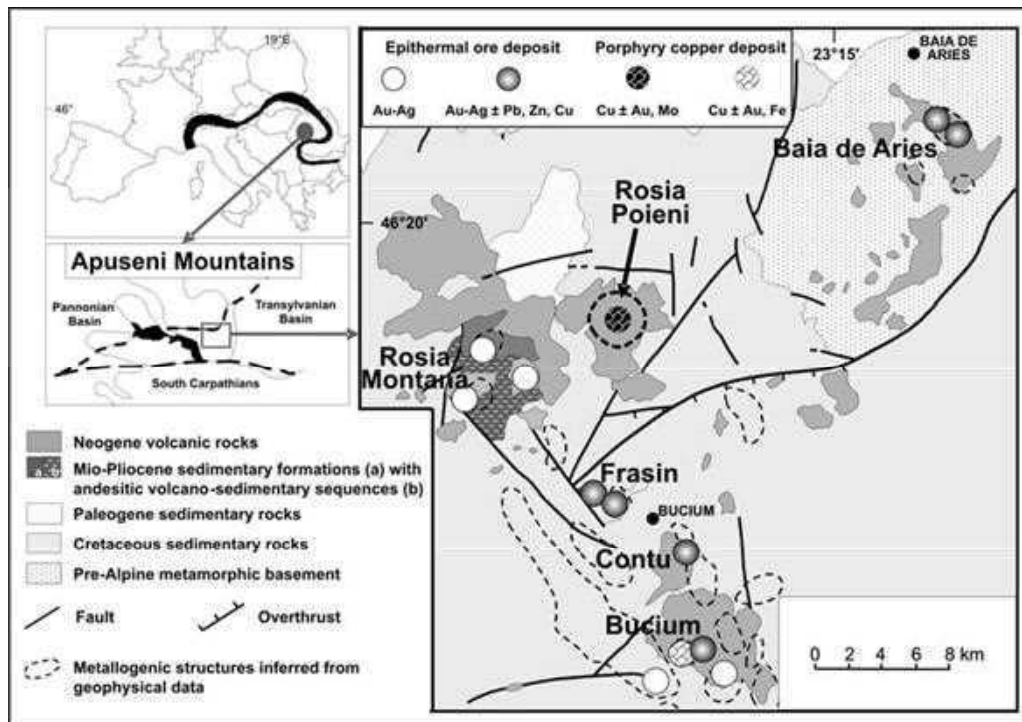


Fig. 2. Geological map of the Bucium-Baia de Arieş metallogenic district and its location in the Apuseni Mountains, Romania (modified from Borcoş et al., 1983).

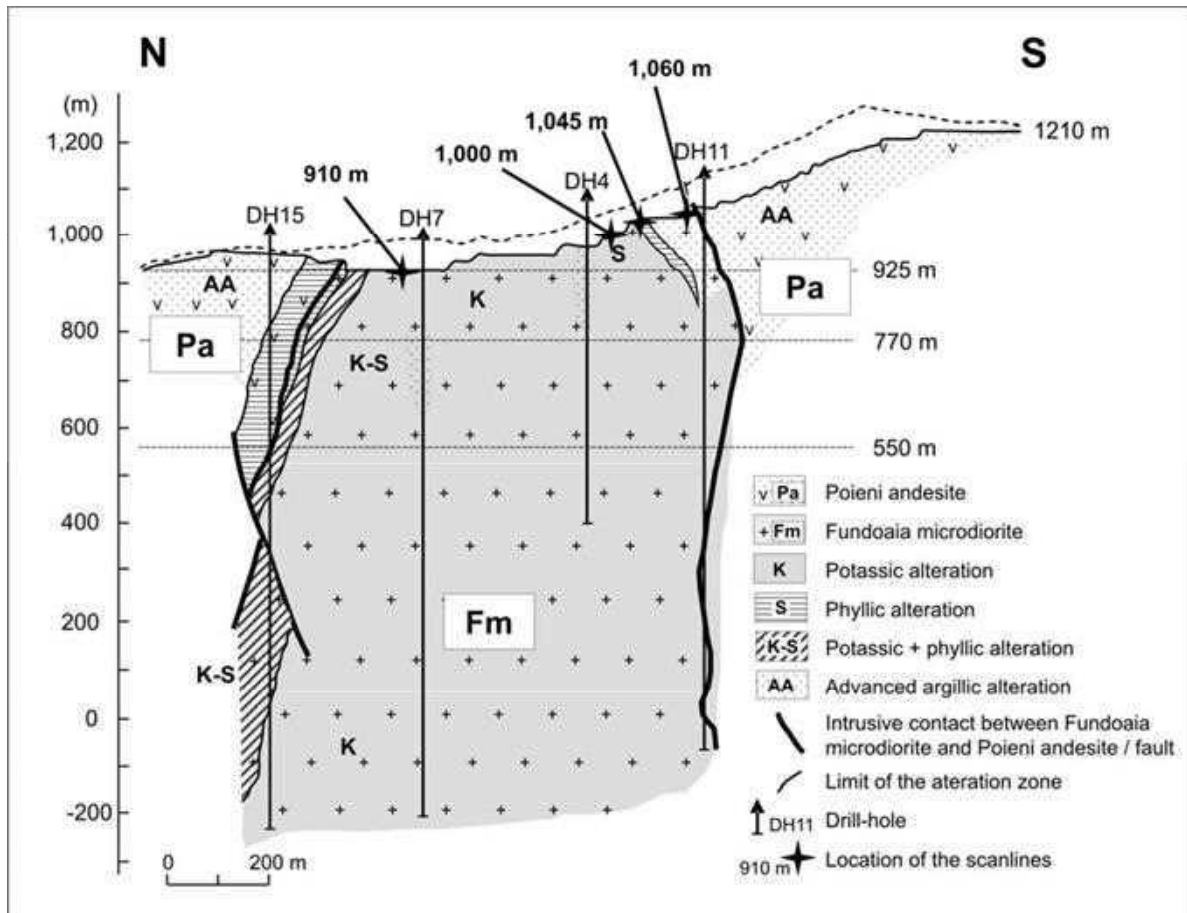


Fig. 3. North-south cross-section of the actual Roșia Poieni open pit (reconstructed from Ionescu, 1974 and Milu, 2004) showing the porphyritic intrusion and host rocks, alteration types and zones. The location of the four scanlines (900, 1,000, 1,045 and 1,060 m) is indicated with stars.

Two types of igneous rocks, both characterized by intense hydrothermal alteration, are exposed in the open pit. The host rocks (the Poieni andesite), are intruded by a subvolcanic body characterized by a porphyritic texture, the Fundoia microdiorite (Fig. 3). Ionescu (1974), Ionescu et al. (1975) and Boștinescu (1984) performed a mineralogical and petrographical study of the rocks from the Roșia Poieni area. They mentioned the disseminated character of the copper mineralization and a zoned pattern of. Recent studies (Milu, 1999; Milu et al., 2004) have shown that the Fundoia subvolcanic body and the Poieni host rocks are characterized by a strong hydrothermal alteration that is spatially related to the Fundoia intrusion.

Fig. 3 presents a schematic reconstruction of the Roșia Poieni volcanic structure and shows the concentric pattern of the hydrothermal alteration from the potassic core to a well-developed advanced argillic zone (Milu et al., 2004). The main high-grade zone ($>0.6\%$ Cu) is located in

the central part of the intrusion (Milu et al., 2004). Outside the potassic zone, the copper grade decreases to 0.1% Cu, except in zones where a phyllic alteration is superimposed on a potassic alteration (Fig. 2). The upper part of the Roșia Poieni structure displays an advanced argillic alteration and is barren ($<0.05\%$ Cu). Gold occurrences greater than 0.02 ppm Au (lower limit of detection) are restricted to the potassic ($\pm$ phyllic) zones and do not exceed a mean value of 0.5 g/t Au (Milu, 1999; 2004). The open pit is now on standby, but the reserves are estimated to be 350 Mt, with an ore grade of 0.36% Cu and 0.29 g/t Au and 0.02 g/t Mo (Borcos et al., 1998). The evolution of the geometrical characteristics of the mineralized stockwork has been described by André-Mayer and Sausse (2007).

The mineralization comprises pyrite, chalcopyrite, magnetite, hematite, molybdenite and bornite with subordinate tetrahedrite-tennantite, enargite and digenite and minor amounts of

pyrrhotite, sphalerite, galena, covellite and chalcocite. Chalcopyrite is the dominant hypogene copper mineral. The mineralization is pervasively disseminated within the altered rock, veins and fractures associated in a well-developed stockwork (Figs. 2 and 3) that is best developed in the potassic zone.

3. Data collection

The present vertical extent of the open pit exposure is 300 m (between altitudes of 910 and 1,210 m; Fig. 3). Four horizontal sampling lines were selected in the open pit at different altitudes (910, 1,000, 1,045 and 1,060 m, Fig. 3). In most mineral exploration situations (trenches, boreholes, drifts, etc.), only this type of 1D sampling is possible despite the 3D organization of both individual veins and networks (Park and West, 2002; Peacock et al.,

2003). Moreover, existing methods allowing a 2D data analysis are still immature, and there is currently no satisfactory method to analyze 3D spatial data (Gillespie et al., 1993; Pickering et al., 1995).

Three of the sampling lines are oriented in an E-W direction (altitudes 910, 1,000 and 1,060 m), whereas the fourth is oriented N-S (altitude 1,045 m, Fig. 3). The 1,060 m scanline is located close to the contact between the intrusive body of Fundoaia and the host andesite of Poieni, which is marked by the onset of the argillic alteration (Fig. 3). For each scanline and each vein, the intersection distance between veins and the spacing between the vein sets and the scanline were measured, and information, such as the width, strike, dip and vein mineral assemblage, was collected (Fig. 1 and Table 1). Vein counts range from 146 to 725 over lines of lengths from 7.9 to 14.4 m, leading to a database of approximately 1,700 fractures (Table 1).

Table 1. Database acquired on the Roşia Poieni stockwork with indications of length, number of measurements, and vein density for each scanline. The 1,000 and 1,060 m scanlines are not continuous because of access difficulties, screes or strong weathering and are thus respectively divided in four and two sections.

Scanline	Section	Length (m)	Number of data		Density N/m (m ⁻¹)
910 m	-	12.7	353		27.5
1,000 m	1	11.3	213	725	18.8
	2	14.4	205		14.2
	3	10.2	219		21.5
	4	7.9	88		11.2
1,045 m	-	11.0	146		13.3
1,060 m	1	13.2	210	468	15.9
	2	8.1	258		31.7

4. Fracture permeability quantification

The study of the fracture permeability based on this type of 1D profile leads to an extrapolation problem because of the visual limitations of the true geometry of the fractures, excluding the reconstitution of the 3D fracture network. Therefore, to model the hydraulic behavior of the fractured media requires the use of equivalent porosity and permeability models cha-

racterizing the fractured rock matrix. The assumption of a negligible matrix, that is crosscutted by a permeable fracture, allows the quantification of local permeability that is directly related to the width of each fracture and to the spacing between fractures.

The permeabilities were first quantified by accounting for the totality of the fractures for a given profile. This method does not consider the chronology of the tectonic events (opening and

reopening of fractures) or the mineralogical filling of each vein, but it does allow the visualization of the evolution of the global permeability distributions at the open pit scale. For each profile, the equivalent hydraulic properties have been quantified using equations 1 to 3, as presented above.

4.1. Evolution of the permeability in the open pit

The mean values of individual directional conductivity distribution developed for each vein and quantified for each profile are shown in Fig. 4. These values are relatively high because the geometrical width measured for the open pit is generally higher than, but proportional to the real hydraulic width of the veins, which is entered as cubic input parameters in Equation 1.

We can observe that the mean permeability decreases from the deeper level (910 m) to the shallower levels (1060 m). The 910, 1045 and 1060 m levels are characterized by very heterogeneous permeabilities, whereas the 1000 m level exhibits much more homogeneous values (Fig. 4).

4.2. Evolution of the permeability at the scanline scale

Permeability heterogeneities have been observed at the open pit scale, considering that each profile is characterized by homogeneous characteristics. However, heterogeneity in the geometrical parameter distributions has been observed in each profile by André-Mayer (2000) and André-Mayer and Sausse (2007), and thus, potential heterogeneities of permeability at the scanline must be addressed.

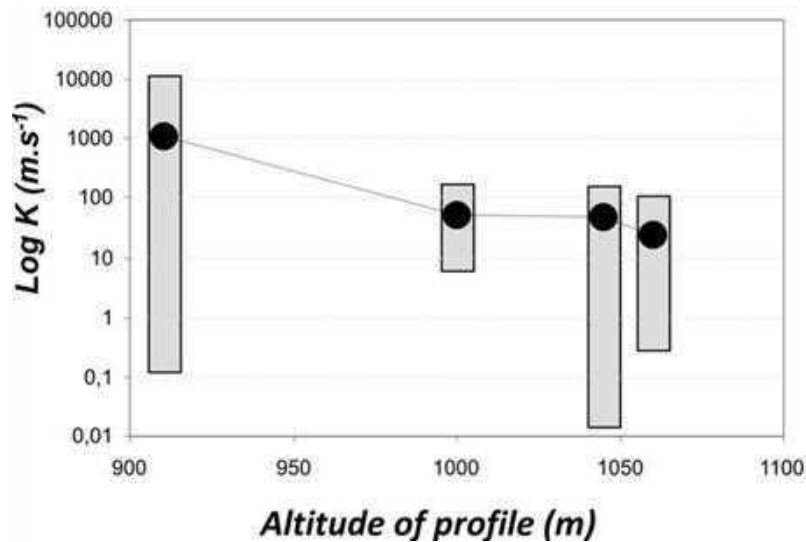


Fig. 4. Mean directional conductivities (black circles) developed by the veins constituting the fractured network in the levels 910, 1000, 1045 and 1060 m. Grey zones shown the extreme values (amplitude between minimum and maximum).

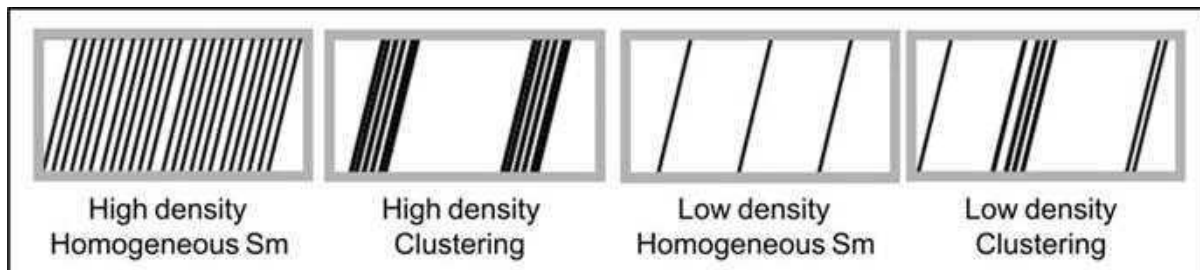


Fig. 5. Theoretical spatial geometries that could be observed along a scanline for different vein densities and spacing standard deviations (André-Mayer, 2000).

Geometrical heterogeneity of the vein network can be indicated using two main parameters: the density of the veins (d) and the interdistance between vein $k+1$ and $k-1$,

which defines the clustering between, veins (S_m). Fig. 5 shows the main geometrical repartition that can be described using the d and S_m parameters.

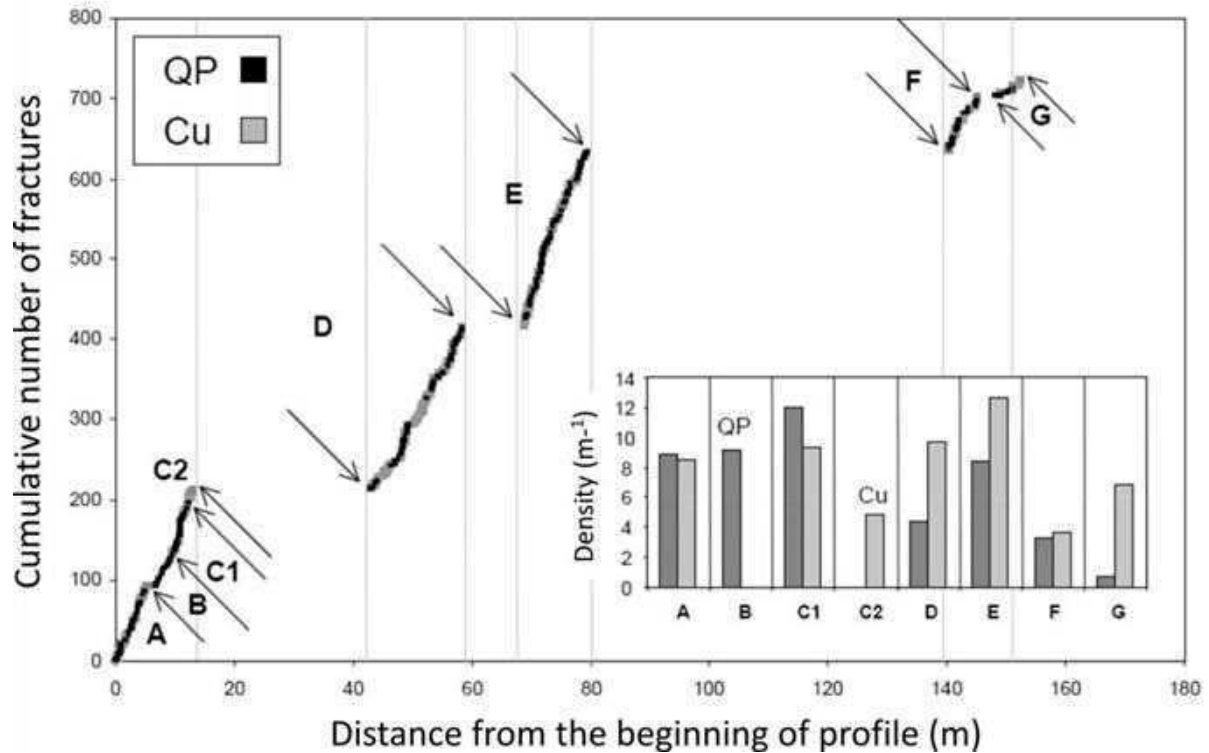


Fig. 6. Discrimination of the 8 zones in the 1000m profile in relation with mineralogical vein fillings or/and vein density differences. B and C2 zones are, for example, characterized respectively by lack of Cu and QP veins.

Table 2. Description of the different zones identified in the three EW profiles. The mean value of spacing S_m and vein thickness w_m and their standard deviations σ_w and σ_s are presented for each zone.

Profile 910 m					Profile 1000 m					Profile 1060 m				
zones	S_m (m)	σ_s (m)	w_m (mm)	σ_w (mm)	zones	S_m (m)	σ_s (m)	w_m (mm)	σ_w (mm)	zones	S_m (m)	σ_s (m)	w_m (mm)	σ_w (mm)
A	0,05	0,05	2,90	3,56	A	0,12	0,07	4,70	4,54	K	0,0593	0,0659	8,72	21,69
B	0,07	0,06	5,90	7,99	B	0,16	0,07	4,90	6,20	J	0,0922	0,0614	5,67	8,14
C	0,06	0,04	4,82	11,43	C1	0,08	0,07	4,20	8,3	I	0,083	0,0866	3,23	3,34
E	0,08	0,05	4,21	4,67	C2	0,09	0,03	11,20	19,25	H	0,32	0,1681	3,25	1,83
F	0,07	0,05	3,45	2,59	D	0,15	0,14	4,39	5,38	G	0,0642	0,0638	3,12	1,55
G	0,04	0,03	2,42	2,43	E	0,10	0,09	3,43	4,84	F	0,0634	0,0474	4,25	10,44
H	0,07	0,07	18,75	25,95	F	0,17	0,18	5,81	9,80	E	0,111	0,088	2,18	2,24
I	0,11	0,06	2,31	0,94	G	0,28	0,25	4,75	2,66	D	0,0929	0,0906	4,4	0,55
J	0,04	0,03	3,80	1,81						C	0,1639	0,1512	3,86	2,27
K	0,04	0,03	1,68	0,81						B	0,0668	0,0486	1,83	0,78
										A	0,0843	0,1234	2,78	1,42

Plotting of cumulative number of fractures in relation to both distance from the profile beginning and mineralogical filling of the fractures (Fig. 6) involves the discrimination of specific zones. For example on the 1000 m profile, 8 zones (named A to G) have been identified in relation to the vein density and the

presence or lack of quartz/pyrite. Table 2 summarizes the obtained zones with this method for each profile, with the mean value of spacing S_m and vein thickness w_m and their standard deviations σ_w and σ_S . Ten, eight and eleven zones were identified for the 910 m, 1060 m and 1060 m profiles, respectively.

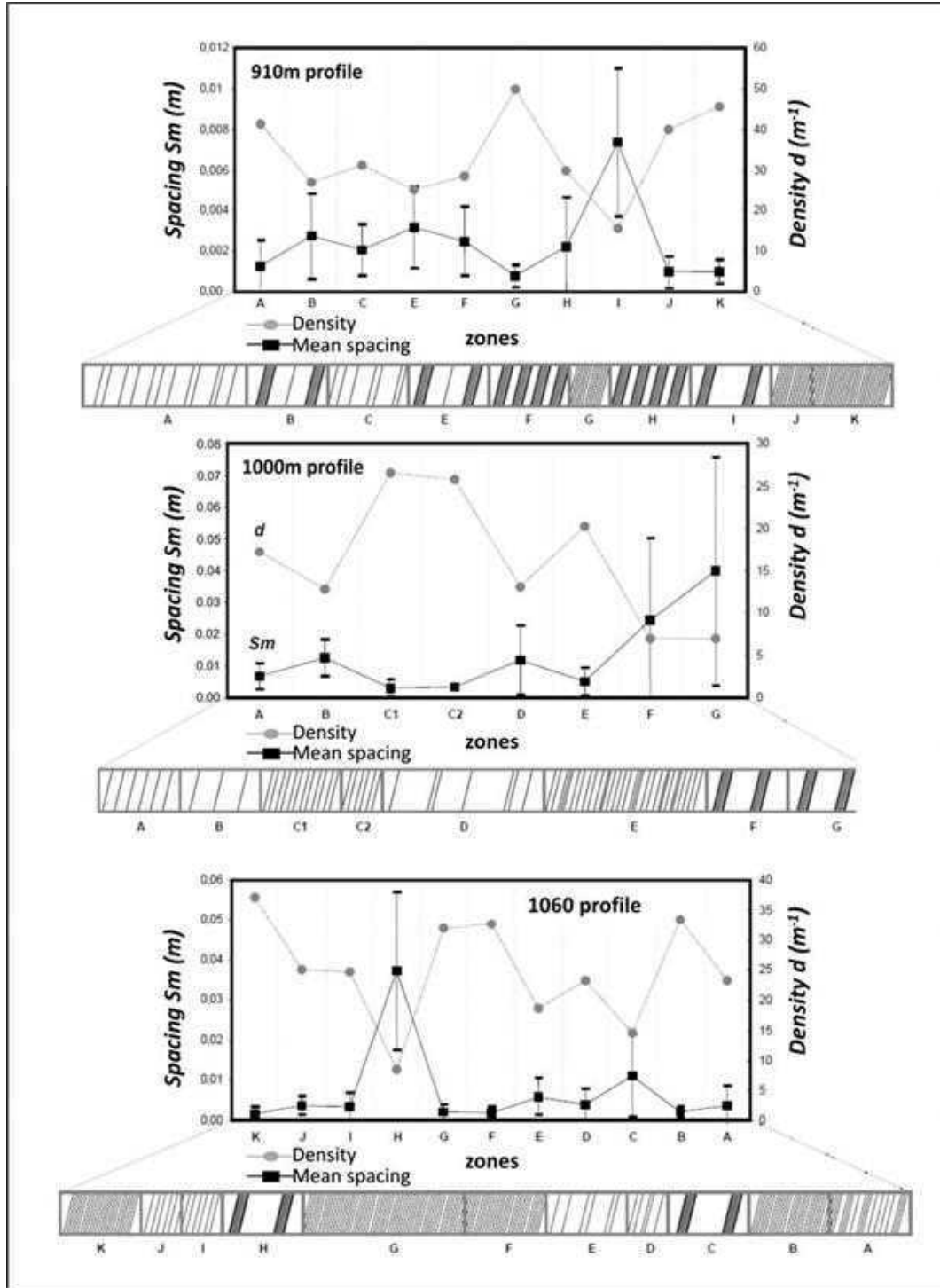


Fig. 7. Diagram showing the mean spacing values S_m (spacing between $k-1$ and $k+1$ vein) with their standard deviation and vein density identified for each zone for the 910 m, 1000 m and 1060 m profiles. A schematic representation of the spatial organization of the veins in each zone is proposed following the description in Fig.5.

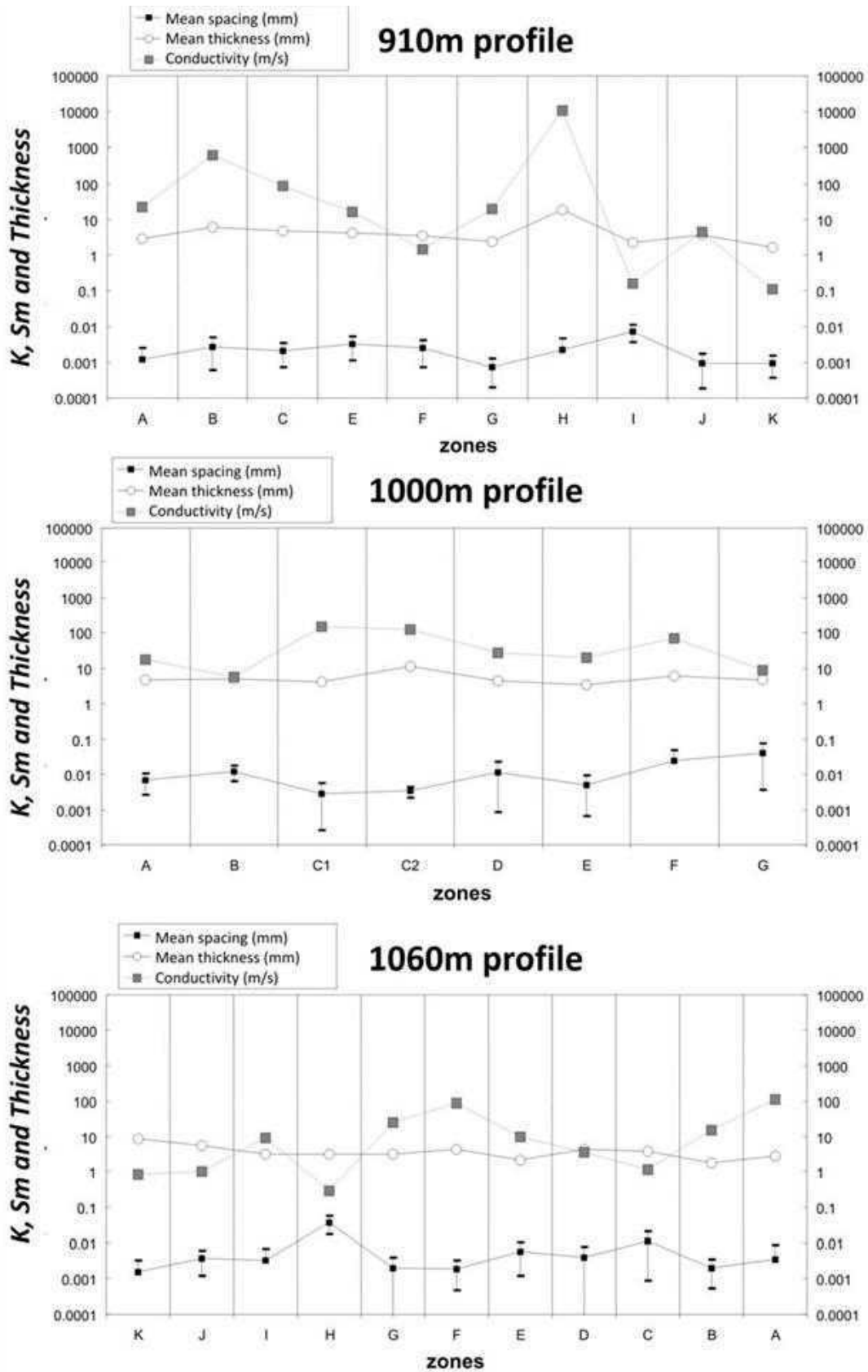


Fig. 8. Mean values of the permeability (grey squares developed by the veins in the 3 profiles with the discrimination of each specific zone. The mean values of spacing (black squares) and thickness (white circles) are also indicated.

Variations in the vein density and spacing for each zone are presented in Fig. 7. In this figure, the spacing values have been normalized to the density for each zone to enable comparison. A schematic representation of the veins' spatial organization is proposed for each profile. Des-

pite a high density, the veins in the 910 m profile have a more clustered distribution than those of the 1000 m and 1060 m levels (Fig. 7).

The mean fissural permeability has been estimated for each zone of each profile using the geometrical characteristics presented above.

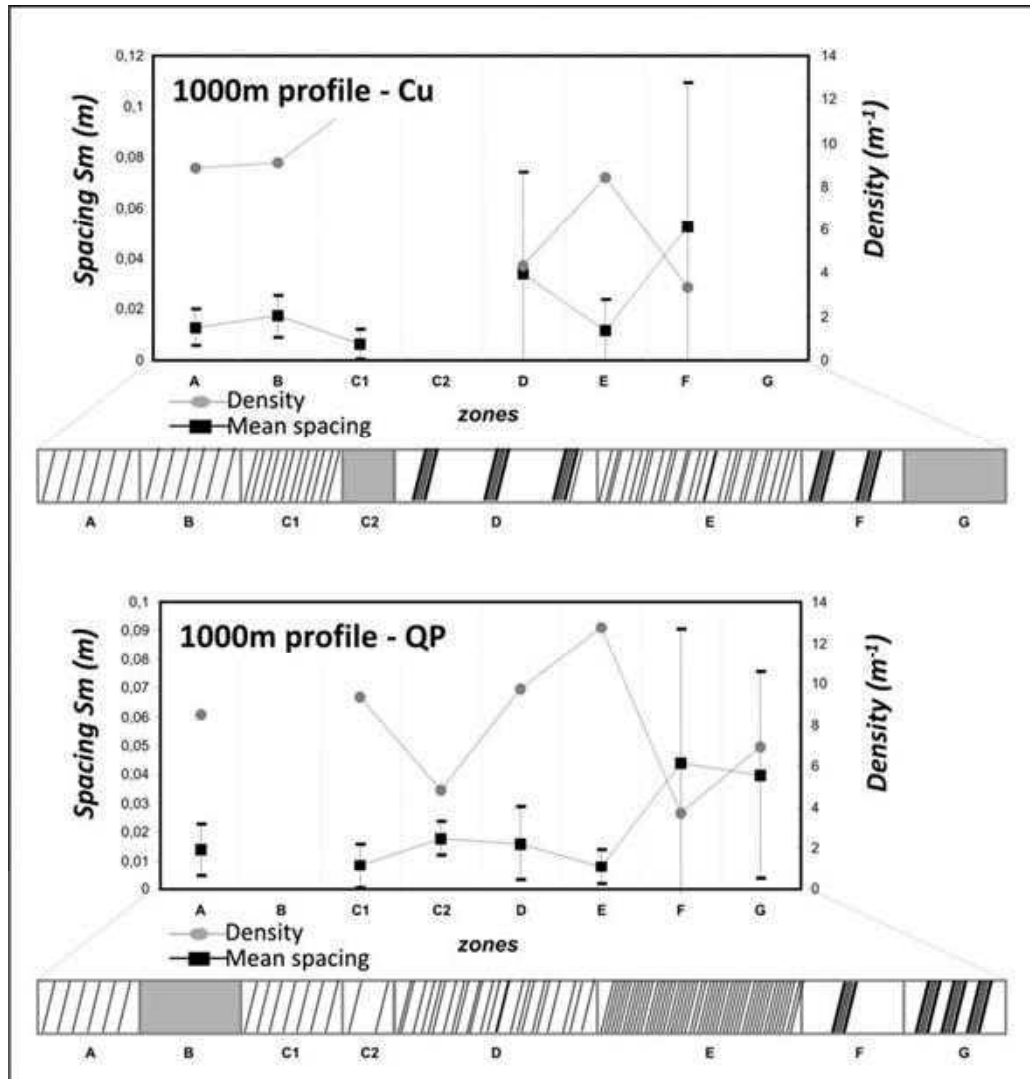


Fig. 9. S_m values, spacing between vein $k-1$ and $k+1$ with the standard deviation (black squares) and density (grey circles) of Cu and QP veins for each discriminated zones of the 1000m profile. A schematic representation of the spatial organization is given following the parameters presented in Fig. 5.

The following results are shown in Fig. 8:

- For the 910 m profile, conductivities are very heterogeneous despite the constant vein thickness. The observed variation in the permeability thus seems not to be linked to variations in the vein thickness (perhaps with the exception of the H zone). The highest permeability seems to be related to a high

density of fractures correlated with a cluster organization. The J and K zones, with high vein densities but not clustering, induce low permeability.

For the 1000 m profile, the estimated permeability seems very homogeneous at first glance. However, this constant value is linked to different parameters: high vein density for specific

zones (C1, C2, E, Fig. 8) or a cluster organization in zones F and G despite low vein density.
 - For the 1060 m profile, the permeability values are the lowest and relatively heterogeneous. In this profile, zones with a clustered organization develop the lowest

permeability values, which should be related to the low vein thickness observed in this profile (Table 1). Thick veins seem to be important for the development of high permeability, even if the veins have a clustered organization.

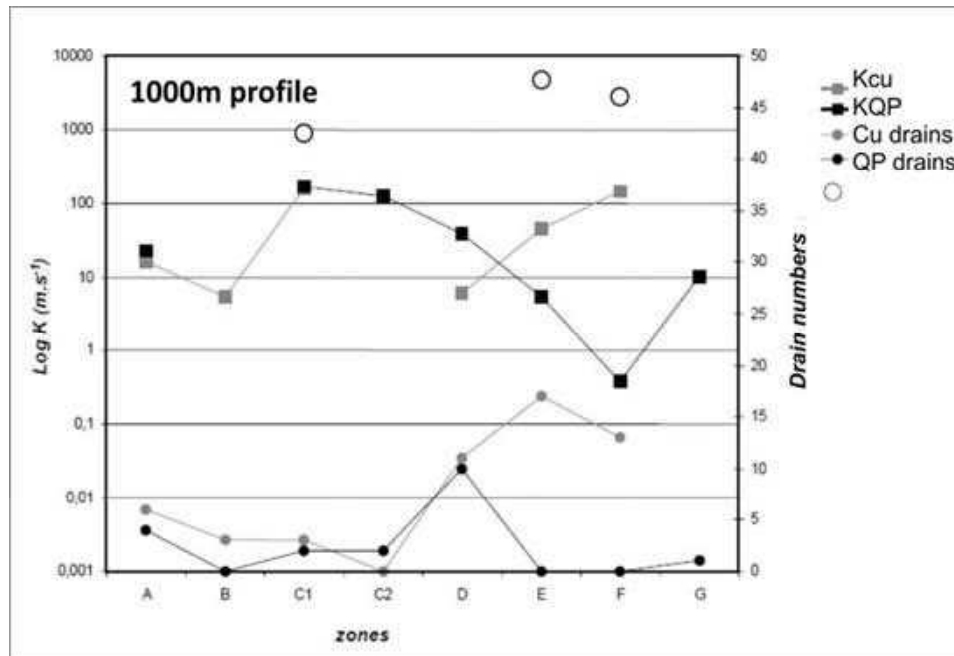


Fig. 10. Comparison of developed permeability of the Cu (grey squares) and QP (black squares) veins.

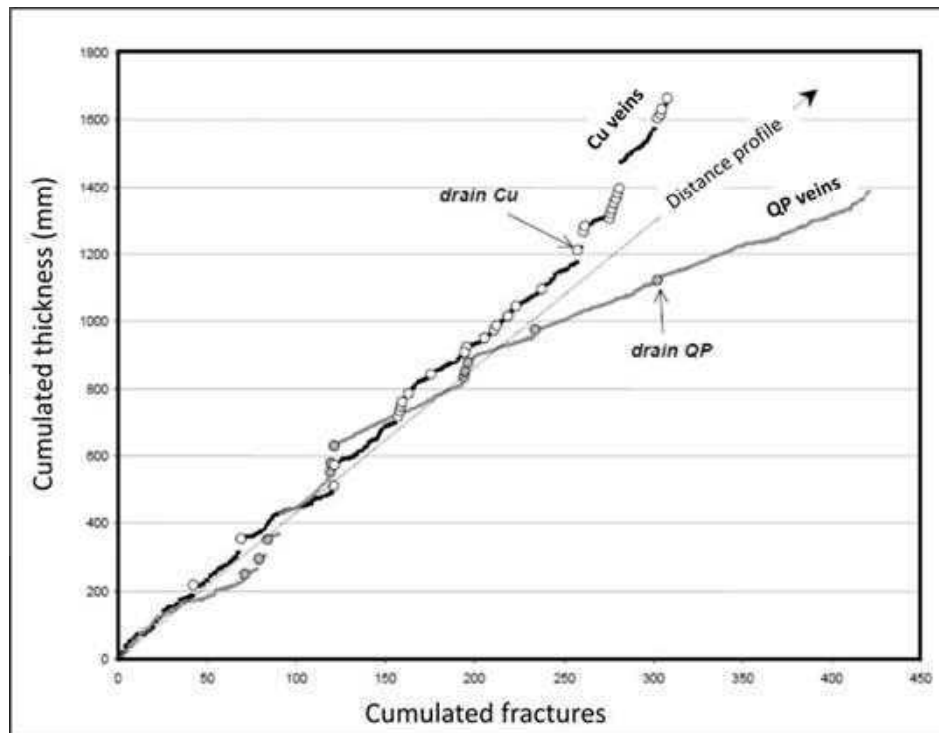


Fig. 11. Cumulated fracture number in function of cumulated thickness.

5. Relation between the fracture permeability and mineralization: the 1000 m level

At the 1000 m level, vein discrimination was achieved using the mineralogical composition of their filling: copper sulfides (Cu) vs. quartz-pyrite (QP). The spatial organization of these two types of veins has been compared to identify the major differences in their hydraulic

behavior. These two types of veins express the same variation, with the end of the profile exhibiting a more clustered repartition of fractures than the beginning. The estimated permeabilities of QP and Cu veins differ along the 1000 m profile (Fig. 10). The Cu vein permeability increases from the D to the F zones, whereas that of the QP vein decreases.

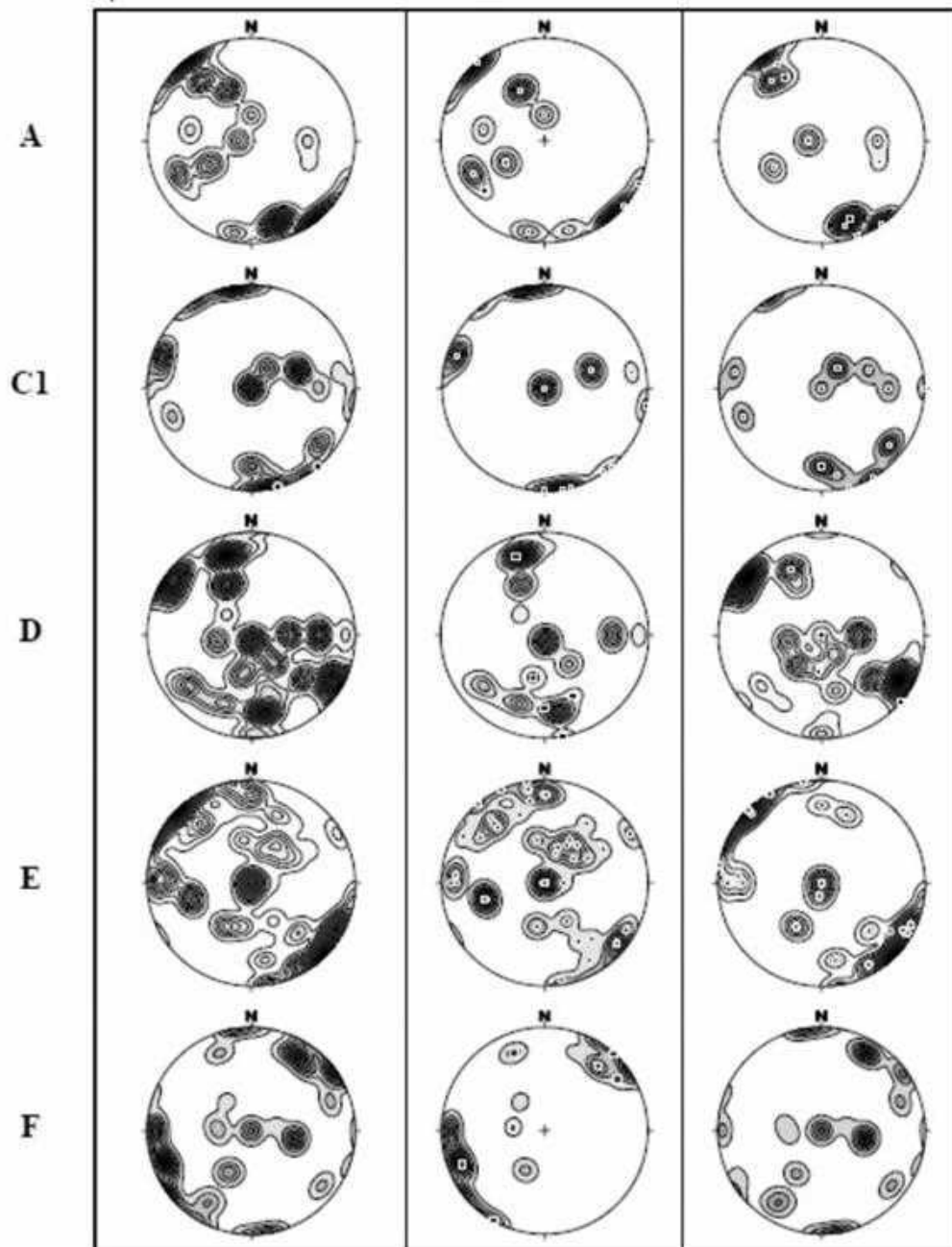


Fig. 12. Stockwork geometry of the 1000m profile in zones A-F, for the entire vein network (left), for the Cu veins (middle) and for the QP veins (right). Small squares express the orientation and dip of vein having thickness greater than 10 mm.

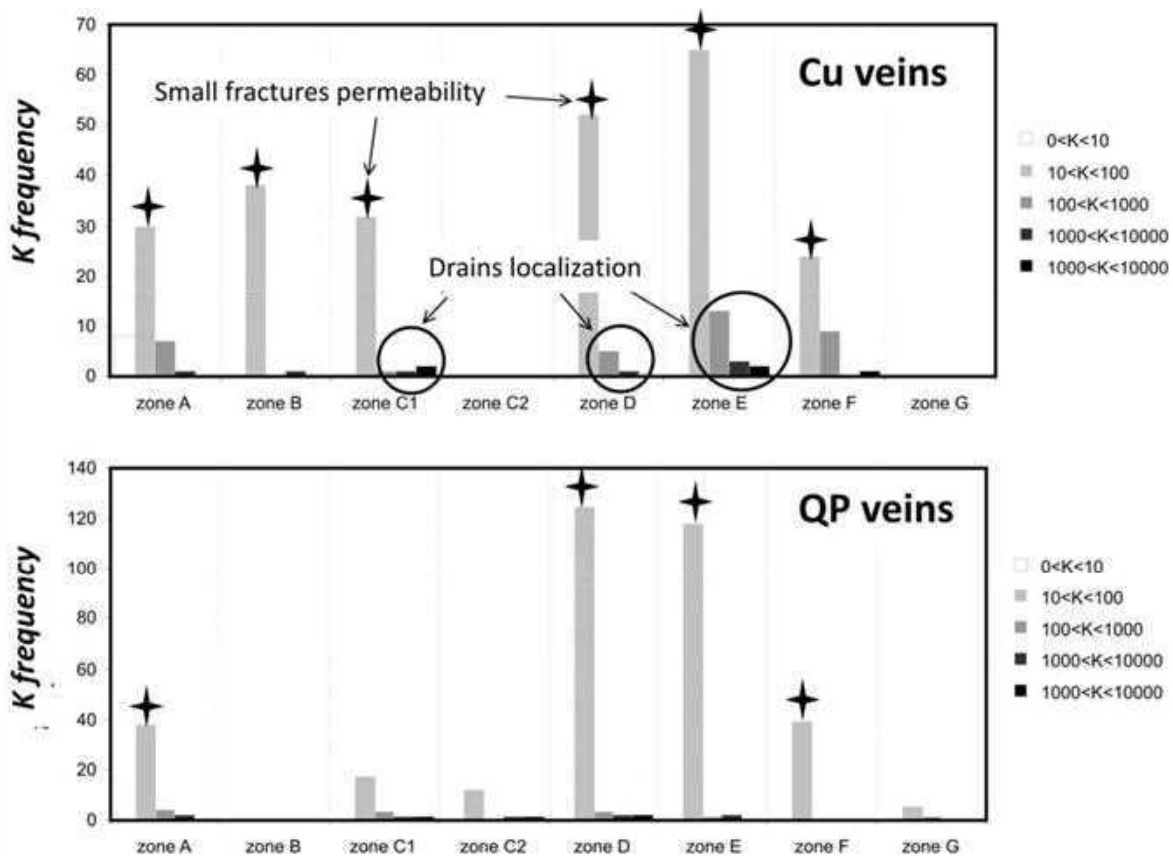


Fig. 13. Frequency of permeability in function of each specific zones.

This different behavior is directly linked to the geometric characteristics of the Cu and QP vein networks:

The cumulated vein thickness related to the cumulated fracture number (Fig. 11) shows the difference in the thickness evolution from the D zone. White and gray circles show the Cu and QP major veins with thicknesses greater than 10 mm, respectively. The higher values are more numerous for Cu veins than for QP veins.

The geometry of the vein network indicates pronounced differences between the QP and Cu veins.

Fig. 12 expresses the stockwork geometry of the 1000 m profile in each zone for the total network (left), the Cu veins (middle) and the QP veins (right). Small squares express the orientation and dip of veins thicker than 10 mm.

These thick veins show much more variable orientations for Cu veins, thus having an increased probability of connection between veins, whereas the orientations of the QP veins are much less variable for both thin and thick veins

Fig. 13 expresses the importance of drains in specific zones of the profile for the Cu veins.

The permeability of Cu veins evolves from a fissural permeability to a “drain permeability,” whereas the QP veins’ permeability is dominated in all the profiles by a fissural permeability without any drain.

Thus, the 1000 m profile, which at first seems to have a homogeneous permeability when the entire vein network is considered, is actually heterogeneous, as shown by the distinct consideration of the mineralogical filling of the veins.

6. Conclusions

Several conclusions can be drawn from this study:

1. The more suitable method for the quantification of permeability in a hydrothermal environment is based on a discontinuous media that treats the rock as a homogeneous media (matrix) + N different elements (fractures), which can be described by a structural analysis. This method uses the cubic law of fracture width and the spacing parameters between fractures within a fracture set. This study indicates the im-

portance of considering S_m in the permeability estimation. A simple cubic law of thickness may yield false homogeneous permeability at the profile scale if the spatial vein organization is not used.

2. The quantification of the permeability expresses the heterogeneity in the hydraulic behavior at the open pit scale: high permeabilities, observed in the lower levels of this hydrothermal system, must be related to major fractures and a clustered organization, whereas permeability becomes more pervasive in the upper level of the system, with thin fractures with relatively homogeneous repartition.
3. The distinction of the QP and Cu veins expresses the differences in their permeability, and a chronological study should be undertaken to understand the correlation between these two vein networks.
4. The systematic relationship between the clustered organization of the vein network and the copper content shows the potential in using the geometrical properties of the mineralized stockwork for exploration.

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Chronology and characterisation of mining development in Romania

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Abstract Archaeological research on the Romanian territory showed that many deposits have been mined and valorized since before the Roman conquest of Dacia (106 A.D.), as pointed out by the ancient works of Herodot, Plinius and Lydus, as well as by more recent articles and books on archaeology and mining history. During the Neolithic and the transition toward the Bronze Age, gold was recovered from alluvial deposits of Crișul Alb river (Apuseni Mountains), and rock salt was extracted from brines at Cacica, Solca and Poduri (northern Moldova). This mining activity was continued during the Bronze Age, when copper and iron began to be extracted mainly in the northern part of the East Carpathians (Baia Mare area), in the western part of the South Carpathians (at Baia de Aramă) and in Dobrogea (at Ceamurlia). In the Iron Age, mining activity was targeted at the same type of resources, i.e. gold, rock salt, iron and copper, with increased production. Historians have estimated that, during the Bronze and Iron Age periods, ca. 250 t of gold and 370 t of silver have been obtained. During the Roman occupation of Dacia (106-272 A.D.), iron mining was developed in the South Carpathians (Poiana Ruscă and Banat), Apuseni Mountains (Trascău Mts.) and Dobrogea. Based on historical data, the gold production during the Roman time was estimated to ca. 498 t. For several centuries after the Roman period, few data are available, consisting of scarce proofs of continuation of gold and rock salt mining. Later, in the 16th Century, iron ores in the South Carpathians (Poiana Ruscă and Banat) and the East Carpathians (Harghita Mts.) have been relatively intensively mined and clay, limestone, quartz and sand began to be used in the manufacturing industry. Scientific studies on minerals and ore deposits can be reported from the XVII-XIX centuries, accompanying the stronger development of mining activities. Increased mineral production (non-metalliferous included) was registered in the period 1849-1918 as the state interests in mining increased, as well. In the period between the two World Wars, the mining activity was strongly developed and diversified. New types of mineral commodities have been mined (Zn, Sb, Sr, Al, Mn, Hg, bauxite, mica, feldspar, gypsum, clays, dolomite, asbestos), with a significant output that induced an increase of industrial entities using domestic ores and non-metalliferous mineral commodities. About 750 such industrial entities have been settled up in the period 1918-1925. The Mining Law in 1929 stated that all mineral resources belong to the Romanian State and enabled a strong development of all kinds of geological research. After the World War II, under the communist administration, all mineral resources in Romania have been intensively mined, in total disregard of the economic profitability and of the environmental impact. Extensive prospecting and exploration programmes led to an increase of localities with mineral resource to over 1, 000, out of which only few proved to be economic (mainly non-metalliferous). Under the consideration of low profitability, in 2008, more than 40 ore mines have been closed, sometimes without the evaluation of the remaining resources. Only four mines (Roșia Poieni and Moldova Nouă – Cu ores, Băița Bihor – Mo, Bi, base metals, W and Au-Ag ores and Crucea – for U) were maintained in operation. The fate of the Roșia Montană gold deposit is still uncertain as there are some unsolved problems of environment protection, preservation of old mining adits (some of them dating back to the Roman time) and other archaeological sites.

Keywords: *mining history, Romania, gold, rock salt, iron, base metal*

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1. Introduction

The Romanian territory was blessed with a great diversity of mineral resources (metallic and non-metallic), and energetic resources (coal, oil, gas, radioactive elements), some of these being

quite abundant. These resources were exploited during thousands of years, being a main reason for several periods of economic and social development. These rich resources also represented, since the pre-Roman ancient times, a permanent attraction and the reason of many wars.

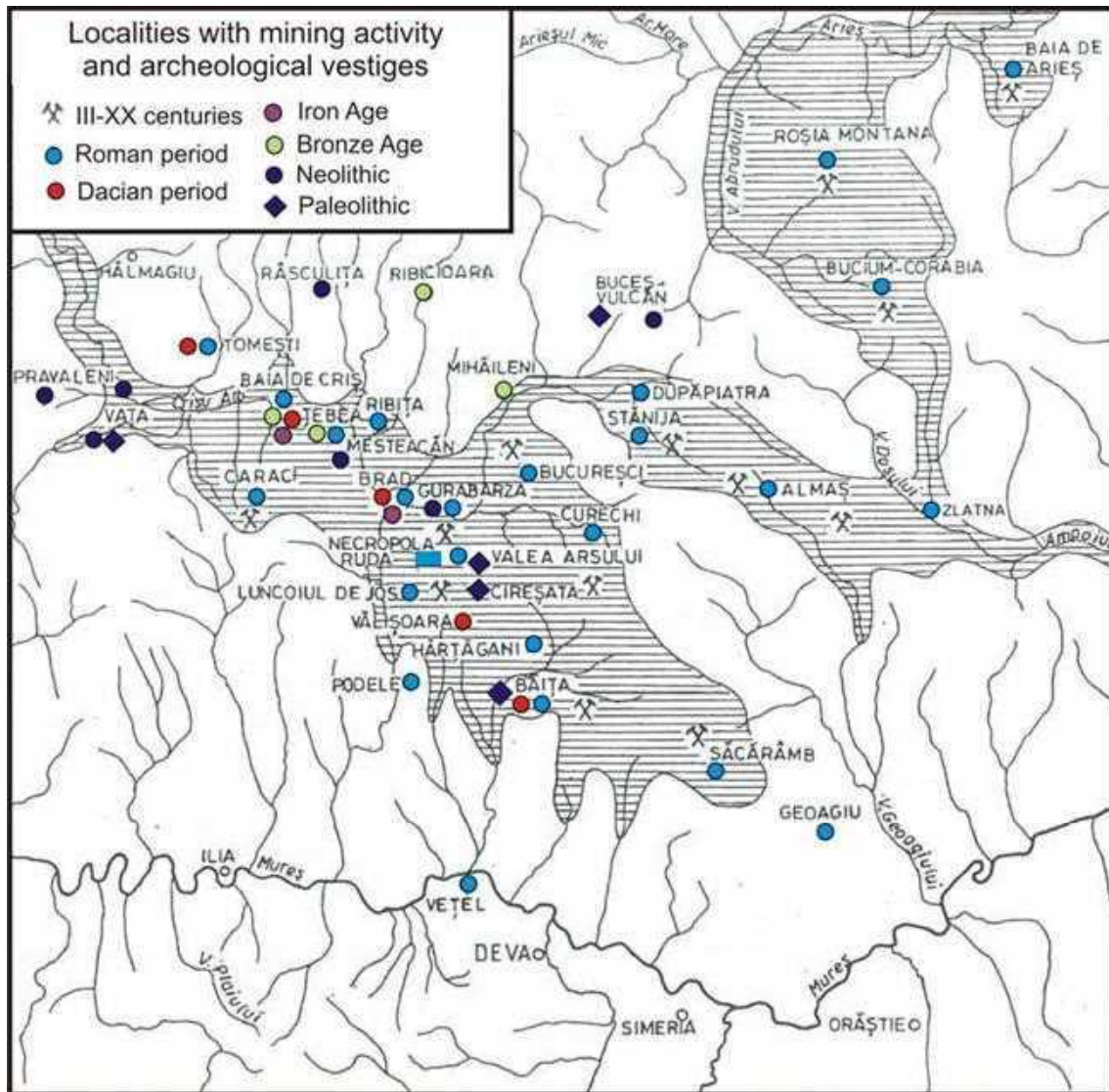


Fig. 1. Area corresponding to the Golden Quadrilateral of the Apuseni Mountains, defined by the localities Brad, Săcărâmb, Zlatna, Stănița, Bucium, Corabia and Baia de Arieș. This area recorded a very long period of gold ore extraction (an interval larger than 4,500 years, from the Neolithic to present, with rare interruptions), with primitive methods or mining works (map from the Gold Museum in Brad).

2. Pre-Roman period

The oldest archeological proofs, most of them indirect, as well as the historical writings, place the first activities of extraction and processing of metals in the late Neolithic, Bronze Age and Iron Age (Herodot, 5th century B.C.; Plinius, 1st century A.D.; Lydus, 1st century

A.D.; Densușianu, 1913; Pârvan, 1926, 1972; Abrudeanu, 1933; Haiduc, 1940; Popescu, 1956; Daicoviciu, 1960; Acker, 1965; Feneșan, 1967; Bujor and Roșu, 1968; Maghiar and Olteanu, 1970; Olteanu, 1971; Zah, 1971; Zsambok, 1985; Csiky, 1985; Sîntimbrean, 1989; Marinescu et al., 1990; Wolman, 1996; Bejan, 1998).

Gold and salt have been extracted since the late Neolithic (the "Decea Mureșului" and "Cucuteni" cultures). Salt was extracted from brines in Moldavia (Cacica, Solca and Poduri). Gold was extracted in Transylvania (Apuseni Mountains) by panning the alluvia from the upstream tributaries of the upper Crișul Alb river (Fig. 1).

During the Bronze Age (1800-1150 B.C.), salt extraction is also attested at Coștiui and Șugatag (in the actual Maramureș County), and in the southern Subcarpathian zone (e.g., Ocnița). Gold extraction from alluvia was widespread in the Metaliferi Mountains (Fig.1). Local copper ore extraction is indicated by vestiges of numerous bronze storage places and bronze objects manufacturing in Apuseni Mountains, in the Baia Mare zone, at Baia de Aramă and in North Dobrogea.

In the Iron Age (1,150 B.C.-160 A.D.), there are proofs of mining for salt, gold, iron and copper (Marinescu et al., 1990). In the Hallstatt period (1,150-650 B.C.) there were at least 8 centers of salt extraction in the Transylvanian Basin. In the South Carpathians, East Carpathians and Dobrogea, iron was extracted in several places from Orăștie Mountains, Harghita Mountains and North Dobrogea, where the mining concerned mainly the oxidation zones of the deposits. In the proximity of these areas, furnaces for iron reduction were found, as well as ore fragments, slug boulders and ceramic fragments, the latter certifying the age of the vestiges. Gold recovery from alluvia continued on the main river courses from the Apuseni Mountains; similar activities took place on the large rivers that drain the mountain slopes in the central-western part of the South Carpathians.

In the second period of the Iron Age - La Tène (650 B.C.-160 A.D.), corresponding to the Dacian (Getic) period, many of the earlier mining sites were kept in operation. Additionally new salt resources were mined in the Transylvanian Basin, new iron resources were discovered in Trascău Mountains, at Rimetea (Mârza, 1962), and gold probably began to be extracted from outcrops, as well, at Roșia Montană (Fig. 1).

The amount of gold and silver extracted in the Bronze and Iron Ages were estimated to 250 t Au and 370 t Ag (Haiduc, 1940; Lăzărescu and Brana, 1972), based mostly on the contemporary information regarding the treasure captured at the Roman conquest of Dacia, which is considered as the greatest capture in the history

of the Roman wars (Lydus, *De Magistratibus*, 1st century). The captured Dacian treasure was estimated to more than four times the value of the public treasure of the Roman empire at the time of the conquest of Dacia (Densușianu, 1913, cf. Plinius). It is possible that the amount of metals mentioned by historians might have been exaggerated, as its assessment was based mostly on comparisons with other war captures (Bejan, 1998).

3. The Roman period

During the Roman occupation in Dacia (106-272 A.D.) the mining activity was organized and administered coherently over the entire occupied territory. Within a short time interval, the richest accumulations of native gold (at Barza, Băița, Caraci, Zlatna, Roșia Montană and Abrud) were identified and partially delimited. These were mined at surface and underground, from veins and stockworks with extremely high grade native gold ore. There are documents mentioning the work schedule, ownership, legislation, organization of mining system, transport and storage of production. The existence of these mining activities is also certified by archeological vestiges (old adits, inclines, transport devices, systems of water removal and aeration and mining tools), as well as inscriptions on monuments (shrines and funeral stones) and wax tablets (Sintimbrean, 1989).

During the Roman period, iron mining extended around the old extraction centers from Poiana Ruscă and from the East Carpathians, known since the Dacian (Getic) period, as well as in many other locations with smaller reserves from the South Carpathians (in the vicinity of Șercaia, Șura Mică and Ilova-Cerna). Iron extraction was archeologically evidenced in Trascău Mountains, at Rimetea, from limonitized siderite (Mârza, 1962), and in Dobrogea, from ferruginous sands, at Boclușea Hill, Dervent Hill and Zmeica Lake, as indicated by fragments of slug and charcoal and by several remnants of furnaces for iron ore reduction (Marinescu et al., 1990). Based on similar arguments, iron and copper extraction is evidenced at Ceamurlia de Sus and Altân-Tepe (in Dobrogea), and especially at Baia de Aramă (a Romanian name for Copper Mine), where copper ore extraction and processing were remnants of smelters with ore fragments, bronze storage

places and dumps are preserved at Bratilov, a neighbouring village. Other mining and processing centers were located in Banat, at Moldova Nouă, Sasca, Bocşa Montană, Ocna de Fier and Dognecea, where, probably, the extraction of gold from base metal ores was performed.

The Roman administration organized exploration expeditions in the Northern Apuseni Mountains, at Băişoara (in Nyerghes perimeter), and also outside the occupied territory, in the southern part of the Gutâi massif (at Săsar, Valea Roşie and Valea Borcutului). In the absence of a systematic record, the production of metals during the Roman period was estimated to 498 t Au and 950 t Ag (Haiduc, 1940; Lăzărescu, Brana, 1972).

Among the non-metalliferous resources, the previously-discovered rock salt accumulations continued to be intensely exploited. New rock salt resources were discovered, the most important being near Turda and Praid (Fig. 2). The salt deposit at Ocnîţa (South Carpathians), known since the pre-Roman times, became one of the main salt sources, with great production. It developed as an important commercial center, from which the salt was exported along the so-called "Salt Road", following the course of Olt river, then across the Danube, in the Balkan area (Tudor, 1968). The number of quarries for construction materials (limestone, marble, sandstone, andesite and shale) increased continuously during the Roman occupation, stimulated by the extensive road construction and maintenance.

4. The post-Roman period

After the Roman retreat from Dacia, during the 4th-10th centuries A.D., successive migrations crossed Dacia from the east to the west, with temporary occupation of the areas with mining activity. Gradually, the systematic extraction of mineral resources diminished and even ceased in many places. Nevertheless, two centers of iron extraction are mentioned in this period: Budureasa, in Apuseni Mountains, and Capidava, in Dobrogea (Maghiar and Olteanu, 1970). Beginning with the 10th century and in the 11th century, the mining activity was resumed, mostly for gold and salt in the Transylvanian Basin, but also for salt in Wallachia and Moldavia (Fig. 2).

5. The Middle Ages period

In the 12th-13th centuries, period corresponding with the early feudalism in Transylvania, the metal extraction was reactivated in the old mining centers (Maghiar and Olteanu, 1970). During this period, several mining centers were attested in official documents: Băiţa Bihor (1270), Baia Sprie, Căvnic and Rodna (1255), which subsequently were among the mines with the greatest productions of lead ore rich in gold and silver (Zsamboki, 1985). In 1327, the mining of gold, silver and copper was attested at Băişoara (Gilău Mountains).

Contemporary documents mention that during the 13th century, the extraction, transport and export of salt, on land and by ship, were considered privileged activities. Salt mines were mentioned in Transylvania (Turda, Dej, Ocna Sibiului), Wallachia (Ocnele Mari) and Moldova (Chilia and Târgu Ocna) (Gusti, 1939).

In the 14th-16th centuries, the lead ores rich in gold and silver from Baia Mare and Rodna Massif were mined. In the same period, the extraction of the iron ore from Poiana Ruscă Massif, Trascău Mountains, Hărgăhita Mountains and Banat took place on a quasi-permanent basis. Furnaces with increasingly greater capacity were built. Similar development involved the exploitation of copper ore in the Altân Tepe deposit.

In Transylvania, the ownership of gold resources was the object of a dispute between the Hungarian kings, the Catholic church and local nobility. The operation of the salt mines in Moldavia and Wallachia was the monopoly of the monarchy. Since the 15th century, the mining of the non-metalliferous resources has begun to develop constantly, stimulated by the more liberal policy towards the manufacturing activities.

6. The industrial period before the World War I

During the 17th-19th centuries, the scientific and technical progress allowed the opening of the deposits to greater depths and increased the capability of extraction and transport. Concomitantly, more effective crushing installations were used and the first metallurgical factories were built for lead ore processing and for the separation of lead from copper and gold and separation of gold from silver.

There are numerous documents concerning mining industry in the 18th and 19th centuries

(mine production registers, acts of privilege granting, acts of ownership, concession, mine closure etc), kept in the state archives from Baia Mare, Cluj-Napoca, Deva and Timișoara. These data show that, in 1841, more than 2,000 gold mines were operating in the Metaliferi Mountains. In the Baia Mare zone, 34 mines were in operation, plus several other in the Rodna Massif. In Banat, in the period 1749-1761 concessions were issued in 128 mining perimeters, which were operated with profit at least until the second half of the 19th century. The annual gold production in Transylvania increased in the interval 1770-1841 from several hundred kilograms to 1,200 kg. Banat province was a main European region with mines of copper and iron, including, as main centers, Oravița, Sasca, Moldova Nouă, Ocna de Fier and Dognecea in Banat. According to the official archives, in 1773-1778, Moldova Nouă deposit produced 4,427 t Cu, 225 t Pb and 124 t Ag; in 1777-1854, Dognecea, produced ca 1,850 kg Ag; at Ocna de Fier, in 1720-1814, at least 100,000 t of iron ore were extracted (Brîndză, 1985). In support of mining industry, railways were constructed in Banat, between Oravița and Buziaș (1854), and on the route Secu-Reșița-Bocșa-Ocna de Fier (1873). Gold separation from alluvia was attested near Bozovici, in 1769, and continued with profit toward the end of the 18th century.

Beside salt, shales, limestones and various construction stones, the quartz-rich sands began to be extensively used in the glass factories, the first of them being established in 1650, at Târgoviște, in Wallachia.

In 1848-1918, most mines for gold and salt from Transylvania were transferred in the state ownership, with good effects for the increase of production. The annual gold production of the Transylvanian mines in the period 1841-1880 increased from 700 kg to 1,570 kg, reaching a maximum of 2,270 kg in the period 1890-1908. In 1864, the Civil Code of the unified Wallachia and Moldavia created more homogeneous and clear regulations for mining activities, with special distinctions regarding the ownership of the surface and underground in the mining areas. In the Romanian Kingdom, the Mining Law adopted in 1885 had the purpose to stimulate the exploitation of mineral resources and to develop the industries related to mining activities. The first version of the "National Map of Mines and Vegetation of the Region between the Rivers Nistru and Tisa" was published in the "Atlas of

Romania and of the neighbouring countries" (Murgoci and Popa Burcă, 1902).

7. The Interbellum

According to the Romanian Constitution in 1923 and the Mining Law in 1924 (modified in 1929), all underground resources were owned by the Romanian state. The Mining Law was meant to ensure a rational exploitation of mineral resources and to regulate the rights of ownership and concession; it also stipulated the creation of mining tribunals.

In the period 1919-1945, Romania had the best strategy of mining development and of harmonisation with the related activities. Actions were taken toward the conservation of mineral resources, both by a more profitable processing of imported raw materials and by Romanian participation to exploration abroad. The national research programs were intensified and amplified in correlation with the development of new industries, leading to the identification of new deposits, containing new commodities (Zn, Sb, Cr, Al, Mn, Hg, Al, mica, feldspar, talc, gypsum, bentonite, kaolin, shales, dolomite, asbestos, strontium, silica). Additionally, the prospecting and exploration increased the reserves of metalliferous and non-metalliferous commodities, as presented in the Mining Indicator of Romania (MIC, 1925), Statistical Yearbook of Romania for the year 1938 (ICS, 1939), and in the Encyclopaedia of Romania (Gusti, 1939), the latter containing the second (revised) version of the National Map of Mining in Romania with information updated to the year 1939, made with the data from the Geological Institute of Romania (founded in 1906).

The first national evaluation of the reserves of Fe, Mn, pyrite and Cr was published by Ghițulescu and Gavăt (1931) and included the following data:

Fe ore: ca. 26 mil. t (25-50% Fe), more than 60% hosted in Poiana Ruscă massif;

Mn + Fe: ca. 9 mil. t (36% Mn, 9% Fe), ca. 60% located in the East Carpathians;

Pyrite: ca. 8 mil. t ($\pm$ Cu, $\pm$ Au), 50% in Baia Mare zone, 25% in Apuseni Mountains,

Cr ore: ca. 2 mil. t (38-52% Cr₂O₃), 100% in Banat.

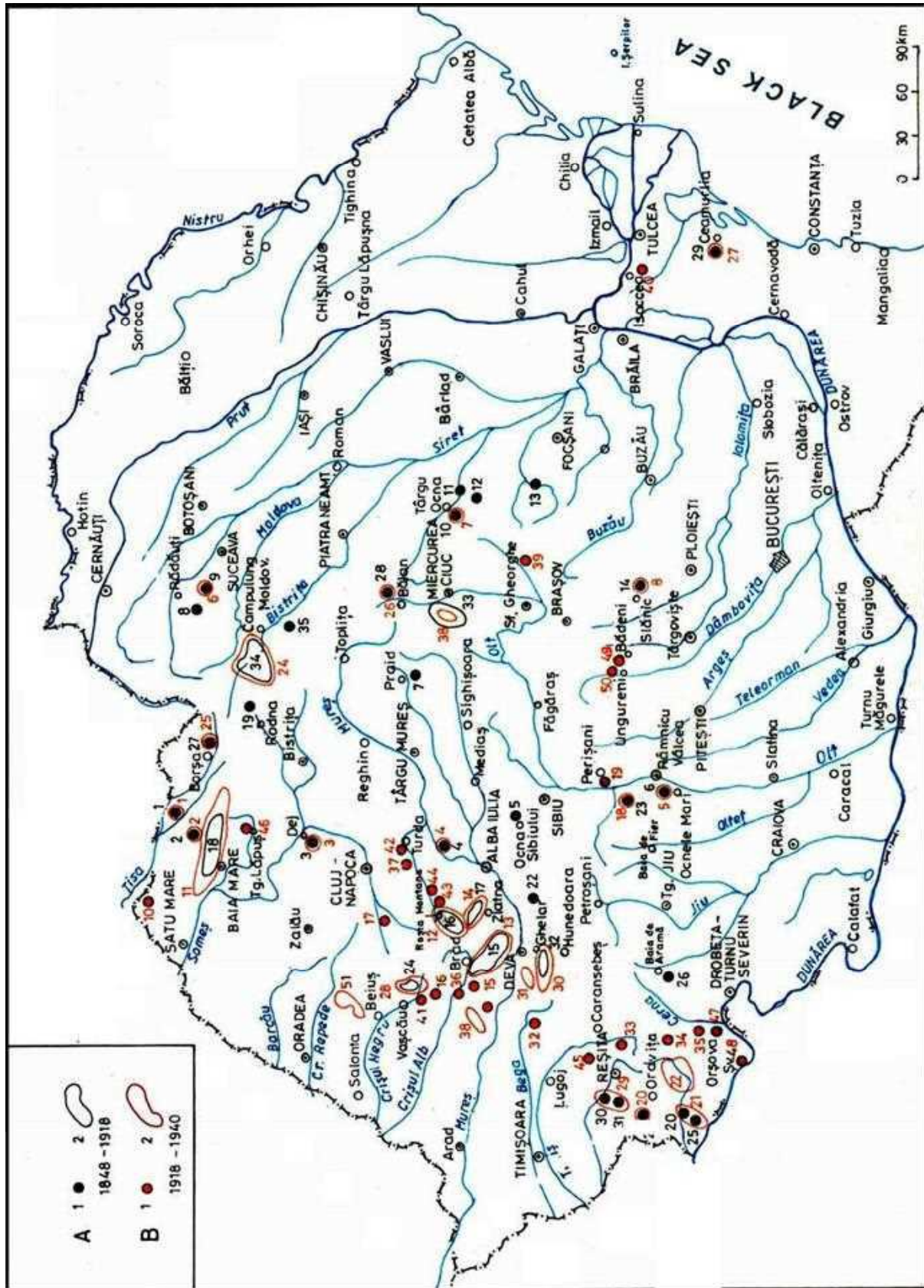


Fig. 2 The location of the mined metalliferous accumulations and salt deposits in Romania. Background map: Romania between 1918 and 1940.

- A: BEFORE 1918:* a. Mining fields and districts; b. Isolated accumulations; SALT (1. Coștiui, 2. Ocna Șugatag, 3. Ocna Dej, 4. Ocna Mureș, 5. Ocna Sibiului, 6. Ocnele Mari, 7. Praid, 8. Solca, 9. Cacica, 10. Tg. Ocna, 11. Dragomirești, 12. Nistoroiaia, 13. Valea Sării, 14. Slănic). Au, Ag-Pb, Zn±Cu ORE (15. Barza - Săcărâmb, 16. Roșia Montană - Bucium - Baia de Arieș district, 17. Zlatna-Amaș+Hg district, 18. Baia Mare district, 19. Rodna, 20. Sasca, 21. Oravița, 22. Pianu, 23. Valea lui Stan). Cu, PyCu ± Au, Ag, Pb ores (24. Băița-Bihor, 25. Moldova Nouă, 26. Baia de Aramă). Pb, Zn, Cu, Py ores (27. Borșa, 28. Bălan, 29. Altîn Tepe+Fe). Fe, Cu ± Pb, Zn, Au, Ag ores (30. Ocna de Fier, 31. Dognecea, 32. Teliuc-Ghelar IRON 33. Mădăraș-Lueta-Vlăhița Fe district). Mn-Fe; Pb, Zn, Cu, Py ores (34. Iacoveni-Vatra Dornei-Pojorâta district, 35. Brusturi Fe)
- B: After 1918:* a. Mining fields and districts; b. Isolated accumulations; SALT (1. Coștiui, 2. Ocna Șugatag, 3. Ocna Dej, 4. Ocna Mureș, 5. Ocnele Mari, 6. Cacica, 7. Tg. Ocna, 8. Slănic, 9. Tuzla). Au, Ag; Pb, Zn±Cu±Au, Ag ore (10. Tarna Mare district Au, Ag±Pb, Zn, Pb, Zn-Au, Ag, 11. Baia Mare district Au, Ag; Pb, Zn±Cu±Au, Ag±Sb, 12. Roșia Montană-Bucium-Baia de Arieș district Au, Ag; PyAu; Pb, Zn, Cu±Au, Ag, 13. Brad-Săcărâmb district Au, Ag ± Te; Pb, Zn-Au, Ag, 14. Almaș district Au, Ag ± Te; Pb, Zn-Au, Ag + Hg, 15. Caraci Au, Ag, 16. Brusturi Au, Ag-Pb, Zn ± Cu, 17. Someșul Rece Au, Ag-Pb, Zn ± Cu, 18. Valea lui Stan Au, Ag, 19. Perișani, Au, Ag). Cu; PyCu; Pb, Zn, Cu±Au, Ag±Bi, Co; Fe ores (20. Oravița Cu, Pb, Zn, Py, Mo, Bi, Co, 21. Sasca-Moldova Nouă district Cu, Pb, Zn±Au, Ag; PyCu, Mo, Bi, 22. Lăpușnicul Mare-Șopotul Nou-Bozovici-Bănia district, Dalboșet Cu; PyCu±Au, Ag; Au, Ag; Fe). PyCu; Pb, Zn; Fe, Mn; Fe ores (23. Drocea district). Py; PyCu; Mn-Fe ores (24. Munții Bistriței district, 25. Borșa Py; PyCu, 26. Bălan, Py; PyCu, 27. Altîn-Tepe PyCu-Fe, 28. Băița-Bihor district Pb, Zn, Cu ± Au, Ag±Mo, Bi). Cu, Pb, Zn-Au, Ag±Mo, Bi, W; Fe ores (29. Ocna de Fier-Dognecea district) Fe ores (30. Teliuc-Ghelar district Fe, 31. Valea Cernei, 32. Gladna-Nădrag, 33. Teregova, 34. Iablanița, 35. Topleț, 36. Vața de Sus, 37. Cacova, 38. Mădăraș-Ciuc-Homorod district, 39. Covasna, 40. Balabancea). Fe, Mn ores (41. Moneasa, 42. Buru, 43. Cioara, 44. Sălcia de Jos, 45. Delinești, 46. Mașca). Cr, Ni, Fe ores (47. Ogradena, 48. Plavișevita, Dubavăț, Tișovița, Eibenthal), Ni (49. Bădeni, 50. Ungureni) BAUXITE (51. Pădurea Craiului Mountains district: Vadul Crișului, Bratca, Remeți, Roșia, Dobrești, Pietroasa)

As a result of the new policy concerning the research on mineral resources, more than 750 factories were established in industrial branches that used commodities from the mining industry. Beside metal-processing activities, a new development was registered in the production of glass, chemical compounds, cement, lime, ceramics, refractory materials etc.

New types of ores were discovered, keeping the pace with the technical advances in metal extraction from the ores (Ni, Co, Bi, Ti, Mo, W, Mg). The Geological Institute of Romania published the first map with quarries for construction stones in Romania, at the scale 1:750,000 (Cantuniari, 1936).

During the Second World War (1939-1944), all deposits (mainly the metalliferous ones) were exploited in emergency regime, with the preferential extraction of the higher grade zones.

During the war, the Cr reserves from Banat were exhausted.

8. The communist period

During the first years after the Second World War, very rich uranium deposits were discovered in the Apuseni Mountains, which, were entirely exhausted in their richest zones before the end of the 1960s, the raw ore being transported to the Soviet Union as war damages. In the same intensive regime were exploited almost all metalliferous deposits, especially the gold deposits from the Metalliferi Mountains. The remaining resources were subsequently managed mostly in conditions of marginal profitability and even at a loss, which led to a gradual decline in mine industry. The policy of industrial development with minimal import of raw materials lead to the

extraction of subeconomic ores, with state subventions up to \$12,000 for \$1,000 production. Examples of subeconomic exploitations are those of native sulphur at Negoilul Românesc, cinnabar at Sântimbru, heavy minerals (Ti-Zr) from the southern Subcarpathian zone, silicates and oxides of Mn in South Carpathians, several porphyry type mineralizations (Cu-Au and Cu-Mo) from Apuseni Mountains and Banat, and several accumulations of base metal sulfides, Fe and Mn-Fe from Apuseni Mountains, East Carpathians and South Carpathians.

The prevalence of the political reasons over the economic criteria led to the neglect of rare elements (Te, Se, Tl, Cd, Ga, In, Sb, Bi, Ge, Sn), although they were found with economic grades in several deposits (Cheşu, 1983; Pandelescu et al., 1986).

Nevertheless, the years 1945-1990 were characterized by the extensive use of the geological, geophysical and geochemical methods, resulting in a comprehensive knowledge on many deposits and their surrounding areas. On the other hand, the extended exploration and mining produced the accumulation of a huge volume of mining and processing waste of more than 1000 mil. m³ (out of which 1/3 are mining waste and 2/3 processing waste), which cover a total area of more than 3,000 ha (Lăzărescu, 1983).

9. The post-communist period

The period 1990-present was marked by a radical revision of the state development strategy, which led to dramatic funding reduction in the geological exploration and mining, thus triggering the collapse of the ore mining industry and of the associated research activity (Rădulescu et al., 1991). Unlike the metalliferous resources, the non-metalliferous commodities began to be better managed after the year 1990, mainly by the elimination of the state subvention and the stimulation of the private companies (both Romanian and foreign) to perform exploration and mining. In 1994, the National Agency for Mineral Resources was created, with the mission of governmental administrator of the mineral resources in Romania. In 1998, the first post-communist Mining Law was enforced. It imposed the use of the UNFC classification in the assessment of resources and reserves, regulated the issue of exploration and exploitation concession licenses

by public auction and implemented clear regulations regarding the responsibility of the license holders for the environmental restoration of the mining sites. Since 1998, more than 2,500 companies were granted licenses (including the ones expired meanwhile) for prospecting, exploration and exploitation of non-metalliferous resources. Numerous abandoned exploitations of kaoline-rich shales, bentonite, brucite, nepheline sienite, quartz sands, feldspars etc began to be mined again (Harosa et al., 1998; Borcoş et al., 2003). In 2007, most metalliferous mines were closed. After 2008, only 4 mining perimeters out of 49 remained in operation: Roşia Poieni and Moldova Nouă for Cu, Crucea for U and Băiţa Bihor for Mo, Bi, W, base metals and Au-Ag. Roşia Montană deposit, ready for exploitation, and the gold-bearing pyrite accumulations in the perimeters Coranda-Hondol and Căinel are the main prospective gold resources.

10. Conclusions

The Romanian territory contained large amounts of mineral resources of varied commodities, which have been extracted for a long time (hundreds to thousands years). In the present, the non-metalliferous mineral resources show the largest volumes and are mined most extensively. Almost all metalliferous deposits are either totally exhausted or preserve only their mineralized parts with subeconomic grade. However, the available data suggest that, given the evolution of metal prices, the profitability of some closed mines should be reconsidered, especially if modern methods of extraction and processing are used. This would allow the capitalization of a relatively large volume of resources of base metals, copper, gold and iron ± manganese. Moreover, the recent success of several exploration programs in Romania and in other countries from the Balkan area suggests that new deposits can be found by the application of modern exploration methods in zones where former projects failed.

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Structural styles and economic potentials of some barite deposits in the Southern Benue Trough, Nigeria

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Abstract: Field and laboratory studies of barite ores and their associated rocks from the Gabu, Alifokpa and Osina barite fields allowed the evaluation of their genesis and economic potential. Results show that ores occur as disseminated nodules, stratabound as well as in two main vein sets, namely: the NW-SE trend that is orthogonal to the main axis of the Benue Trough and a comparatively younger, N-S trend, crosscutting the other one, with steep dips (generally $>80^\circ$). These structural arrangements not only predispose the ore deposits to vertical stripping using open pit method but also indicate epigenetic and epithermal processes for the ore formation that relate closely with basin tectonic history. Ore has the following geotechnical properties: specific gravity is 3.1-4.5, porosity is 0.1-0.5%, water absorption capacity ranges from 2 to 12% and uniaxial compressive strength varies between 11 and 43 N/mm². Geochemical investigations indicate the following composition: BaO = 37.23-97.54%; Fe₂O₃ = 1.06-37.98%; CaO = 0.01-1.09%; SrO = 0.11-2.17%; Hg = 0.01-0.019 ppm and Cd = 0.042-0.1 ppm. These properties, when compared to standard industrial specifications, indicate that some barites are of high grade while others are of low quality. Ore quality appears to depend on such factors as mining depth, presence of certain gangue minerals and location within the barite field.

Keywords: barite ore; structural styles; geotechnical properties; geochemical properties; mineral variability; economic geology; Nigeria

1. Introduction

The importance of barite (BaSO₄) as an economic mineral is seen from its invaluable utilization as a major constituent of drilling mud, in production of industrial wares such as glass, radiation shields and as source for barium-based chemicals. The efficient and economic use of barites in any of these civil and industrial applications necessitates that the ores possess desirable qualities and properties comparable to generally acceptable specification standards. Furthermore, its inherent properties and geologic disposition are pertinent considerations to justify economic exploitation.

The Benue Trough of Nigeria hosts more than eleven barite fields (Fig.1) that are being

scavenged by small-scale and artisanal miners. Despite the earlier reports on the presence of these barite fields, (e.g. Farrington, 1952), their field characteristics, grade, dimensions, and abundance are not known and documented well enough. The paucity of these pertinent data on the mineral deposits continues to constitute a major impediment towards their economic exploitation and utilization. Indeed, almost all previous works on sulphate ore mineralization in the Benue Trough have, hitherto, largely focused on the origin/genesis of the ore veins or their associated saline waters or igneous bodies (Ezepue, 1984; Akande and Abimbola, 1987; Akande et al 1989; Akande and Mucke, 1989; Uma, and Leohnert, 1992; Ekwere and Ukpong, 1994; Tijani, et al, 1996). In this

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paper, we present the results of field studies and laboratory analysis of barite ores from Gabu, Alifokpa and Osina fields. The structural patterns relative to ore host rocks and inherent properties of the barite ores are used to evaluate and highlight their economic potentials and ore genesis.

2. Physiographic framework

The Benue Trough is often described as an elongate shaped, intracratonic basin which is underlain by a thick succession of Cretaceous sedimentary rocks that is punctuated by economic ore veins, volcanic and intrusive rocks and deposited on undulating basement (Cratchley and Jones, 1965; Uma and Leohnert, 1992). The basement rocks are exposed in the south-eastern and north-western extremities of the trough where their peaks generally coincide with the major hydrologic boundaries. The principal area of investigation, Gabu- Osina-Alifokpa barite field, falls within the southern part of the Benue Trough and is located in Yala Local Government Area of Cross River State. It is bounded by latitudes 6°45'N and 6°57'N, and longitudes 8°40'E and 8°55'E (Fig. 2) and spans an approximate area of about 250 km².

The topography, which is part of the major regional land form, the Cross River plains (Bygott and Money, 1975), shows a gradual ascent from the plains, to the south, to highlands, in the central north. Whereas the northern part of the area is characterized by gentle sloping high lands that can attain up to 300 m to 500 m above sea level, the terrain in the southern part of the area is dominated by low lands with elevation well below 300 m. In terms of the climate, the area falls within the geographical region that experiences two major distinct seasons namely, the rainy and dry seasons (Duze and Ojo, 1993). Balogun (2000) noted that the climate of the area is controlled by the seasonal movement of the inter-tropical convergence zone (ITCZ) that leads to contrasting dry and wet seasons. The dry season which extends from November to February, is characterized by high mean temperatures of about 32°C (Gates, 1978) with the period of December to January witnessing some large diurnal variation in temperature due to the harmattan, during which temperatures may fall below 28°C. The mean temperatures may vary widely from the month of January reaching a maximum about May.

The rainy season usually commences in the month of March and ends in October. Rainfall is heaviest during the months of June to September, with dry spell of two to three weeks duration in late July to early August (Monamu, 1975). The annual total rainfall in the region ranges between 1375 mm and 2560 mm (Gates, 1978). The area is drained by four major rivers, namely, Konshisha, Oruaba, Okpauku and Kpa and their tributaries. These major rivers and their tributaries show dendritic and occasionally trellis drainage patterns and flow into the Cross River drainage system. Whereas some of the streams are perennial, others are ephemeral and usually dry up during the dry season. They commonly originate as surface flows from springs that indent the relatively higher elevations from the north towards the southern part of the area. Igbozurike (1975) attributed the drainage pattern to structural inequalities in rock hardness/texture, recent diastrophism and geologic/geomorphic history.

The vegetation is typically mangrove overlapping into the rain forest belt (Gates, 1978) and is characterized by scattered trees with low covering shrubs and grassland. It can be classified as grassland and rainforest types (Igbozurike, 1975) or Guinea savanna (Iloeje, 1965). The vegetation is generally controlled by geology, climatic condition and the distribution of rivers in the area. Areas that are underlain by unconsolidated sandstone and shale are covered by giant green trees and plants, while areas that are underlain by consolidated sediments are characterized by grasses and shrubs. Tall trees and evergreen plants follow the pattern of the major and minor river channels in the area. Surficial soils in the area classify as interior zone of laterite soils following the classification scheme of Iloeje, (1965). The soils are deeply ferruginized with colour grading from dark grey to mottled red. They are generally sticky when wet with some zones of alluvium.

3. Geology

The Benue Trough of Nigeria occupies the major re-entrant into the West African continental margin and its extent is about 80-150 km wide and 800 km long (Uma, and Leohnert, 1992). It extends in the NE-SW direction from the Niger delta, Nigerian part of the Gulf of Guinea to Chad basin in the interior of the West African Precambrian shield. The trough has often been described as an elongate, partly fault-

Barite deposits in the Southern Benue Trough, Nigeria

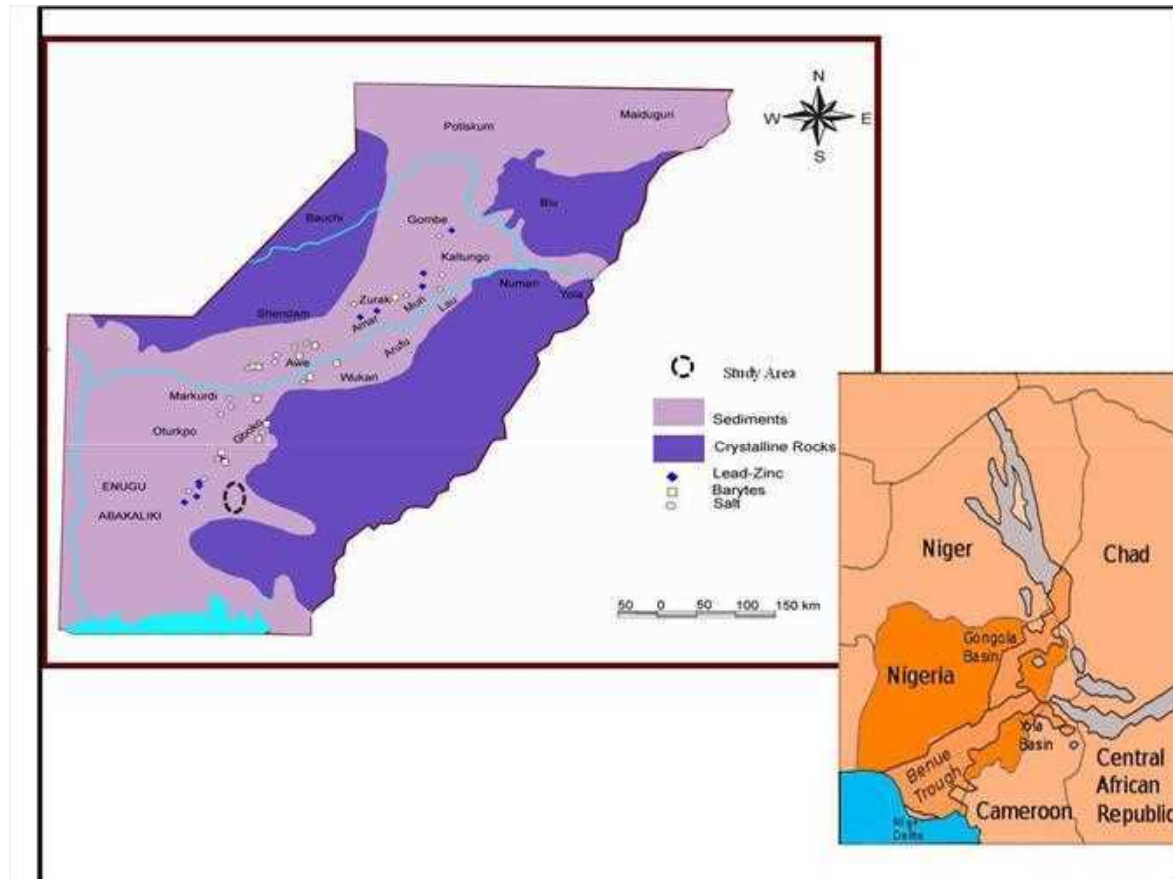


Fig. 1. Distribution of lead-zinc-barite and salt mineralization along the Benue Trough (Modified after Cratchley and Jones 1965). Insert sketch map of Nigeria showing the Benue Trough)

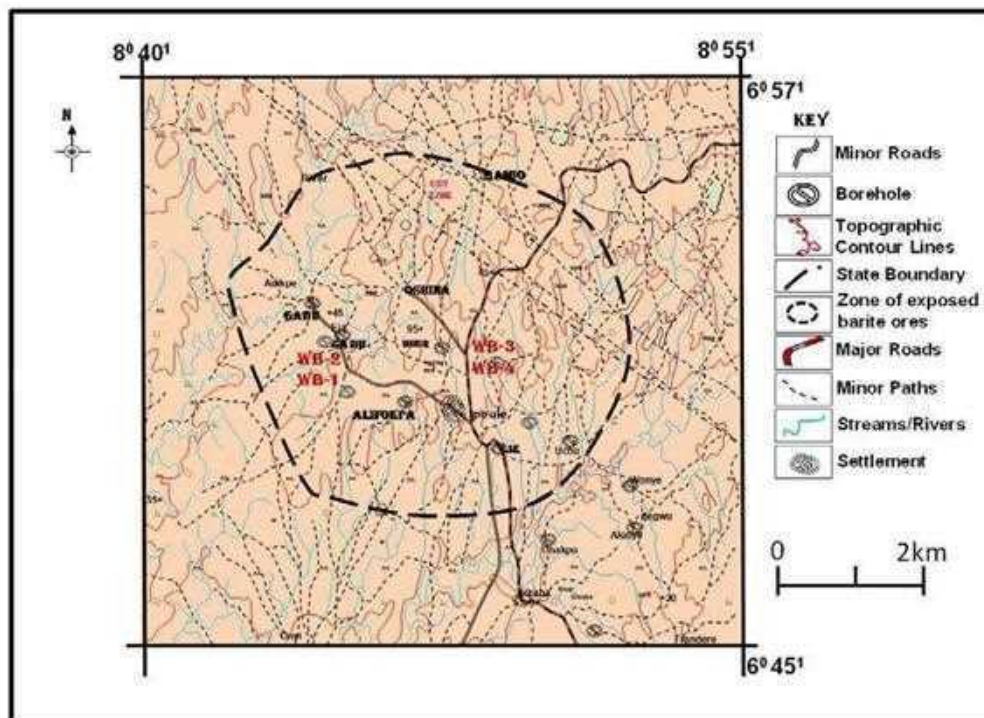


Fig. 2. Geomorphic Map of Gabu – Alifokpa – Osina barite field.

bounded depression occupied by up to 6000 m thick marine and fluvio-deltaic sediments (Benkhelil, 1989). The sediments in the trough are thought to have been compressional or extensionally folded in a non-orogenic shield environment (Wright, 1976, Ugwuonah and Obiora, 2008). Offodile (1989), apparently for the purpose of clarity and convenience, subdivided the Benue Trough into lower/south, middle/central and upper/north geomorphic regions. In the southern (lower) region, there are many structural and depositional elements that include the Abakaliki anticlinorium, Afikpo and Anambra synclines or rift systems. These structural elements inadvertently control and dominate the geologic evolution and litho-stratigraphic architecture of the southern part of the trough.

The tectonic evolution and stratigraphic history of the Benue Trough have been extensively studied, reviewed and discussed and are fairly well known. The discussion that follows attempts to synthesize the presentations of Reyment (1965), Grant (1971), Burke et al. (1972), Olade (1975), Wright (1976), Offodile (1976), Uzuakpunwa (1980), Benkhelil (1989), Kogbe (1989), Ojo (1992), Akande and Mucke (1989) and Akande (1999). These authors agree that the application of Y-shaped triple rift model (RRR) to the break-up of the Afro-Brazilian plate, in early Cretaceous times, best explains the final configuration of the Benue Trough. They argue that the Benue Trough originated as an aulacogen consequent upon the separation of the African and South American plates sequel to the opening of Southern Atlantic in the early Cretaceous. The tectonics and the accompanying depositional processes led to the on-land reactivation of the equatorial oceanic fracture zones, particularly the Chain and Charcort zones, which probably generated successive tensional and compressional stresses in the basement rocks as well as in the overlying sedimentary successions.

During the Santonian, such crustal instability is believed to have been accompanied by widespread and spectacular magmatism, folding, and faulting, which resulted in the creation of the relatively elevated Abakaliki anticlinorium (in the Southern Benue Trough), flanked by two synclines: the Anambra to the west and Afikpo to the east. The Abakaliki anticlinorium then became a positive geomorphic feature, and sedimentation was transferred laterally to the adjoining synclines. The sedimentary infill in the

lower Benue Trough (Fig. 3) (the Abakaliki Basin) records the first tectono-sedimentary cycle that inundated the southeastern Nigeria rift basins. The cycle which probably started with the Alptian-Albian transgression deposited the oldest marine sediments with shelf environment, designated as the Asu River Group (ARG). The ARG consists of mainly thick laminated shales, arkosic sandstones, and subordinate limestones and later with associated volcanics, intrusions and pyroclastics. It represents the first cycle of shallow marine and brackish water terrigenous clastic sediments that lie unconformably over the Precambrian to lower Paleozoic Basement Complex rocks. The Cenomanian is characterized by a regressive continental to transitional marine environment with high clastic sediment influx (e.g., Keana/Markurdi Sandstone in the central Benue Trough), which may have created a wide coastal plain predominantly to the west of the basin (Anambra syncline). This may have led some earlier workers (e.g., Reyment, 1965; Nwachukwu, 1972) to suggest the absence of Cenomanian deposits in the Abakaliki basin. The subsequent Turonian transgressive phase is dominated by the Benue-Trans-Sahara Seaway (Tethian) and accounts for the deposition of extensive fossiliferous black to grey shales, the Ezeaku Group, which consists of marine shales with subordinate limestone (Benkhelil, 1989), as well as several sand bodies ranging from fluvial to marine (Hoque and Nwajide, 1985). The Abakaliki basin is said to have evolved into a (stable) platform with subsidence localized to the western shelf during the period of Turonian transgression. The localized subsidence and gradual migration of the basin axis to the west, coupled with continuous rise in sea level into the Coniacian, led to the creation of the Anambra rift with the deposition of thick upper Turonian-Coniacian Agwu Group. The Santonian folding, and uplifting of the Abakaliki basin displaced the axis of sedimentation and limited the deposition of Santonian to Maastrichtian sediments to the Anambra and Afikpo basins.

The sedimentary packages in the Benue Trough have been affected by, at least two major sets of tectonic features, namely, the pre-Turonian and Santonian. In the lower Benue Trough, the latter tectonic set involved compressional movement along an established NE-SW trend. This may have resulted in the folding, faulting and uplift of the Abakaliki Anticlinorium. The tectonic episode was not

Barite deposits in the Southern Benue Trough, Nigeria

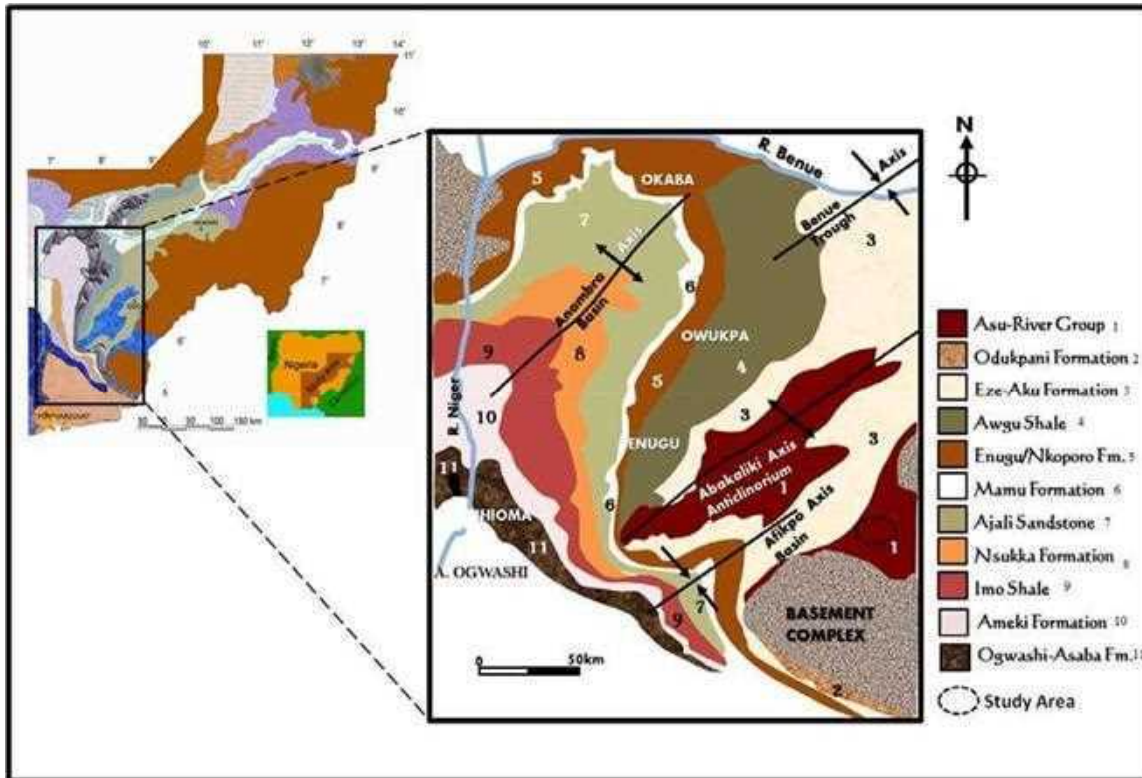


Fig. 3. Regional stratigraphic sequences in Southeastern Nigeria (Modified from geologic map of Nigeria, 1994)

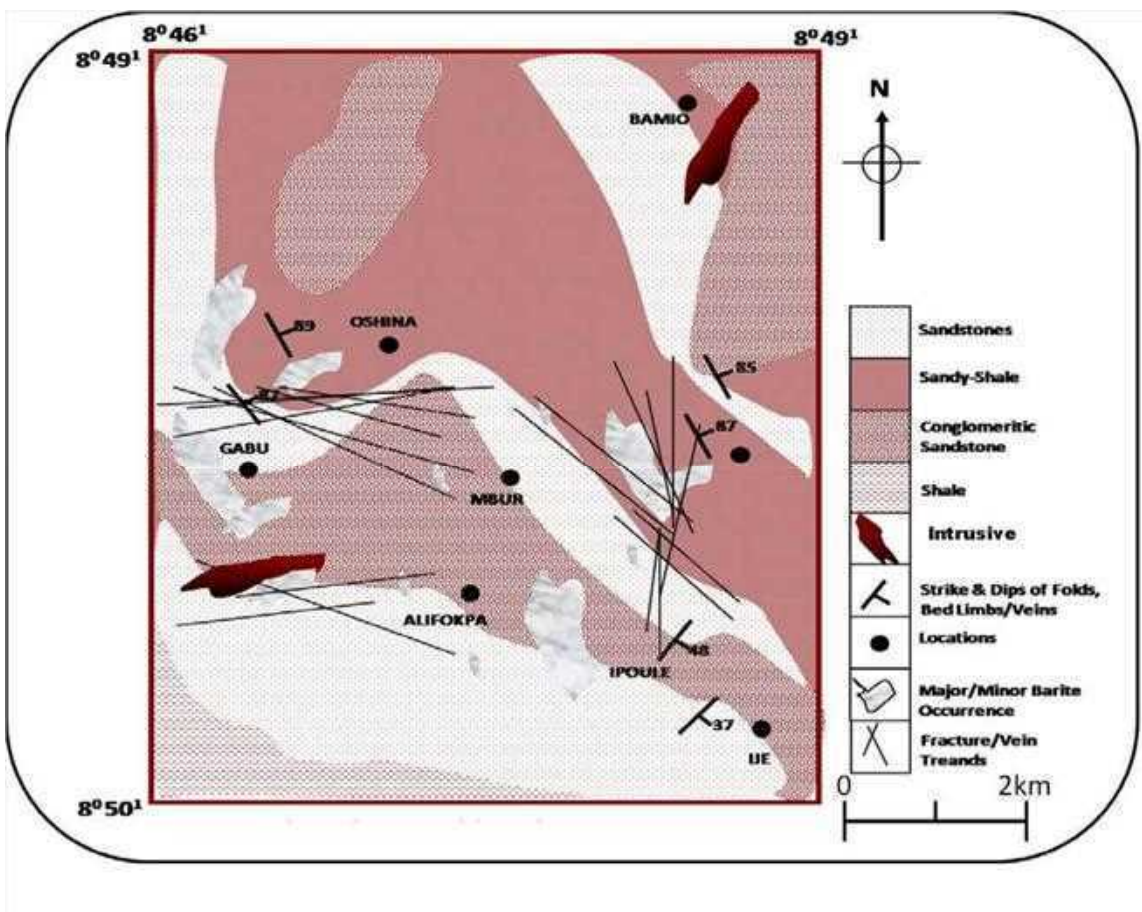


Fig. 4. Geologic map of the study area

only characterized by contemporaneous subsidence of the Anambra platform and displacement of the depositional axis westwards, but also is thought to have been marked by a number of minor intermediate intrusions as well as associated lead-zinc mineralization. The folded and fractured ARG rocks (shales, limestone, sandstone, and their equivalents in the central and northern Benue Trough) are hosts for associated lead-zinc, and fluorite-barite mineralization.

4. Methodology

Field mapping of the area was conducted and it involved the location, description, measurement and sampling of various exposed ores and the associated lithologies, geologic structures (fractures) along mine pits and river channels. Barite ore samples were subjected to geotechnical and geochemical tests in the laboratory to evaluate the grade of the barite ore. The geotechnical tests included specific gravity, porosity, water absorption capacity, and uniaxial compressive strength while the geochemical investigations involved major and trace element analysis of the ore samples. The geotechnical tests were carried out in accordance with the procedure outlined in British standard (BS, 1378, 1990). The geochemical analyses were performed by inductively coupled plasma mass spectrometry techniques (ICP-MS) using Perkin Elmer SCIEX ELAN 6100 instrument and calibration with certified reference materials from National Institute of Standards and Technology (NIST, USA). Appropriate sample preparation for the analysis was carried out on pulverized ore samples by digestion and fusion with aqua regia and wet acid ($\text{HNO}_3 + \text{HCl}$) treatment following the procedure given in Balcerzak (2002).

5. Results and discussion

5.1. Structural styles

Detailed field studies of the barite field show that three major rock types coexist with barite veins in the area (Fig. 4). These rocks include arkosic, conglomeratic sandstone, intrusive bodies and fractured weathered shales. The sandstone is truncated by intrusives and saline water pools are restricted in the shale regions. The contact between barite ore and the sandstone host rock is sharp and devoid of observable wall-rock alterations. There is a notable absence of

the host rock fragments in the barite ore samples, despite their crosscutting field relations. The lithologic association of the different rock types and barite ores indicates that barite veins, as well as stratabound concordant barite beds are widely distributed in the indurated sandstone (Fig. 5A-E). The ores are also occasionally disseminated as swarms of veins and/or globular concretions in weathered/fractured shales. In terms of their spatial and temporal appearance, veins occur predominantly in swarms but occasionally depict block profile and their lateral extents vary widely ranging between 10 cm and 8 m with mean of 1 m. The barite ores occur in association with other minerals such as quartz, galena, sphalerite and feldspar (Fig. 6).

Detailed records of the mineralogy and geochemistry of igneous bodies and of the composition of the saline waters that are known to associate with base metal mineralization in the lower Benue Trough are contained in the works of Ezepue (1984) and Uma and Leohnert (1992). Whereas the felspathic, conglomeratic coarse-grained sandstone, which trends on 310° - 130° azimuth with dip of about 43° , plays host to the barite ore veins in the Gabu-Alifokpa-Osina fields, mineralization of predominantly lead-zinc sulphides and fluoride ores have been reported to occur in shales of the ARG in similarly close geographic association with saline waters and intrusives (Akande and Mucke, 1989 and Akande, 1999). The lithofacies association and occurrences in the ore fields tend to suggest that ore genesis is intricately related to both magmatic activities and hydrodynamics of the associated hydrothermal fluids. A closer examination of the veins relative to the host rock indicates that open (cavity) space filling as opposed to replacement phenomena is the dominant mode of ore precipitation and concentration, a situation that explains the absence of the host rock fragments in the barite ores, despite their crosscutting field relations. Back and Clark (1993) thought that the paucity of host rock fragments in ores found in such host rocks predicates the near surface position of the host prior to deep burial during ore genesis. A further support to the open (cavity) space filling is the fact that the ore-host rock contact is devoid of observable wall-rock alterations. According to Paradis et al (1998), the seemingly minimal nature or complete non-existence of wall-rock modification obviously points to the low thermal conditions that prevailed during ore mobilization and precipitation from hydrothermal fluids

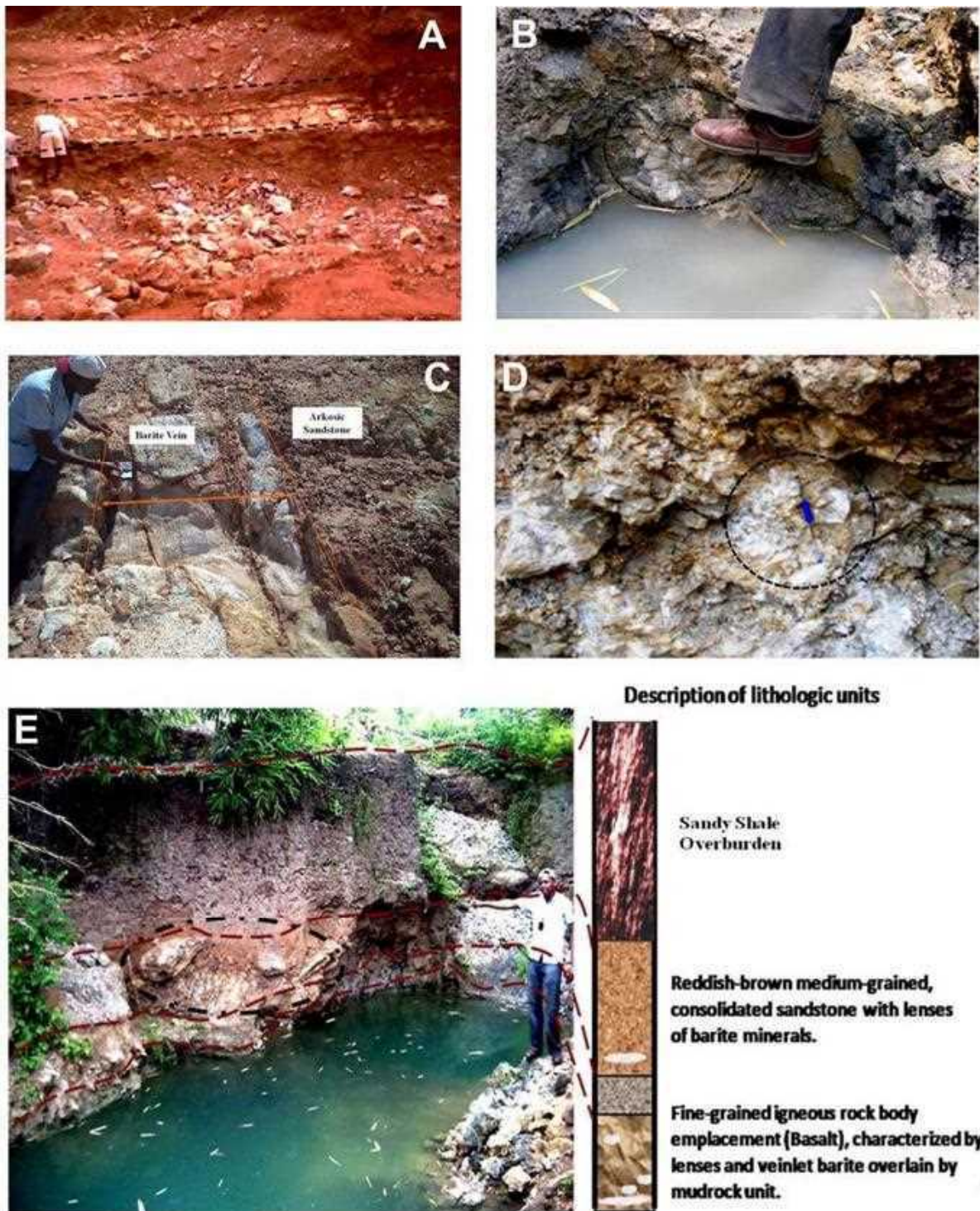


Fig. 5. A. Fractured stratabound barite ores (delimited by the dash lines) occurring as 2-5 m thick beds in sandy shale at Alifokpa (active mine pit). B. Fractured barites and barite nodules (dashed circles) occurring in shales at Oshina (abandoned mine pit due to groundwater flooding). C. Mining pit face showing fractures and associated barite-mineralized vein in the conglomeratic sandstone unit at Alifokpa site; the first author is measuring the attitude of the ore vein and fractures. D. Mining pit face showing fractures and associated barite-mineralized vein in the conglomeratic sandstone unit at Alifokpa site. E. Composite photograph and lithologic section in an abandoned mine pit at Gabu. Note the disseminated barite ores in the basal sandstone units (in dash circles).



Fig. 6. Barite ore samples from Gabu/Alifokpa/Osina fields; Cross-River State, Nigeria. (A). Bladed barite matrix on sphalerite from Gabu mine, Ogoja. (B) compact yellowish dark brown metallic crystals of sphalerite intermixed with galena and quartz crystals; Gabu mine, Ogoja. (C) Tabular, clear/transparent crystals of barite from Alifokpa mine, Ogoja. (D) Tabular crystal of barite characterized by well-defined cleavage planes; Gabu/Osina mine, Ogoja. (E1) crystals of barite occurring in association with quartz and galena and (E2) crystals of barite occurring in association with feldspar.

Trend analysis of the ore veins (Fig. 7A-B) reflects two dominant fracture direction sets, namely, NW-SE trend that is orthogonal to the main axis of the Benue Trough, and a comparatively younger, crosscutting N-S trend. The NW-SE-trending ore veins predominate over N-S trending veins, and the dips of the barite veins are generally steeper than 80° . The near vertical dip of the veins predisposes the ore deposits to manual extraction and vertical stripping using open pit method. Similar observation of the spatial distribution of rock-types and association, vein morphology and presence of brines has been noted by other workers, who studied other fields in the Trough. These observations as well as fluid inclusion studies have generated a lot of controversy in the past with respect to ore or saline water genesis and two contrasting hypotheses have emerged.

The first one (e.g. Orajiaka, (1964); Ezepeue, (1984); Uma and Leohnert, (1992); Tijani, et al, 1996) believes that igneous intrusions caused changes in the geochemistry of the surrounding rocks, releasing important masses of volatiles. Changes in pressure and temperature conditions

probably entrained precipitation and concentration from fluids to form the ore. The authors supporting this hypothesis apparently came to this conclusion in an attempt to explain the presence of igneous bodies in close vicinity to the ore veins. The proponents of the second hypothesis, such as Olade and Morton (1985), Ekwere and Ukpong (1994), Akande and Abimbola (1987), Akande et al. (1989), think that the brine was contemporaneous with deposition of the host rocks, prior to the magmatic processes and to the concentration and precipitation of ores and that ore concentration and precipitation occurred in response to basinal fluid expulsion induced by gravity compaction and hydromechanics and subsequent fluid interaction with the arkosic sandstone which provided the source of base metal concentrates. While these authors accept the epigenetic formation model, they absolutely disagree with the magmatic hydrothermal brine source. They, however, opine that much of the brine was contemporaneous with the deposition of the host rocks, prior to the magmatic processes and concentration and precipitation of ores. Such

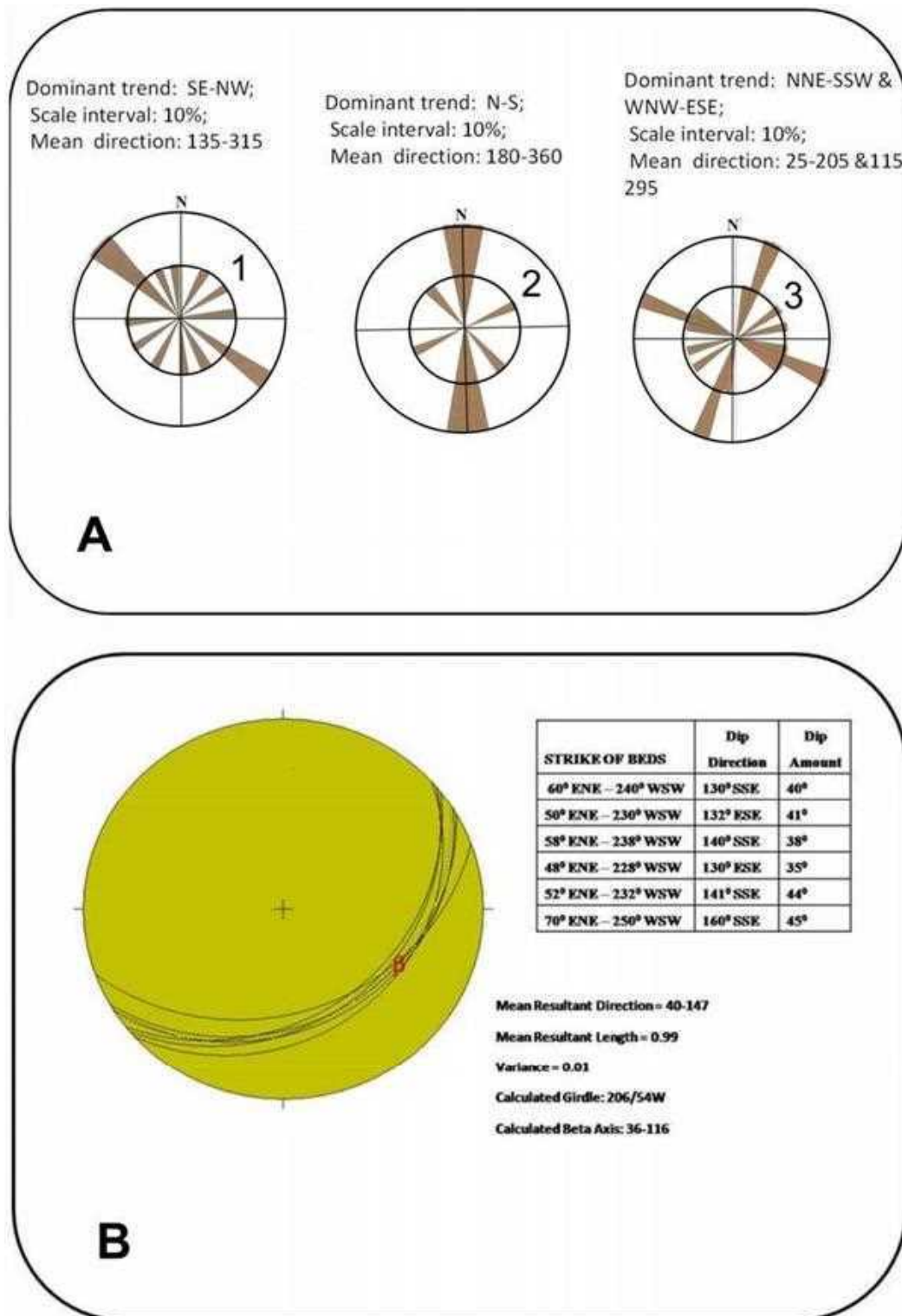


Fig. 7. A Rose diagram plots of barite veins and major fractures (joints) for: (A) Gabu, (B) Alifokpa and (C) Osina barite fields. B. Stereonet plot of ore veins and major fractures (joints) attitudes for Gabu, Alifokpa and Osina barite fields

opinion attempts to invoke plausible explanation for the continuous presence of brines within the geographic location of the veins.

From this study, it is adduced that the intrusive rocks are not in themselves mineralized and the migration pathway of the supposed magma-derived hot fluid is yet to be resolved. Also, there has not been any known igneous body directly correlated or linked to a particular mineralization. Again, there seems to be disagreement, even among the proponents of the magmatic-hydrothermal concept on the provenance and migration pathway of ore-forming fluid. For example, Farrington (1952) believes that intrusive rocks induced the formation of the ore directly, while Reymont (1965) thinks that volcanic rocks generated the hydrothermal fluids from which the ore was deposited and Oraziaka (1962) considers either of the two rock types (sedimentary and magmatic) as possible sources of fluid. It can be considered that the basinal expulsion genesis model is more favoured. However, the indications of magmatic processes, such as the presence of igneous bodies, add yet another question to the model. Indeed, the contending models to explain the mineralization in Benue Trough reflect the complex history of basin formation and dynamics. It follows, therefore, that a holistic genetic model, however, has to account for the coexistence of the mineralization with igneous bodies and brines and the varied morphology and mode of occurrence of the ores. Such model has to be in line with the tectono-sedimentological evolution of the southern Benue Trough. It is then concluded that the Santonian tectonic event mobilized the base metals contained in the Asu River Group sediments due to magmatic intrusions that generated hot brines. The flow of the hydrothermal fluid through fissures and fractures, which may have been created by pre-Santonian tectonic events concentrated and precipitated the ores. Alternatively, the magmatic activity may have, as well, served as the source of the base metals, which, in the case of flow through the rocks witnessed ionic differentiations that led to the concentration and formation of the ores. The remnant hydrothermal fluids generated the brine ponds. Irrespective of the source of the base metals, the present structurally-controlled occurrences of the ores in varied vein trends as well as stratabound mineralization, are indications that ore genesis may have followed closely the tectonics and

deformations of basin evolution in the Cenomanian (Benkhelil, 1989; Ojo, 1992), Turonian and Santonian (Nwachukwu, 1972).

5.2. Economic potential

The physical appearance of the ore samples show that the ores exhibit several colours, ranging from colourless-white, for samples that occur at greater depths (more than 30 m from the ground surface), to light yellow and greenish, for those found closer to the ground surface. In addition, some ore samples show variegated colours (Fig 6) apparently due to the presence of the accompanying minerals such as galena, feldspar and quartz. On the other hand, the colours shown by samples closer to the surface could be attributed to the oxidation and alteration of the sulphates. All samples generally display typical transparent, glassy luster.

Results of the geotechnical testing of the ore samples are presented in Table 1. From the table, it can be seen that the specific gravity, porosity, and water absorption capacity vary from 3.1 to 4.5 g/cm³, 0.1 to 0.5%, and 2 to 12%, respectively. The uniaxial compressive strength is in the range 11-43 N/mm². It is obvious from Table 1 that there is a general increase in the value of the specific gravity of the ores with increasing depth of occurrence. Such contrasting property at increasing mining depth, perhaps demonstrates the occurrence of higher grade ores as mining proceeds to greater depth. It can also be noted from Table 1 that the uniaxial compressive strength (UCS) of the ores varies widely. However, there is a decreasing trend in values with increasing depth, which is obviously contrasting with the specific gravity trend. A closer look at the high strength barites indicates that they are admixed with associated minerals (Fig 3) and occur closer to the ground surface in the veins, while the low strength types occur at greater depths and are devoid of gangue minerals. The higher strength recorded for barite samples that are richer in the gangue minerals is attributed to increased frictional resistance to compression due to co-recrystallization of barite with relatively harder minerals such as quartz. It would appear therefore, that ore quality (when judged from the point of view of specific gravity values and presence/absence of gangue), is higher at increasing depth in the barite fields.

The results of the geochemical tests (Table 2) lend support to the improvement of barite quality with increasing depth in the veins. For example,

Barite deposits in the Southern Benue Trough, Nigeria

Table 1. Geotechnical properties of barite ores from Gabu-Alifokpa barite fields

Sample location	Vein depth(m)	Specific gravity	Porosity (%)	Water absorption capacity (%)	Moisture content (%)	Uniaxial compressive strength(N/mm ²)
Gabu						
A	0-32	3.1*	0.3	8	0.3	43
B	33-36	3.2	0.2	2	0.8	38
C	37-45	3.6	0.2	7	0.5	23
D	46-48	4.2	0.1	6	0.2	20
Alifokpa						
A	0-29	2.9	0.5	12	0.4	40
B	30-34	3.3	0.3	10	0.6	34
C	35-43	3.9	0.2	9	0.4	14
D	44-47	4.5	0.1	2	0.2	11
Osina						
A	0-24	3.1	0.2	5	0.24	39
B	25-35	3.7	0.4	11	0.34	31
C	36-37	3.7	0.4	11	0.45	18
D	38-39	4.3	0.3	4	0.21	12

*Average measured value of five samples for the depth range

Table 2. Geochemical properties of barites ores from Gabu-Alifokpa barite fields

Sample location	BaO(%)	Fe ₂ O ₃ (%)	SiO ₂ (%)	F(%)	Al ₂ O ₃ (%)	CaO(%)	SrO(%)	Hg(ppm)	Cd(ppm)
Gabu									
A	49.64	24.54	3.43	19.8	0.87	1.07	0.12	0.01	0.1
B	53.45	34.87	1.05	9.67	0.56	0.04	0.19	0.03	0.1
C	79.76	5.76	0.45	13.56	0.51	0.01	0.13	0.02	0.4
D	75.54	8.06	1.38	12.98	0.43	0.01	0.13	0.05	0.2
Alifokpa									
A	37.23	27.98	3.09	23.93	8.34	0.05	0.16	0.012	0.12
B	43.56	35.13	2.36	13.87	1.99	0.03	2.17	0.018	0.23
C	65.65	15.65	0.04	17.46	1.43	0.01	0.12	0.016	0.37
D	74.87	10.53	0.07	13.73	0.52	0.02	0.11	0.011	0.25
Osina									
A	44.51	35.51	3.06	10.45	2.88	1.09	2.13	0.01	0.22
B	54.75	22.43	2.31	14.61	5.65	0.03	0.12	0.017	0.19
C	62.56	24.56	0.45	9.67	1.67	0.03	0.13	0.019	0.34
D	73.26	20.08	0.07	6.98	0.43	0.01	0.15	0.014	0.42

some of the chemical components, such as BaO, mimic the trend of specific gravity values showing consistent increase in grade with increasing depth. In contrast, the contaminants, such as Fe₂O₃ and SiO₂ show decreasing contents with increasing depth similar to the trend exhibited by UCS values. The increase in the specific gravity of the ore samples with increasing depth of occurrence is thus, attributed

to the increasing content of barium (atomic mass: 137) and other heavy elements and decreasing content of comparatively lighter elements such as silicon (atomic mass: 28) and iron (atomic mass: 56).

The industrial specifications for different use of barites vary widely (Table 3), even among manufacturers or industries. In order to assess the quality of barite ores from the studied fields,

the measured geotechnical and geochemical properties are compared with some generally acceptable standards for processed barite as there are no standards for mineral ore. The comparison shows that the geotechnical properties of some of the ore samples compare reasonably well with the American Petroleum Institute (API) specification standard for barite used as a weighting agent in oil well drilling mud, which requires at least a specific gravity of about 4.2. For example, the ore samples that occur at greater depths in the veins have specific gravity above 4.2 and thus, are suitable for oil

well drilling applications. In addition, the ores with high specific gravity show lower concentration of heavy metal contaminants. For example Hg (0.01-0.019ppm) and Cd (0.042-0.1 ppm) are well within the permissible limits for barites used in drilling mud. In contrast, some of the ore samples particularly those that occur close to the ground surface and/or are contaminated with gangue minerals, fall short of this requirement as their specific gravity values fail to attain the recommended values of 4.2, and hence, are of low quality.

Table 3. API and ASTM general specification standards for various uses of barite ores.

Standard	Specific gravity	BaSO ₄ (BaO)	Soluble alkaline content (Sr and Ca)	Heavy metals	Fe ₂ O ₃	SiO ₂	Al ₂ O ₃	Moisture Content
API(mud)	4.2(min)	92%(min)	250ppm max	N/A	N/A	N/A	N/A	N/A
ASTM:								
Glass	N/A	95%(min)	N/A	N/A	0.15% (max)	1.5% (max)	0.15% (max)	N/A
Pharmaceuticals	N/A	97.5% (min)	<0.01ppm	<0.001ppm (max)	N/A	1.5% (max)	N/A	N/A
Paint	N/A	95% (min)	0.2% (max)	N/A	0.05% (max)	N/A	N/A	0.5%(max)
Chemicals	N/A	92% (min)	1% (max)	N/A	1% (max)	N/A	N/A	N/A

N/A – Not applicable

The results of the geochemical analysis lend support to the varied grades of ores in the barite field as shown by the wide variations in the concentrations of the geochemical facies. Whereas the ore with low concentrations of water soluble elements: CaO (0.01-1.09%), SrO (0.11-2.17%), and heavy metals: Hg (0.01-0.019ppm) and Cd (0.1-0.42ppm) have desirable qualities of good grade barites for drilling mud and others that have higher concentration of these elements need beneficiation to attain acceptable limits. Also, the concentrations of BaO (37.23-97.54%) and Fe₂O₃ (1.06-37.98%) indicate that some of the barites classify as high grade when compared to the ASTM specification for barites utilized in pharmaceuticals, paints, glass and chemicals and therefore have good potential for use as source of barium-based chemicals and in the production of industrial goods. The low grade ores, which fail to meet API and other specifications, occur close to the

surface and are characterized by the presence of gangue and ore minerals such as galena, quartz, feldspar and sphalerite. It would appear that ore quality is dependent on mining depth and is impaired by the presence of the associated minerals. In addition to mining depth, ore grade distribution within the barite field varies with location as higher grade ores appear to predominate towards the northern part of the prospect around Gabu/Osina area while ores of comparatively lower quality and contaminated by the associated minerals occur southwards, in the Alifokpa area (Fig 2 and Table 1).

6. Conclusions

Field studies and laboratory analysis of ore samples from Gabu, Osina, and Alifokpa barite field allowed for the assessment of ore morphology and occurrence, as well as the

industrial properties in order to evaluate the economic potential and usability of the ores. From this study, the following conclusions are drawn:

1. Barite occurs predominantly in steeply dipping NW-SE trending veins and less frequently in N-S trending veins as well as concordant stratabound deposits spread in arkosic sandstone and shales.
2. The barite ore shows the following geotechnical properties: specific gravity: 3.1-4.5, porosity: 0.1-0.5%, water absorption capacity: 2-12% and uniaxial compressive strength: 11-43 N/mm².
3. The analyses of ore chemistry indicate the following concentrations: CaO (0.01-1.09%), SrO (0.11-2.17%), Hg (0.01-0.019 ppm), Cd (0.1-0.42 ppm), BaO (37.23-79.76%) and Fe₂O₃ (5.76-35.51%).
4. Whereas the geotechnical and geochemical properties of some of the barites compare fairly well with API and other industrial specifications implying that the ores are of good grade, others fail to meet the requirements and hence, will require processing and beneficiation.
5. Ore quality is controlled by the depth of occurrence, presence or absence of associated minerals and the geographic location of the vein within the ore field.

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Identification of a fossil trunk from Achlada Mine (Florina, Greece) newly installed at the entrance of the coal mine

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Abstract: Huge trunks were discovered in the open cast lignite mine of Achlada, which is situated in the late Neogene intramontane basin Florina-Ptolemais-Servia, northwestern Greece. One stump of six meters tall found there was removed and installed in front of the administrative building of Achlada mine, while the remained part was preserved in situ. By palaeoxylotomical study on a sample of the exposed stump, the original tree was identified as *Taxodioxylon taxodii*, which represents the fossil equivalent of the extant *Taxodium distichum*. A few considerations on the palaeoenvironment in which the Achlada lignite deposits were formed are added.

Keywords: Late Miocene, xylite, *Taxodium*, palaeoenvironment, Achlada, Greece

1. Introduction

Achlada is an important Neogene coal deposit (lignite deposit) in Greece, beside of Florina, Ptolemais, Megalopolis and Drama, formed within an intramontane basin (Florina-Ptolemais-Servia Basin), which extends over a distance of 120 km from Bitola (Macedonia) to Servia in northwestern Greece (Fig. 1), generated by the neotectonic activity within this area.

The basin developed in the Late Miocene in the Pelagonian Zone, the westernmost part of the Internal Hellenides, in response to NE–SW extension, and was subsequently fragmented, during Pleistocene, resulting in some sub-basins, *i.e.*, Florina, Ptolemais and Servia (see Pavlides and Mountrakis, 1987). The Neogene-Quaternary sediments that fill the Achlada basin unconformably overlay both Palaeozoic metamorphic rocks and the Mesozoic crystalline limestones (Oikonomopoulos et al., 2008). This stack of sediments can be divided into three lithostratigraphic units starting with coarse sediments (basal conglomerates with elements of metamorphic rocks), followed by a coaly sequence of marls, sandy marls, sands, clays and lignite layers. The age of this unit has been

defined by fossils (Velitzelos and Petrescu, 1981) as latest Miocene (Pontian) to early Pliocene (see Fig. 2). The middle unit is entirely a coaly one and contains thick lignite beds in alternation with clays, marls, sandy marls, and sands. The Ruscian age for this unit was obtained by pollen analysis and microfaunal and macrofaunal study (Weerd, 1979; Koufos, 1982; Ioakim, 1984). The upper unit is terrigenous and represented by Quaternary terrestrial and fluvio-terrestrial conglomerates, lateral fans and alluvial deposits (Pavlides and Mountrakis, 1987). The tectonic structure of Achlada coaly sediments respects the tectonic style of the basin, defined by normal faults with two main directions: a NE–SW direction and a NW–SE one, the latter being almost perpendicular on the first one (see Pavlides and Mountrakis, 1987).

Anyway, Oikonomopoulos et al. (2008) remarked the scarcity of detailed researches, especially palaeobotanical, and started a multidisciplinary study (palaeocarpological, palynological, coal-petrographic, mineralogical and sedimentologic) trying to interpret the palaeoenvironmental conditions associated with Achlada lignite deposits, as well as the role of the palaeoclimate in the evolution of the

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particular lignite deposits. Maybe palaeontological studies would still be useful, even if few animal remains like fossilized cocoons and coprolites were observed but no shells were found. Anyway, Weerd (1979) described Early Ruscinian rodents and lagomorphs (Mammalia) from the lignites near Ptolemais.

Even if the study of Oikonomopoulos et al. (2008) cover only a part of the coaly sequence, it is obvious that the palaeobotanical, coal-petrographic, sedimentologic, and mineralogical features of the Achlada lignite deposits suggest that the general area was a floodplain environment in the Neogene to Quaternary, which was traversed by a principal meandering river system. The lignite-bearing sequences are marked by matrix-dominated lithotype, or by mixed xylite-rich/matrix lithotypes and xylite horizons and constitute the main characteristic of the Achlada coal deposits, and occur in both sections with either autochthonous or alloch-

thonous origin. Both length and frequency of the xylite horizons indicate the occurrence of many independent xylite bodies that significantly contributed to peat formation (Oikonomopoulos et al., 2008).

Palynological and palaeocarpological analyses revealed that the peat-forming vegetation was represented by a forest of mixed type, typical for reedmoor in the lower unit to open water in the upper unit. Peat was accumulated in a wet forest swamp under telmatic to limno-telmatic conditions and increased inorganic influx of clay minerals which prevail in all the samples, with illite-muscovite being the dominant phase, followed by kaolinite and chlorite. The combination of mineralogical composition and the plant communities identified in the studied area, suggests a humid and warm climate conditions during the Achlada peat accumulation (Oikonomopoulos et al., 2007).

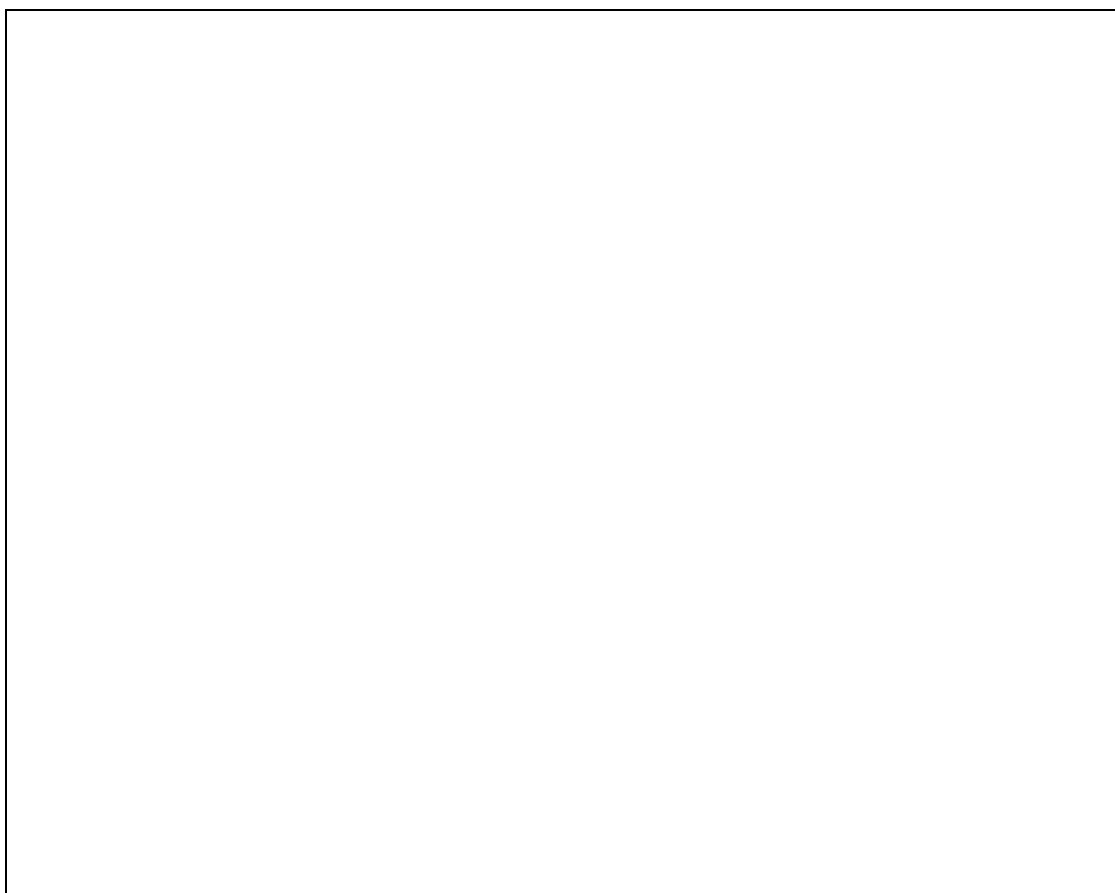


Fig. 1. Geological map of the Florina Basin with the location and generalized lithological column of (a) the study area (after Pavlides and Mountrakis, 1987) and (b) Achlada lignite deposits, NW Greece (reproduced from Oikonomopoulos et al., 2008).

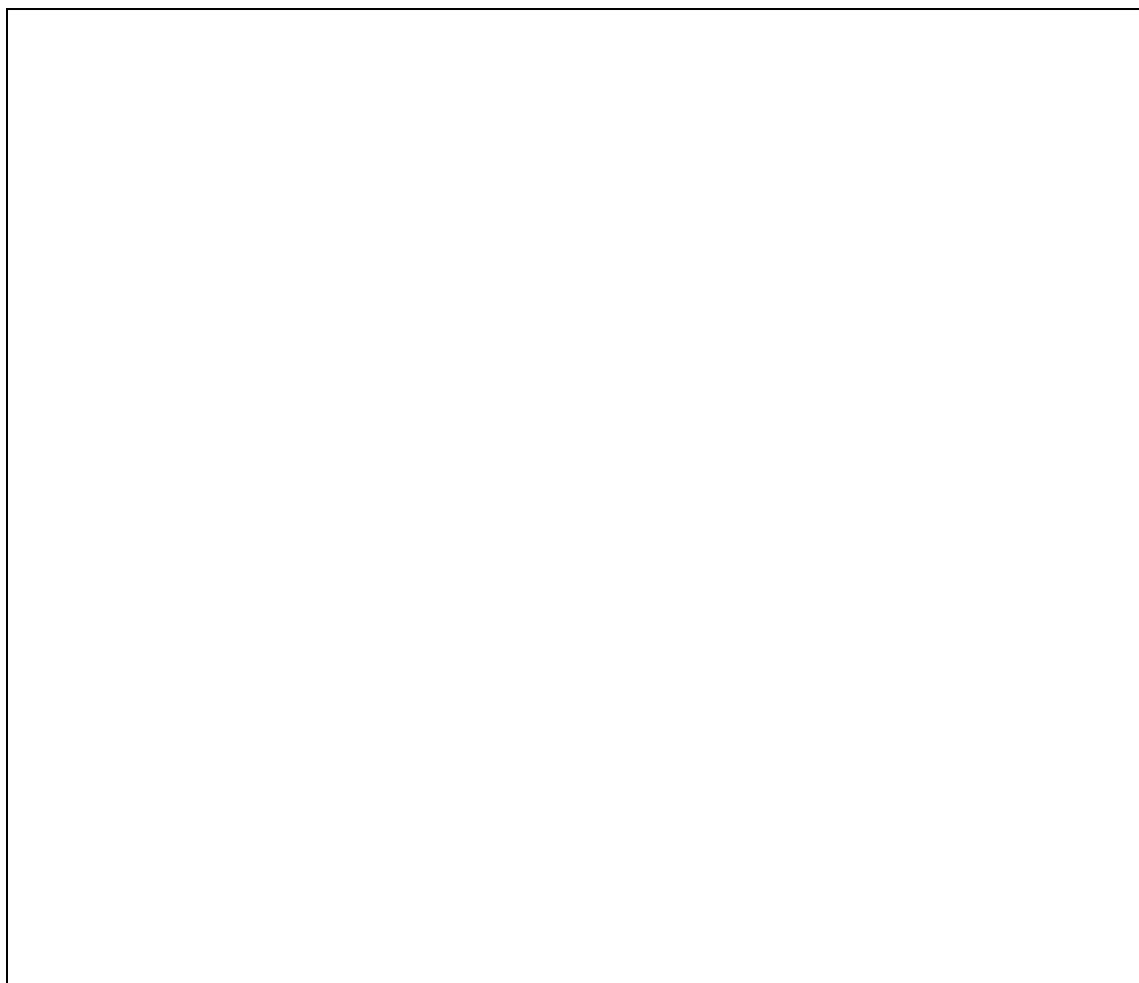


Fig. 2. Latest Miocene-Pliocene chart (from Harzhauser and Piller, 2007, modified according to Walker and Geissman, 2009).

As a whole, the coal-generating flora is poor in species and specimens. A preliminary palynological analysis at Achlada lignite seams shows the following association: Phytoplankton (*Ovoidites*) and pollen of Conifers (Taxodiaceae, Polypodiaceae, Pinaceae), of Angiosperms Monocots (*Sparganiaceae*, *Gramineae*, *Cyperaceae*, *Palmae*) and Dicots (Magnoliaceae, Cyrillaceae, Compositae, *Fagus*, *Acer*, *Tilia*, *Myrica*, *Carpinus*, *Alnus*, *Betula*, *Ulmus*). The palaeocarpological analysis, also informal, showed: Epiphytes, Algae, Nymphaeaceae, *Potamogeton*, *Batrachium*, *Sparganium*, *Decodon*, *Epipremnites*, *Glyptostrobus*, *Actinidia*, *Meliosma*, *Pterocarya*, *Rubus*, *Sambucus*, *Symplocos* (Oikonomopoulos et al., 2008).

The presence of lignified wood fragments of various sizes is well known in this coal deposit and Oikonomopoulos et al. (2008) gave particular emphasis “to the presence of xylite, which is found in large amounts in the Achlada lignite deposits” as “well-preserved horizontal

xylite layers of various thickness (from 2-12 cm up to 15 m) or isolated xylite parts present in the lignite mass” as fragments of branches of various size, stems or roots with well-preserved tissue. Big stems, sometimes in growing position in a huge inorganic clayey bed, often continues into the overlying lignite horizon, suggesting an autochthonous character.

In such conditions, a big fossil trunk of about 10 m long and 2 m in diameter was recuperated from this area and a portion of 6 m was installed upright just in front of the administrative building of Achlada mine (Fig. 3). The remained part of 4 m was preserved *in situ*.

2. Materials and methods

The found fossil wood is weakly coalified (it could be better considered mummified) and probably an effort of conservation will be necessary since, exposed to air, temperature and wetness variations, it will be destroyed.



Fig. 3. The stump exposed in front of Achlada mine (left) and detail (right)

As it was specified, a portion of ca. 6 m of the found fossil trunk was installed as a stump, like a trophy, in front of the administrative building of Achlada mine, while the other portion, of about 4 m, remained in the original place. A palaeoxylotomical study was necessary at least to know the original tree.

After field observations and photos, a small sample from the outer part of the trunk was taken by one of the authors for palaeoxylotomical study. From it, standard oriented sections were obtained by two methods: firstly a small piece was boiled in water to become softer, then cut with a razor blade and mounted in glycerine jelly; secondly, oriented polished surfaces were impregnated with a synthetic resin (araldite) and thin section have been made by the classic petrographic method.

The microscopic study was made in transmitted light using a Nikon Eclipse 400 microscope, and a camera. The anatomical descriptions were made in accordance with the IAWA standard for softwood (IAWA Committee, 2004). The thin sections described herein are deposited in

Velitzelos Collection, NKUA. (National and Kapodistrian University of Athens).

3. Results

The investigations of the fossil trunk, both macroscopic and at the microscope (Plate I, Plate II) led to the following diagnostic:

Family **Cupressaceae** RICH. ex BARTLING, 1830

Subfamily **Taxodioideae** ENDL. ex K. KOCH 1873

Genus ***Taxodioxylon*** (HARTIG) GOTHAN 1905

Taxodioxylon taxodii GOTHAN 1906

3.1 Origin and macroscopic description

The massive trunk (originally more than 10 m in length, at least 2 m in diameter) was found in the coaly formation from Achlada open cast

mine, in a level of Late Miocene age (equivalent to Pontian). The studied sample represents a small fragment taken from the outer part of the huge trunk. It is grey in color, fibrous in texture and with conspicuous annual rings, suggesting a coniferous wood. The wood seems not to be very lignified, still keeping a typical coniferous smell.

3.2 Microscopic description

The growth rings are distinct, marked by 2-6 thicker-walled cells flattened in the final wood, in contrast with the early wood constituted from bigger lumened thinwalled cells, which are very compressed and folded. The rings are 24-35 or more cells wide and the transition from early to late wood is gradual. Normal resin ducts are totally absent.

The tracheids are polygonal in cross section, usually rectangular in latewood, showing rounded lumina with radial/tangential diameters of 30-50(75)/40-60 μm , more tangentially flattened in the late wood where the radial diameter is less than 40 μm . The cellular wall is 3-6 μm thick and the double wall is thicker (7-10 μm) in the final wood. There are 1-6 radial regular rows between two rays. Their frequency on square mm is difficult to count because of the folded structure. The tangential pitting is present and is represented by small bordered pits of 8-12 μm in diameter, with apertures of 3-4 μm , spaced and slightly irregularly arranged in a single vertical row. The radial pitting is badly preserved, is represented by abietineous pits 12-16 μm in diameter and round apertures of 4-5 μm in diameter, are spaced arranged in 1 vertical row on the tracheids of transitional and late wood, when are smaller than 12 μm in diameter. On the tracheids of early wood, the pits are bigger, arranged in 1-2 vertical rows and closer to each other but not contiguous. *Crassulae* were not observed, most probably these are absent.

The axial parenchyma is hardly discernible in cross section, since it is similar to the tracheids but sometimes is filled with resin. However, it is fairly frequent in the longitudinal sections, as vertical strands of rectangular cells with 1.5-3 μm -thick double walls. The transverse end walls are 1-3 μm thick or even thicker, smooth or nodular, with 3-5(7) nodules. The resin content appears as small or large balls and plugs with rounded endings.

The rays seen in cross section are folded in the early wood, but rectilinear in the latewood

and are constituted by rectangular cells sometimes crushed laterally. Vertically, they are uniseriate, rarely with short biseriations, and are of 1-27 or, more frequently, 3-19 cells tall. Seen tangentially, the ray cells are circular or vertically elliptic and usually empty. The density is of 3-6-10 rays on tangential mm. The rays are homogeneous, the parenchyma ray cells are 15-20 μm high, with the marginals sometimes taller (20-24 μm); the horizontal walls are relatively thick (2-4.5 μm double wall), smooth and probably pitted, the marginals with outer wall slightly wavy. Their tangential walls are vertical, oblique or slightly arcuate and thin (0.5-1.5 μm), smooth to slightly rugose or delicate nodular. Indentures are indistinct or absent. The cross fields are usually blurred, pitting is very indistinct and hardly can be guessed 1-2(3) blurred taxodioid pits of 6-7 μm in diameter, with rounded to oval borders and small apertures. They are arranged in horizontal or slightly diagonal pairs.

4. Discussions

The xylotomical study of this specimen allowed the identification of a combination of features, typical for the Cupressaceae of "taxodiaceous" type. By comparison with other extant or fossil "taxodiaceous" taxa, we found that the investigated specimen shows most affinities with the genus *Taxodium* Rich, respectively of *Taxodioxydon* as a fossil genus, because of the presence of some typical structural details: no resin ducts, tracheidal pitting, taxodioid cross fields, presence of axial parenchyma. And, despite some supposed confusion inside the *Taxodioxydon* as fossil morphogenus concerning the modern affinities (Dolezych and van der Burgh, 2004), the generic identification is correct.

Although the wood is badly preserved, the studied specimen shows the presence of uniseriate, rarely biseriate radial pitting and typical taxodioid cross fields. The parenchyma presents the transverse end walls ("horizontal walls") relatively thick and nodular, with 3-5(7) nodules, even if sometimes are smooth, details which are typical also to the extant *Taxodium*, a genus with few species usually living in wet habitats. *Taxodium distichum* (L.) RICH. is living now in Southern USA, Mexico and Guatemala, in riparian and wetland habitats, usually occupying swampy regularly-flooded

forests from the south-eastern area of the USA. Another species *T. mucronatum* TEN. (Montezuma Cypress) prefers not too wet environment and is known better as an element of mountain swampy forests from Mexico (Teodoridis and Sakala 2008).

The comparison with fossil taxodiaceous woods already described showed that there are some affinities with *Taxodioxydon cuddalorensis* RAMANUJAM, 1960, but this one presents bi- and triseriate radial pitting on the tracheids. Most similar seems to be *Taxodioxydon taxodii* GOTHAN (as was described by Zalewska 1953, Greguss 1967, Gottwald 1966, 1992, Iamandei et al. 2001a,b,c), a perfect equivalent of the extant *Taxodium distichum* (L.) RICH., considering the radial pitting, the cross fields and other details.

The fossil wood studied here is devoid of resin ducts, has diffuse parenchyma with nodular horizontal walls and tracheidal pitting uni- to biseriate without crassulae, uniseriate rays and typical taxodioid pits cross fields with 1-2 pits per field, and based on this is identified to *Taxodioxydon taxodii* GOTHAN 1906.

5. Conclusions

The big trunk found in Achlada mine area and partially exposed as a stump of 6 m tall was identified as *Taxodioxydon taxodii* GOTHAN, a fossil tree equivalent to the extant *Taxodium distichum* (L.) RICH. as nearest living relative, named today the Bald Cypress living in Central America and the south-east area of North America, but which, during the Miocene, was one of the largely spread “taxodiaceous” taxa in Europe, very implied in the Neogene carbogenesis (Țicleanu, 1992; Teodoridis, 2001, 2002; Kvacek et al., 2004).

The late Miocene flora in Achlada area had not too numerous taxa. Analyzing all the palaeobotanical findings i.e. pollen, seeds, charcoal as fusinite, fungal perithecia and other plant debris, the palaeoenvironment in which the Achlada lignite deposits were formed was reconstructed (Oikonomopoulos et al., 2008).

Frequent cysts of Zygnemataceae (*Ovoidites*), some green algae indicating a reedmoor environment also confirmed by the presence of *Phragmites communis* and divers Sparganiaceae accompanied by Polypodiaceae (spores) indicating semi-fresh reedmoor environment similar to recent peat deposits on eastern

American shoreside areas (see Riegel quoted by Oikonomopoulos et al., 2007).

The pollen of Taxodiaceae, Pinaceae Aceraceae, *Tilia*, *Myrica* and *Betula*, indicates a mixed forest environment living in warm and humid conditions.

This kind of vegetation allowed to suppose for the area a wet and warm climate with average temperatures of the coldest winter month +6 °C for *Taxodium* and +5 to +10 °C for *Glyptostrobus*, while for the warmest months temperature could reach +30°C (Kloosterboer-van Hoeve quoted by Oikonomopoulos et al., 2007), since these taxa were found in other coal deposits from the basin.

New researches in progress may evidence a rhythmic deposition of lignite-grey clay and the distinction of six sedimentological cycles, which probably correlate with orbital forcing – Milankovitch Cycles (Oikonomopoulos et al., 2007).

Anyway, thoroughly systematic palaeopalynological and palaeobotanical (palaeoxylological included) studies could give a more complete reconstruction of the palaeoenvironment in which the Achlada lignite deposits were formed.

Acknowledgements

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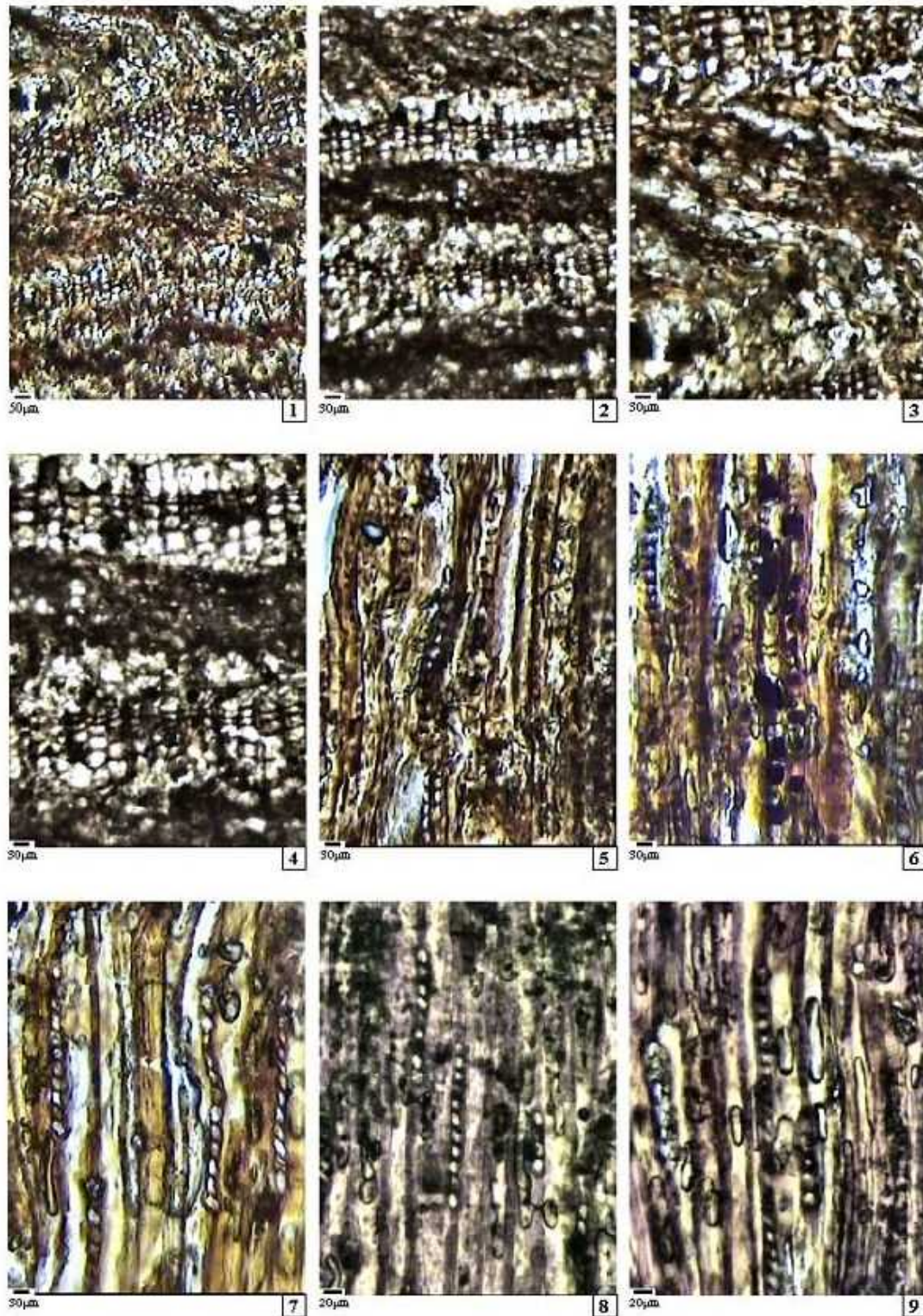


Plate I. *Taxodioxylon taxodii* GOTHAN 1906 (graphic scale). 1-4. Cross section, polygonal tracheids folded in early wood and disperse parenchyma cells. 5-9. Tangential section, small pitted tracheids, uniseriate rays medium tall, knotted horizontal parenchyma wall.

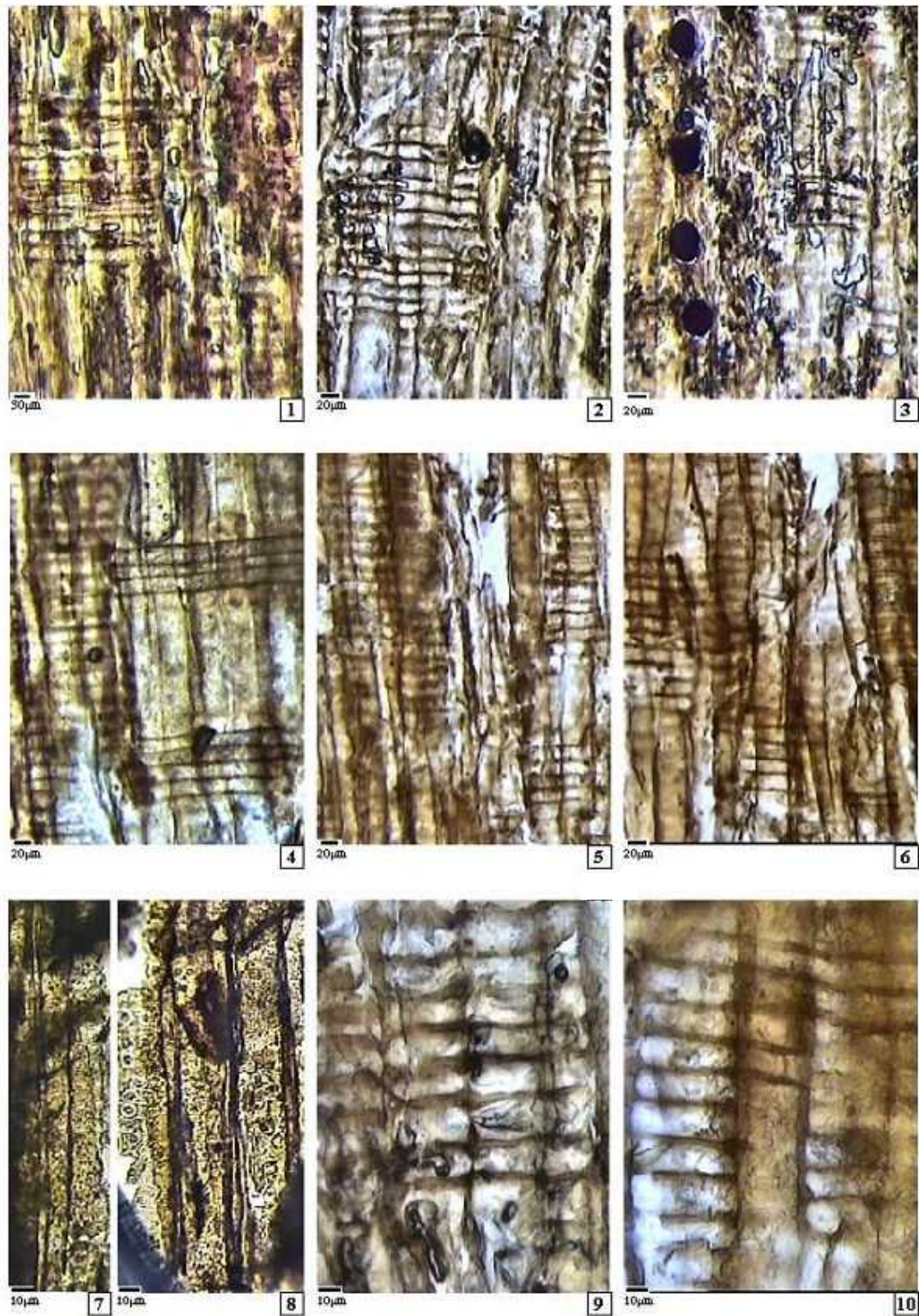


Plate II. *Taxodioxylon taxodii* GOTHAN 1906 (graphic scale). Radial sections, 1-2-radial pitting, parenchyma with globular resin deposits, taxodioid cross fields.

Exhumation of the South Carpathians (Romania) and their block structure

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Abstract The present paper is dealing with the post-collision, disjunctive tectonics of the South Carpathians. The exhumation of this mountain chain started by the end of the Middle Cretaceous, but this process ended during the Upper Cretaceous, culminating with the Getic Nappe thrust. During this uplifting process and especially after it, during the Early Tertiary, the raising Carpathian Chain was affected by some longitudinal, transverse and diagonal faults, which separated it into several blocks.

Keywords: South Carpathians, exhumation, fault, block tectonics, Romania.

1. Introduction

A not-advised geologist who would have a look at any geological map of South Carpathians would wonder why in the western part of this mountain chain the Danubian Autochthon occurs from under the Getic Nappe and in the eastern part it is still covered by this huge nappe. Moreover, while the Semenic-Almăj, Poiana Ruscă and Sebeş-Cibin massifs look like lower plateaus, the axial massifs of the South Carpathians, such as Cerna, Retezat, Vulcan, Parâng, Căpăţâna and Făgăraş are comparatively higher, being separated one another by faults or systems of faults. The greatest difference of altitude between the axial zone of the South Carpathians and the surrounding areas occurs between the Făgăraş Mountains and the Transylvanian Basin, which is of about 2000 meters. Besides, within the South Carpathians, intramontane sedimentary basins were formed, and outliers of the Getic Nappe are still preserved over the outcropping Danubian Autochthon, sometimes together with Mesozoic sedimentary deposits. All these aspects suggest that the different segments of the South Carpathians moved vertically with different amplitudes, generating a block structure of this mountain chain (see also Savu, 2005).

Contributions to the post-collision young tectonics of the South Carpathians have been presented by numerous authors: Berza and Drăgănescu (1988), Ratschbacher et al. (1993), Stănoiu (2004), Moser et al. (2005), Stănoiu et al.

(2006), Schmid et al. (2008) and others. Nevertheless, a detailed approach of the block structure of South Carpathians was not published, yet. In the present paper I try to show the main blocks into which the South Carpathians are separated and to mention the fractures along which these blocks moved.

2. Alpine evolution of the South Carpathians

By the end of Paleozoic and the beginning of the Mesozoic the European continent was undergoing a strong peneplanation process, owing to which the Variscan orogens have been intensely eroded. This erosion process affected the South Carpathians, too. Nonetheless, by the end of Triassic, an incipient continental rifting process started almost parallel with the eroded Variscan sutures (Savu, 2009). By the beginning of the lower Jurassic (ca. 180 Ma), along that continental rifting zone, a process of spreading started, so that the Carpathian Ocean opened along the rifting zone.

The spreading of the Carpathian Ocean lasted up to the end of Jurassic, when its closure started, with the convergence of the Moesian Plate and the Transylvanian Plate. Consequently, the crust of this ocean was shortened either by subduction or by obduction, the latter generating the olistostromes from the Carpathian ophiolitic suture, as for instance, that from the Mehedinţi

Plateau. The closure of the Carpathian Ocean was accomplished in the Middle Cretaceous, when the suture formed by the Severin ophiolites were covered by the Getic Nappe (Săndulescu, 1984). At the end of the upper Cretaceous, the huge Getic Nappe was thrust over the Danubian Autochthon, covering the formations and structures of the Carpathian suture. The resulted Carpathian Chain was covered by sedimentary deposits; subsequently, the main exhumation stage of the South Carpathians took place.

3. Exhumation of the South Carpathians

Exhumation of South Carpathians can be inferred for the Middle Cretaceous period, when a slow uplift of the old basement induced erosion, as shown by the exotic blocks of granitoid and crystalline schists present in the conglomerate deposits from two areas: the Mureș Couloir and the Hațeg Basin.

In the Barremian-Aptian deposits situated near the Tisa Village from the Mureș Couloir, exotic blocks of crystalline schists and large porphyritic granites with microcline megacrysts occur in a red horizon of conglomerates (Savu *et al.*, 1996). The porphyritic granites remind of the shoshonitic granites from the Voineasa area (Savu, 2008). These exotic blocks most probably came from the Sebeș crystalline schists, in which such granitoid occur south of Orăștie (I. Stelea, oral communication), a place situated near the Mureș Couloir. Berbelec (1968) also reported exotic blocks of granitoid rocks from the Cretaceous conglomerates occurring southeast of the Metaliferi Mountains, north of the Orăștie area.

Another area with exotic blocks of granitoids occurs in the Hațeg Basin, where such blocks are present in the upper Cretaceous conglomerates (Berza, 2004). These blocks must have come from the granitoid plutons of the Danubian Autochthon (which is exposed in the proximity of this sedimentary basin) during the postcollision exhumation and erosion.

The above observations show that there were uplifting phases of the crystalline basement of the South Carpathians as early as the Barremian-Aptian period, i.e. before the Getic Nappe thrust. The collision of the Moesian Plate with the Transylvanian Plate determined the thrust of the

Getic Nappe (shear nappe) over the Danubian Autochthon (Murgoci, 1905; Codarcea, 1940; Săndulescu, 1984). By the end of the Cretaceous, after the Getic Nappe thrust, owing to some fault systems of east-west and northeast-southwest orientations, in the Carpathian Chain *sensu stricto*, the main exhumation stage began (see also Matenco and Schmid, 1999 and Moser *et al.*, 2005), thus exposing the granitoid plutons to erosion, as the granitoid blocks from the Hațeg conglomerates showed. It is possible that the big granitoid plutons from the Danubian Autochthon, the mass of which was lighter than that of the surrounding crystalline schists, would have had an important role in the uplifting process. During this process or later, in the Early Tertiary, because of several diagonal and transverse fault systems (Savu, 2004-2005; Stănoiu, 2004; Stănoiu *et al.*, 2006), the South Carpathians chain was divided into several blocks (Fig. 1) and several intramontane sedimentary basins were generated, as well.

4. The main faults and blocks from the South Carpathians

First of all, regarding the structure of South Carpathians, it must be underlined again that west of the Ciunget – Polovragi fault (7, in Fig. 1) the Danubian Autochthon occurs from under the Getic Nappe and that east of this important fault the Danubian Autochthon is still covered by the Getic Nappe. This shows that the western segment of the South Carpathians was uplifted with respect to the eastern one, so that its tectonic cover consisting of the Getic Nappe was eroded, except for some outliers of this nappe (Savu, 2005).

Another feature of the South Carpathians structure is given by the Rusca Montană-Lotru-Loviște fault (9), which extends east-west on almost the entire length of the South Carpathians. The Rusca Montană-Lotru-Loviște divides the South Carpathians into a northern domain, mostly including the outcropping area of the Getic realm, and a southern domain, which mostly contains the outcropping area of the Danubian realm (Savu, 2005), as shown in Fig. 1. Moreover, along this fault, three intramontane sedimentary basins (Rusca Montană, Petroșani and Loviște (Brezoi-Titești) line up. The northern and southern

domains are generally formed of pre-Variscan and Variscan medium- to high-grade and low-grade metamorphic rocks. More information on these rocks and their outcropping areas are contained by the Geological Map of Romania, scale 1:1,000,000, which constituted the base on which the sketch-map from Fig. 1 was drawn. Yet, the map does not show clearly the block structure, excepting for the intramontane sedimentary basins, which formed as an effect of the post-collisional fault systems. I evidenced myself, alone or to-

gether with co-workers, some of these faults, such as the Lotru fault (Savu, 1968), Vârciorova-Craiu fault (Savu et al., 1973), Isverna fault (Savu, 1985), Ilova fault (Savu et al., 1973) and Muntele Mic fault (Savu, Hann, 1982), and checked in the field the presence and geometry of other major faults identified by other authors. An important contribution to the knowledge of the fault systems from the western part of the South Carpathian had Stănoiu (2004) and Stănoiu et al. (2006).

Fig. 1. Sketch-map showing the major faults and blocks of the South Carpathians. *Faults*: 1, Kujazevac – Piro; 2, Cuptoare-Secul – Sichevița; 3, Oțelu Roșu – Orșova; 4, Vârciorova – Buta (it was figured by dash line because it crosses the Retezat-Godeanu block, thus, it does not mark a block boundary); 5, Mehadia – Isverna; 6, Cerna; 7, Ciunget – Polovragi; 8, Valea lui Stan; 9, Rusca Montană – Lotru – Loviște; 10, North Făgăraș; 11, Olt Valley; 12, East Leaota; 13, North Sebeș; 14, Cărpiniș. *Blocks*: CP, Căpățâna; CZ, Cozia; DA, Danubian (Vulcan-Parâng); F, Făgăraș; L, Leaota; OL, Ocna de Fier – Locva; PF, Porțile de Fier; PR, Poiana Ruscă; RG, Retezat – Godeanu; SA, Semenice – Almăj; SC, Sebeș – Cibin;. *Intramontane basins*: B, Bozovici; CM, Caransebeș – Mehadia; LO, Loviște (Brezoi-Titești); OH, Orăștie – Hațeg; P, Petroșani; RM, Rusca Montană; RM₁, Reșița – Moldova Nouă.

4.1 Blocks from the northern domain of the South Carpathians

In the northern domain of the South Carpathians there are fewer blocks, but they are larger than those from the southern domain. In the westernmost part of this domain there is the Poiana Ruscă block (PR). This block is surrounded

by the Cretaceous and younger sedimentary deposits of the Rusca Montană Basin (RM) and Orăștie – Hațeg Basin (OH). At the north it gets in contact with the Mureș Couloir deposits, and at south it is delimited from the Retezat – Godeanu block (RG) by the long fault Rusca Montană – Lotru – Loviște (9). This block consists of low-grade and medium-grade metamorphic rocks, inclusively

crystalline limestones and greenstones (Mureşan, 1973). Numerous east-west trending secondary faults are crossing this block (Kräutner *et al.*, 1969).

East of the Poiana Ruscă block there is the Orăştie – Haţeg Basin (OH), formed due to some local faults (see also Rotschbacher *et al.* 1993), in which Mesozoic and Tertiary sedimentary deposits accumulated. East of Orăştie – Haţeg Basin, there is the large block of Sebeş – Cibin (SC), which is surrounded by the sedimentary basins Orăştie – Haţeg at the west, and the Transylvanian Basin at the northeast. At its south margin it gets in contact with the blocks of the southern domain of the South Carpathians, from which it is delimited by the Rusca Montană – Lotru – Lovištea fault (9). At the east, the Sebeş – Cibin block is delimited from the Făgăraş Block by a system of north-south trending faults (11), as shown by Dumitrescu and Săndulescu (1970), which extends along the Olt Valley zone. A branch of this fault system is running along the Vasilatu (Valea lui Stan) Valley, and continues southward as the Vasilatu (Valea lui Stan) fault (8), which is associated with gold mineralizations. These faults are reverse faults, steeply dipping eastward (see the map by Savu *et al.*, 1977). Therefore, they support the idea of some thrusting, despite the fact that they are perpendicular on the dominantly E-W trend of the East Carpathians structures. Like the Poiana Ruscă block, the Sebeş – Cibin block was also affected, along its northern part, by some east-west trending tectonic accidents. One of them is the Căpâlna incipient shear zone, where the crystalline schists and the hosted dykes of acid rocks have been deformed. In 1960, I observed south of the Cărpiniş Village an east-west trending fault in the crystalline schists, along which, the rocks had been transformed into a fault argillite with a thickness of about 1 meter.

The easternmost block of the northern domain is the Făgăraş block (F). This block is delimited by the fault (10) from the Transylvanian Basin, along which, the greatest difference of altitude (2000 m) occurs, as mentioned above. At the south, this block is contiguous with the Lovištea sedimentary basin, and is separated from the Cozia block (CZ) by the Rusca Montană – Lotru – Lovištea fault (9). The northeastern extremity of this block is delimited from the Jurassic and Cretaceous sedimentary deposits of the Braşov Basin by an

almost north-south trending system of reverse faults dipping west (12), as shown by Savu and Schuster (1971). The limit between this block and the Leaota block (L) is represented either by faults or by normal lithological contacts. This limit could represent the northeastern extension of the longitudinal fault Rusca Montană – Lotru – Lovištea (9), but this possibility was not investigated so far. Transverse faults are crossing this block (Stelea *et al.* 2004; 2009, unpublished reports).

4.2 Blocks from the southern domain of South Carpathians

The southern domain of South Carpathians contains several blocks (Fig. 1). The westernmost block from this domain is the Ocna de Fier – Locva block (OL), which is separated from the rest of the South Carpathians by faults generated since the end of the Upper Cretaceous, like the Kujazevac – Pirot fault (1), which runs from Romania southward through Serbia (Giuşcă *et al.*, 1966). Along this main fault, important banatitic intrusions, like those of Ocna de Fier and Moldova Nouă, were emplaced. Toward the west, this block is contiguous to the Pannonian Basin. The block consists of low-grade crystalline schists (Maier, 1974), partly covered by sedimentary deposits.

East of the Ocna de Fier – Locva block, there is the Reşiţa – Moldova Nouă Basin (RM₁), which subsided between the Kujazevac – Pirot fault (1) and the Cuptoare-Secul – Sicheviţa system of faults (2). Therefore, this basin looks rather like a graben, in which Paleozoic and Mesozoic sedimentary deposits were preserved (Răileanu *et al.*, 1959). It is worth showing that some of these faults formed also since the end of Upper Cretaceous, because along them, dykes of banatitic acid rocks were intruded in the Văliug area.

The Semenice – Almăj block (SA) follows toward east, which is located between the last sedimentary basin and the Caransebeş – Mehadia Basin (CM). This block includes rocks from both the Getic Nappe and the Danubian realm. In this block, the Getic Nappe consists of the medium to high-grade crystalline schists ± manganese silicates and crystalline dolomites of the Sebeş-Lotru Group and numerous granitoid (granitic and trondhjemitic) plutons (Savu, 1970; 2008; Stan

and Tiepac, 1996). The Danubian Autochthonous consists of crystalline schists, granitoid plutons and ultramafic rocks (Codarcea, 1940; Bercia and Bercia, 1975; Anastasiu, 1976; Stan et al., 1985). In the southern part of this block, the Bozovici Neogene sedimentary basin (B) was formed along local faults.

The Caransebeş – Mehadia Basin (CM) (Fig.2), containing Neogene sedimentary deposits, extends, on the north-south direction, between Semenici–Almăj block (SA) and the Retezat – Godeanu block (RG). The Oțelu Roșu – Orșova fault system (3) marks the limit between the Caransebeş – Mehadia Basin and the Retezat – Godeanu block. It is obvious that the faults from the Oțelu Roșu – Orșova system, and probably from other associated systems, have been open faults, because, along them, Paleozoic formations (Savu, Hann, 1982), Mesozoic limestones (Gherasi, Savu, 1968) and the Paleozoic formations from the granitoids near Orșova occur as tectonically-included wedges. An example in this respect is the Piatra Ilovei (Ilova Stone), which is a limestone wedge tectonically included in the pre-Variscan granitoid pluton of Muntele Mic (see the picture in Savu et al., 1973). On the western margin of the basin, it seems that its Neogene deposits transgressively lie on the crystalline schists from the Semenici–Almăj block, although, near the Slatina Timiș, some north-south short faults are separating the two structural units.

The Retezat – Godeanu block (RG), which follows toward the east, in the Danubian Autochthon area, exhibits more complicated aspects, both by position and structure. It is delimited, toward the west, by the Oțelu Roșu – Orșova fault system (3), toward the southeast, by the Cerna fault (6), and toward the north, by the Rusca Montană – Lotru – Loviștefa fault (9). According to Berza and Drăgănescu (1988), the Cerna fault extends further away, toward the northeast. The Mehadia – Isverna fault (5) intersects the Cerna fault at its southern extremity. Around the crossing of the two faults there are the famous Băile Herculane thermal water springs, known since the Roman period. In the structure of the Retezat – Godeanu block, an infrastructure and a superstructure can be distinguished. The infrastructure is formed by the Jurassic sedimentary formations of the Arjana Zone (Codarcea, 1940), which include alkali volcanics,

and of Cretaceous sedimentary deposits, over which the superstructure of the Godeanu outlier of the Getic Nappe is placed (Gherasi, 1937; Bercia, 1975), being formed of the crystalline schists from the Sebeș-Lotru Group. The west–east trending Vârciorova–Buta fault (4) cuts the Retezat – Godeanu block. Along this fault, the Muntele Mic granitoids have been strongly sheared.

Southeast of this block, the Danubian (Vulcan-Parâng) block (DA) follows, which extends from the Isverna – Mehadia fault (5) round up to the Ciunget – Polovragi fault (7), which delimits at the same time the two big segments of the South Carpathians (Savu, 2005).

On the old geological maps of Romania, the Ciunget-Polovragi fault (7) was represented as a margin of the Getic Nappe, thus part of the Getic thrust. If it had been so, the Danubian Autochthon should have extended like a ‘V’, downstream, along the Lotru Valley, which is a dip valley. Nevertheless, the representation of this tectonic zone on the geological maps as an almost straight line, indicates a very steep dip to the east, thus its fault nature, with a north-south strike. Therefore, the Ciunget-Polovragi fault (7) cuts the South Carpathians chain, so that the western Danubian (Vulcan-Parâng) block (DA) was uplifted and the eastern Căpățâna block was subsided, the latter still preserving the Getic Nappe formations (see also Savu, 2005). At the same time, it determined the displacement toward north of the Danubian (Vulcan-Parâng) block (DA) and of the plane of the long fault (9), as shown in Fig. 1.

The Danubian (Vulcan-Parâng) block (DA) consists of the Lainici-Păiuș and Drăgășan metamorphic rock series (Manolescu, 1933) and numerous granodioritic-granitic, thondhjemitic and shoshonitic plutons (Savu, 2008) and Mesozoic sedimentary deposits over which some Getic Nappe outliers are lying. Between this block and the Retezat – Godeanu block is situated the intramontane sedimentary basin of Petroșani (P), which includes coal deposits among its formations.

The possible involvement of east-west longitudinal faults at the transition from the South Carpathians to the Moesian Platform it is very difficult to check because of the foredeep cover. Nevertheless, in some places, such as Stănceti (Savu et al., 1972) and Mănăstirea Crasna (Savu et al., 1973) such faults are visible. These are reverse

faults dipping northward, which catch along them deposits of the Schela Formation.

At the southwestern extremity of the Danubian (Vulcan-Parâng) block (DA) the Porțile de Fier block (PF) is located, which extends from the Mehadia-Isverna fault (5) down to Danube. It consists of Mesozoic sedimentary deposits overlain by the Getic Nappe outlier of Porțile de Fier. This block and the Mehadia-Isverna fault (5) are situated in front of the Wallachian spur (Stille, 1953) of the Moesian Plate, which advanced from the east.

East of the Danubian (Vulcan-Parâng) block (DA) and the Ciunget-Polovragi main fault (7), the Căpățâna block (CP) is located. It is delimited at the north by the Rusca Montană – Lotru – Lovișteea fault (9) and by the Vasilatu (Valea lui Stan) fault (8) at the east. This block consists of the medium- to high-grade metamorphic rocks of the Sebeș-Lotru Group. The Cozia block (CZ) follows to the east and extends eastward to the Leaota block (L). The Cozia block is delimited from the Făgăraș block by the Rusca Montană – Lotru – Lovișteea fault (9) and from the Leaota block (L) by the northwest-southeast trending Câmpulung fault system. It consists of the medium- to high-grade crystalline schists of the Făgăraș Mountains, the Cozia ocular migmatites and by Eocene sedimentary deposits (Dimitrescu, 1962; Savu, Schuster, 1971; Balintoni, 1975). The easternmost block of the southern domain of South Carpathians is the Leaota block (L). This has a more complicated structure compared with the other blocks of the southern domain, consisting of metamorphic rocks (Gherasi, Dimitrescu, 1974) and patches of Mesozoic sedimentary deposits, which extend eastward to the Bucegi Mountains, from which the Leaota block is delimited by the fault system (12) consisting of reverse faults with westward dip (Savu and Schuster, 1971), in the opposite sense to the fault system (11) from the western boundary of the Făgăraș block (F), which are dipping eastward. Both these structural features support the bilateral uplifting of the Făgăraș block (F).

Like the other blocks from the southern domain of the South Carpathians, the Leaota block is in contact with the young sedimentary deposits of the Getic Depression.

5. Conclusions

The exhumation of South Carpathians took place along some longitudinal, marginal faults and its inception goes as far back as the end of the Middle Cretaceous, as shown by the exotic blocks of Carpathian granitoids occurring in the sedimentary deposits of this age. The most important uplift of the South Carpathians occurred during the Upper Cretaceous, culminating with the Getic Nappe thrust. By the end of the Upper Cretaceous and during the Early Tertiary, besides the mentioned longitudinal faults, transverse and diagonal faults played an important role, determining the separation of the South Carpathians into several blocks. It seems that, during the exhumation of the South Carpathians, owing to the inner longitudinal Rusca Montană – Lotru – Lovișteea fault (9), the Carpathian Chain was divided into a northern domain and a southern domain. Then, the transverse fault Ciunget-Polovragi (7) separated the southern domain into a western segment and an eastern segment, displacing the western segment northward. The western segment was more uplifted than the eastern one, so that its cover of Getic Nappe was almost completely eroded, save for some outliers, while over the subsided eastern segment the Getic Nappe is still present. Concomitantly with this process and after it, the mentioned fault systems achieved the separation of South Carpathians into several blocks. This study points to the importance of faults, beside thrust nappes, in the construction of the present edifice of the South Carpathians.

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In Memoriam

Gheorghe Munteanu-Murgoci
(1872-1925)



In 2012, on the 20th of July, we celebrate the 140th anniversary of Gheorghe Munteanu-Murgoci's birthday. Gheorghe Munteanu-Murgoci is one of the most brilliant figures in the Romanian geology. He was born in 1872, in the south-eastern part of Romania, near Brăila, while his parents were moving with their sheep herds to the grazing lands. The name Munteanu was added by his teacher in the primary school. In 1895, he graduated from the University of Bucharest (Faculty of Sciences). After doctoral studies in Vienna and Munich under the supervision of professors Tschermak, Becke, Berwerth, Groth and Weinschenk, in 1900, Gheorghe Munteanu-Murgoci got his Ph.D. degree from the University of Munich. His teaching activity was shared between high school and Academia (University of Bucharest, Polytechnics School, University of Cluj). In the period 1906-1925, Gheorghe Munteanu-Murgoci played a leading role in the research activity of the Geological Institute of Romania. He was member of the Romanian Academy and of the French Academy for Agriculture, president of the International Committee of Soil Mapping (1923) and director of the Pedologic Map of Europe (1924). He died prematurely on the 5th of March, 1925.

Gheorghe Munteanu-Murgoci published more than 200 articles and books covering very diverse subjects: mineralogy and petrology, tectonics, regional geology, geological mapping, geography, mineral resources, hydrogeology and pedology (e.g., Murgoci and Mrazec, 1897; Murgoci, 1901a,b, 1903, 1905a,b,c,d,e, 1906, 1910, 1911a,b, 1912a,b,c,d, 1914, 1915, 1916, 1917a,b, 1922a,b, 1923a,b,c, 1924a,b,c,d, 1925a,b, Murgoci and Osiceanu, 1908, Murgoci et al., 1925), but his most outstanding contributions are in the fields of mineralogy, tectonics and pedology.

Mineralogy. Gheorghe Munteanu-Murgoci contributed to the knowledge of the amphiboles (Murgoci, 1905d, 1906, 1922a,b), investigating the relations between the chemical composition and the optical properties of this group of minerals. Murgoci (1922a,b) published the optical characteristics of the amphibole barroisite (named after the French geologist Charles Barrois, 1851-1939), a valid IMA species.

In his Ph.D. Thesis, published in the Bulletin of the Society of Sciences of Bucharest (Murgoci, 1901), he described the pumpellyite in the Southern Carpathians, under the name "lotrite" (Coombs, 1953), giving its chemical composition and optical properties. The data on the lotrite produced by Murgoci were also published in the Bulletin de la Société Française de Minéralogie (Gaubert, 1901). Lotrite was subsequently identified in the greenstones from the Calaveras Valley, California (Murgoci,

1906; Eakle, 1914, page 132). However, the information on the lotrite seems not to have been spread enough and the mineral was reported as pumpellyite by Palache and Vassar (1925), a few months after Murgoci's death.

Murgoci (1903, 1925a) studied the amber deposits in Romania, showing that their chemical properties and paleogeographic conditions of formation are similar with those of the amber deposits in Scandinavian area, United Kingdom and Crimea.

Tectonics. Gheorghe Munteanu-Murgoci (1905b,c,d) described the Getic thrust nappe of the Southern Carpathians, two years after the first report of thrust nappes in the Carpathians (Lugeon, 1903, in the Western Carpathians). The cartographic outlines of the Getic front drawn by Murgoci were generally confirmed by the subsequent research.

Gheorghe Munteanu-Murgoci studied the geology of Dobrogea, with inferences on the role played by North Dobrogea in the Cimmerian orogeny (e.g., Murgoci, 1912b, 1914, 1915, 1925b).

Pedology. Gheorghe Munteanu-Murgoci contributed to the development of soils classification and mapping and investigated the influence of climate, relief and vegetation on soil composition (Murgoci, 1911a,b, 1923b, 1924a,b,c,d). He issued the first soil map of Romania (Murgoci, 1924b) and also published studies on soils from Ukraine and Turkey (Murgoci, 1924a,c).

In the extension of these very brief biographical notes, we give the translation of that part of Gheorghe Munteanu-Murgoci's Ph.D. Thesis dedicated to the characterization of the new mineral lotrite (pumpellyite), a text which was not published in English language, so far. We also add the map "Dunărea Albastră" ("Blue Danube"), presented by Gheorghe Munteanu-Murgoci at the International Geological Congress in Stockholm, in 1910, which contains the first graphical representation of the Getic Nappe.

Marian Munteanu

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The hornfelses with vesuvian and garnet from the serpentinites of the Parâng massif¹

Gheorghe Munteanu-Murgoci

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West of the Mărghilele Cărbunelui, towards Căldarea Petresii, an interesting contact between the serpentinites and greenschists or calcschists is visible. Beside the common epidote schists and epidotites, zoisitites also were generated. Aside from outcrops, there are many blocks of rocks with epidote and zoisite (zoisite is dominant), whose aspect suggests they are enclaves in serpentinite.

The blocks show envelopes of compact chlorite, which send thin veinlets inside the reddish-whitish or greenish-grey hornfelses. The limit between the chlorite envelope and the hornfels is not sharp. Commonly, there is a gradual transition from the chlorite to a mixture of chlorite and zoisite, then to the zoisite mass. The zoisite mass is very compact, white, greenish-grey or spotted, with irregular fracture, and, in some places, it produces effervescence with hydrochloric acid (veins and nests of calcite). The white spots are monominerallic zoisite. At the microscope, the hornfels is made up of an aggregate of small zoisite columns with yellowish epidote spots and indented brown orthite grains, which, sometimes, show yellowish epidote rims.

The greenish-grey or spotted hornfelses are aggregates of clinozoisite intergrown with chlorite or with a new mineral, the *Lotrite*. The spotted aspect is given either by the greenish Lotrite, which generates

¹ Ph.D. Thesis defended at the University of Munich, in 1900, and published in 1901, with the title "Granat und Vesuvianfels aus dem Serpentin von Parângu" in "Buletinul Societăţii de Ştiinţe din Bucuresci" (Bulletin of the Society of Sciences of Bucharest), vol. 9, issues 5 and 6, p. 568-612 and 764-831.

veinlets and irregular spots in the white zoisite groundmass, or by the reddish clinozoisite, which forms veinlets and spots in the translucent greenish mass of the Lotrite.

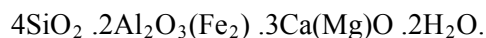
At the microscope, the greenish mass is an aggregate of small columns and blades of a strongly refringent mineral ($N = 1.67$), with a distinctive cleavage parallel with the elongation. It is weakly birefringent ($\gamma - \alpha$ is maximum 0.014), the extinction angle measured to cleavage is 28° , $2E = 30^\circ$, $2V = \text{ca. } 18^\circ$, and the optical sign is positive. The plane of the optical axes is transversal to the elongation and perpendicular to the cleavage, $\gamma \parallel b$.

An analysis on a mineral separate yielded the following composition:

		Molecular ratio	
SiO ₂	38.02	6.33	
Al ₂ O ₃	} 30.90	0.303	
Fe ₂ O ₃			
FeO	0.33		
CaO	23.56	0.421	} 0.491
MgO	2.80	0.70	
LOI	6.24	0.346	
Total	101.85		

Specific gravity: 3.23. Hardness: 7.5.

Considering the analysis, this is a hydrated aluminous-calcian silicate, with the approximate formula



Characteristic for this mineral is the silica-water ratio $\text{SiO}_2 : \text{H}_2\text{O} = 2 : 1$, a ratio that does not correspond to any known aluminous-calcian hydrous silicate.

Neither its optical properties allow its identification with any of the known minerals.

Therefore, it is a new mineral, which, because of its occurrence in the valley of Lotru River, I gave the name *Lotrite*.

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The mineral that I named *Lotrite* can be found only in the compact rocks which occur as enclaves of varied size, in serpentinites, near the contact from Mărgmilele Cărbunelui.

It participates, alone or with the clinozoisite, to the make-up of the hornfelses that are separated from the serpentinite by a zone of chlorite. At the margins of the enclaves, toward the chlorite envelope, the clinozoisite dominates, while the lotrite develops only toward the center of the enclave. There are very many nodules formed by clinozoisite (alone or together with chlorite), especially near the contact, where minerals such as epidote, chlorite etc are present in a higher proportion; some enclaves are made up of chlorite-epidote schists still not transformed.

The lotrite occurs as fine blades up to 15 mm in length, grey to dark green, colorless in thin sections. It has the properties already mentioned and is characterized by extinction angle of 28° and weak birefringence ($\beta - \alpha$) of ca. 0.002. The angle of the optical axes is small ($2V$ maximum 20°), with positive optical sign. In some respects, the lotrite resembles the prehnite, but differs from it by the higher

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refrindex and lower birefringence ($N = 1.67$; γ - α maximum 0,014); the density and hardness of the lotrite are much greater; moreover, the chemical composition indicates it is different from the prehnite.

Lately, I performed a new analysis on a higher-purity separate, which yielded the following composition:

SiO ₂	39.44	
Al ₂ O ₃	28.33	Includes Fe ₂ O ₃
CaO	22.21	
MgO	3.20	
Na ₂ O	0.93	
H ₂ O	6.58	
Total	100.69	

Specific gravity: 2.229.

This analysis does not show great difference from the one presented earlier, therefore, the formula $4\text{SiO}_2 \cdot 2\text{Al}_2\text{O}_3(\text{Fe}_2) \cdot 3\text{Ca}(\text{Mg})\text{O} \cdot 2\text{H}_2\text{O}$, determined initially, seems to be correct. The *Chlorastrolite*, very similar to the lotrite from the chemical point of view, proved to be an impure prehnite.

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The tectonic sketch map "Dunărea Albastră" ("Blue Danube"), containing the first graphical representation of the Getic Nappe. The map was presented at the International Geological Congress in Stockholm, in 1910. The original map is kept at the University Politehnica of Bucharest, Faculty of Material Science and Engineering. By courtesy of Dr. Mircea Ionuț Petrescu, Vice Dean of the Faculty of Material Science and Engineering.

